Holocene Climate Variability on Centennial-to-Millennial Time Scales: 1. Climate Records from the North-Atlantic Realm

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Abstract: Holocene oxygen isotope data from the GISP2 ice core reveal temperature oscillations in Greenland with a periodicity of ~900 y, which can be correlated to climate perturbations in northern and central Europe. We suggest that the 900-y climate fluctuations are generated within the climate system, and are probably triggered by negative salinity anomalies in the North Atlantic. A simple template is used to show that two such triggering events centered at ~8.3 and 4.7 ky BP are required to explain temporal evolution of 900-y climate cycles between ~3.5–8.5 ky BP as sequence of damped oscillations. Although pacing of the 900-y cycles by changes in the Earth's orbit cannot be ruled out, we regard this scenario as unlikely. We show that the existing paleoceanographic evidence for ~1400–1500-y climate oscillations during the Holocene is questionable. Instead we suggest that deep-sea records from the North Atlantic may be reconciled with 900-y climate oscillations during this period.

Introduction: The Seemingly Stable Holocene

Recent results from ice-cores, drilled at the summit of the Greenland ice sheet, boosted paleoclimate research since they provide clear evidence for recurrent and abrupt climatic changes during the last glacial period. In particular oxygen isotope ($\delta^{18}O$) data, which are assumed to monitor air temperature, reveal that large and rapid temperature fluctuations (warming by up to 10°C within a few decades) dominated climate in Greenland between ~11–74 thousand years before present (ky BP; all ages are reported as calendar years) (Dansgaard et al. 1993; Grootes and Stuiver 1997; Fig. 1). Grootes and Stuiver (1997) demonstrated that warm peaks (so called Dansgaard-Oeschger interstadials) occurred with a fundamental period of 1470 years (y). The significance of this millennial-scale climate variability is corroborated by similar climate variations on a global scale which can be correlated to the Dansgaard-Oeschger interstadials (e.g. Sirocko et al. 1993; Behl and Kennett 1996; Schulz et al. 1998; Sarnthein et al. 2001). Accordingly, the ice-core $\delta^{18}O$ series at least recorded climate in the North Atlantic region (Grootes and Stuiver 1997) and to some extent global climate variability (Blunier et al. 1998) during the last glacial period.

A salient feature of the Greenland $\delta^{18}O$ record is its relative small variation during the Holocene, compared to the preceding time interval, implying that no large temperature variations occurred in Greenland during the last 10 ky (Fig. 1). However, this finding contrasts with data from various Euro-

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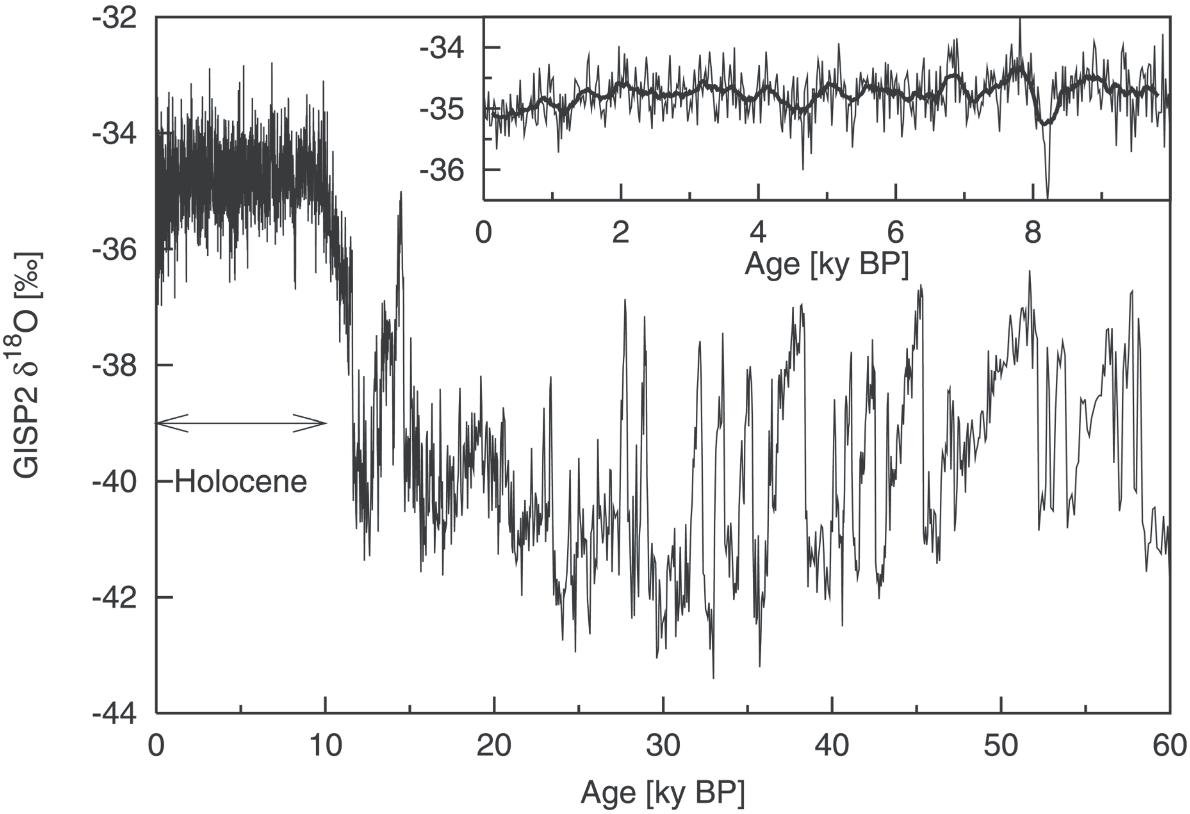

Fig. 1. Oxygen isotope ($\delta^{18}O$) time series from Greenland (GISP2 ice core; Grootes and Stuiver 1997). The inset shows expanded Holocene portion (thin line) and 300-y running mean (bold line).

pean paleoclimatic archives, which reveal marked climate variations during the Holocene, as indicated for example by fluctuations of glaciers and tree-lines (e.g. Karlén 1993; Dahl and Nesje 1996) as well as net precipitation (Harrison and Digerfeldt 1993; Schettler et al. 1999). Although the dating of these events is not without problems, it appears that European climate varied on multicenntennial-to-millennial time scales during the Holocene. These natural climate fluctuations were probably strong enough to affect early human societies in Europe. Insights into the temporal evolution of Holocene multicentennial-to-millennial scale climate variability are not only important for interpreting archeological data, but also because of the potential interference with today's man-made climate perturbation.

