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The Preboreal oscillation around the Nordic Seas: terrestrial and lacustrine responses

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ABSTRACT: The occurrence of an early Preboreal climatic cooling/oscillation (PBO) in lacustrine and glacial records from northwest Europe, Iceland and Greenland is reviewed and documented. The often subtle response of the proxy records to this oscillation, in combination with its short duration, make it difficult to detect. Owing to its chronostratigraphic position between the 10 000–9900 and 9600–9500 ^{14}C plateaux (c. 11 300–11 150 calendar yr BP) it is also difficult to ^{14}C date with precision. We find that the vegetation response to the PBO varies between sites and regions. In contrast to the pioneer vegetation in Iceland and southern Sweden, the expanding birch–pine forest in Germany–Denmark was more susceptible to deteriorating growing conditions. The combined lacustrine, tree-ring and glacial records imply that the PBO was characterised by cool and humid conditions throughout northwestern and central Europe. This is documented by vegetation changes, decreased aquatic production, increased soil erosion, increased ^{2}H and ^{13}C content in tree-rings, readvances or stillstands of the ice sheet in Norway and Finland, and ingression of brackish water into the Baltic. Icelandic proxy records from lake sediments and glacial moraines imply cooler conditions than during the previous Preboreal period, but not as extreme as during the Younger Dryas. Greenland records suggest that the early Preboreal was characterised by ice readvances, as an effect of cool climate and increased precipitation (in relation to the Younger Dryas). It was not until the end of the PBO that climate was warm enough to melt the land-based ice sheet. This Preboreal oscillation, found on both sides of the Nordic Seas, is interpreted as an effect of increased freshwater forcing on the thermohaline circulation in the Nordic Seas, which is implied by a simultaneous and distinct rise in the atmospheric ^{14}C/^{12}C ratio. A slow-down of the thermohaline circulation may temporarily have pushed the Polar Front further south. © 1997 by John Wiley & Sons, Ltd.

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Preboreal oscillation; thermohaline circulation; lacustrine records; glacial records; chronology.

Introduction

If changes in thermohaline circulation and ocean ventilation rates are the main driving force for the deglacial climate oscillations (e.g. Broecker, 1990; Broecker *et al.*, 1990; Bjorck *et al.*, 1996), it might be expected that there would be a climate response to the large freshwater influx caused by the warming at the onset of the Preboreal, i.e. meltwater peak IB (MWP IB: Fairbanks, 1989; Bard *et al.*, 1996). This would be expressed most likely as a cooling in regions where a decreased thermohaline circulation would have a direct influence on local/regional climate. The occurrence of a fairly distinct Preboreal oscillation (PBO) in large parts of the North Atlantic region has also been pointed out by, for example, Lowe *et al.* (1994) and Bjorck *et al.* (1996). This was based on new data as well as a thorough review of older observations. Bjorck *et al.* (1996) also noted that the biotic response to a cooling in recently deglaciated regions where pioneer biota were dominant was, in some respects, less clear than in areas with a more varied vegetation cover, but that it can be found even in sites very close to a cooling ice-sheet, if the right type of proxy records are examined in detail. To check the significance, character and geographical extent of this oscillation, we therefore make a comparison between different regions and different sets of data from around the Nordic Seas. (Fig. 1A).

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Some previous reports of a Preboreal oscillation in lacustrine sediments

Since the 1960s, and onwards, detailed pollen studies of early Holocene lake sediments in central Europe (e.g. Welten, 1958; Zoller, 1960), northern Germany (e.g. Behre, 1966, 1978), Great Britain (e.g. Simmons, 1964; Walker *et al*., 1993), Denmark (Iversen, 1973) and The Netherlands (Van Geel *et al*., 1981) have implied that the early Preboreal (PB) warming was soon followed by a short climatic revertance. When Behre (1978) summarized European pollen data for this assumed cooling he found evidence for it in numerous pollen diagrams from Holland and France in the west, central Russia (1800 km east-northeast of St Petersburg) in the east, North Italy in the south and Denmark in the north. He named it 'jungste Dryaszeit', i.e. Youngest Dryas, as a small-scale analogue for the Younger Dryas (YD). In Switzerland it had been named the Piottino oscillation (Zoller, 1960), in Denmark the Friesland (Fig. 1A) oscillation (Iversen, 1973), which was actually the name Behre (1966) had given to the short warm period between the Younger Dryas and the cooling, whereas Bjorck *et al*. (1996), for practical reasons, simply called it the Preboreal oscillation (PBO).

