High-resolution analyses of an early Holocene climate event may imply decreased solar forcing as an important climate trigger

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ABSTRACT

Early Holocene lacustrine, tree-ring, ice-core, and marine records reveal that the Northern Hemisphere underwent a short cooling event at 10 300 calendar yr B.P. (9100 ^{14}C yr B.P.). The records were compared on a common high-resolution time scale and show that the event lasted less than 200 yr, with a cooling peak of 50 yr, and the event coincides with a distinct Holocene thermohaline disturbance recognized in the North Atlantic Ocean. In spite of well-known freshwater forcings at the time of the event, the negligible difference between the modeled Δ^{14}C record, based on the GISP2 (Greenland Ice Sheet Project 2) ^{10}Be data, and the measured values, does not allow for detectable Δ^{14}C changes related to global ocean ventilation. We can, however, show that the onset of the cooling coincides with the onset of one of the largest Holocene ^{10}Be flux peaks. This finding may imply that the climate system is more sensitive to solar-related changes than previously thought and that such changes may be an important underlying mechanism for sub-Milankovitch climate variability.

Keywords: Holocene, sub-Milankovitch cooling event, lake sediments, marine records, ice-core proxies, tree rings, Δ^{14}C, ^{10}Be, ^{14}C modeling, solar forcing.

INTRODUCTION

The presence of sub-Milankovitch climate variability in a 1000–2000 yr cooling cycle has been established in the North Atlantic region (Bond et al., 1997). Changes in North Atlantic Deep Water production may be the most likely explanation for North Atlantic cooling events, and sudden freshwater forcings are obvious candidates for being the cause of sudden slowdowns of the thermohaline circulation during glacial and deglacial periods. We performed a multiproxy, high-resolution study on the time period when the Holocene ice-rafted debris (IRD) cooling event 7 (Bond et al., 1997) is inferred to have occurred—i.e., 10 300 calendar yr B.P. (9100 ^{14}C yr B.P.).

ICE-CORE AND MARINE RECORDS

The Greenland Summit ice cores (Fig. 1) show a δ^{18}O minimum at ca. 10 300 cal. yr B.P. (Dansgaard et al., 1993; Stuiver et al., 1995). Correlations to other archives are facilitated by the late Preboreal Saksunarvatn Ash (Waagstein and Johansen, 1968), an Icelandic basaltic tephra horizon (Grönvold et al., 1995). It was dated to 10 180 cal. yr B.P. in the GRIP (Greenland Ice Core Project) ice core (Grönvold et al., 1995), now revised to 10 240 cal. yr B.P. (see Figs. 2 and 4), to 10 275 cal. yr B.P. in the GISP2 (Greenland Ice Sheet Project 2) ice core (Zielinski et al., 1997), and to 9100–9000 ^{14}C yr B.P. (10 250–10 200 cal. yr. B.P.) in different Nordic lake records (Mangerud et al., 1986; Björck et al., 1992).

IRD event 7 (Bond et al., 1997) began at 10 400–10 300 cal. yr B.P. In core VM 29-191 (Fig. 1), it is characterized by the most distinct Holocene δ^{13}C depletion in benthic foraminifera, displaying a distinct thermohaline disturbance signature (Bond et al., 1997). Other events, which may be related to reduced ocean ventilation, occurred at approximately the same time: the sudden 5–10 m drainage of the Baltic Ancylus Lake stage (Björck, 1995) and significantly increased Δ^{14}C values (Stuiver et al., 1998). The former is a candidate for a North Sea freshwater forcing, and the latter could be a response to such a forcing; the freshwater flux may have disturbed the ocean-atmosphere ventilation, thereby decreasing the ventilation of old ^{14}C from the ocean to the

