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Changes in meridional temperature and salinity gradients in the North Atlantic Ocean (30°-72°N) during the last interglacial period

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Abstract. Eight deep-sea sediment cores from the North Atlantic Ocean ranging from 31° to 72°N are studied to reconstruct the meridional gradients in surface hydrographic conditions during the interval of minimum ice volume within the last interglacial period. Using benthic foraminiferal $\delta^{18}O$ measurements and estimates of Sea Surface Temperature (SST) and Sea Surface Salinity (SSS), we show that summer SSTs and SSSs decreased gradually during the interval of minimum ice volume at high-latitude sites (52°-72°N) whereas they were stable or increased during the same time period at low-latitude sites (31°-41°N). This increase in meridional gradients of SSTs and SSSs may have been due to changes in the latitudinal distribution of summer and annual-average insolation and associated oceanic and atmospheric feedbacks. These trends documented for the Eemian ice volume minimum period are similar to corresponding changes observed during the Holocene and may have had a similar origin.

1. Introduction

Societal interest in the character of possible future climate variation and lingering uncertainty regarding the stability of the climate during the last interglacial period (the Eemian in Europe and isotopic substage 5e in marine records [*Mangerud et al.*, 1979]) have stimulated renewed interest in natural climate variability during warm intervals. Interest in the last interglacial period, in particular, was peaked by initial results from the deeper portions of the Greenland Ice Core Project (GRIP) suggesting that the climate of Eemian was not uniformly warm [*Climate: Long-Range Investigation Mapping and Prediction (CLIMAP) Project Members*, 1984] but experienced rapid climatic coolings [*Dansgaard et al.*, 1993]. It is now well established that ice originally thought to correspond to the Eemian in GRIP has been disturbed [*Fuchs and Leuenberger*, 1996]. Nevertheless, several subsequent studies in continental and marine environments have shown that Eemian climate was not as stable as previously suggested [*Adkins et al.*, 1997; *Cortijo et al.*, 1994; *Field et al.*, 1994; *Fronval and Jansen*, 1996; *Keigwin and Jones*, 1994; *Seidenkrantz et al.*, 1995]. These underscore the continued need to evaluate the nature and origin of climate variability during the last interglacial period and other warm intervals.

In this study we use faunal and stable isotope analyses of foraminifera in eight sediment cores from 30° to 72°N in the North Atlantic Ocean to reconstruct and evaluate changes in surface hydrography during the ice volume minimum period within marine isotopic substage 5e. Although there is little evidence of dramatic variability in SST or SSS during the ice volume minimum period, we show that latitudinal gradients in SST and SSS along the Gulf Stream current increased markedly during the interval. This may have been a response to gradual changes in the latitudinal distribution of solar radiation and/or a gradual reduction in the strength of the large-scale overturning thermohaline circulation.

2. Core Locations and Methodology

Cores used for this paper are located along a SW-NE transect in the North Atlantic Ocean (Table 1 and Figure 1) and generally follow the present-day path of the Gulf Stream-North Atlantic Drift from subtropical latitudes (31°N) to polar latitudes (72°N). The exception is core CH69-K9 at 41°N, located in the area alternatively bathed by the Gulf Stream and the Labrador Current [*Worthington*, 1962] but outside the region influenced by cold- and warm-core rings [*Nof*, 1986; *Saunders*, 1971; *Tchernia*, 1969].

Because *Cibicides wuellerstorfi* was not present in all samples, the chronology is based on $\delta^{18}O$ measurements of three different benthic foraminiferal species[1]: *C. wuellerstorfi*, *Oridorsalis tener*, and *Uvigerina peregrina*. Benthic foraminiferal $\delta^{18}O$ results have therefore been corrected to account for species-dependent isotopic fractionation (+0.64‰ for *C. wuellerstorfi* [*Shackleton and Opdyke*, 1973] and +0.37‰ for *O. tener* [*Streeter et al.*, 1982]; see Table 1). Planktic species used for $\delta^{18}O$ analysis are *Neogloboquadrina pachyderma* sinistral (left-coiling) in core V27-60, *N. pachyderma* dextral (right-coiling) in

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[1] Supporting data are available electronically at World Data Center-A for Paleoclimatology, NOAA/NGDC, 325 Broadway, Boulder, Colorado (e-mail: paleo@mail.ngdc.noaa.gov; URL: http://www.ngdc.noaa.gov/paleo).

Table 1. Location of the Cores Used in This Study and the Benthic Foraminiferal Species Analyzed

Core	Latitude	Longitude	Depth, m	Reference	Benthic Species
V27-60	72°11'N	8°35'E	2525	1	*Cibicides wuellerstorfi* (+0.64)[4] *Uvigerina peregrina*
JPC8	61°N	25°W	1917	2	*C. wuellerstorfi* (+0.64)[4]
NA87-25	55°11'N	14°44'W	2320	1	*C. wuellerstorfi* (+0.64)[4] *U. peregrina*
NEAP18K	52°46'N	30°21'N	3275	this study	*C. wuellerstorfi* (+0.64)[4]
CH69-K9	41°N	47°N	4100	this study	*C. wuellerstorfi* (+0.64)[4]
SU90-03	40°30'N	32°03'W	2475	3	*C. wuellerstorfi* (+0.64)[4]
MD95-2036	33°41'N	57°34'W	4461	this study	*C. wuellerstorfi* (+0.64)[4] *Oridorsalis tener* (+0.37)[5]
JPC37	31°41'N	75°25'W	2972	this study	*C. wuellerstorfi* (+0.64)[4]

References given on the table: 1, *Cortijo et al.* [1994]; 2, *Oppo et al.* [1997]; 3, *Chapman and Shackleton* [1998]; 4, *Shackleton and Opdyke* [1973]; and 5, *Streeter et al.* [1982].

