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The Preboreal climate reversal and a subsequent solar-forced climate shift

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ABSTRACT: Accurate chronologies are essential for linking palaeoclimate archives. Carbon-14 wiggle-match dating was used to produce an accurate chronology for part of an early Holocene peat sequence from the Borchert (The Netherlands). Following the Younger Dryas–Preboreal transition, two climatic shifts could be inferred. Around 11 400 cal. yr BP the expansion of birch (*Betula*) forest was interrupted by a dry continental phase with dominantly open grassland vegetation, coeval with the PBO (Preboreal Oscillation), as observed in the GRIP ice core. At 11 250 cal. yr BP a sudden shift to a humid climate occurred. This second change appears to be contemporaneous with: (i) a sharp increase of atmospheric ^{14}C; (ii) a temporary decline of atmospheric CO_2; and (iii) an increase in the GRIP ^{10}Be flux. The close correspondence with excursions of cosmogenic nuclides points to a decline in solar activity, which may have forced the changes in climate and vegetation at around 11 250 cal. yr BP. Copyright © 2004 John Wiley & Sons, Ltd.

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KEYWORDS: Preboreal; solar forcing; climate change; peat; wiggle-match dating; ^{14}C; ^{10}Be; ^{18}O.

Introduction

Lake sediments and peat deposits are valuable archives for the study of past climate change. Calendar year chronologies are of crucial importance for linking the climatic signals from different locations. Precise dating is essential for the study of leads and lags and the potential causes and effects within the climate system. In this paper we compare inferred climatic change as recorded in a European terrestrial sequence from the Borchert (from the east of The Netherlands) with Greenland ice-core data (Fig. 1). The Borchert deposits accumulated in an abandoned channel of the former Dinkel River drainage system. In this sequence, the early Holocene is recorded at high-resolution, which makes this record unique for the North Atlantic region. The detailed palynological–palaeoclimatological record from the Borchert was published by van Geel *et al*. (1981). Sedimentation started under lacustrine conditions during the later part of the Younger Dryas and continued to ca. 3400 yr BP. Originally the sequence was ^{14}C dated by the conventional method using a limited number of bulk peat samples. However, as a consequence of the atmospheric ^{14}C variations in the past, no precise calendar age chronology was obtained for the Late-glacial–early Holocene part. This was mainly owing to the presence of two large radiocarbon plateaux within this time interval, at 10 000–9900 and 9600–9500 ^{14}C BP. A more accurate chronology has now been obtained from AMS ^{14}C wiggle-match dating (WMD) of selected macrofossils. This chronology enables comparison with other chronologies of climate proxies, in particular the Greenland ice-core record. Furthermore, relationships are suggested between climate changes (terrestrial and Arctic), solar variability as evidenced by cosmogenic isotope excursions (both ^{14}C and ^{10}Be), and fluctuations in the atmospheric CO_2 concentration.

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Methods

A new series of macrofossil samples (including fruits, seeds, catkin scales, bud scales, leaves and leafy stems of mosses) was selected from the original Borchert sediments (Table 1). All samples were pretreated in order to remove contaminants (Mook and Streurman, 1983). Core slices were boiled in 5%