In northern Germany the PBO is characterized by decreasing *Pinus* pollen and increasing values of herbs as well as *Juniperus* and *Empetrum* (Behre, 1966). Behre (1966, p. 158) also pointed out that the vegetational effects of this assumed cooling were most obvious in 'transition zones of different vegetational units'. The vegetation in areas with a very recent immigration of plants should be more susceptible to a cooling if the vegetation is not completely out of balance with climate. In the earliest part of the Preboreal a birch–pine forest had begun to be established in northern Germany, whereas the vegetation in, for example, southern Scandinavia (Fig. 1A and C) still had partly the character of a pioneer flora as a remanence of the YD cold phase. However, if

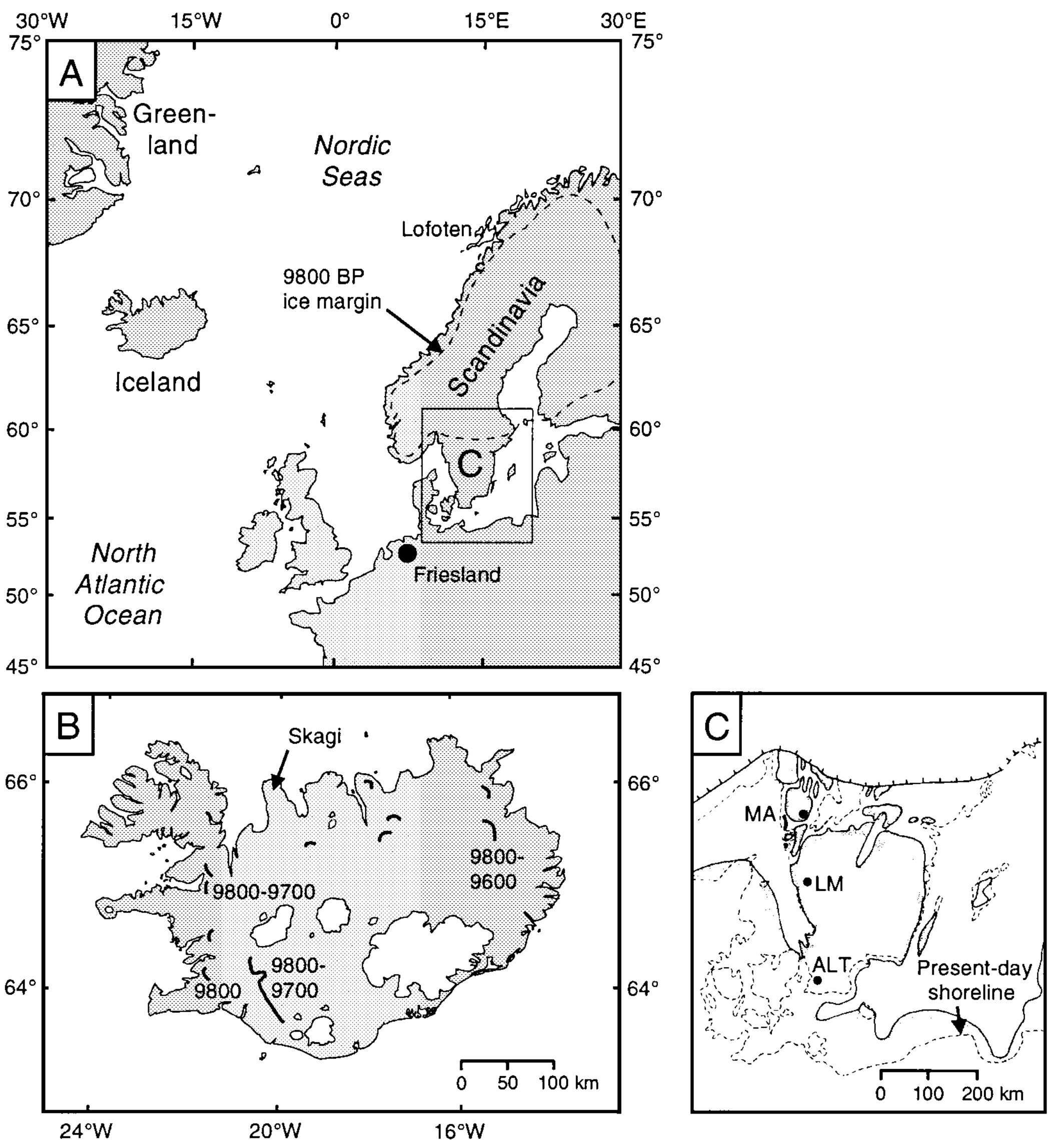

Figure 1 (A) Map of the Nordic Seas region between northwest Europe and eastern Greenland. Friesland in northern Germany is also indicated, as well as the position of the Scandinavian Ice Sheet margin at the time of the Preboreal oscillation. (B) Map of Iceland with black lines marking proposed early Preboreal ice-marginal positions (Ingolfsson, 1988; Hjartarson and Ingolfsson, 1988; Norðdahl, 1991; Sæmundsson, 1995; Norðdahl and Hjort, 1993; Norðdahl and Asbjornsdottir, 1995). Ages are in ^{14}C yr BP. (C) Palaeogeographic map of the southern Baltic region (cf. Fig. 1A) at the time of the Preboreal oscillation according to Bjorck (1995). The three Swedish sites (cf. Fig. 3) are indicated. Note the narrow western straits, which allowed only eastward intrusion of saline water during the short-lasting Preboreal oscillation.

detailed pollen studies from the early Preboreal of southern Sweden (e.g. Berglund, 1966; Bjorck and Digerfeldt, 1982, 1986, 1991; Svensson, 1989) are scrutinised, it is evident that the same vegetational pattern as described by Behre (1966) can be discerned, often including an increase in *Hippophae* pollen values. As this cooling also seems to have influenced areas with a more open pioneer-like vegetation than further south, its impact was possibly more widespread than thought previously. Bjorck *et al.* (1996) also showed that various lithological variables (carbon and allochtonous carbonates) seem to respond to the PBO.