core EW9302-JPC8 [*Oppo et al.*, 1997], *Globigerina bulloides* in cores SU90-03, NEAP18K, NA87-25, and CH69-K9, and *Globigerinoides ruber* (white variety) in cores MD95-2036 [*Adkins et al.*, 1997] and KNR140-JPC37. Isotopic measurements for cores V27-60, NA87-25, and CH69-K9 were performed at the Laboratoire des Sciences du Climat et de l'Environnement, Gif-sur-Yvette, France, using a Finnigan Mat 251 isotope mass spectrometer equipped with a "Kiel device" for automated individual acidification of samples. The mean external reproducibility of powdered carbonate standards is ±0.06‰ for oxygen. Isotopic measurements for cores EW9302-JPC8 (referred to hereafter as JPC8), MD95-2036, and KNR140-JPC37 (referred to hereafter as JPC37) were performed at Woods Hole Oceanographic Institution, Woods Hole, Massachusetts. Results for JPC8 and MD95-2036 were obtained using a Finnigan Mat 252 with a Kiel device. The mean external precision for the powdered carbonate standard National Bureau of Standards (NBS)-19 is ±0.08‰. Results for JPC37 were obtained using a semiautomated VG903 mass spectrometer fitted with a small volume inlet and a high sensitivity source. The mean external precision for the powdered carbonate standard NBS-19 is better than ±0.10‰. Isotopic measurements for cores SU90-03 and NEAP18K were performed at the Godwin Laboratory, University of Cambridge, Cambridge, England, United Kingdom. Planktic samples were analyzed using a VG SIRA II mass spectrometer. Benthic foraminiferal results were obtained using a VG PRISM mass spectrometer. Both mass spectrometers are fitted with the VG isocarb common acid bath system. Analytical precision of laboratory carbonate standards is better than ±0.08‰ for $\delta^{18}O$. All isotopic data are reported versus Pee Dee Belemnite (PDB) standard after calibration with NBS-19 [*Coplen*, 1988].

Sea surface temperature (SST) estimates were based on planktic species counts of at least 300 individuals. The modern analog technique was used to estimate paleotemperatures by identifying the five most similar core-top samples in the North Atlantic database. The database contains 615 core-top samples between 5° and 80°N (i.e., the northern part of the database used by *Pflaumann et al.* [1996]). Summer and winter SSTs were

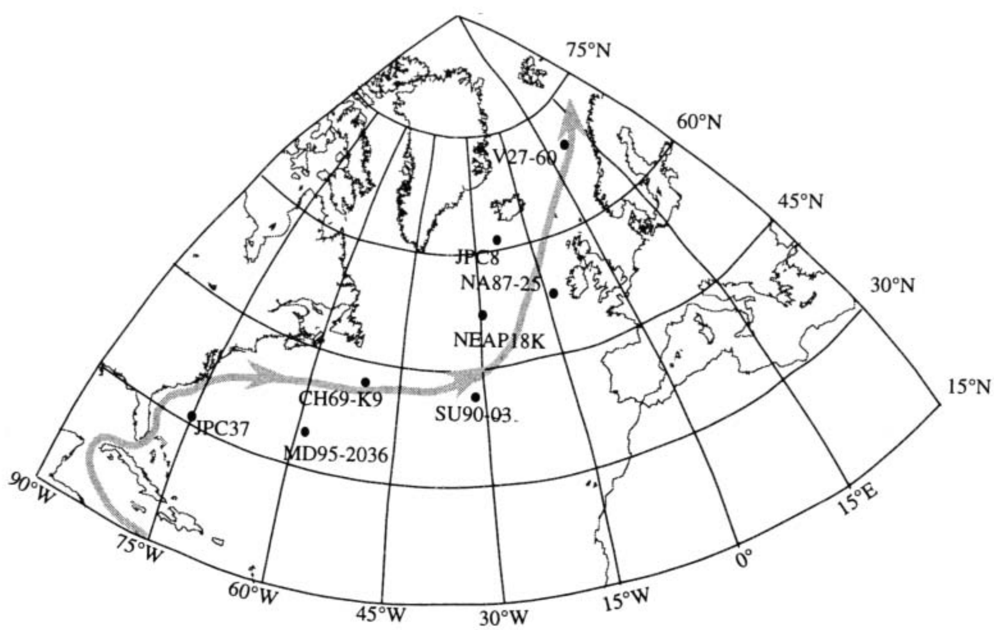

Figure 1. Location of cores used in this study.