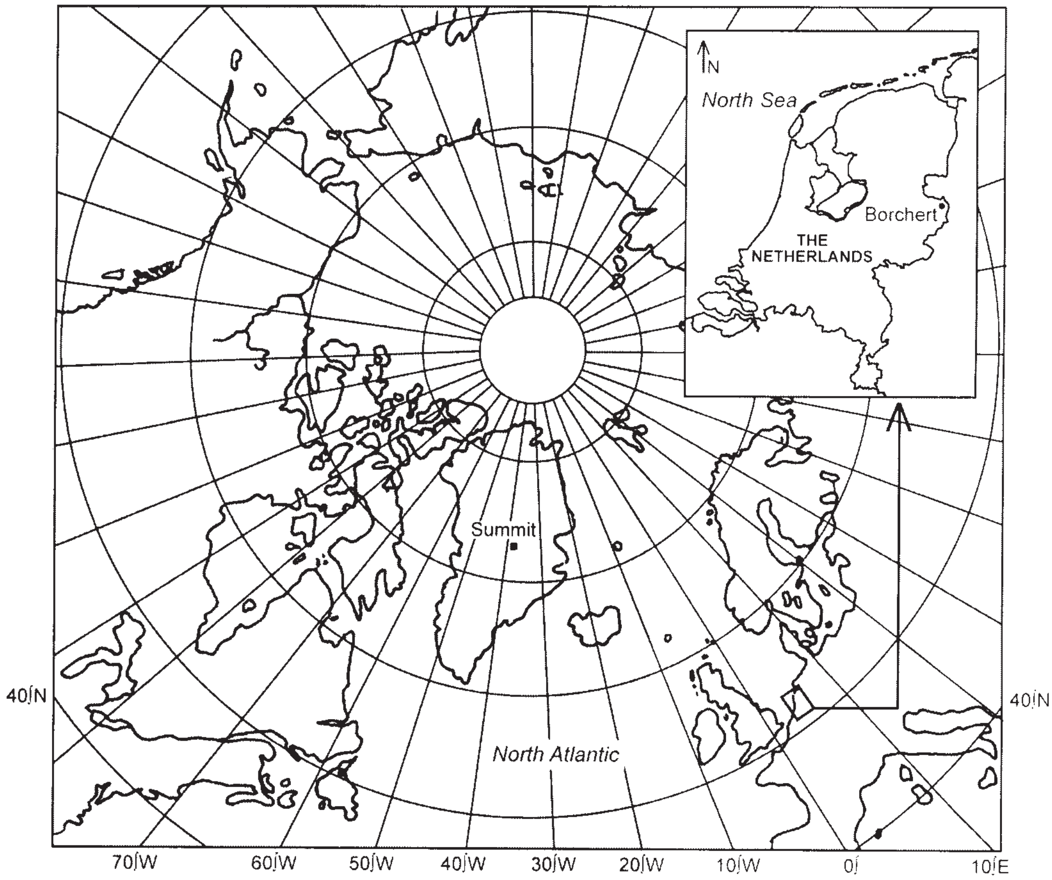

Figure 1 Map of the Northern Hemisphere, with the location of the Borchert (The Netherlands) and the Summit ice-cores GRIP/GISP2 (Greenland)

KOH for 10 min; the macrofossils were washed with deionised water over a 125–140 μm sieve, until excess humic and fulvic acids were removed from the sample material. Macrofossils were picked out manually, selected for ^{14}C dating, and stored at a temperature of 4 °C in a very dilute solution of HCl (Kilian *et al.*, 2000). Samples were then combusted and purified to CO_2 by an automatic CN analyser. The CO_2 produced was trapped cryogenically, and transferred into graphite by reduction under an excess of hydrogen gas (Aerts *et al.*, 2001). The ^{14}C content of the graphite was measured by the Groningen AMS facility (van der Plicht *et al.*, 2000), producing a radiocarbon age for every sample. Radiocarbon ages are reported in BP, defined using the international oxalic acid standard, the conventional half-life of 5568 yr, and correction for isotope effects

Table 1 AMS radiocarbon dates from the early Holocene peat sequence from the Borchert, The Netherlands. The WMD numbers correspond to the numbers in Fig. 2C. The material dated includes selected macrofossils such as fruits, seeds, catkin scales, bud scales and leafy stems of mosses

WMD number	Sample number	Depth (cm)	Material dated	Laboratory number	$\delta^{13}C$	^{14}C age (BP)
23	62	491.2	*Pinus* + *Betula*	GrA-17590	−25.70	9550 ± 60
22	60	492.8	*Pinus* + *Betula*	GrA-17588	−26.31	9480 ± 60
21	58	494.4	*Pinus* + *Betula*	GrA-17578	−26.32	9520 ± 60
20	56	496.0	*Pinus* + *Betula*	GrA-17577	−26.76	9530 ± 70
19	54	497.6	*Betula*	GrA-17456	−27.05	9600 ± 60
18	52	499.2	*Betula*	GrA-17576	−26.72	9490 ± 60
17	50	500.8	*Betula*	GrA-17455	−26.52	9610 ± 50
16	48	502.4	*Betula*	GrA-17453	−26.70	9600 ± 50
15	45	504.8	*Betula*	GrA-17575	−27.44	9530 ± 60
14	42	507.2	*Betula* + *Populus*	GrA-1715	−26.14	9650 ± 50
13	40	508.8	*Betula* + *Populus*	GrA-1717	−26.74	9700 ± 50
12	38	510.4	*Betula* + *Populus*	GrA-1718	−25.99	9790 ± 50
11	34A	513.6	Mosses	GrA-2623	−28.28	9900 ± 60
10	34B	513.6	*Betula*	GrA-2654	−26.83	10,200 ± 200
9	33	514.4	*Betula* + Mosses	GrA-1711	−25.69	9830 ± 50
8	31	516.0	*Betula*	GrA-1714	−25.78	9860 ± 50
7	29	517.6	*Betula* + Mosses	GrA-1712	−25.72	9740 ± 50
6	28A	518.4	Mosses	GrA-2621	−28.09	10 050 ± 60
5	28B	518.4	*Betula*	GrA-2643	−27.21	10 070 ± 100
4	26B	520.0	*Betula*	GrA-2792	−27.23	10 400 ± 800
3	26A	520.0	Mosses	GrA-2620	−32.89	9990 ± 60
2	24	521.6	*Betula*	GrA-1716	−25.98	9900 ± 50
1	22	523.2	*Betula*	GrA-1713	−27.54	9860 ± 50