Moreover, recent paleoceanographic proxy data from the North Atlantic (Bond et al. 1997; Bianchi and McCave 1999) suggest millennial-scale climate variability with a periodicity at ~1400–1500 y during the Holocene, implying that Dansgaard-Oeschger events, although attenuated, may have occurred during the last 10 ky, too. The lack of similar events in the $\delta^{18}O$ data from Greenland (Fig. 1, inset) raises the question if the ice-core data can be regarded as climate indicator for the *entire* North-Atlantic realm during the Holocene.

A Holocene 900-y Temperature Cycle in Greenland

Despite its random appearance, the Holocene bidecadal $\delta^{18}O$ record from the Greenland Ice Sheet Project Two (GISP2; Grootes and Stuiver 1997) contains a distinct low-frequency component, which becomes clearly visible after applying a 300-y running mean filter to the data (Figs. 1, 2). The running mean varies between approximately

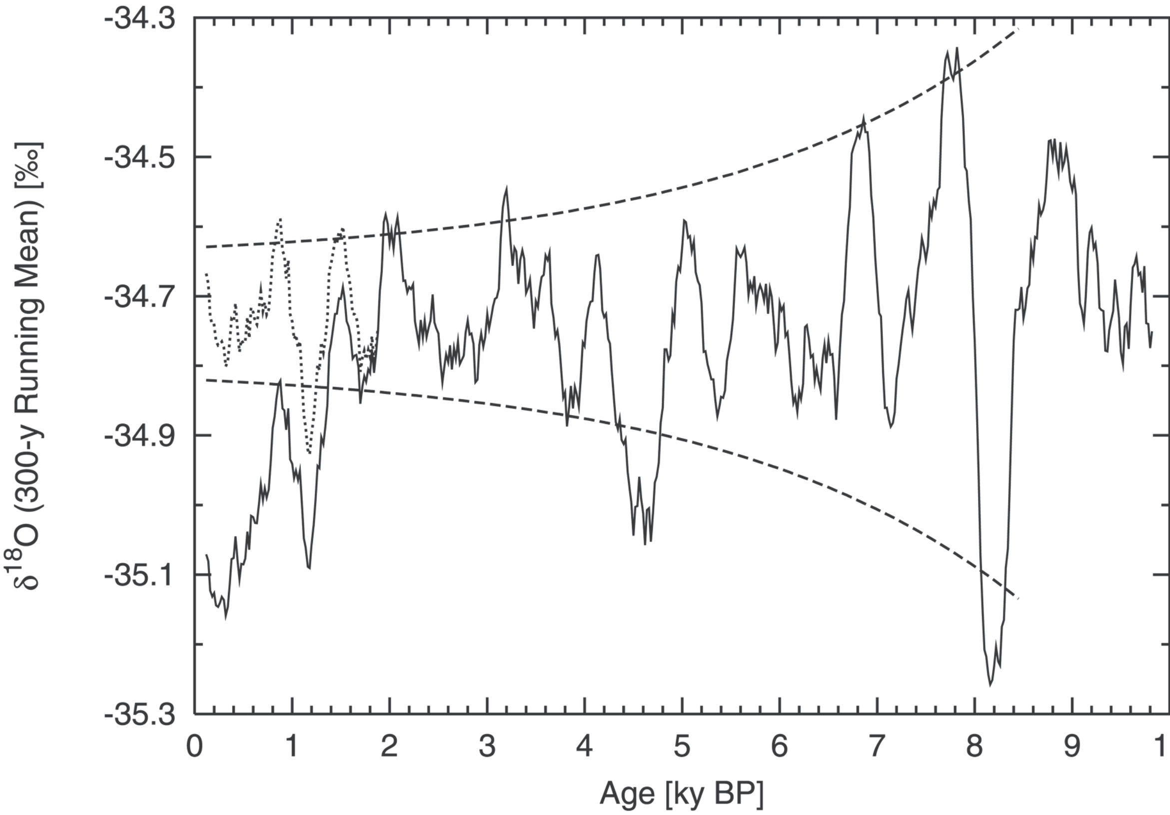

Fig. 2. Smoothed $\delta^{18}O$ time series (300-y running mean [solid line]; as in Fig. 1, inset). For subsequent analyses, linear trend between 0-2 ky BP was subtracted (dotted line). Exponential envelopes (dashed lines) depict general temporal evolution of $\delta^{18}O$ minima and maxima.

± 0.5‰ to ± 0.1‰ around its average value (Fig. 2). It appears that the amplitude of the running-mean fluctuations decreases exponentionally from a maximum around 8.2 ky BP. This tendency is, however, somewhat obstructed by a trend in the youngest part of the record (<2 ky BP). Most extreme $\delta^{18}O$ minima occur approximately every 3.4 ky (at 1.2, 4.7 and 8.2 ky BP; Fig. 2). Between 3–8.5 ky BP the fluctuations of the running mean appear to be almost periodic, with a recurrence time of ~900 y. Harmonic analysis of the unsmoothed $\delta^{18}O$ data (Fig. 3) indicates that the oscillations are indeed contained in the time series and are not an artefact of the running-mean filter (cf. similar periodicity in Grootes and Stuiver 1997).

To test the robustness of the above inference (i.e. a damped ~900 y oscillation), one can use the $\delta^{18}O$ record from the North GRIP ice core (Hammer this volume). Between 5–9 ky BP the temporal evolution of the smoothed $\delta^{18}O$ time series from North GRIP (not shown) is similar to the GISP2 data (Fig. 2). Therefore, the North GRIP data corroborate the existence of damped ~900-y temperature oscillations in Greenland in the early Holocene. However, the North GRIP record shows no evidence for a cooling at ~4.7 ky BP. Instead a trend towards lighter $\delta^{18}O$ values dominates the last 5 ky in this data set. Possible causes for this unexpected discrepancy between the two ice-core records remain open for debate (cf. Hammer, this volume).