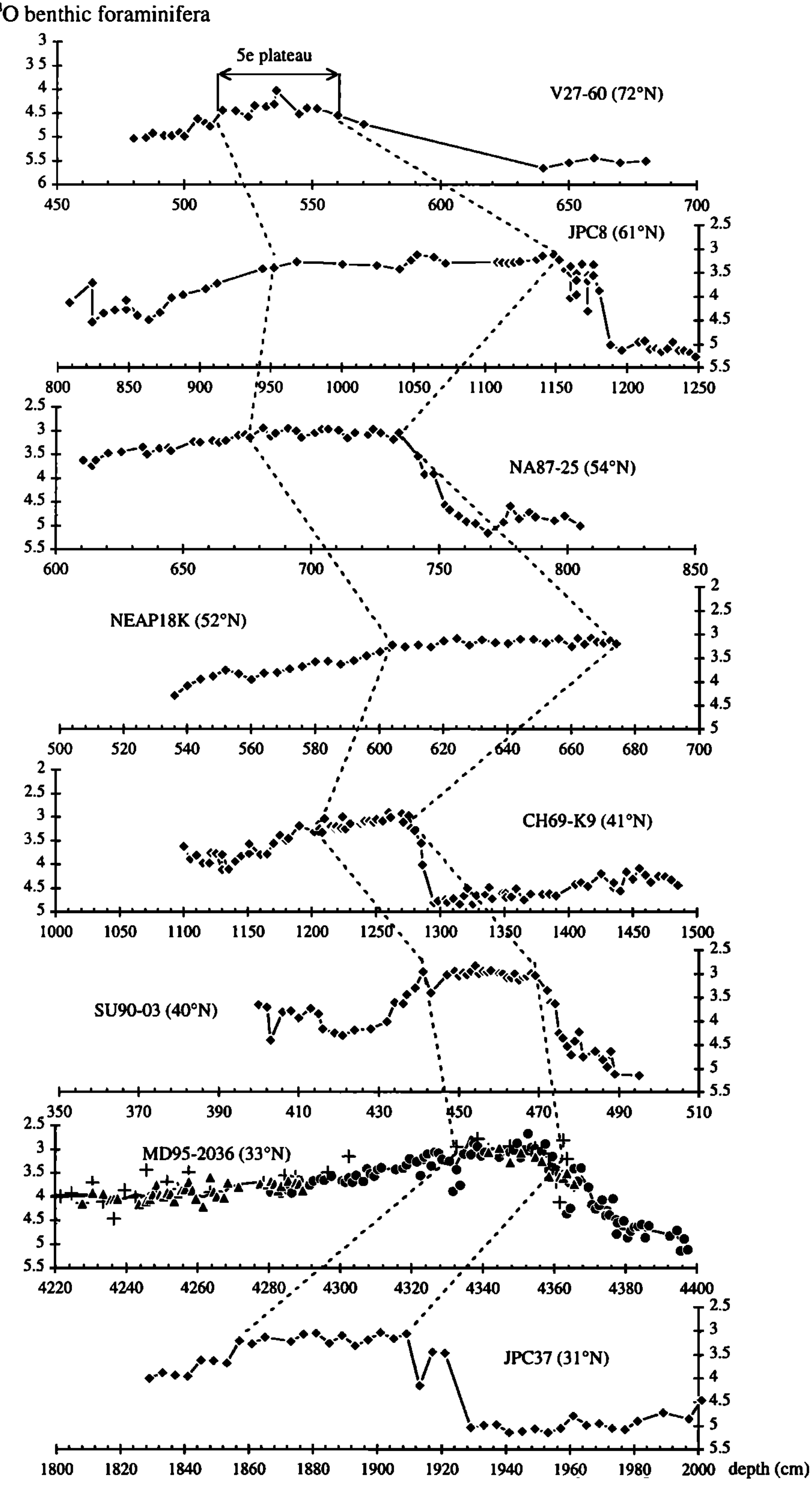

Figure 2. Benthic foraminiferal $\delta^{18}O$ records in all cores showing the 5e plateau versus depth in each core. In core MD95-2036, benthic foraminiferal $\delta^{18}O$ data are from three sources: VG *Cibicides wueslIerstorfi* (triangles), Finnigan *C. wueslIerstorfi* (plus signs) and *Oridorsalis tener* (circles). A line is drawn through the data as a three-point running average in core MD95-2036 to smooth the signal. The *C. wuellerstorfi* and *O. tener* $\delta^{18}O$ values are corrected by +0.64‰ [*Shackleton and Opdyke*, 1973] and by +0.37‰ [*Streeter et al.*, 1982], respectively to take into account the departure from isotopic equilibrium.

estimated by averaging summer and winter SSTs associated with the most similar core tops [*Prell*, 1985]. Similarity between sample and core-top assemblages was calculated for all 32 planktic taxa using the squared chord dissimilarity measure. Uncertainty in the SST reconstructions corresponds to the root mean square error of the top five analog temperatures. In each case all of the five most similar core-top samples were valid modern analogs for the studied fossil sample (i.e., having a dissimilarity coefficient <0.2 [*Overpeck et al.*, 1985]). Error bars on temperature reconstructions are between 0.5°C and 2°C for all cores except MD95-2036 and JPC37, where standard deviations are sometimes as great as 3.5°C, probably because of the paucity of nearby core tops in the reference database. Two of the studied cores (CH69-K9 and MD95-2036) were collected from more than 4000 m water depth and are subject to mild dissolution during glacial periods which might influence associated SST estimates, but there is no evidence of dissolution in interglacial sediments.

In order to estimate changes in sea surface salinity (SSS) we use the measured planktic foraminiferal $\delta^{18}O$ and the reconstructed summer SST to solve the paleotemperature equation [*Shackleton*, 1974] for the $\delta^{18}O$ of seawater. SSS is reconstructed from seawater $\delta^{18}O$ using the global average $\delta^{18}O$/salinity relationship for modern surface water of 0.5/1 [*Craig and Gordon*, 1965]. As salinity-related changes in seawater $\delta^{18}O$ during the transition into and out off marine isotope substage 5e are very difficult to quantify because of changes in ice volume, we restrict our calculation of SSS to the interval of minimum ice volume within the substage.

The approach described above assumes that planktic foraminifera have produced shell calcite in isotopic equilibrium with seawater, as determined by growth temperature. The relation between the growth temperature and summer SST has been calibrated for *N. pachyderma* (left-coiling) and *G. bulloides* species [*Duplessy et al.*, 1991] in the North Atlantic and for *G. ruber* (white variety) in the eastern tropical Atlantic [*Wang et al.*, 1995]. However, we also use *N. pachyderma* (right-coiling) in core JPC8, a species for which the growth temperature has not been calibrated, and *G. ruber* in two cores coming from the western Atlantic (JPC37 and MD95-2036) where oceanographic characteristics differ substantially from those in the eastern Atlantic (in the Sargasso Sea, *Deuser* [1987] has shown that *G. ruber* white was close to the equilibrium, within 0.2‰). The accuracy of the salinity reconstructions is highly dependent on the temperature estimates: an average error of 2°C in SST will be translated into an error of 1 in SSS. For these reasons and because the $\delta^{18}O$/salinity relationship of seawater during the last interglaciation may have differed from the modern one and not