to $\delta^{13}C = -25‰$. $\Delta^{14}C$ denotes the atmospheric ^{14}C content expressed as the per mil deviation of the ^{14}C content of the oxalic acid standard, after correction for radioactive decay and fractionation. The stable carbon isotope (^{13}C) concentrations are expressed in $\delta^{13}C$, defined as the $^{13}C/^{12}C$ ratio of the sampled gas, in per mil difference to the $^{13}C/^{12}C$ ratio of the CO_2 from the international carbonate standard VPDB (Mook and van der Plicht, 1999).

Individual radiocarbon ages need to be calibrated to obtain calendar ages. However, calibration often results in a probability distribution in calendar age encompassing a relatively long period, and moreover, the age distribution may show various maxima and minima. This is caused by natural variations in the atmospheric ^{14}C concentration, resulting in the so called 'wiggles' in the calibration curve (Suess, 1970). In radiocarbon sample series from high-resolution peat sequences, these ^{14}C wiggles of the calibration curve have been recognised and used for optimising the time control (van Geel and Mook, 1989; Kilian *et al.*, 2000; Speranza *et al.*, 2000; Mauquoy *et al.*, 2002). In this method a series of uncalibrated AMS radiocarbon dates can be matched to the ^{14}C calibration curve, using the stratigraphical position of the ^{14}C dated samples. The WMD method is especially used for the steep parts of the calibration curve, which correspond to periods with a strongly variable $\Delta^{14}C$. The recommended radiocarbon calibration curve INTCAL98 (Stuiver *et al.*, 1998) was used for WMD of the radiocarbon samples of the Borchert sequence. Absolute ages are reported in cal. yr BP, i.e. calibrated or calendar age relative to 1950.

Botanical analysis of the Borchert sequence has been carried out previously by van Geel *et al.* (1981), who produced highly detailed diagrams showing both numerous microfossils and macroremains. For the pollen diagram of the Borchert, percentages of plant taxa were calculated based on a pollen sum including arboreal pollen (AP), and pollen of upland herbs including Poaceae and Cyperaceae (NAP, non-arboreal pollen). For the present paper a summary pollen diagram was constructed, showing the main trends in the groups of trees (e.g. *Betula* and *Pinus*) and shrubs, Poaceae, Ericales and other upland herbs.

Concentrations of CO_2 were reconstructed by stomatal frequency analysis from fossil birch leaves by Wagner *et al.* (1999). Stomatal frequency is expressed in terms of stomatal density and stomatal index (SI). The stomatal index (SI) = [stomatal density/(stomatal density + epidermal cell density)] $\times$ 100 was used as a proxy for past changes in atmospheric CO_2 concentration. The analysis of stomatal parameters of leaf cuticles was performed with a computer-aided imaging system. For more details, we refer to Wagner (1998).

The ^{10}Be flux, measured in Greenland ice (GISP2), was earlier published by Muscheler *et al.* (2000). The ^{10}Be flux is considered to be a good measure of changes in the production rate of the cosmogenic isotopes ^{14}C and ^{10}Be.

Results

Palaeobotanical data

The palynological record (Fig. 2E) shows several prominent changes in the AP/NAP ratio and dominant arboreal taxa (*Betula* and *Pinus*). The sequence starts with a transition from a predominantly open landscape to a birch (*Betula*) dominated forest. This is the transition between the Younger Dryas and the Preboreal biozones. During the Younger Dryas period, sandy gyttja was deposited at the sample site (Fig. 2D).