With a $\delta^{18}O$-surface temperature gradient of 0.5‰ per °C (Jouzel et al. 1997 and refs. therein), the estimated changes in the running mean imply temperature fluctuations, ranging from ±1°C in the early Holocene to only ±0.2 °C in the youngest part of the record. Compared with temperature variations associated with glacial Dansgaard-Oeschger

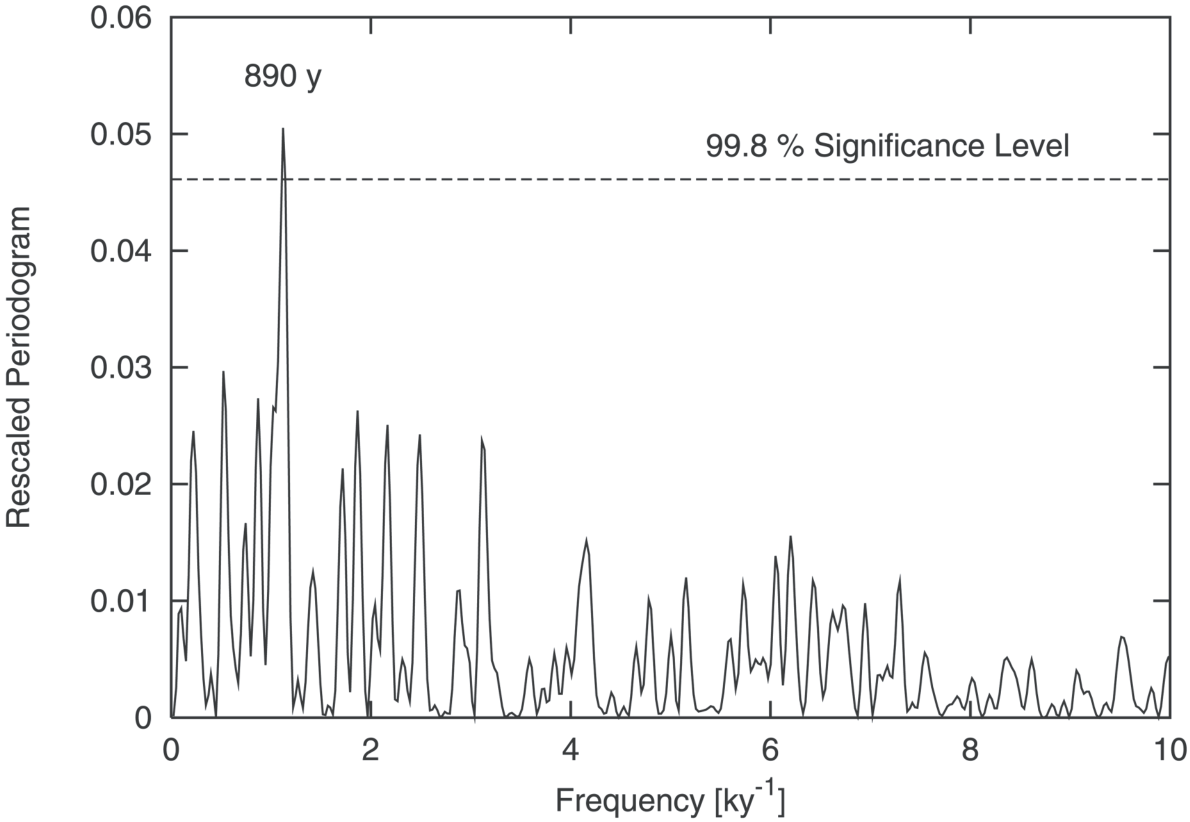

Fig. 3. Harmonic analysis (Fisher 1929; Schulz and Stattegger 1997) of unsmoothed $\delta^{18}O$ time series between 0-10 ky BP. The time series can be distinguished from white noise at the (1-1/N) × 100% = 99.8% significance level (cf. Thomson 1990; N = number of data points). The sharp peak at periodicity of 890 y is highly significant and indicates the presence of a harmonic signal component.

interstadials (up to ~10°C; cf. Fig. 1), the multicentennial temperature perturbations during the Holocene are rather small. Based on the premise that $\delta^{18}O$ fluctuations preserved in Greenland icecores are at least significant for the North Atlantic region (Grootes and Stuiver 1997), one has to ask if these seemingly small temperature oscillations are relevant to climate variations in Europe.

Climate Variations in Northern and Central Europe

Dahl and Nesje (1996) estimated Holocene summer temperature in south-central Scandinavia based on variations of the altitude limiting pine-tree growth (Fig. 4). Furthermore, these authors reconstructed changes of the equilibrium-line altitude (ELA) of glaciers in southern Norway (Fig. 4) during the Holocene. ELA variations provide a measure for shifts in net glacier growth.

A comparison of the Scandinavian summer temperature with the smoothed $\delta^{18}O$ record from Greenland (Fig. 4) reveals a good correspondence between extremes of these two climate proxies. (It should be noted that the stratigraphies of the records are *not* tuned to each other.) Before 6.6 ky BP cold events in Greenland (negative $\delta^{18}O$ anomalies) coincide with cold summers in Scandinavia, whereas between 4–6.6 ky BP cold conditions in Greenland correlate with warm summers in northern Europe. This distinct change of the phase relationship between the temperature re-