The timing of the Preboreal oscillation

Based on a thorough examination of European Preboreal pollen data, ^{14}C dates and sedimentation rates, Behre (1978) concluded that the cooling began some hundreds of years after the end of the Younger Dryas and placed it between ca. 10 000 and ca. 9600 yr BP. Through three sets of closely AMS dated terrestrial macrofossils in lake sediments and pollen stratigraphical correlations, Bjorck *et al.* (1996) found that the cooling began when the 10 000–9900 ^{14}C yr plateau ended, i.e. 250 yr after the end of the Younger Dryas (Fig. 2). In the annually laminated Polish Lake Gościąż (Goslar *et al.*, 1995), a clear minimum in both elm pollen and authigenic carbonates can be seen 300–350 yr after the end of the Younger Dryas. In addition, closely spaced AMS measurements on a high-resolution core at the Troll site to the west of Norway (Haflidason *et al.*, 1995), show the distinct peak (ca. 40%) of the cold water form *Neogloboquadrina pachyderma* (s) in the planktonic foraminiferal fauna occurred 300 yr after the end of the Younger Dryas.

As Bjorck *et al.* (1996) pointed out, the rising Δ^{14}C values during the PBO make it difficult to assign it a well-defined ^{14}C age: it covers a time period of 100–150 calendar yr corresponding to 300–400 ^{14}C yr (Fig. 2). As it occurs between the two ^{14}C plateaux at 10 000–9900 and 9600–9500 yr BP, one would expect the most accurate age to be 9800–9700 ^{14}C yr BP. Large sigma values, or presence of slightly older or younger material with respect to the short PBO can, however, easily result in ages around 10 000 or 9500 ^{14}C yr BP. This dating problem may also explain the range of ages that have been proposed for different sets of data for a Preboreal cooling around the North Atlantic (Bjorck *et al.*, 1996). However, in terms of calendar years, Bjorck *et al.* (1996) dated the PBO to about 11 200–11 050 calendar yr BP. A recently effected overlap between the German oak and pine chronologies (Bernd Kromer and Marco Spurk, pers. comm.) has, however, shown that the calendar age of the pine chronology should be c. 80 yr older than suggested by Bjorck *et al.* (1996). This alteration places the PBO at c. 11 300–11 150 calendar yr BP.