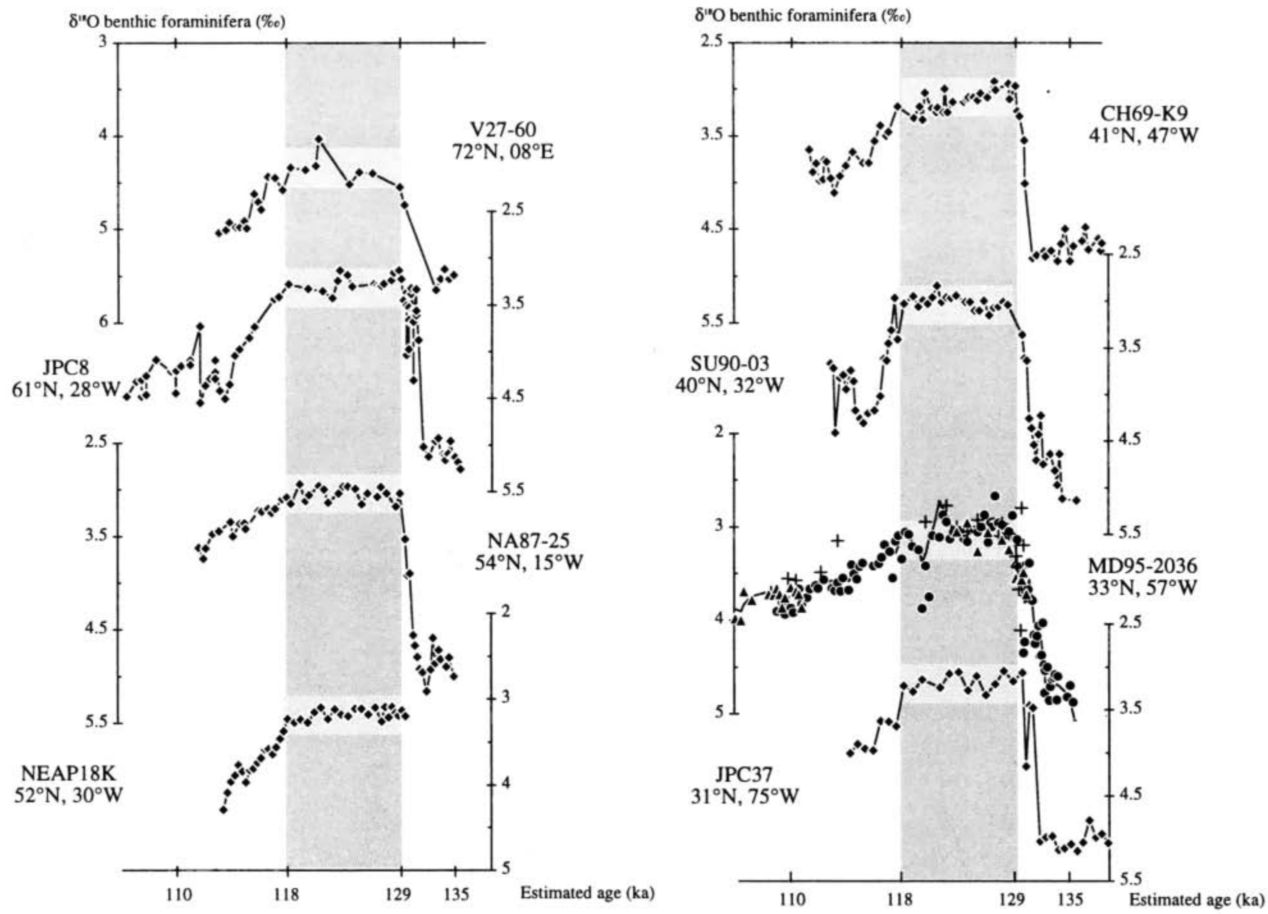

Figure 3a. Benthic foraminiferal $\delta^{18}O$ in all the cores studied in this work put on the same age scale using four tie points (caption for core MD95-2036 is the same as in Figure 2).

be constant through the North Atlantic Ocean [*Rohling and Bigg*, 1998] we will not consider absolute salinity values, and we will only discuss relative changes in estimated SSS.

3. Construction of an Age Scale and Identification of Isotope Substage 5e

The chronology is established using benthic foraminiferal $\delta^{18}O$ records. All studied cores except NEAP18K show the large decrease in benthic foraminiferal $\delta^{18}O$ associated with the transition between isotope stage 6 and substage 5e. However, the high levels of ice-rafted debris input and the presence of *N. pachyderma* left-coiling at the base of the NEAP18K (>670cm), signals that also recorded in the 6/5e transition of cores NA87-25 and JPC8, indicate that the bottom of core NEAP18K corresponds to the 6/5e transition. The substage 5e-5d transition is much more difficult to identify with confidence (Figures 2 and 3a). However, in focusing on the interval of minimum ice volume within substage 5e we are able to side step some of the uncertainties in identifying isotopic transitions within the different cores. Whenever possible, we used the following four tie points to constrain the age scale: the isotopic maxima of stage 6, the isotopic maxima of substage 5d, and, when possible, the local isotopic minima at the beginning and end of the interval of sustained low values (events 5.53 and 5.51, respectively, in the stacked and orbitally tuned timescale of *Martinson et al.* [1987] (Table 2)). We refer hereafter to the latter interval as the "5e plateau". In some cases the bounding isotopic minima were not obvious. In these cases, as, for example, in core CH69-K9, we chose to define the 5e plateau on the basis of the interval of lowest average $\delta^{18}O$ value ±0.2‰ (to account for natural isotopic and temperature variability during the interval of minimum ice volume).

All the cores studied in this work were placed on a common age scale following the method described above and using the age scale established by *Adkins et al.* [1997] as reference (Table 2). Sediment thicknesses for stratigraphic sections corresponding to the 5e plateau (Table 2) range from 25 (core SU90-03) to 192 cm (core JPC8). These thicknesses equate to a range in average sedimentation rate of 2.5 to 19.2 cm kyr^{-1} with the *Adkins et al.* [1997] timescale (or from 9.6 to 73 cm kyr^{-1} if we use the definition of the 5e plateau by *Martinson et al.* [1987]). Although the temporal resolution of sampling for core V27-60 is not as great as in the other cores, we include it here to provide needed constraint on possible changes in SST and SSS at polar latitudes. Unfortunately, there are as yet no better records available from this part of the Norwegian Sea with SST reconstructions.

The precise duration of the Eemian period is not fully resolved [see, e.g., *Adkins et al.*, 1997; *Kukla et al.*, 1997; *Martinson et al.*, 1987]. However, neither the duration nor the precise timing