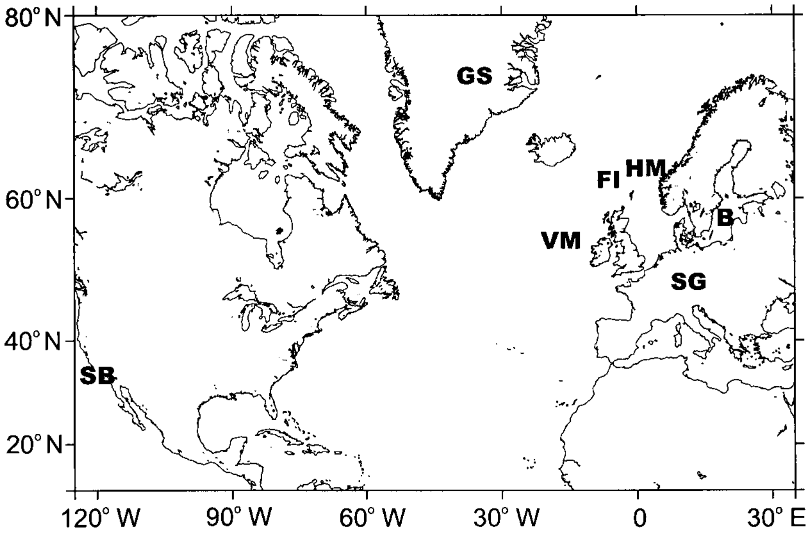

Figure 1. Location of some of Northern Hemisphere sites discussed in text. B—Baltic Sea, SG—southern Germany, HM—marine site HM79-6 (Koç Karpuz and Jansen, 1992), FI—Faroe Islands, VM—marine site VM29-19 (Bond et al., 1997), GS—Greenland Summit, and SB—Santa Barbara Basin.

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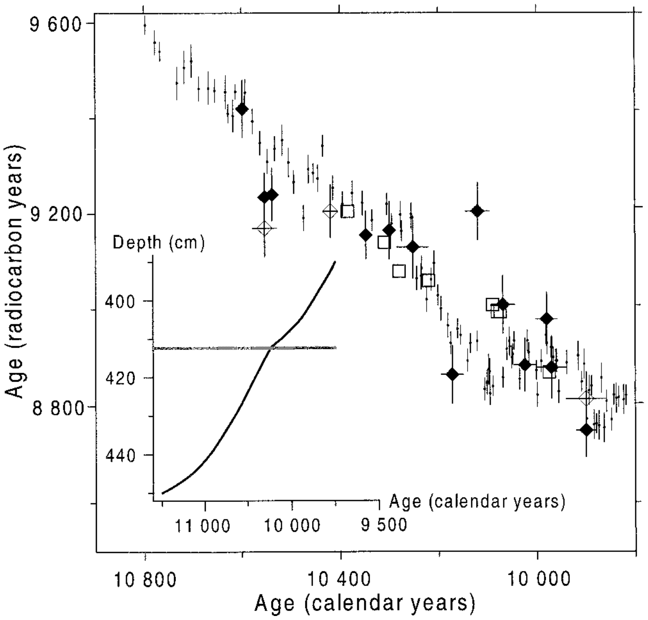

Figure 2. ^{14}C calibration to calendar years (Stuiver et al., 1998) between 10 900 and 9800 cal. yr B.P., based on ^{14}C dates of dendrochronologically dated German pine and oaks (shown as vertical lines displaying ^{14}C uncertainty). Other data are eight ^{14}C dates on pine stumps and peat (squares) found in final raised beach of Ancylus transgression in Estonia (Veski, 1998) and matched ^{14}C dates from Lake Starvatn (solid diamonds—bulk dates, open diamonds—moss dates). Matching was performed with age vs. depth curve to lower left, showing position of Saksunarvatn Ash as gray line at 412 cm depth. Curve is based on GRIP age of Saksunarvatn Ash, dendrochronological age of Younger Dryas–Preboreal transition (Björck et al., 1997; Spurk et al., 1998), calibrated ages of onset and end of 9200–9300 ^{14}C yr plateau, and final ^{14}C age decline at 8800 ^{14}C yr B.P.

atmosphere. We therefore gathered the youngest ^{14}C dates on peat and pinewood found in the final raised beach of the Ancylus Lake (Fig. 2). These dates give the maximum age of 8870 $\pm$ 85 ^{14}C yr B.P. (10 160–9790 cal. yr B.P.) for the drainage and imply that the lake drained after the onset of the sudden ^{14}C age decline, which marks the beginning of the Δ^{14}C rise. Therefore this freshwater forcing is unlikely to have triggered the rising Δ^{14}C values. In addition, eastern Lake Agassiz outlets opened at ca. 9100 ^{14}C yr B.P. and increased the St. Lawrence River flux by 0.05 Sverdrup (Licciardi et al., 1999). This less well dated event is thus an additional candidate for a thermohaline circulation disturbance at the time of the 10 300 cal. yr B.P. event.