Bio- and lithostratigraphical indications of a Preboreal oscillation in Norden

From a northwest European viewpoint the PBO seems to display some common pollen stratigraphical changes as described above. However, in individual pollen diagrams from Scandinavia, this event is not expressed in a uniform way. Iversen (1973) showed that its main character in

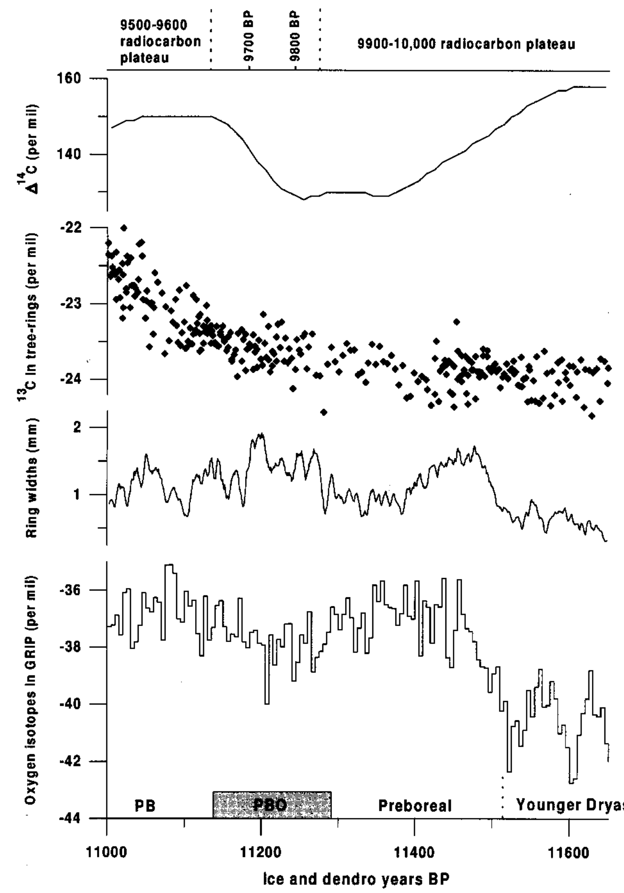

Figure 2 Four different high-resolution records from late Younger Dryas and Preboreal related to the correlations presented by Bjorck *et al.* (1996), with the addition of c. 80 years as a result of a recent overlap between the German oak and pine chronologies (B. Kromer and M. Spurk; personal communication). Based on Bjorck *et al.*'s (1996) reasoning this implies an almost perfect agreement between the dendro and GRIP chronologies. From top to bottom: the absolute chronostratigraphic positions for the two distinct Younger Dryas-Preboreal radiocarbon plateaux, situated on both sides of the Preboreal oscillation, as well as for ^{14}C ages of 9800 and 9700 BP. The Δ^{14}C values (‰) from the German oak-pine dendrochronology (Kromer and Becker, 1993) are according to Bjork *et al.* (1996) with the additional 80 years. The data were low-pass filtered by FFT smoothing. The δ^{13}O values (‰) are from the German pines (Becker *et al.*, 1991) and have a resolution of 1 to 5 years. Ring widths of German pines (Leuschner and Spurk, in prep.; Bjorck *et al.*, 1996) are the mean of all measured trees and rings with a 5 year running average. The δ^{18}O values (‰ in relation to standard mean ocean water) from GRIP are 5-year mean values (Johnsen *et al.*, 1992; Bjorck *et al.*, 1996) and the chronology is related to A.D. 1950. The absolute chronostratigraphic positions of Younger Dryas, Preboreal and PBO are marked in the bottom.