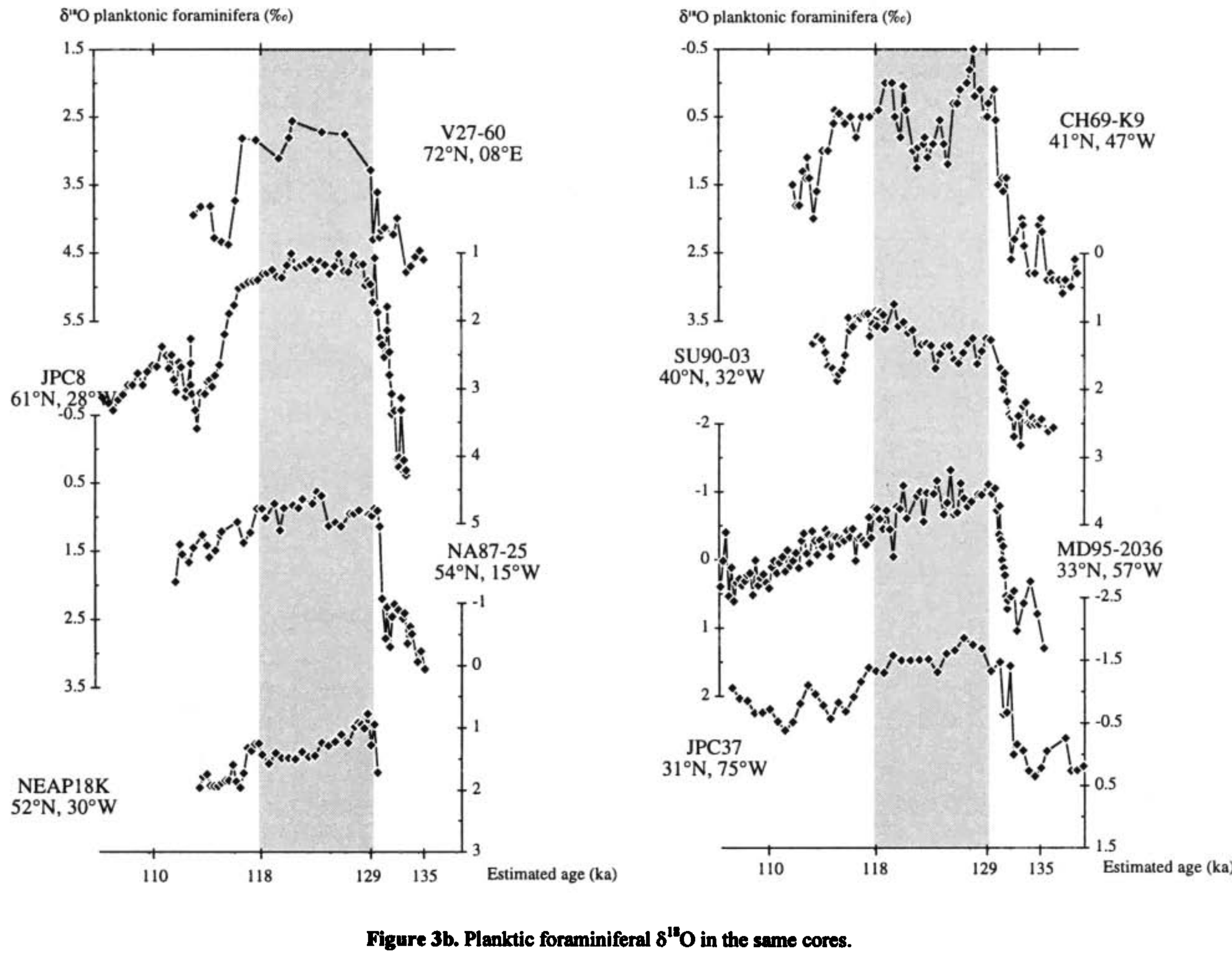

Figure 3b. Planktic foraminiferal $\delta^{18}O$ in the same cores.

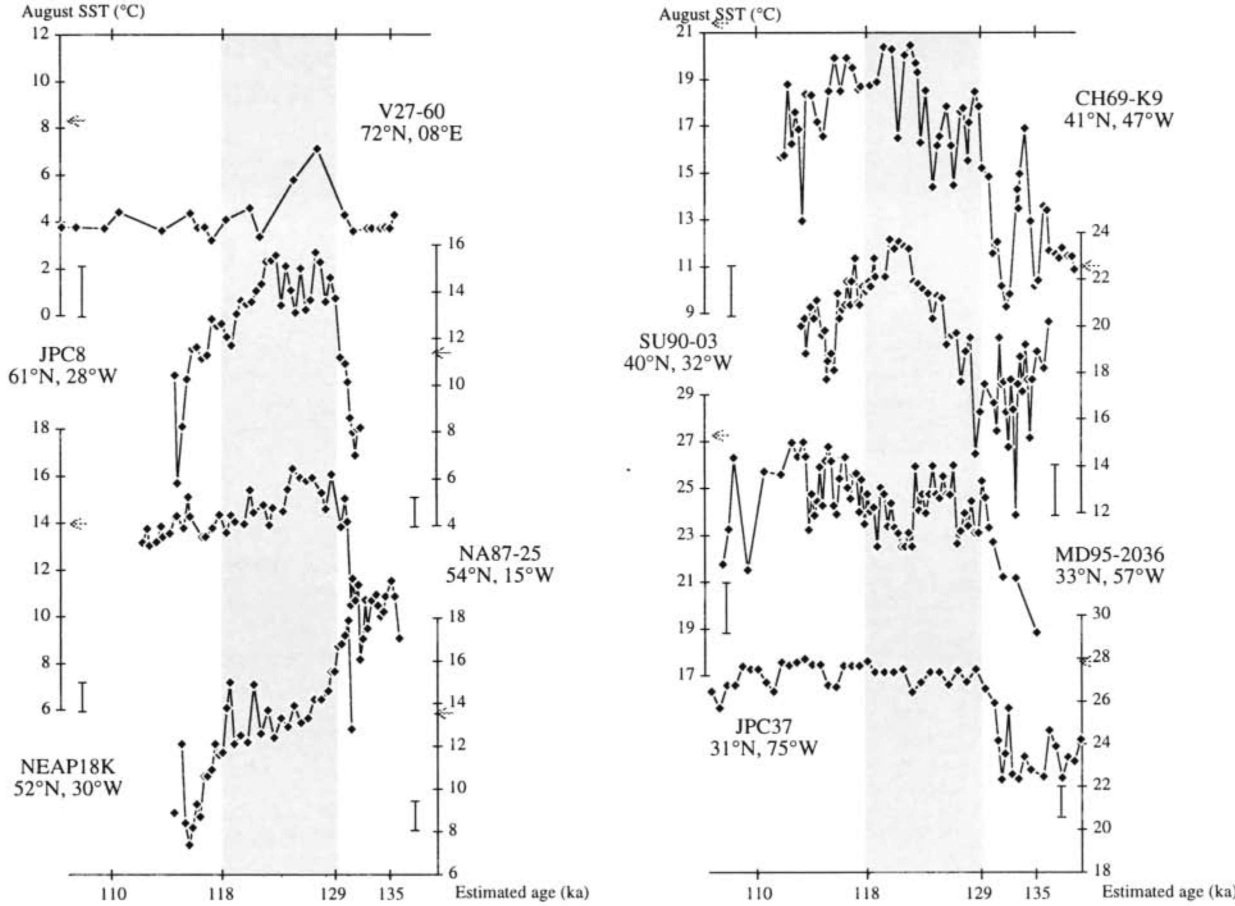

Figure 3c. Summer sea surface temperature (SST) reconstructions using the modern analog method (the arrow indicates modern summer SST [from *Levitus*, 1982] at the site of each core).

of the beginning and end of the 5e plateau significantly affect our arguments regarding climatic forcing.