HIGH-RESOLUTION LACUSTRINE RECORDS FROM THE FAROE ISLANDS

The Faroe Islands (Fig. 1), within the main arm of the North Atlantic Drift, are in a uniquely sensitive area for registering Holocene cooling events. In colder periods, when the North Atlantic Drift current weakens or its main branch occupies a more southern position, a tongue of polar water from the East Iceland branch of the East Greenland Current may approach the Faroe Islands from the northwest (Humlum, 1998).

Age control for lacustrine records on the Faroe Islands is provided by the occurrence of the Saksunarvatn Ash. The 10 500–10 000 cal yr B.P. deposits in Lake Starvatn (long 6°41′30″W, lat 62°04′30″N, 100 m above sea level) were analyzed in detail for ^{14}C, pollen, biogenic silica, and other proxy records. On the basis of the depth versus age curve (Fig. 2), our 16 ^{14}C dates display the ^{14}C and dendrochronological age structure well. We can thus relate the Starvatn data to a calendar-year time scale (Figs. 2 and 3). The most notable pollen change occurred at 10 350–10 300 yr B.P.; declines in percentage and influx values of birch (*Betula*) and total pollen concentrations are accompanied by a clear rise in grass (Poaceae) and herb pollen (Fig. 3). This anomalous vegetation succession was found below the Saksunarvatn Ash at other Faroe Islands sites (Johansen, 1982) and was interpreted as a late arrival of grasses. However, our data show that grass was abundant before the change and that a perturbation changed the vegetation composition. The other records (Fig. 3) also show clear changes; mineral matter increases, and biogenic silica, organic matter, and $^{13}C_{org}$ decrease. The data suggest decreased aquatic productivity (biogenic silica) and increased soil erosion (δ^{13}C and mineral matter) for the 50–100-yr-long period before the deposition of the ash. Interference from the tephra, shown by magnetic susceptibility, makes it difficult to assess whether these conditions continued after the pure-ash deposition, although severe conditions for vegetation and aquatic productivity appear to have persisted for at least another 100 yr.

NORTH ATLANTIC PALEOCLIMATIC SYNTHESIS

Our record suggests that winter climate in the Faroe Islands became harsher slightly before 10 300 cal. yr B.P., making it harder for birch vegetation to survive in exposed settings and thereby favoring grass expansion. Longer periods of lake ice resulted in decreased aquatic productivity, and longer periods of snowfall would cause increased surface runoff during spring, resulting in soil erosion and decreased δ^{13}C values. Furthermore, increased time resolution of diatom records from core HM79-6 (Fig. 1) in the Norwegian Sea (Koç Karpuz and Jansen, 1992) shows a distinct, but brief, sea surface temperature (SST) minimum 0.5 cm below the Saksunarvatn Ash; mean sum-