Denmark is a period of less *Betula* and *Filipendula* pollen and increased grass and sedge pollen, and he regarded it as a period during which the warmth-requiring tree species were halted in their northerly expansion. In southern Norway the signs of an early Preboreal climatic revertance are very subtle in pollen diagrams, but the detailed studies of Paus (1989a,b) show a short period of decreased pollen concentrations combined with decreased *Betula* and increased grass pollen frequencies. On Lofoten, northern Norway, Moe (1982) shows a very clear early Preboreal decline in tree-birch pollen and increased abundances of shrubs and light demanding plants such as *Koenigia islandica* and *Montia lamprosperma*. This assumed cooling was correlated to reports of glacial advances further inland (Andersen, 1975).

Generally speaking the PBO can, however, be difficult to detect, unless a detailed set of reliable ^{14}C dates is available from the end of the 10 000 ^{14}C yr plateau to the beginning of the next plateau at 9600 ^{14}C yr BP, between which the PBO occurs (Fig. 2).

The PBO is difficult to detect in the sediments unless its carbon and carbonate contents are analysed in detail. In Mjallsjon and Torreberga, and to a lesser extent Madtjarn, the carbon curves show a clear break during the PBO, preceded and followed by rising trends (Fig. 3). Carbonates occur only in one of the basins (ALT), but the rise in the content of detrital (allochtonous) carbonates, detected by stable isotope studies, seems to be the most clear evidence for a significant change of the sediments and their source during PBO (Fig. 3).

In addition, we present here three pollen diagrams (Fig. 4) along a north–south transect in southwestern Sweden (Fig. 1C) to show the somewhat diffuse character of this oscillation. The northernmost site, Lake Madtjarn (Fig. 4), is a 4500 m^2 small oligotrophic lake and was situated in a very sheltered position on a granite island not very far from the ice margin during the time of the PBO (Bjorck *et al.*, 1996). The pollen diagram shows some subtle, but still clear, changes in pollen composition: pollen values of *Betula alba* and *Pinus* decrease, whereas *Betula nana* pollen values display a clear increase and *Hippophae, Rumex, Empetrum* and *Filipendula* values increase slightly. In addition there is a clear minimum in pollen concentrations. There is, however, no clear increase in *Juniperus* pollen frequencies. A decrease in *Pediastrum* values occurs at the onset of PBO.

Lake Mjallsjon (0.1 km^2) lies in a narrow valley surrounded by gneiss bedrock. Here the PBO is shown as low *Pinus* pollen values, rising values of *Betula* undiff. (possibly *B. nana*), *Filipendula* and *Artemisia*. The *Juniperus* frequencies show a rise after a general decline. The PBO is also characterised by a *Pediastrum* minimum.

In ancient Lake Torreberga the pollen signals are slightly different from the other two lakes (Fig. 4). The setting of this site is also very different. This was a fairly large lake (3–4 km^2) in the Preboreal (Berglund and Digerfeldt, 1970) surrounded by clayey, carbonate-rich soils. To the north lies a flat plain and in the south the lake is bordered by a hummocky kettle-hole landscape, which may have contained dead ice through the Younger Dryas. The most distinct pollen-stratigraphical signal is the break in the rise of *Betula* pollen (which later resume high values) and total pollen concentrations, in combination with rising herb (mainly grass and sedge pollen) and pine pollen values. The latter is possibly an effect of the declining *Betula* frequencies. *Pediastrum* values are low after a distinct maximum.

We can conclude that the vegetation response in Swedish pollen diagrams is not clear-cut (Fig. 4). Data do, however, imply that tree vegetation experienced a short set-back during its initial expansion in the early Preboreal. This seems to have temporarily favoured some more light-demanding plants, but not the typical tundra plants we know from the Younger Dryas, such as, for example, *Artemisia* or Chenopodiaceae. There may have been a brief halt in the immigration wave of the shading trees and the woodland became slightly less dense as a consequence. This suggests that the cooling was less severe than during the Younger Dryas. The decreased carbon values imply that the biological production in the lakes decreased, perhaps as a result of longer seasons of ice-cover. The increased input of allochtonous carbonates indicate enhanced surface run-off as a consequence of increased precipitation and/or unstable soil conditions, which may have made light conditions in the lake worse and disfavoured *Pediastrum*.