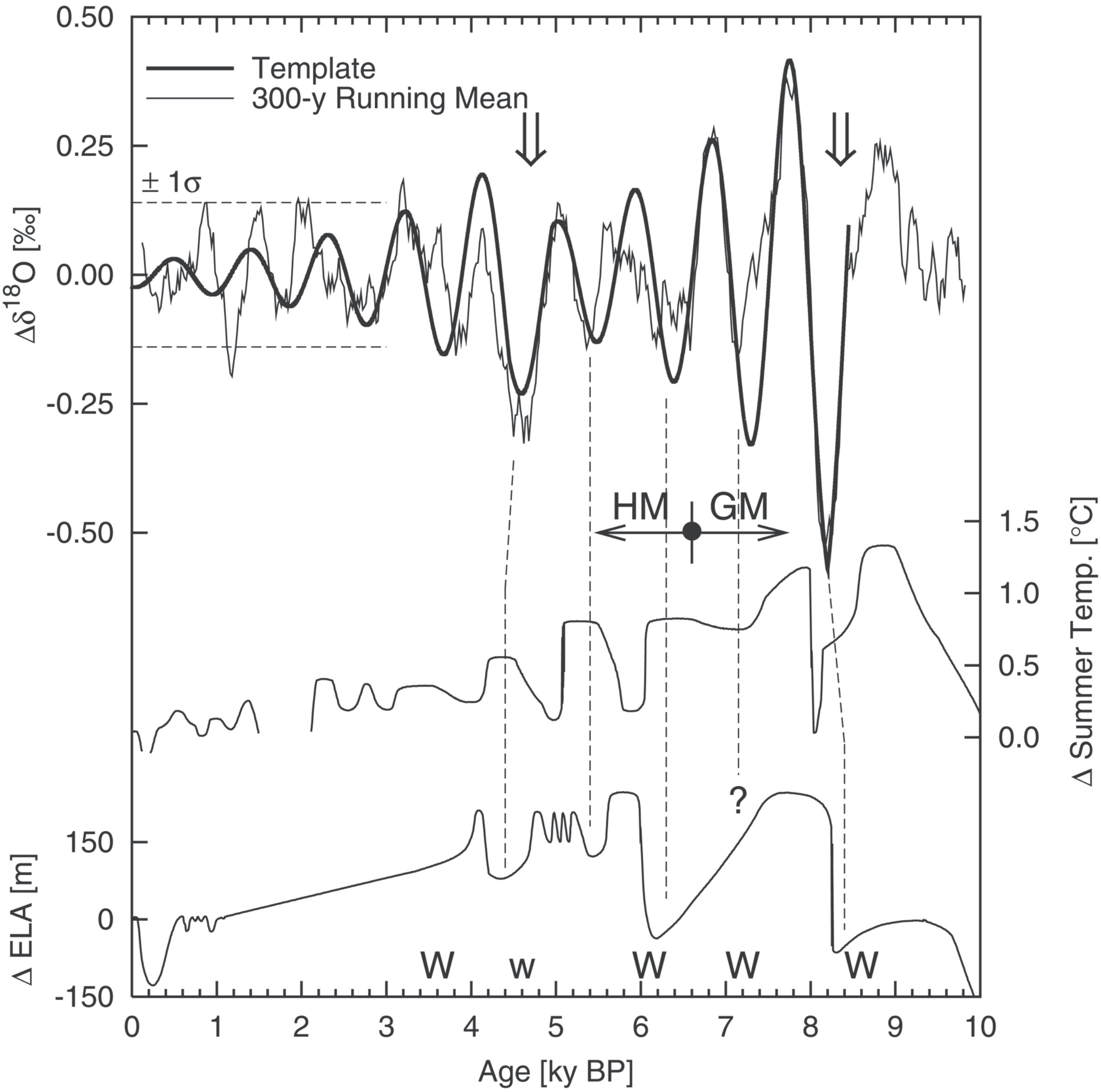

Fig. 4. Top: Smoothed $\delta^{18}O$ record (as in Fig. 2, but after subtraction of mean value) and output of the template. Arrows indicate forcing of the template by cooling phases. Dashed horizontal lines between 0-3 ky BP depict uncertainty of $\delta^{18}O$ data (1σ confidence interval). Center: Deviation of summer temperature from modern value in Scandinavia (Dahl and Nesje 1996). Bottom: Departure of equilibrium line altitude from modern position in Scandinavia (Dahl and Nesje 1996). Negative values indicate glacier advance. 'W' denote wet periods in central Europe (Schettler et al. 1999). Dashed vertical lines depict inferred correlation between climate records. The inferred transition from glacial (GM) to Holocene mode (HM) is indicated. See text for further details.

cords around 6.6 ky BP is, however, not seen between $\delta^{18}O$ and ELA. Instead, glacier advances (ELA minima) go along with cold temperatures in Greenland between 4–8.5 ky BP. Moreover, between ~3.6–8.6 ky BP cold spells in Greenland occur together with periods of enhanced rainfall in central Germany (Schettler et al. 1999), except for the cold period at 5.4 ky BP (Fig. 4).

For ages younger than ~3.6 ky BP, climate indicators from Europe and Greenland show no clear match on a multicentennial time scale. Since the amplitudes of the climate perturbations are smaller in the late Holocene than during the preceding time interval, the lack of correspondence between the records may in part be caused by a reduced signal-to-noise ratio.

The Icelandic Low: Climate Interface Across the North Atlantic

Observations, the earliest going back to the second half of the eighteenth century, indicate that mild (severe) winters in western Greenland go along with severe (mild) winter conditions in northern Europe (Barlow et al. 1997 and refs. therein). This temperature „seesaw“ across the North Atlantic is closely linked to the variability of the Icelandic Low. Warm winters in Europe coincide with a strong Icelandic Low, located between Iceland and Greenland (Fig. 5b). This pressure distribution steers warm and humid air masses from mid-latitudes eastward into northern and central Europe, whereas polar air masses reach Greenland. In contrast, when the Icelandic Low is weak and located near the southern tip of Greenland (Fig. 5a) the flow of cold polar air to Greenland is reduced, whereas more cold polar air reaches Europe.

The anti-phase relationship of temperature anomalies between Greenland and Scandinavia between 4–6.6 ky BP (Fig. 4) can thus be explained in terms of the temperature „seesaw“ mediated by the Icelandic Low. This inference is corroborated by the simultaneous occurrence of warm and humid periods in Europe; the latter is thought to lower the ELA by increased snowfall (Dahl and Nesje 1996). Accordingly, we surmise that the 900-y climate oscillations between 4–6.6 ky BP may be linked to multicentennial variations in strength and location of the Icelandic Low.

Today, extremes of the temperature „seesaw“ between Greenland and northern Europe are linked to positive and negative phases of the decadal North Atlantic Oscillation (e.g. Kushnir 1999; Fig. 5). Accordingly, minima and maxima of the 900-y temperature cycles may be an expression of a modulation of North Atlantic Oscillation intensity, once the modern atmospheric mode was established at 6.6 ky BP. This conjecture is supported by driftwood records from the Arctic (Dyke et al. 1997), which are thought to indicate shifts in the mean state of the North Atlantic Oscillation (Tremblay et al. 1997).

A different picture emerges for times prior to 6.6 ky BP when the temperature „seesaw“ was apparently inoperative, resulting in synchronous temperature variations in Greenland and Europe. This situation is reminiscent of a glacial climate mode, with the Icelandic Low located further south (e.g. Harrison et al. 1992), allowing an almost unrestricted flow of cold polar air into the entire North-Atlantic realm. Based on the phase relationship between temperature variations in Greenland and Europe, we can only speculate that the glacial climate mode operated in this region until 6.6 ky BP. The transition to today's climate mode at ~6.6 ky BP is consistent with a marked change in the prevailing wind direction from northeast to west in south-western Sweden between 6–7 ky BP (Mörner 1980). However, this timing of the mode transition contrasts results of Stager and Mayewski (1997), who suggested a global transition from glacial to postglacial mode between 7.8–8.2 ky BP, that is, simultaneous to the final decay of the North American ice sheet (Licciardi et al. 1998).

Making 900-y Climate Oscillations: Internal Feedbacks *vs.* Orbital Control

The existence of 900-y cycles in climate records does not provide direct information about the origin of these cycles. We will consider two mechanisms, which are not mutually exclusive: (1) Internal oscillations of the climate system and (2) external forcing by secular changes of the Earth's orbital parameters.