Based on changes in the AP/NAP ratio and dominant arboreal taxa, the Preboreal biozone has been further subdivided into a Friesland Phase, a Rammelbeek Phase and the Late Preboreal. The Friesland Phase (Behre, 1966) is characterised by a strong rise in the values of *Betula*. The macrofossil record of the Borchert (van Geel *et al.*, 1981) indicates that it was mainly *Betula pubescens* that caused this increase, although *Betula nana* was still present within the area. The Friesland Phase was a period of rising mean summer temperatures (van Geel *et al.*, 1981). The *Betula* increase is coincident with a change from a minerogenic to organic dominated sediment (Fig. 2D).

During the following Rammelbeek Phase (Wijmstra and de Vin, 1971), forest expansion was interrupted and grasses (Poaceae) dominated the regional vegetation. The contemporaneous occurrence of thermophilous plants, such as *Nymphaea alba, Ceratophyllum* and representatives of the Zygnemataceae, suggests relatively dry and warm summers with mean July temperatures around 13–15 °C (van Geel *et al.*, 1981), although winter temperatures may have been low (van Geel and Kolstrup, 1978). In climatological terms, a dry continental phase was initiated and regionally the wet birch, *Betula pubescens*, gave way to a steppe-like vegetation rich in grasses. A similar expansion of herbaceous vegetation also has been recognised in other pollen records from The Netherlands (Wijmstra and de Vin, 1971; Hoek, 1997) and northwest Germany (Behre, 1967). Locally, at the Borchert, the small, shallow lake filled up with *Drepanocladus* (moss) peat (Fig. 2D).

During the Late Preboreal, birch forest expanded again and the macrofossil record (van Geel *et al.*, 1981) shows that *Betula pendula* occurred in the vegetation surrounding the site. At the transition from the Rammelbeek Phase to the Late Preboreal there is a sudden transition from *Drepanocladus* peat, through a brief *Menyanthes–Utricularia* phase, to *Sphagnum* peat (Fig. 2D). The local presence of *Sphagnum* and *Assulina* suggest relatively oligotrophic conditions and a rainwater-fed local vegetation. *Pinus* immigrated during the later part of the Late Preboreal. The transition to the Boreal biozone is characterised by the immigration of *Corylus*.

CO_2 data

Stomatal frequency signatures of fossil birch leaves in the Borchert sequence, published by Wagner *et al.* (1999), reflect a sharp rise in atmospheric CO_2 during the Friesland Phase (Fig. 2F). In this phase, CO_2 concentrations rose from ca. 260 ppmv to 327 ppmv, followed by a more gradual increase during the Rammelbeek Phase, to ca. 336 ppmv in the early part of the Late Preboreal. After the start of the Late Preboreal there was a gradual decline to a minimum of ca. 301 ppmv. This was followed again by a strong increase to ca. 348 ppmv, before the CO_2 concentrations stabilised to between 333 and 347 ppmv.

Radiocarbon data

Owing to the absence of datable material, ^{14}C measurements have not been made from the Younger Dryas interval of the Borchert sequence. However, a new series of 23 AMS ^{14}C dates (Table 1) have been measured from the Preboreal part. For the material dated, the ^{14}C age (in BP $\pm 1\sigma$) and the $\delta^{13}C$ values (in ‰ VPDB) are listed for each depth in Table 1. The ^{14}C dates were wiggle-matched on the INTCAL98 ^{14}C calibration curve (Stuiver *et al.*, 1998). The relevant part of the INTCAL98 calibration curve is shown as red lines (1σ envelope) from ca. 11 950 to 10 650 cal. yr BP in Fig. 2C. To each ^{14}C date, a