4. Results

The 5e plateau, or ice volume minimum interval, is defined by a succession of minimum benthic foraminiferal $\delta^{18}O$ values with little or no trend within marine isotope substage 5e (Figure 3a). The exceptions are cores CH69-K9 and MD95-2036 where values increase by 0.2‰-0.3‰ during the same interval. Detailed benthic Cd/Ca results from core MD95-2036, which lies in the present-day mixing zone between lower North Atlantic Deep Water (NADW) and deep waters of southern source, indicate little change in the proportion of these water masses during the interval of gradually increasing benthic foraminiferal $\delta^{18}O$ at the two sites [*Adkins et al.*, 1997]. Thus one possible explanation for the changing benthic foraminiferal $\delta^{18}O$ values is the cooling of high-latitude surface waters involved in the formation of lower NADW during the interval of minimum ice volume, as seen in the planktic foraminiferal $\delta^{18}O$ and SST records of Norwegian Sea core V27-60.

The planktic foraminiferal $\delta^{18}O$ records in the different cores display a variation of ±0.5‰ during the 5e plateau, most likely because of changes in SST and SSS (Figure 3b). However, the *G. bulloides* $\delta^{18}O$ record in cores CH69-K9 and SU90-03 show a peculiar positive event, with foraminiferal $\delta^{18}O$ values increasing by 1‰, in the middle of isotopic substage 5e. The event is associated with a decrease in both estimated SSS and SST, suggesting that both changes may have resulted from the advance of the relatively cold freshwater of the Labrador Current over the site.

Reconstructed summer SSTs follow different trends, depending on the latitude of the core location (Figure 3c). Northern cores V27-60, JPC8, and NA87-25 show a decrease, primarily within the second half of the 5e plateau, during which summer SST values decline by 2°C-4°C. The southern cores CH69-K9, MD95-2036, and JPC37 generally show an opposite trend during the same interval, with summer SST values rising (within the error bar) by 0.5°C-4°C. However, in the case of core JPC37 the apparent rise does not exceed the root-mean-square error of the estimate.

SSS reconstructions for the 5e plateau display trends that are similar to those already described for summer SST (Figure 3d). The northern cores exhibit a decrease during the second half of the 5e plateau, while the southern cores exhibit an increase (or a stability) during the same period. Table 3 summarizes the calculated slope of the linear regression between the beginning and the end of the 5e plateau as a function of both depth in core and estimated age.

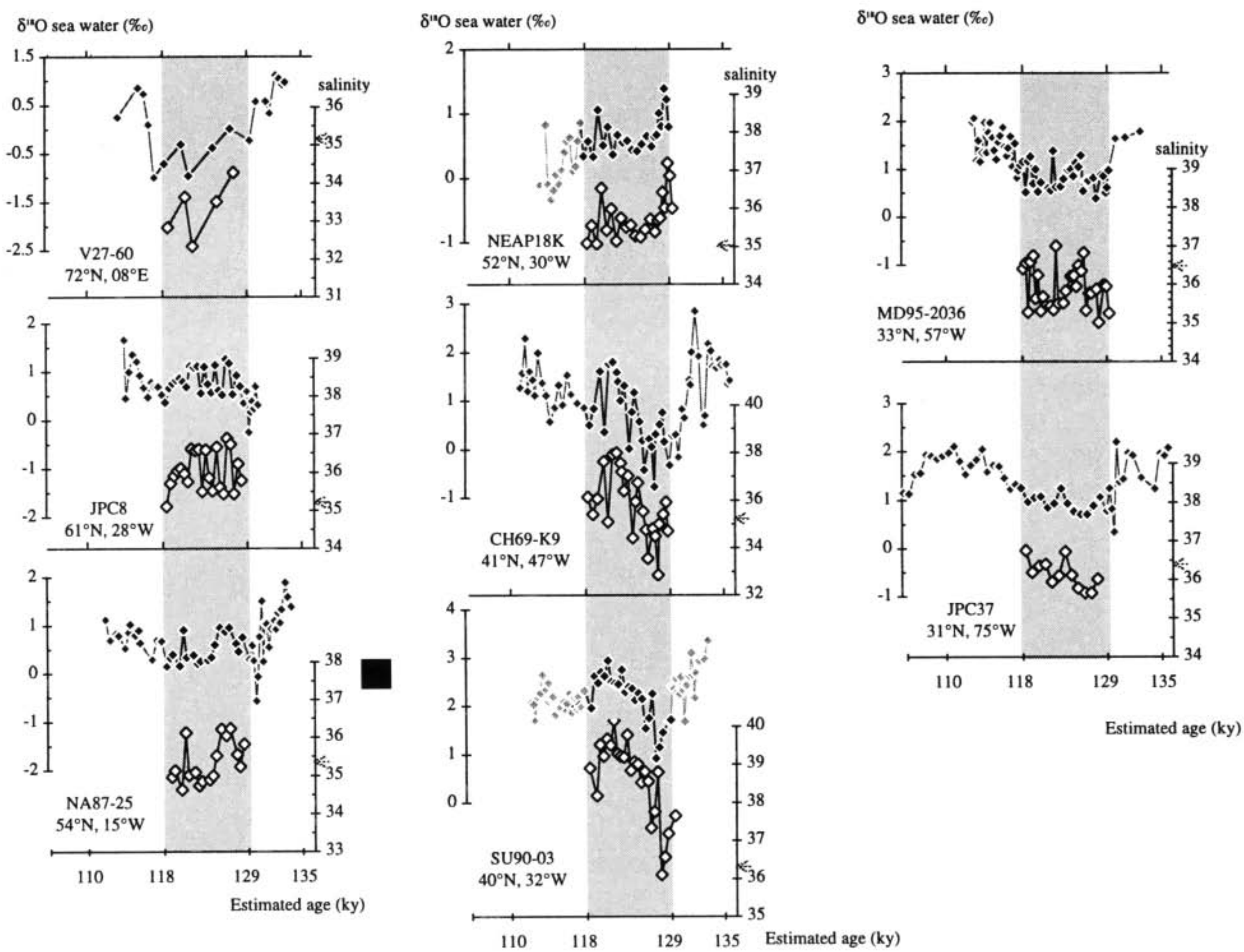

Figure 3d. Seawater $\delta^{18}O$ (solid and shaded diamonds) for the whole interval and sea surface salinity (SSS) reconstructions (open diamonds) for the 5e plateau only. Because the seawater$\delta^{18}O$ values have not been corrected by the global ice volume variations, only the 5e plateau is interpretable in terms of trends of salinity (solid diamonds). Modern SSS values [from*Levitus*, 1982] are indicated by an arrow, but because of the uncertainties in calculating paleo SSS, these values are not really comparable.