Internal origin: Based on the finding that the amplitude of the 900-y temperature oscillations decays almost exponentially (Fig. 2), one can build simple mathematical templates to mimic this behavior. We use the term template in order to distinguish the following excercise from modeling efforts which treat the involved physical processes explicitly. One possibility for such a template is to express

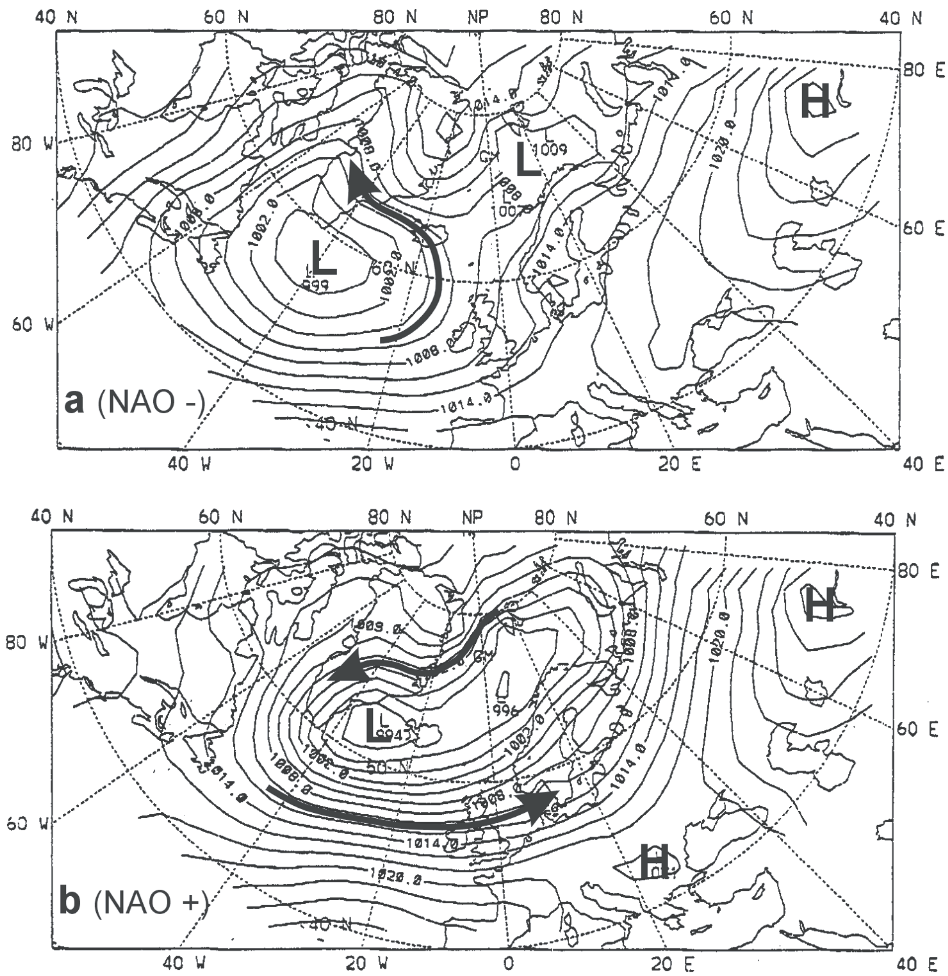

Fig. 5. Mean sea-level pressure in the North Atlantic realm for opposing modes of the temperature „seesaw“ (modified from Barlow et al., 1997). **(a)** Southwesterly position of the Icelandic Low causes mild Greenland winters and severe temperatures in Europe and corresponds to the negative phase of the North Atlantic Oscillation (NAO). **(b)** Opposite situation with cold Greenland winters (positive NAO phase). Main centers of high (H) and low (L) pressure are indicated.

the temperature anomaly in Greenland (δT in units of $\delta^{18}O$) as function of the average value of δT during the preceding τ years, denoted by $\langle \delta T \rangle_\tau$:

$$\frac{d\delta T}{dt} = a \langle \delta T \rangle_\tau + c(t), \qquad (1)$$

with t being time, a = -0.009 ‰/y, τ = 415 y and

$$c(t)\ [‰/y] \begin{cases} -0.0025 \text{ between 8.20-8.45 ky BP} \\ -0.0007 \text{ between 4.45-4.95 ky BP} \\ 0.0 \text{ else} \end{cases}$$

With a less than zero, the „memory“ of the system (provided by $\langle \delta T \rangle_\tau$) determines the current evolution of δT; a positive (negative) value of the average temperature anomaly during the past 415 y will cause a decrease (increase) of the current temperature. The combination of $a < 0$ and „memory“ for past states results in a delayed negative feedback, which in principle allows the template in eq. (1) to produce temperature oscillations. However, without an initial perturbation the template cannot produce any deviations from its equilibrium ($\delta T = 0$). Such external forcing is therefore included into eq. (1) by the time-dependent parameter c; setting $c < 0$ for some time interval results in a cooling event. The observation that the $\delta^{18}O$ anomalies go beyond the lower exponential envelope at ~ 4.7 and ~8.2 ky BP (Fig. 2) suggests cooling events at these times. Actual values for $c(t)$ and a were found after some numerical experiments by matching the template to the observed $\delta^{18}O$ anomalies („fit by eye").

The template agrees well with the observed $\delta^{18}O$ anomalies between 3–8.5 ky BP (Fig. 4). Between 0–3 ky BP, the match degrades, possibly due to the low signal-to-noise ratio of the data. The template describes the 900-y cycles as damped oscillations, triggered by two cooling events at ~8.3 and 4.7 ky BP, with the magnitude of the first cooling event being ~4-times larger than the mid-Holocene cooling. The periodicity of the predicted temperature cycles is set by the length of the system „memory" (τ). To assess the appropriateness of the template to real world processes, it is important to identify the possible origin of the delayed negative feedback, in particular, to locate the „memory" of the system.

Today, the large-scale circulation of the Atlantic Ocean is dominated by deep-water formation in the northern North Atlantic, where surface water gains density, through heat and freshwater exchange with atmosphere, and sinks to greater depth. The southward flow of newly formed deep water is compensated by a northward return flow of near-surface and intermediate water masses (e.g. Schmitz 1995). Experiments with ocean general circulation models (England 1995) indicate that ventilation ages (i.e. the time since the last contact of a water parcel with the atmosphere) of intermediate and deep waters in the Atlantic Ocean measures up to several hundred years. Hence, it should be possible to propagate a hydrographic anomaly from the sea surface near the deep-water formation sites in the North Atlantic southward, with the deep-water flow. Subsequent upwelling of the anomaly and its entrainment into northward flowing water masses, which feed deep-water formation, returns the (attenuated) anomaly to its origin. The reappearance of the anomaly at its source after several hundred years constitutes a delayed feedback loop. A smorgasbord of processes within the feedback loop can be envisioned to produce an overall negative feedback. Careful modeling studies are required to assess possible delayed negative feedback loops within the Atlantic thermohaline circulation (cf. Paul and Schulz, this volume).