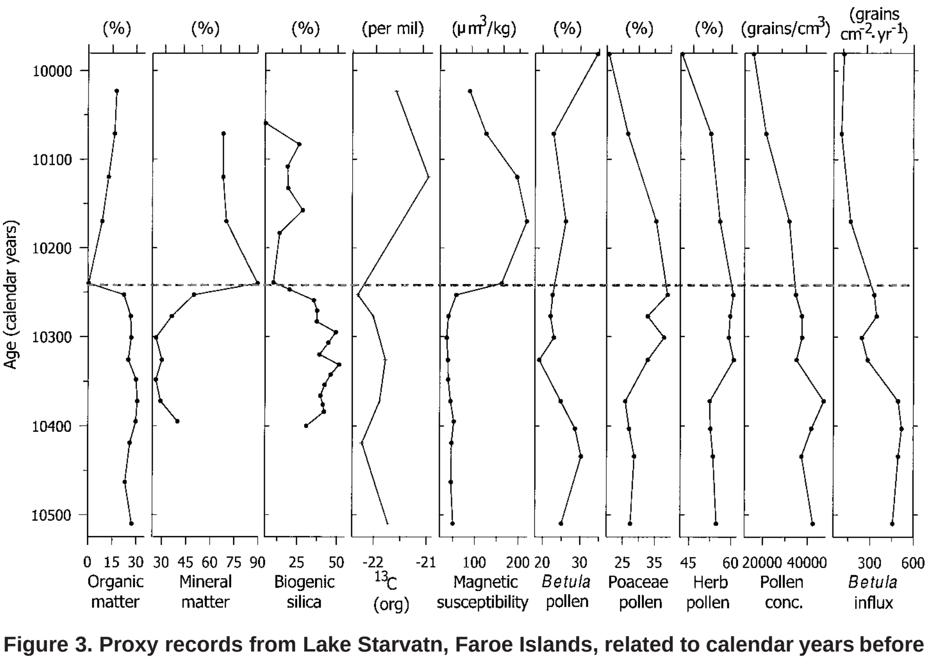

Figure 3. Proxy records from Lake Starvatn, Faroe Islands, related to calendar years before present (Fig. 2). Organic and mineral matter and biogenic silica are related to percentage of dry weight, and $^{13}C_{org}$ values are related to PDB (Peedee belemnite) standard. Pollen percentages are expressed as percentage of all terrestrial pollen grains. Gray dashed line at 10 242 cal. yr B.P. indicates Saksunarvatn Ash.

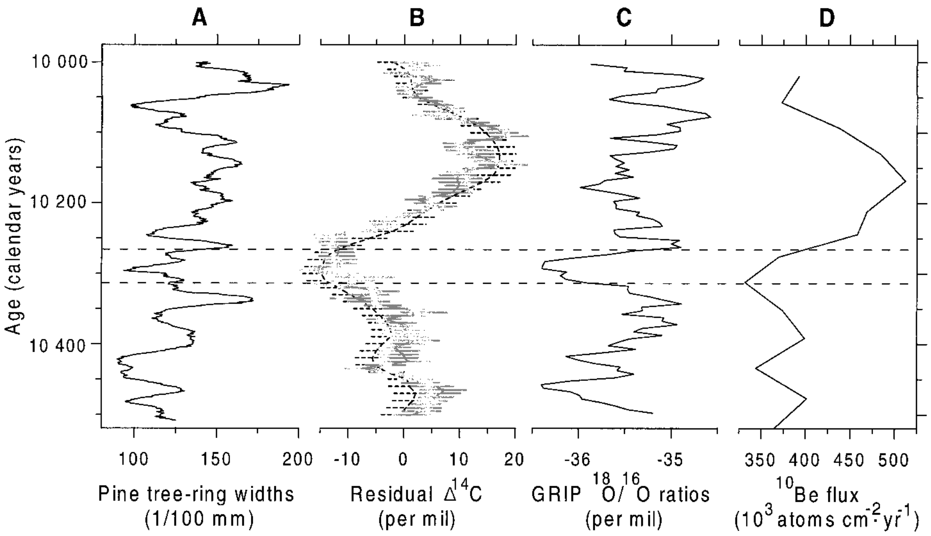

Figure 4. High-resolution records from 10 500 to 10 000 cal. yr B.P.; dashed lines mark peak of 10 300 yr event. A: Tree-ring widths of German pines were filtered to exclude aging effects. B: Residual $\Delta^{14}C$ denotes atmospheric $^{14}CO_2$ activity, expressed as per mil deviation of ^{14}C activity from NIST (National Institute for Standards and Technology) ^{14}C standard, after decay and fractionation (Stuiver and Polach, 1977) and after subtraction of a 500 yr running average to remove long-term trend. Measured $\Delta^{14}C$ values (gray line with gray error bars) are based on ^{14}C dates on German pines (Stuiver et al., 1998), and modeled $\Delta^{14}C$ values (dashed line), where errors (dashed bars) are based on measurement errors of ^{10}Be data, were calculated according to Muscheler et al. (2000). C: Oxygen isotope ($^{18}O/^{16}O$) ratios from GRIP (Dansgaard et al., 1993) are 15 yr running-mean values. D: ^{10}Be fluxes are derived from ^{10}Be concentrations in GISP2 ice core (Finkel and Nishiizumi, 1997) and calculated accumulation rates (Johnsen et al., 1995). Accumulation rates were averaged on time intervals for which ^{10}Be values are available. Ice-core time scales were synchronized by using Saksunarvatn Ash as a time marker for correlation. GRIP ice core is assumed to be more or less synchronous with German dendrochronology (Björck et al., 1997; Spurk et al., 1998) and is here presented with slightly revised early Holocene chronology compared to original one by Dansgaard et al. (1993).