5. Discussion

Summer insolation decreased at all northern latitudes by ~30 W m^{-2} (Figure 4a) during the second part of the interval of minimum ice volume and lead eventually to renewed growth of terrestrial ice sheets [*Imbrie et al.*, 1984; *Paillard*, 1998]. However, the latitudinal distribution of these changes was strongly asymmetric and may have contributed to the latitudinal patterns of SST and SSS change documented for the 5e plateau and then ultimately led to the entrance in the glacial period [*Berger et al.*, 1992]. For example, Figure 4c shows that annually averaged insolation decreased by ~4.5 W m^{-2} at 65°N and

Table 2. Depth and Age of the Tie Points Used in the Different Cores

Core	Depth of Stage 5d (~110 kyr), cm	Depth of the End of 5e Plateau (~119 kyr), cm	Depth of the Beginning of 5e Plateau (~129 kyr), cm	Depth of Stage 6 (~135 kyr), cm
V27-60	480	525	560	680
JPC8	864	968	1160	1248
NA87-25	610.5	671.5	734	805
NEAP18K	536	604	670	>674
CH69-K9	1100	1190	1275	1485
SU90-03	421	447	472	495
MD95-2036	4301.5	4324.5	4358.5	4394.5
JPC37	1829	1857	1909	2005

The ages attributed to the beginning and the end of the 5e plateau are from *Adkins et al.* [1997].

Table 3. Estimation of the Slope Variation in the Summer Sea Surface Temperatures and Seawater $\delta^{18}O$ During the 5e Plateau

Core	Latitude	Slope of Summer SST, °C cm^{-1}	Slope of SST, °C kyr^{-1}	Slope of Seawater $\delta^{18}O$, ‰ cm^{-1}	Slope of Seawater $\delta^{18}O$, ‰ kyr^{-1}
V27-60	72°N	0.0476	-0.1517	0.0181	-0.0576
JPC8	61°N	0.0067	-0.1166	0.0006	-0.0096
NA87-25	55°N	0.0200	-0.1137	0.0060	-0.0345
NEAP18K	52°N	0.0555	-0.3331	0.0062	-0.0374
CH69-K9	41°N	-0.0454	0.3512	-0.0163	0.1259
SU90-03	40°N	-0.3136	0.7127	-0.0507	0.1152
MD95-2036	33°N	-0.0017	0.0056	-0.0082	0.0260
JPC37	31°N	-0.0140	0.0663	-0.0059	0.0279

The values are calculated between the beginning and the end of the 5e plateau, versus the depth in core JPC8 and versus time. A positive slope value versus depth indicates a decreasing trend in temperature or seawater $\delta^{18}O$ record and corresponds to a negative slope versus time, while a negative slope versus depth indicates an increasing trend in temperature or seawater $\delta^{18}O$ record and corresponds to a positive slope versus time.

increased by ~2 W m^{-2} at 10°N during the interval of minimum ice volume. Although the annually averaged changes are small, they may produce significant heating or cooling at the sea surface because the upper ocean integrates direct insolation forcing over several years. The SST response at high latitude may be amplified further by changes in the seasonal extent of sea ice and associated albedo feedbacks. For example, summer insolation receipts reached maximum values in the middle of the interval of minimum ice volume but then declined rapidly. If sea ice cover became more extensive in summer as a result of the decline in summer radiation, less energy would be available for surface heating. This, together with the larger change in annually averaged insolation at 65°N, may explain why the decrease in SST at high latitudes was considerably greater than the associated increase at low latitudes (Table 3). Because of the strong dependence of evaporation rate on temperature (~5% °C^{-1}), the same mechanism may also explain the associated (positively correlated) changes in estimated SSS. The cores we studied are located along the Gulf Stream system, which is the major energy supplier to the northern latitudes. Changes in the location of the core will result in a less clear signal, like in core MD95-2036, which is just on the boundary of the main current.

The same forcings and feedbacks may have also operated during the Holocene (cf. Figure 4d), though with less amplitude as the orbital configuration is different and then the insolation changes are smaller than during the Eemian. For example, an SST reconstruction for 6 kyr B.P. indicates that the high latitudes in the North Atlantic were slightly warmer than today, while lower latitudes were cooler [*Ruddiman and Mix*, 1993], although the differences from the present climate are almost always within the error estimates. A similar SST pattern has been produced in simulations using atmospheric general circulation models with ocean heat fluxes fixed at control values and forced by 9 and 6 kyr B.P. insolation (*Liao et al.* [1994] and *Mitchell et al.* [1988], respectively). These patterns may have been reinforced by changes in equatorial upwelling. Other Holocene proxies, like the gray-scale reflectance from Cariaco basin cores [*Hughen et al.*, 1996], are also in favor of stronger equatorial upwellings and therefore cooler tropics at the beginning of the Holocene, leading to a progressive warming of the tropics as the winter insolation increased between 10 and 0 kyr B.P. (Figure 4d). Thus both data and models suggest that a slight cooling over the ocean at high latitudes and a slight warming at lower latitudes may have developed during the course of the Holocene. Our data suggest this was also the case for the Eemian but most likely with a larger amplitude because of the larger amplitude of the insolation changes during the last interglaciation.