The timing of the inferred cooling events in the template coincide with negative salinity anomalies in the North Atlantic Ocean, suggesting a potential physical mechanism: Deep-water formation in the North Atlantic results in the transport of heat from low to high latitudes (e.g. Peixoto and Oort 1992). Modeling studies (e.g. Bryan 1986) suggest that freshening of surface-waters in the North Atlantic could attenuate or even suppress deep-water formation, and as a consequence, reduce northward heat transport. Since the final disintegration of the North American ice sheet delivered meltwater into the Labrador Sea around 8.2 ky BP (Barber et al. 1999), it is likely that this freshwater influx suppressed deep-water formation in this area and thereby caused the inferred cooling in the North-Atlantic realm. However, the lack of deglacial meltwater influx makes this scenario inappropriate for the mid-Holocene cooling at ~4.7 ky BP. A change in the freshwater budget of the North Atlantic seems nevertheless a reasonable assumption, since a major decrease in sea-surface salinity occurred at ~5 ky BP in the northeastern Atlantic Ocean (Duplessey et al. 1992). Since the cause of this salinity perturbation is not yet known, we can only surmise that it may have weakened deep-water formation in the North Atlantic and induced the mid-Holocene cooling event in the North-Atlantic realm. This inference is supported by benthic isotope data which suggest indeed a weakening of deep-water production at this time (Maslin et al. 1996; Knaack 1997).

Orbital forcing: Loutre et al. (1992) identified a ~900-y periodicity in eccentricity variations of the Earth's orbit, which modulates incoming solar radiation. Accordingly, it might be possible that the observed 900-y climate fluctuations are linked to this external forcing. The amplitude of the ~900-y component of summer insolation at 65°N is less than 1 mW m^{-2} (Fig. 6), that is, two orders of magnitude smaller than the estimated radiative forcing resulting from solar output variability

(~300 mW m^{-2}; Schimel et al. 1996). The effect of the latter on Earth's climate is probably rather small (Schimel et al. 1996), making it unlikely that the extremely small 900-y insolation variations should affect climate directly. Nevertheless, one could envision that internal oscillations of the climate system are paced by the orbital variations, resulting in a phase-lock between forcing and climate variations.

Between ~3.5–8 ky BP the 900-y component of northern summer insolation shows indeed a good correlation with the inferred long-term temperature variations in Greenland (Fig. 6). However, close inspection of the phase relationship between the two time series reveals that the phasing varies in time: While there is no discernable phase-lag between 7.6–8.4 ky BP, warming trends seen in the $\delta^{18}O$ running mean show an increasing delay to changes in insolation from ~0 y at 8 ky BP to 300 y at 4.4 ky BP. This suggests that the frequency of the insolation variation slightly exceeds the frequency of the climate oscillations; a finding which is difficult to reconcile with a phase-lock between the two signals. Dating uncertainties of the ice-core $\delta^{18}O$ record amount to 1–2% during the Holocene (Meese et al. 1997) and are thus too small to account for the increasing phase lag. Furthermore, between 1–3.5 ky BP and prior to 8.4 ky BP we observe more mismatches than matches between peaks of the insolation and $\delta^{18}O$ curve. Finally, we note that the exponential envelope of the $\delta^{18}O$ data (Fig. 2) contrasts the more linear decrease of the amplitude of the 900-y insolation anomaly (Fig. 6). Although we cannot rule out orbital pacing of the 900-y climate oscillations, the above findings neither support direct orbital forcing nor a simple phase-locked behavior. Whether the good correlation between insolation forcing and climate

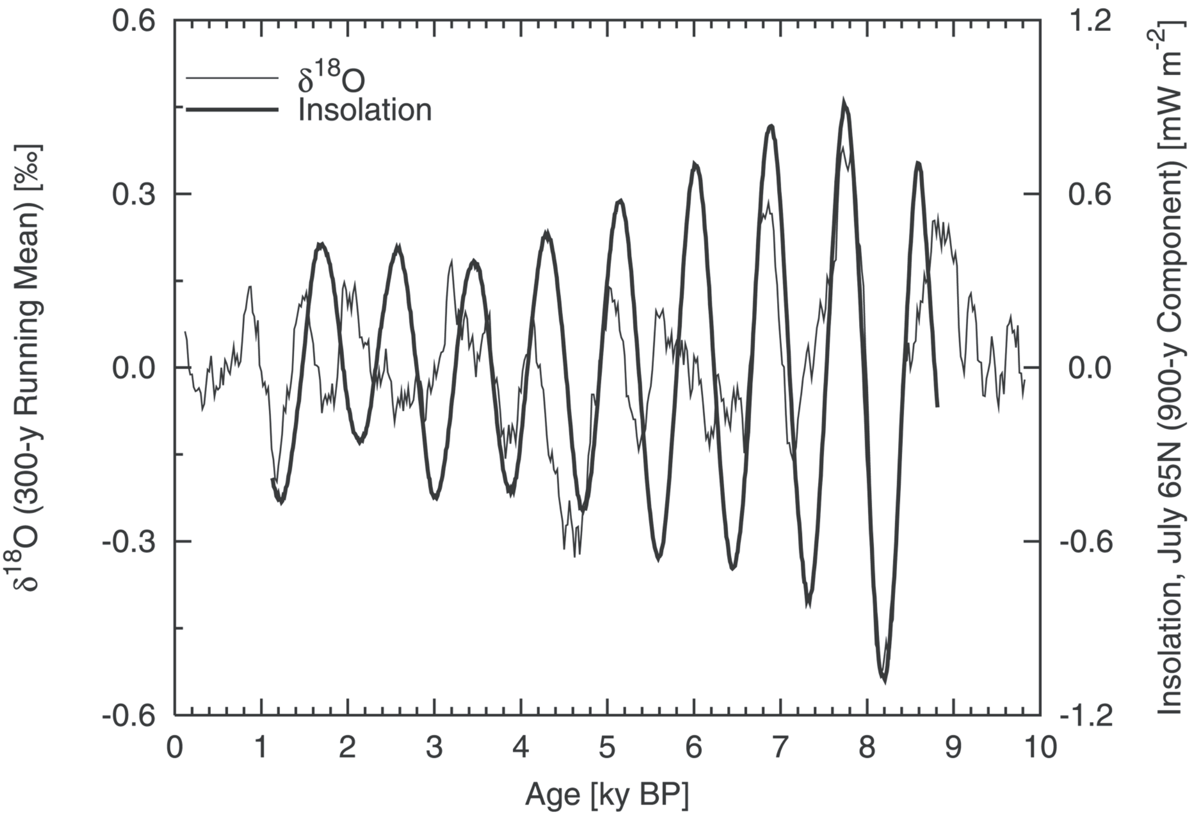

Fig. 6. Smoothed $\delta^{18}O$ time series (as in Fig. 4) together with 900-y signal component in summer insolation at 65°N (Loutre et al. 1992), which was estimated by a harmonic-filtering algorithm (after Ferraz-Mello 1981). Prior to filtering a polynomial trend was subtracted from the insolation time series.

change between 7.6–8.4 ky BP points to a causal mechanism or occurred by happenstance remains an open question.