mer SST dropped 2.7 °C, and winter SST dropped 2.1 °C. Assuming a constant accumulation rate during the period in question, the cooling occurred at ca. 10 300 cal. yr B.P. and lasted less than 200 yr. This Norwegian Sea cooling, harsher conditions on the Faroe Islands, the distinct $\delta^{13}C$ depletion in benthic foraminifera west of Ireland, the $\delta^{18}O$ minimum in the ice cores, and the tree-ring width minimum in German pines (Fig. 4), indicate southward displacement of polar waters and weakening of the North Atlantic Drift, possibly explained by reduced thermohaline circulation and reduced northward heat advection in the North Atlantic.

A GLOBAL EVENT?

The marine varve record from the Santa Barbara Basin shows a peak of oxygenation starting at 10 300 cal. yr B.P. and lasting 150–200 yr (Behl and Kennett, 1996). Such events, related to phases of cold surface water such as during the Younger Dryas, are thought to be local responses to reduced North Atlantic Deep Water production and caused by an increase in proximally derived waters at intermediate depths in the Santa Barbara Basin (Fig. 1). In addition, the ice-core record from the Guliya ice cap in Tibet (Thompson et al., 1997) shows a $\delta^{18}O$ minimum at ca. 10 300 cal. yr B.P.

We conclude that convincing evidence exists in the Northern Hemisphere for a distinct but short climatic event centered at ca. 10 300 cal. yr B.P. In the Southern Hemisphere, very few existing high-resolution records span this time interval; thus, it is difficult to establish whether the event had a global impact.

CAUSES FOR $\Delta^{14}C$ AND ^{10}Be CHANGES AND COMPARISONS OF RECORDS ON A COMMON TIME SCALE

Two important questions to address are: what causes the $\Delta^{14}C$ anomalies, and how are the $\Delta^{14}C$ and ^{10}Be peaks (Fig. 4) related to the observed cooling? The ^{14}C is produced in the atmosphere by the interaction of the cosmic ray flux with constituents of the atmosphere. Shortly after the production, ^{14}C becomes oxidized to $^{14}CO_2$ (Lal and Peters, 1967) and hence is involved in exchange processes between the atmosphere, biosphere, and ocean as part of the carbon cycle. Therefore, $\Delta^{14}C$ variations are caused by changes in the carbon cycle (Siegenthaler et al., 1980) or in ^{14}C production rate due to solar and/or geomagnetic modulation of the cosmic ray flux (Lal and Peters, 1967). ^{10}Be records from ice cores provide information on the production-rate history of ^{14}C to disentangle the different causes for $\Delta^{14}C$ changes. The ^{10}Be is produced similarly to ^{14}C, but becomes attached to aerosols and is removed from the atmosphere after a mean residence time of 1–2 yr (McHargue and Damon, 1991). Therefore, ^{10}Be records provide a direct signal of changes in the production rates of the cosmogenic radionuclides.

Assuming that the ^{10}Be flux to Greenland is proportional to the global ^{10}Be production rate (Muscheler et al., 2000), we derived the ^{14}C production rate using recent production-rate calculations (Masarik and Beer, 1999). A box-diffusion carbon cycle model (Siegenthaler, 1983) was used to calculate $\Delta^{14}C$ from the reconstructed production rate. To obtain the pure production signature, no changes in the carbon cycle were assumed during the model run. Monte Carlo simulations with 100 model runs were performed to assess uncertainties due to the ^{10}Be measurement errors (Finkel and Nishiizumi, 1997). The modeled $\Delta^{14}C$ curve was detrended by subtracting a 500 yr average.