We also note that the SST and SSS patterns we have reconstructed for the Eemian ice volume minimum period are characteristic of those associated with weakening of the large-scale overturning thermohaline circulation in numerical models. The replacement of newly formed and exported NADW is associated with a divergence of heat at the surface of the low-latitude North Atlantic and a corresponding convergence in high latitudes. If the rates of NADW formation and large-scale overturning are reduced, the ocean heat transport is relaxed, leading to cooling at high latitudes and warming in low latitudes. There are also pronounced feedbacks in the hydrologic cycle and in surface salinity. The dominant feedback is one of freshening at high latitudes and salt build up at low latitudes, due simply to the increasing residence time of surface waters in latitudes of net precipitation (>40°N) and evaporation (<40°N) [*Bryan*, 1986]. A long-term weakening of the overturning circulation may have been set in motion by insolation-induced freshening at high latitudes, as discussed above. This thermohaline feedback would serve to reinforce the documented latitudinal trends in SST and SSS. However, it must be pointed out that available proxy records of NADW show no clear evidence for a gradual decrease in strength during the interval of minimum ice volume within substage 5e. Low-resolution studies tend to show that the deep water circulation was always active during the isotopic substage 5e, but the resolution of these studies are not precise enough to follow the behavior of the system during the 5e plateau [*Duplessy and Shackleton*, 1985]. There are still too few high-resolution

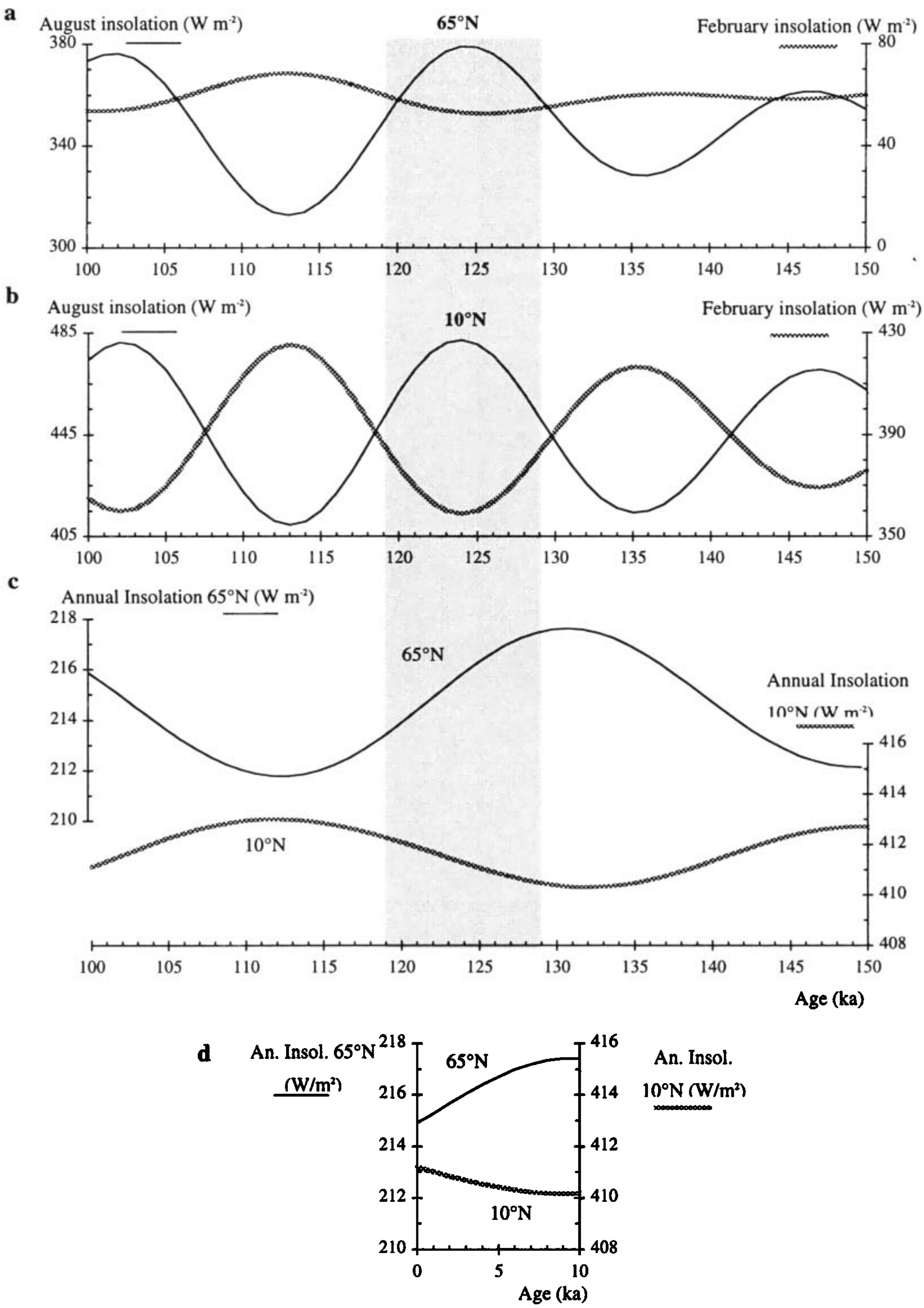

Figure 4. February (February 1-28) and August (August 1-30) insolation variations at (a) 65°N and (b) 10°N between 150 and 100 kyr. Annual (January 1 to December 30) insolation variations between (c) 150 and 100 kyr and (d) 10 and 0 kyr. The shaded area give the boundaries of the 5e plateau as defined by the *Adkins et al.* [1997] age scale.

proxy records (like signals of *C. wuellerstorfi* $\delta^{18}O$ and $\delta^{13}C$) to follow in great detail the paleocirculation through the 5e plateau. The gradual insolation-induced changes in the SST and SSS fields may have lead ultimately to a dramatic and sudden weakening of the overturning circulation, such as has been documented at the end of the 5e plateau in several recent studies [*Adkins et al.*, 1997; *Oppo et al.*, 1997].

6. Conclusion

We have estimated temperatures and salinity for the last interglacial period on a meridional transect of the North Atlantic from 31° to 72°N. During the minimal ice volume time interval the northern temperatures and salinities show a decreasing trend, while the southern ones increase. These changes are mainly gradual but appear to be rather rapid in the middle of substage 5e, at least in some locations. Such changes in sea surface hydrology are somewhat similar to the changes already observed during the Holocene. They may be linked to changes in insolation and in atmospheric circulation. They are also consistent with a reduction of the North Atlantic thermohaline circulation, though more high-resolution deep water records would be required to reach such a conclusion. More ocean-atmosphere coupled model experiments would be necessary to evaluate the real impact of the annual insolation changes on the glaciation incept.

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