Apart from the type of evidence presented above there are no records in southern Sweden that can be clearly related to a Preboreal oscillation. The ice-marginal deposits in Varmland, western Sweden, are certainly of Preboreal age (Lundqvist, 1988), but neither the age nor the character of any of these can assign them to a distinct ice readvance of a PBO origin. Further north, in southwest Norway, there are, however, reports of distinct moraines that seem to have been formed by Preboreal ice advances (e.g. Andersen, 1968; Anundsen, 1972; Vorren, 1973) and the ^{14}C ages of these readvances clearly fall in between the 10 000 and 9500 ^{14}C year plateaux.

Many other indications of a Preboreal oscillation are of circumstantial character. For example, Morner (1969) correlates a 2.5 m calculated sea-level regression to a Preboreal cooling, and lake-level studies in Europe (e.g. Digerfeldt, 1971; Gaillard, 1984; Bohncke and Wijmstra, 1988; Magny, 1995) suggest that the YD–PB transition was characterised by low lake levels followed by high levels (PBO?) before the marked drop in lake waters in the late Preboreal (Digerfeldt, 1988). Magny (1995) termed this high level the Remoray phase. It is thus likely that the PBO in large parts of Europe was cool and humid, which is supported by the increasing ring widths and fairly low isotope values in the German pines of this period (Becker *et al.*, 1991; Bjorck *et al.*, 1996), as shown in Fig. 2.

Correlations to the Swedish varved clays may also be circumstantial but it is undoubtedly an archive that may

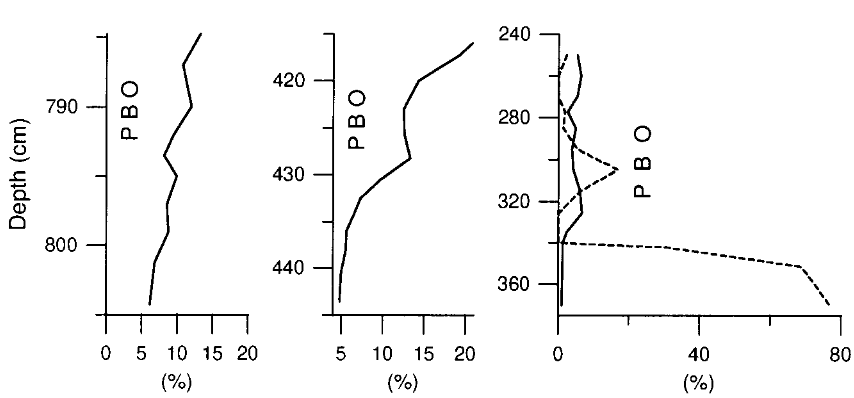

Figure 3 Organic carbon content in per cent dry weight in the late Younger Dryas and early Preboreal of the three lakes in Fig. 1C. From left to right: Lake Madtjarn (MA), Lake Mjallsjon (LM) and ancient Lake Torreberga (ALT). The dashed line in ALT shows the content of detrital carbonates, which was obtained by a mass balance calculation (see Bjorck *et al.*, 1996). The approximate stratigraphical position of the PBO is indicated.