Our lake proxy records can be tied to a common calendar-year chronology (Figs. 2 and 3) and show a climate event centered at 10 300 cal. yr B.P. Furthermore, ice-core and tree-ring data show distinct minima at 10 290 cal. yr B.P., preceded by a declining trend and followed by rising values during a 50-yr-long period, and can be related to $\Delta^{14}C$ changes and ^{10}Be fluxes (Fig. 4). Because modeled and measured $\Delta^{14}C$ values agree within error bars, $\Delta^{14}C$ and ^{10}Be flux at 10 300 cal. yr B.P are dominated by production-rate changes. We conclude that the increase in $\Delta^{14}C$ following the 10 300 cal. yr event is related to a sharp decrease in solar and/or geomagnetic forcing shown by the rising ^{10}Be flux between 10 310 and 10 170 cal. yr B.P. (Fig. 4).

DISCUSSION

A perfect synchronization between the records is impossible to achieve, but the onset of the cooling trend does, broadly speaking, coincide with rising ^{10}Be fluxes—i.e., decreased solar or geomagnetic forcing. To our knowledge, no large magnetic field variation that could have caused this event has been found. We therefore postulate that the ^{10}Be maximum was caused by distinctly reduced solar forcing.

Linkage between global climate and solar forcing is a controversial topic. It may, however, be argued that two processes related to solar forcing—variations in UV radiation (related to solar-irradiance changes) and cosmic-ray flux (related to solar-wind changes)—can trigger more efficient climatic feedback mechanisms. For example, changes in UV radiation may substantially change the heat budget of the stratosphere (van Geel et al., 1999; Shindell et al., 1999; Haigh, 1996), altering the atmospheric circulation. Solar-wind variations, on the other hand, may be amplified by changing the atmosphere's optical parameters and radiation balance (Ney, 1959; Pudovkin and Raspopov, 1992), but are also related to cloudiness (Svensmark and Friis-Christensen, 1997), an important climatic feedback mechanism. However, the culmination of the ^{10}Be flux and cold-event peaks in the ice cores do not coincide (Fig. 4). Perhaps the complete event was not registered over Greenland, but is of longer duration in other records, which is indicated by some of the already described data, or the synchronization is not perfect. It is also likely that a straightforward relationship between solar forcing and climate does not exist.

The rise in ^{10}Be flux starting at 10 300 cal. yr B.P. is one of the most prominent Holocene ^{10}Be anomalies and is comparable in magnitude to an early Preboreal peak (Finkel and Nishiizumi, 1997). The onset of this latter ^{10}Be peak, 300 yr after the onset of the Holocene, coincides with the Preboreal oscillation (Björck et al., 1997). Furthermore, the onset of the Younger Dryas is also coeval with a rise in ^{10}Be flux (Goslar et al., 2000; Andresen et al., 2000), as well as with a very distinct Δ^{14}C rise (Björck et al., 1996). However, Muscheler et al. (2000) found that some of this Δ^{14}C rise might be explained by decreased ocean ventilation.

We conclude that a short cooling event occurred in the Northern Hemisphere at ca. 10 300 cal. yr B.P, and the onset of the cooling coincides with one of the two most prominent Holocene rises in ^{10}Be flux. The event may correspond to cold surface waters in the Santa Barbara Basin and reduced thermohaline circulation in the North Atlantic, but the ^{10}Be record can explain the entire Δ^{14}C rise from 10 250 to 10 100 cal. yr B.P. This implies that any decreased ocean ventilation, caused by large freshwater forcings, was not strong enough to show up in the global Δ^{14}C signal. Our calculations also imply that a 200-yr-long 10% reduction of global ocean circulation is within our error bars (Fig. 4).

So we are still left with some key questions on the relation between climate and solar forcing: (1) Is it merely an accidental circumstance that both the 10 300 cal. yr B.P. event and the Preboreal oscillation coincide with the two possibly most distinct Holocene rises in ^{10}Be flux? (2) Could decreased solar forcing have played some role in triggering and/or reinforcing climate-related feedback processes, in some cases perhaps leading to significant cooling events? (3) Is it possible that certain modes or states of the climate system are particularly sensitive to solar-related changes? To answer some of these questions, it may be fruitful to study and model the initial climate conditions of such events in the greatest possible detail.

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