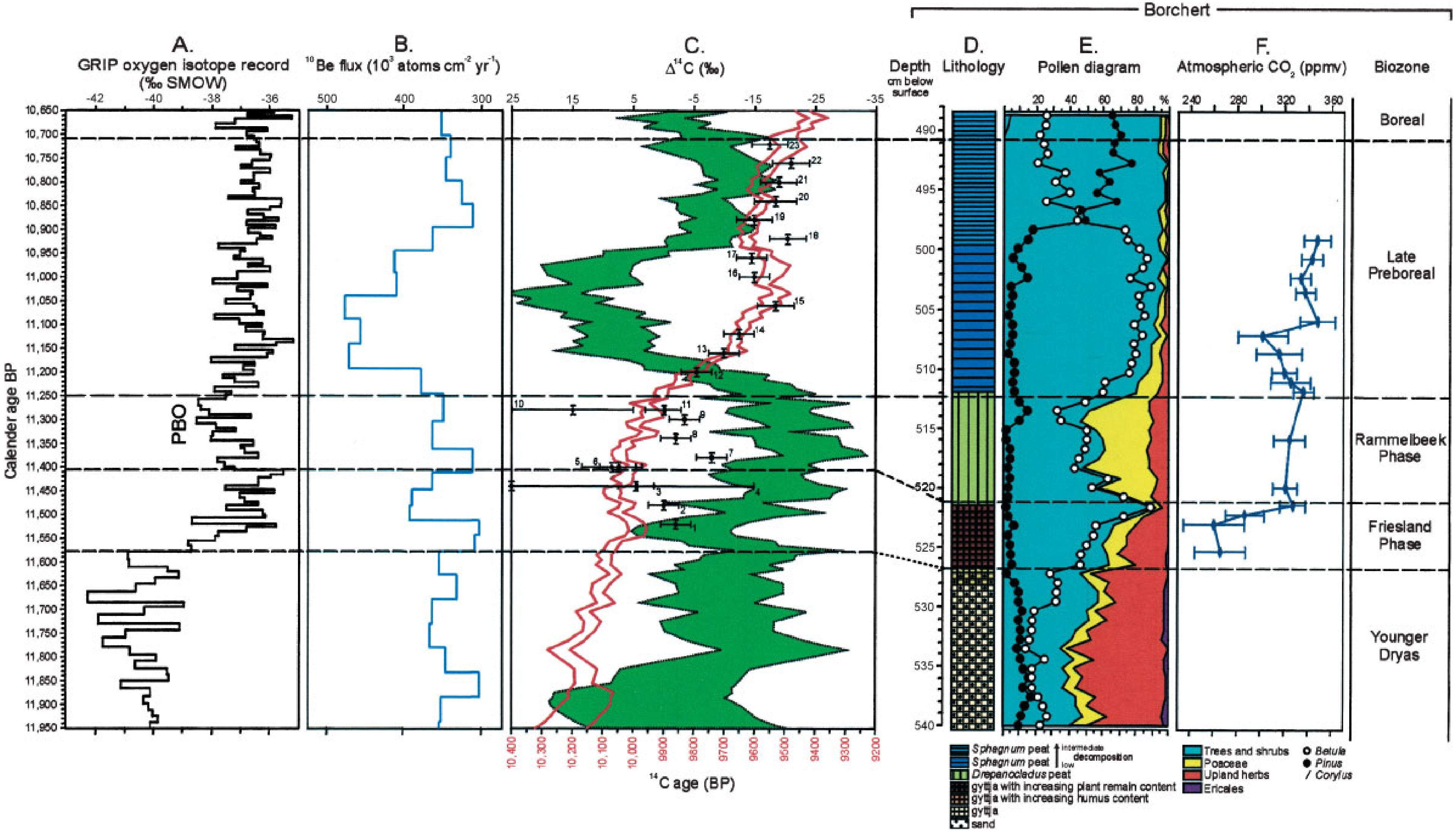

Figure 2 Comparison of the Borchert sequence with other proxy records. Note that the vertical scale of A, B and C is a linear time-scale, whereas the vertical scale of D, E and F is a linear depth scale. The dotted and shortly dashed lines between C and D are assumed corresponding levels. (A) The GRIP oxygen isotope record (Johnsen *et al.*, 1997) in cal. yr BP (calendar years before 1950) with, around 11 570 cal. yr BP, the Late-glacial–Holocene transition. Between 11 400 and 11 250 cal. yr BP a negative excursion is observed in the $\delta^{18}O$ isotope values: the Preboreal Oscillation (PBO). (B) The ^{10}Be flux (measured in the GISP2 ice-core) is given, here displayed on the GRIP time-scale (see Muscheler *et al.*, 2000). The GISP2 time-scale was transferred to the GRIP time-scale by synchronising the end of the Younger Dryas and the 8200 cal. yr BP event. In between, to avoid matching every small $\delta^{18}O$ wiggle, synchronisation was accomplished after low-pass filtering both GRIP and GISP2 $\delta^{18}O$ and synchronising the common structure (Muscheler *et al.*, 2000). (C) The high-resolution series of ^{14}C AMS dates from the Borchert sequence (D–F) are wiggle-matched to the INTCAL98 ^{14}C calibration curve (in red, Stuiver *et al.*, 1998). Ice-core years (based on the GRIP ice-core chronology) and calendar years (based on dendrochronology) are taken as synchronous. The ^{14}C age values are reported in yr BP (horizontal scale). The attribution of a cal. yr BP time-scale to the upper part of the Borchert sequence is based on ^{14}C wiggle-match dating. Note that before 11 250 cal. yr BP the absolute time-scale of D–F is uncertain. However, the Younger Dryas–Holocene transition at 11 570 cal. yr BP can be matched to the transition between the Younger Dryas and Friesland Phase in the pollen diagram (dotted line between C and D). The short dashed line between C and D connects the start of the Preboreal Oscillation (A) with the start of the Rammelbeek Phase (D–F). In C the ^{14}C record of INTCAL98 is also shown as detrended Δ^{14}C values (green shaded area, corresponding to 1σ uncertainty). Lithology (D), simplified pollen diagram (E) and biozones, distinguished in the Borchert sequence, range from the late Younger Dryas to the early Boreal. The CO_2 curve (F) is based on the stomatal index of birch leaves from the same Borchert sequence (Wagner *et al.*, 1999)