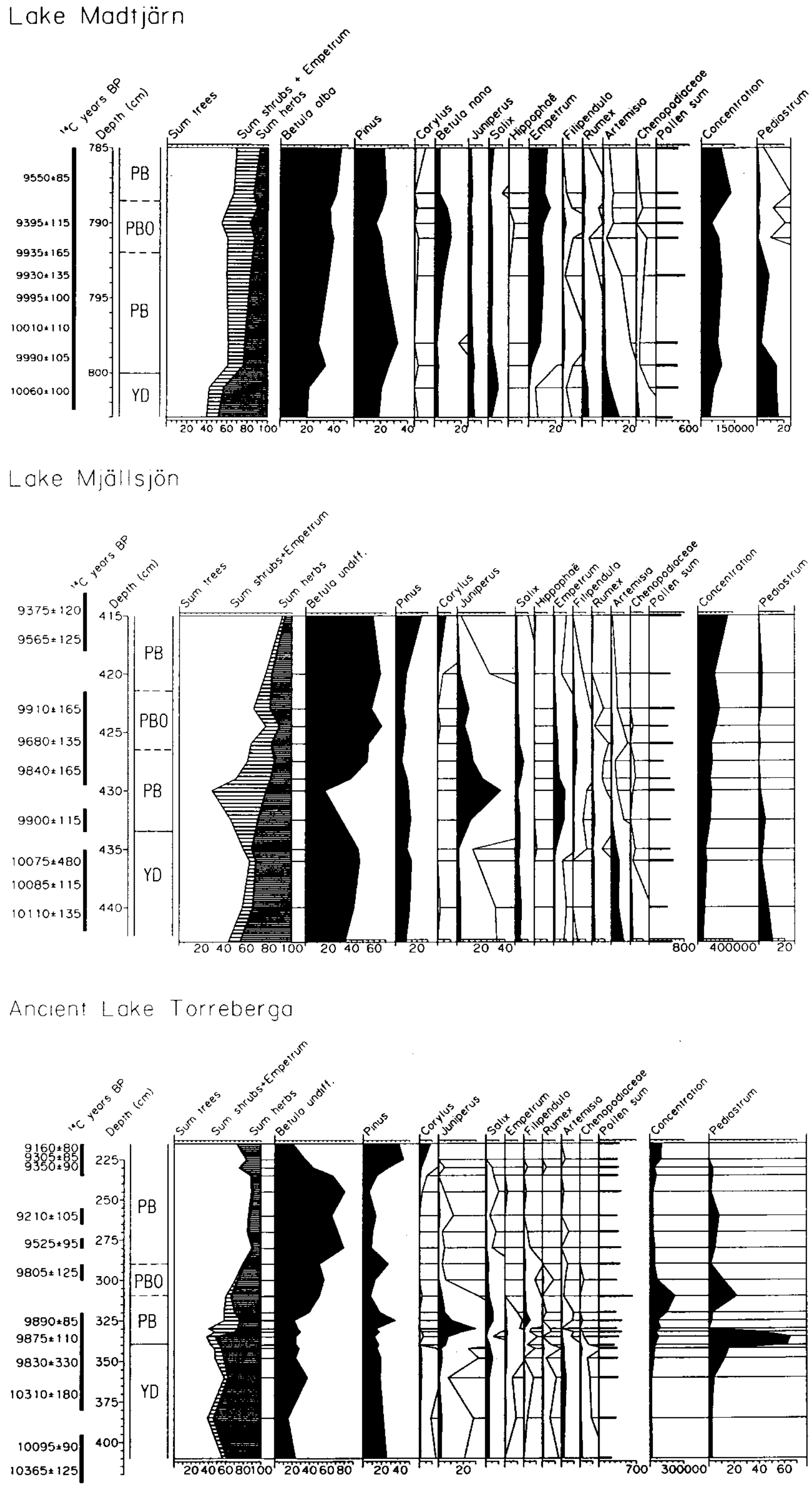

Figure 4 Pollen diagrams of late Younger Dryas–Preboreal from the three Swedish sites, showing percentage values of the most diagnostic pollen types, total pollen concentrations and *Pediastrum* frequencies. From top to bottom: Lake Madtjarn (MA), Lake Mjallsjon (LM), and Ancient Lake Torreberga (ALT). All ^{14}C dates with >1 mg C are presented (Bjorck *et al.*, 1996) and all AMS measurements were performed on identified terrestrial plant remains. The climatostratigraphical boundaries between Younger Dryas (YD), Preboreal (PB) and the Preboreal oscillation (PBO) are shown. Note the very different sedimentation rates between the sites. The sediments are as follows. Lake Madtjarn: 829.5–821.5 cm, brown clay gyttja, rich in mosses, upper boundary (UB) is rather sharp; 821.5–800 cm, brownish-grey clay gyttja with occasional mosses, UB is very gradual; 800–793 cm, clayey fine detritus gyttja, UB is very gradual; 793–785 cm, dark brown fine detritus gyttja. Lake Mjallsjon: 445–435 cm, light brown clay gyttja, UB is rather sharp; 435–429.5 cm, brownish-grey clayey fine detritus gyttja, UB is rather sharp; 429.5–418.5 cm, greenish brownish-black clayey fine detritus gyttja, UB is rather sharp; 418.5–415 cm, black fine detritus gyttja. Ancient Lake Torreberga: 410–341 cm, grey, slightly organic clay with thin sand laminae, UB is very gradual; 341–330.5 cm, brownish-grey, slightly calcareous gyttja clay, UB is rather sharp; 330.5–280 cm, beige and dark brown laminated calcareous clay gyttja, UB is rather sharp; 280–274.5 cm, yellow-beige lake marl, UB is rather gradual; 274.5–223 cm, blackish-brown and light-beige laminated lake marl, UB is very gradual; 223–215 cm, black fen peat.