The Deep-Sea Record of Holocene Climate Oscillations

The deep-sea record of continental rock fragments transported by icebergs are thought to be a sensitive indicator of past climate change. Bond et al. (1997) identified periods in the Holocene during which the input of ice-rafted debris (IRD) from Greenland and Iceland to a location west of Ireland increased (Fig. 7). Bond and coworkers consider ocean-surface coolings as immediate cause of these IRD events, which occur with a recurrence time of 1370 ± 500 y (1-σ confidence interval) during the Holocene. We note that the large uncertainty of the estimated recurrence time does not exclude a 900-y pacing. Considering chronostratigraphic uncertainties, the average sampling interval of the two IRD time series (~220 y in the Holocene), possible smoothing due to bioturbation and a non-linear response of IRD abundance to climatic cooling in the North-Atlantic realm, the IRD events may be tentatively correlated to $\delta^{18}O$ minima in the ice-core record, that is, to air-temperature minima in Greenland (Fig. 7). Given the above limitations of the IRD records, it cannot be ruled out that the 900-y climate cyclicity, observed in the atmospheric records, is also contained in the IRD time series but was conceived by Bond et al. (1997) as longer cycle. Hence, the important question regarding the cyclicity of North Atlantic IRD

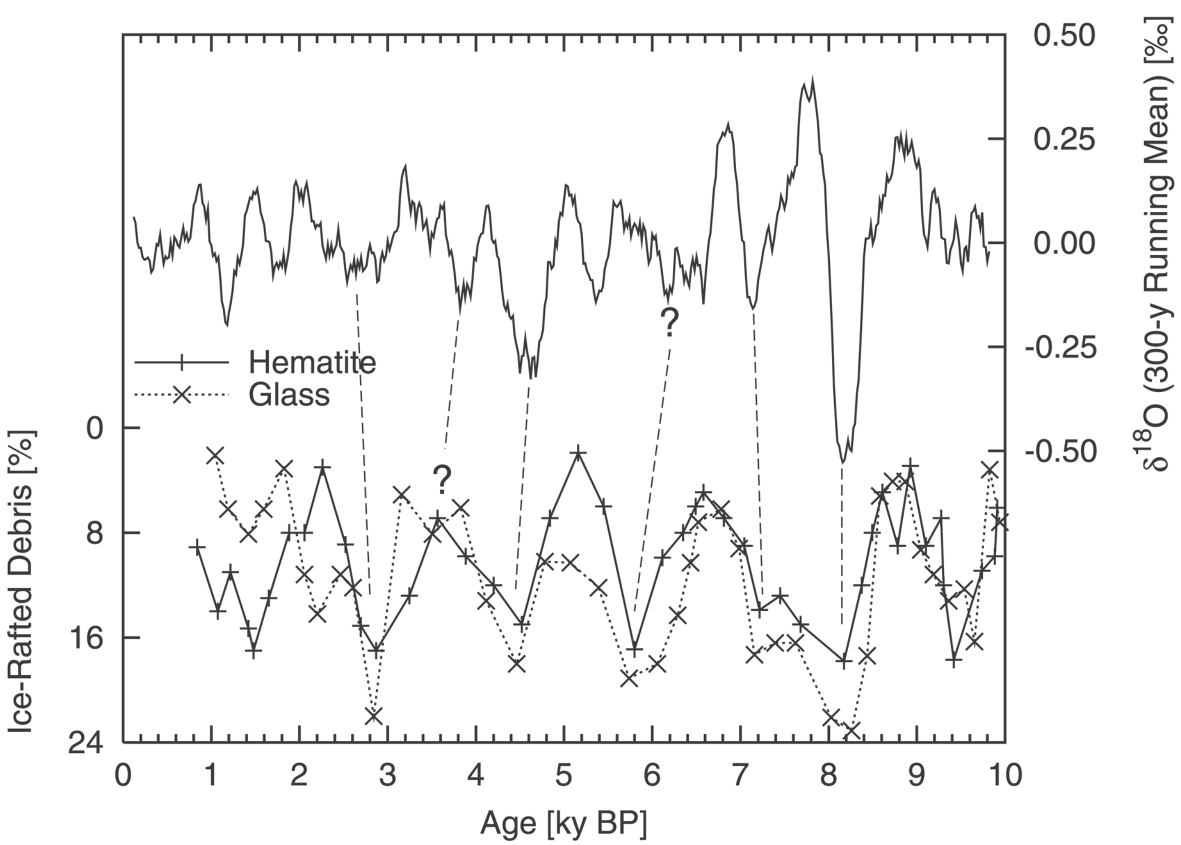

Fig. 7. Smoothed $\delta^{18}O$ time series (as in Fig. 4) together with two ice-rated debris records from the North Atlantic (Bond et al. 1997). Hematite-coated sediment grains stem from East Greenland or Svalbard whereas glass fragments originate from Iceland. Note inverted scale for ice-rafted debris; high values are thought to be caused by cool time intervals (Bond et al. 1997). Dashed vertical lines depict possible correlation between the records.

events in the Holocene cannot be answered unequivocally until deep-sea records with a better temporal resolution become available.

Bianchi and McCave (1999) inferred changes in deep-water flow in the North Atlantic from sedimentary grain-size data. According to their results, velocity of deep-water flow varied with a periodicity of ~1500 y during the Holocene, implying concomitant changes in oceanic heat transport. This finding contrasts the existence of 900-y climate oscillations and seems to be more in line with the 1370 ± 500 y IRD cycle. However, times of reduced deep-water flow do not coincide with the IRD events described by Bond et al. (1997; not shown). Furthermore, harmonic analysis reveals that the time series cannot be distinguished from white noise at a 95% significance level (Fig. 8). This finding most likely explains why the deep-water flow record shows no consistent correlation to the IRD or smoothed $\delta^{18}O$ time series.

Conclusions

Proxies of atmospheric temperature and humidity from Greenland and northern/central Europe show evidence for 900-y climate oscillations between 3-8.5 ky BP. The magnitude of the climate perturbations in Europe was probably large enough to affect human societies, especially since they occurred during the important transition from hunting-gathering life style to sedentary agriculture (e.g. Willis and Bennett 1994).