WMD number was assigned (numbers 1–23). Figure 2C shows that a WMD based chronology for the sequence was obtained from ca. 11 250 to 10 700 cal. yr BP. Before 11 250 cal. yr BP, WMD fails and thus the absolute time-scale of the Borchert sequence is uncertain. The same phenomenon was also observed in another record in The Netherlands; paradoxically, however, there are also records that seem to give a good WMD within the same time interval (Bos *et al.*, in preparation). Possible causes for this phenomenon are currently under investigation.

The $\Delta^{14}C$ data, plotted in green (Fig. 2C), have been calculated from the INTCAL98 data set. Here we use the residual or detrended $\Delta^{14}C$ signal, i.e. the geomagnetic component has been subtracted (Damon and Peristykh, 2000). The green shaded area corresponds with the 1σ uncertainty of the data. The $\Delta^{14}C$ curve shows no major fluctuations during the early Preboreal period. There is a tendency, however, to more negative values during the Friesland Phase, when the residual $\Delta^{14}C$ values change from ca. -5‰ to ca. -20‰. During the Rammelbeek Phase these values stabilise around -22‰. At the beginning of the Late Preboreal, the residual $\Delta^{14}C$ values rise very rapidly from ca. -15‰ to positive values and reach a maximum of ca. $+15$‰ around 11 000 cal. yr BP. Later, during the Late Preboreal, the residual $\Delta^{14}C$ values decrease again to around -10‰.

Comparison with other proxy records

Using the dating accuracy based on WMD, the Borchert sequence was considered 'absolutely dated' from ca. 11 250 to ca. 10 700 cal. yr BP. This means that we are now able to compare the Borchert record directly with other proxy records, such as the $\delta^{18}O$ ratio and ^{10}Be flux in Greenland ice-core records (Fig. 2A and B).

The ^{10}Be flux was measured originally in the GISP2 ice-core. In Fig. 2B the GISP2 time-scale was transferred to the GRIP time-scale by synchronising the end of the Younger Dryas and the 8200 cal. yr BP event. To avoid matching every small $\delta^{18}O$ wiggle (which often cannot be allocated unambigiously), the synchronisation was achieved by low-pass filtering of both GRIP and GISP2 $\delta^{18}O$, and synchronising the common structure (Muscheler *et al.*, 2000). As pointed out by Lowe *et al.* (2001), there are differences between the Greenland GRIP and GISP2 ice-core chronologies. However, the official (ss09) as well as the revised GRIP (ss08c) chronology suggest synchronous changes in ^{10}Be-derived ^{14}C (Muscheler *et al.*, 2000) and measured $\Delta^{14}C$ around 11 000 cal. yr BP, indicating that the GRIP time-scale agrees well with the absolute tree-ring time-scale (Friedrich *et al.*, 2001).

In Fig. 2A, B and C the vertical scale is the absolute age in cal. yr BP (calibrated years or calendar years relative to AD 1950). This time-scale is valid for the ice-core records and the ^{14}C plots. The vertical scale for the palaeoecological plots of the Borchert sequence (Fig. 2D–F) is a depth scale.

The Younger Dryas–Holocene transition, which could not be dated by AMS, was present clearly in both the pollen diagram of the Borchert sequence, and the sedimentological record by a lithological change from predominantly minerogenic to organic deposits (LC; see Björck *et al.*, 1996). This transition in the Borchert pollen diagram (Fig. 2E, for details see van Geel *et al.*, 1981) was synchronised (dotted line between Fig. 2C and D) with the Late-glacial–early Holocene transition in the GRIP ice-core at ca. 11 570 cal. yr BP (Johnsen *et al.*, 1997). Furthermore, it is possible to connect the start of the Rammelbeek Phase (short dashed line between Fig. 2C and D) with the start of the Preboreal Oscillation in the GRIP ice-core (PBO, Fig. 2A). This is based on the assumption that the accumulation rate of the well preserved *Drepanocladus* peat, which accumulated during the Rammelbeek Phase, was considerably higher than the accumulation rate of the lacustrine deposit, which was formed during the Friesland Phase. The Rammelbeek Phase occurred during a period of lower $\Delta^{14}C$ values and maximum solar insolation at 65 °N at ca. 11 000 cal. yr BP (Berger and Loutre, 1991). Botanical evidence (see previous section) points to relatively dry and warm summer conditions.

From ca. 11 250 cal. yr BP onwards, the Borchert chronology is based on WMD and the ice core and Borchert time-scales (see Fig. 2) are regarded as synchronous. Figure 2B and C shows that the transition of the Rammelbeek Phase to the Late Preboreal at 11 250 cal. yr BP coincides with the start of sharp rises of $\Delta^{14}C$ and the ^{10}Be flux. Locally a transition from *Drepanocladus* to *Sphagnum* peat was recorded. This change indicates a hydrological transition from groundwater-fed to predominantly rainwater-fed local vegetation. Regionally the birch forest was able to expand again at the expense of open grassland vegetation. The analysis of stomatal frequencies in *Betula* leaves from these levels showed a temporary decline of atmospheric CO_2 levels after the start of the Late Preboreal (Fig. 2F). The contemporaneous abrupt changes of the local and regional vegetation at the transition from the Rammelbeek Phase to the Late Preboreal are interpreted as the effects of a sudden shift towards wetter climatic conditions. The Late Preboreal atmospheric CO_2 decline apparently followed this climate change (Fig. 2F).

Discussion

Given the dating accuracy provided by WMD, we are now able to compare climate change (as inferred from changes in local and regional vegetation composition) with atmospheric ^{14}C changes, and with Greenland ice-core data (Johnsen *et al.*, 1997; Fig. 2A). The two cosmogenic nuclides ^{14}C and ^{10}Be, show similar, contemporaneous fluctuations during the Preboreal period (Muscheler *et al.*, 2000). These radionuclide records provide information on past solar activity, as the galactic cosmic ray flux, which produces ^{14}C and ^{10}Be in the atmosphere, is modulated by geo- and heliomagnetic shielding. Whereas ^{14}C enters the global carbon cycle in the form of $^{14}CO_2$, which exchanges with the biosphere and the oceans, ^{10}Be attaches to aerosols deposited mainly by precipitation (McHargue and Damon, 1991). The close correspondence between the ^{10}Be flux (Fig. 2B) and the residual $\Delta^{14}C$ record (Fig. 2C) shows that a changing production rate was the major factor influencing the fluctuations in both cosmogenic nuclide records, most probably caused by changes in solar activity (Beer *et al.*, 1988).

In the early Preboreal part of the Greenland ice-cores (Fig. 2A), a negative excursion is recorded in the $\delta^{18}O$ isotope values between 11 400 and 11 250 cal. yr BP, which is interpreted as a cool climatic phase (Kapsner *et al.*, 1995; Björck *et al.*, 1996), i.e. the Preboreal Oscillation (PBO). In the ice cores the PBO is a phase of diminished snow accumulation. It has been attributed to a meltwater pulse, caused by the melting of the Scandinavian ice-sheets, including the drainage of the Baltic Ice Lake (Björck *et al.*, 1997; Hald and Hagen, 1998) and drainage of Lake Agassiz (Fisher *et al.*, 2002). This was suggested to result in a temporary decrease of the thermohaline circulation in the North Atlantic.

In the North Atlantic terrestrial record the PBO is positioned between the 10 000–9900 and 9600–9500 ^{14}C yr BP plateaus

(Björck *et al.*, 1996, 1997). It is synchronous with a significant rise in atmospheric $\Delta^{14}C$ as shown in Fig. 2C. Björck *et al.* (1996, 1997) interpreted the fluctuations in the Greenland ice-cores and North Atlantic terrestrial data sets as one event, the PBO. However, the Borchert record suggests that the early Preboreal may have been characterised by two distinct climatic events, caused by different processes. Initially there is a phase, which in the Greenland ice-cores was recorded as a minimum in the oxygen isotope record, i.e. the defined Preboreal Oscillation. This was possibly caused by a large meltwater flux that resulted in a temporary decrease of the thermohaline circulation. The northwest European terrestrial equivalent of this phase may have been dry and continental, i.e. the Rammelbeek Phase. The second phase, characterised by the rise in cosmogenic nuclides ^{14}C and ^{10}Be and by $\delta^{18}O$ rising again, starts at 11 250 cal. yr BP, and correlates with the 'European terrestrial PBO' as defined by Björck *et al.* (1997). This shift to more humid conditions in northwest Europe at 11 250 cal. yr BP could have been triggered by a sudden decline of solar activity leading to a sharp rise of cosmic ray intensity, which is shown by both the $\Delta^{14}C$ record and the ^{10}Be flux (Fig. 2B and C).