Based on the temporal evolution of the temperature proxy from Greenland, we regard it as more

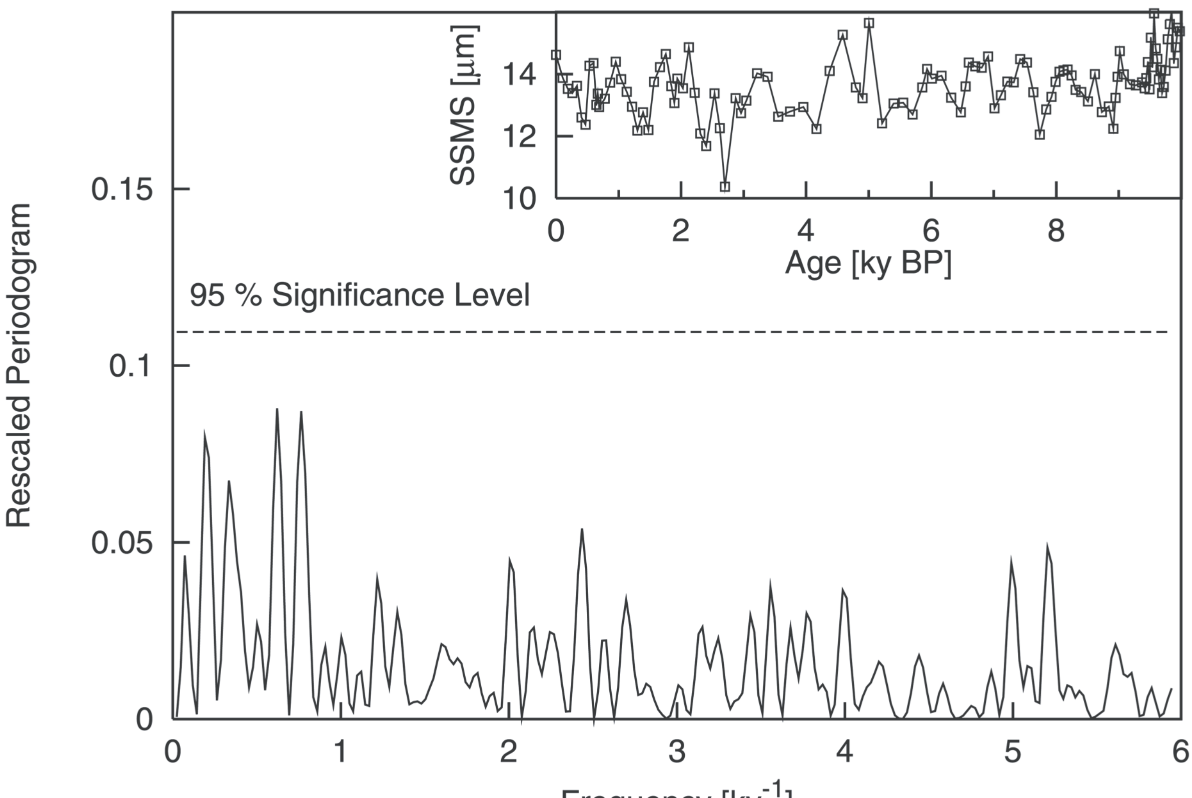

Fig. 8. Harmonic analysis (Fisher 1929; Schulz and Stattegger 1997) of sedimentary grain-size data from the North Atlantic (Bianchi and McCave 1999). The time series (inset; SSMS = sortable silt mean size) cannot be distinguished from white noise at the (1-1/N) × 100% = 99.2% significance level (cf. Thomson 1990; N = number of data points) and not even at the 95% significance level.

likely that the climate oscillations were generated within the climate system than by external orbital forcing. Using a simple template, the climate cycles can be described as damped oscillations, triggered by cooling events. We suggest that negative salinity anomalies in the North Atlantic are the immediate cause of these cold spells, by means of reduced northward heat transport via thermohaline circulation. However, uncertainty remains regarding the ultimate cause of the salinity anomalies. While the first event at ~8.2 ky BP may be linked to the final decay of glacial ice sheets, the cause of the younger event remains unknown. Claussen et al. (1999) simulated an abrupt transition in the hydrological cycle as highly nonlinear response to slow variations in orbital forcing in the mid-Holocene. Although the timing of this transition occurs at ~5.4 ky BP in the model, it may nevertheless point to a potential mechanism, capable of perturbing the salinity field in the north Atlantic.

So far, evidence for the existence of Dansgaard-Oeschger cycles in the Holocene depends largely on the identification of the characteristic *glacial* Dansgaard-Oeschger periodicity (~1470 y) in Holocene climate records (cf. Bond et al. 1997; Bianchi and McCave 1999). Our analysis shows that the evidence for ~1470 cycles in deep-sea records from the North-Atlantic is, however, contestable and that these time series do not contradict 900-y climate oscillations during the Holocene. Considering the current lack of a generally accepted, comprehensive theory for Dansgaard-Oeschger cycles, we cannot specify the possible relation between the Holocene 900-y cycles and the glacial 1470-y Dansgaard-Oeschger cycles. However, the persistence of the 900-y climate oscillations, across the inferred transition from glacial to Holocene mode of atmospheric circulation at ~6.6 ky BP, suggests that the underlying physical mechanism might also work under glacial boundary conditions. Accordingly, the oscillations might be linked to recurrent glacier surges from Greenland, which are in turn generated by feedbacks in the ocean-atmosphere-cryosphere system as proposed by Sarnthein et al. (2001) for glacial Dansgaard-Oeschger cycles.

Until further paleoclimatic data become available, the 900-y cycles should be considered as a regional signal of the North Atlantic realm. Moreover, it is conceivable that different components of the climate system varied on different centennial-to-millennial time scales during the Holocene. For example Wang et al. (1999) identified a ~780-y periodicity in the East Asian monsoon regime. Episodes of enhanced dust deposition occurred in Greenland approximately every 2.6 ky, starting at ~8.3 ky BP (O'Brien et al. 1995). The lack of an obvious correlation between these dust events and inferred temperature perturbations in Greenland (not shown) suggests a different origin of the two signals. Analysis of modern air-mass trajectories (Kahl et al. 1997) suggest that dust sources are mainly located in northern Eurasia and North America. It is beyond the scope of this study to investigate possible links between climate records from these source regions and the dust accumulation in the Greenland ice cores. We note however that times of increased dust deposition seem to coincide with glacier advances in North America (Denton and Karlén 1975) and increased East Asian summer monsoon (Wang et al. 1999).

The smoothed $\delta^{18}O$ data as well as the template suggest only small-amplitude multicentennial climate variability between 0–3 ky BP, making it difficult to predict any future interference with a possible man-made climate perturbation. However, it is possible that the climate system is still capable of generating 900-y climate oscillations. A decrease in North Atlantic sea-surface salinity, predicted by climate models for a future greenhouse climate (e.g. Wood et al. 1999), may therefore restart the climate oscillator – a potential hazard for human societies in Europe.

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