More evidence for abrupt changes to wetter climatic conditions in the temperate zones *at the onset* of sudden increases of $\Delta^{14}C$ have come from raised bog studies (van Geel *et al.*, 1998) and lake-level studies (Magny, 1993). Major climate shifts date to 2750 cal. yr BP (van Geel *et al.*, 1998) and the 'Little Ice Age' (Mauquoy *et al.*, 2002). For these middle and late Holocene climatic events it was postulated that they were caused by variations in solar activity. In earlier studies (Denton and Karlén, 1973; Stuiver and Braziunas, 1989; Magny, 1993; Björck *et al.*, 1997) it was suggested that rises of atmospheric ^{14}C, contemporaneous with cooling phases during the Late Glacial and early Holocene, were the result of a decreased thermohaline circulation, the consequence of episodic outbursts of large meltwater reservoirs into the North Atlantic.

Based on the Borchert Preboreal record, a different, less 'oceanocentric', but 'heliocentric' conclusion is proposed (van Geel *et al.*, 2003): decreased solar activity was responsible for the sharp increases of the cosmogenic nuclides, and may also have been the trigger for the climate shift (cf. Björck *et al.*, 2001; Bond *et al.*, 2001). In principle, the $\Delta^{14}C$ changes may also have been caused by changes in ocean circulation. However, the $\Delta^{14}C$ signal lacks sensitivity to changes in the thermohaline circulation. Carbon cycle models indicate a change of 40‰ or less during the Younger Dryas (see e.g. Marchal *et al.*, 2001). This change may have lasted for approximately 1200 yr. Most probably the ocean ventilation changes at the time of the PBO are smaller in magnitude and of shorter duration compared with the Younger Dryas. Therefore, the ocean circulation related $\Delta^{14}C$ change around the PBO is expected to be much smaller and not detectable in the ^{14}C record. Furthermore, the synchronicity between the ^{10}Be and ^{14}C records and the quantitative agreement of both ^{10}Be- and tree-ring-based $\Delta^{14}C$ (Muscheler *et al.*, 2000) confirms that the rise of atmospheric ^{14}C was caused by an increased production of cosmogenic nuclides. This change is most likely the result of a change in solar activity. Therefore, these changes suggest a cause of the recorded climate shift: although reduced solar activity may have triggered the climate change, a contemporaneous decrease in solar wind increased the cosmic ray flux impinging on the upper atmosphere, resulting in an increased production of ^{14}C and ^{10}Be. Nevertheless, oceanic and CO_2 changes can amplify the solar-induced changes.

There are additional mechanisms that could amplify climate changes owing to a decline in solar activity. It has been proposed that reduced solar UV intensity may cause a decline of stratospheric ozone production and cooling as a result of less absorption of solar radiation. This might influence atmospheric circulation patterns (extension of Polar Cells and equatorward relocation of mid-latitude storm tracks (Haigh, 1996; van Geel *et al.*, 2001). Increased cosmic ray intensity might also influence climate via stimulated formation of low clouds and precipitation (Svensmark and Friis-Christensen, 1997; Carslaw *et al.*, 2002).

Conclusion

Previously, accurate dating and comparison of Preboreal climatic oscillations in terrestrial records was problematic. However, by using the dating precision based on WMD, it is now possible to compare palaeoecological data in great detail with the fluctuations of cosmogenic isotopes, and with the oxygen isotope records in Greenland ice-cores. This study demonstrates that the early Preboreal climate history of northwestern Europe was complex; a continental climate with warm, dry summers and cold winter conditions of the Rammelbeek Phase was coeval with cooling (PBO) in Greenland. The onset of a more humid period at 11 250 cal. yr BP in northwest Europe coincides with a return to normal interglacial conditions over Greenland as well as with a significant decline in solar activity. The data presented form an indication for a link between changes in solar activity and climate change. Based on the data available, it is also concluded that the atmospheric CO_2 fluctuations, as reconstructed from the stomatal index of birch leaves in the same peat deposit, followed climate change.

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