have registered the PBO. Bjorck *et al*. (1996) suggest that this cooling may have been responsible for the short ingression of brackish water, characterising the mid-part of the Yoldia Sea, because cooler conditions may have decreased the melting of the Scandinavian Ice Sheet. This decrease in meltwater flux to the Baltic could have allowed salt water to penetrate through the narrow straits in south-central Sweden, situated between Skagerrak in the west and the Baltic proper in the east (Freden, 1988; Bjorck, 1995). The varve age of this event is dated to 10 430 varve yr BP., i.e. some 300 varve yr after the final drainage of the Baltic Ice Lake (Brunnberg, 1995). This drainage is supposed to have occurred shortly before the onset of the YD III pollen zone (Bjorck and Moller, 1987), corresponding to Berglund's (1966) YD–PB transition zone, which starts at the Younger Dryas–Preboreal boundary (Bjorck *et al*., 1996). Although the absolute varve ages of the Swedish Time Scale (Cato, 1987) for these events seem 700–900 yr too young, the time difference between the two events fits very well with other estimations. Correlations based on varves and ice recession lines over the Baltic Sea make a connection between the PBO, the brackish ingression and the third Salpausselka ridge in Finland plausible because it has been suggested that the Swedish Yoldia ingression occurred in connection with the formation of Salpausselka III (Brunnberg, 1995).

Records of a possible Preboreal oscillation on Iceland

Iceland and the Faeroe Islands hold key positions in the North Atlantic, being situated within the present area of North Atlantic deep water formation. Therefore, ocean ventilation changes may be expected to exert a strong influence on the regional climate.

The Younger Dryas glaciation in Iceland was extensive, with ice fronts reaching the coastal areas around most of the island (Ingolfsson and Norðdahl, 1994). In the fjord landscapes of eastern, northern and northwestern Iceland, ice fronts pushed into the marine environment in most fjords. This glacier advance culminated at about 10 600 yr BP. Glacial retreat from the Younger Dryas maximum was underway by ca. 10 300 yr BP. Relative sea-level was high in connection with the ice retreat. Although the post-Younger Dryas deglaciation of Iceland was rapid, and possibly completed by ca. 9000 yr BP, the ice retreat was delayed in several areas by early Preboreal readvances or temporary halts (Fig. 1B). In south-central Iceland, the inland ice-sheet margin stood at the Budi morainal complex, a more than 50-km-long ice-marginal deposit of moraine ridges and stratified outwash deposits, at 9800–9600 yr BP (Hjartarson and Ingolfsson, 1988; Ingolfsson and Norðdahl, 1994). Ice-marginal deposits of proposed early Preboreal age in other parts of the country are primary morainal ridges in tributary valleys above the marine limit as well as ice-contact deltas in the valley and fjord landscapes, deposited at or below the marine limit (Ingolfsson, 1988; Norðdahl, 1991; Sæmundsson, 1995; Norðdahl and Hjort, 1993). This glacial event has been dated to 9800–9600 yr BP (Hjartarson and Ingolfsson, 1988; Norðdahl and Ásbjornsdottir, 1995; Ingolfsson *et al*., in press.). An early Preboreal glacial event is also implied by a small-amplitude transgression at Skagi, northern Iceland (Fig. 1B), a few hundred years after the Younger Dryas–Preboreal transition (Rundgren *et al*., in press), as well as by a transgression in connection with an ice readvance in central western Iceland (Norðdahl and Ásbjornsdottir, 1995). This is strong evidence for a change in the mass balance of the Icelandic ice-sheet, and may be related to climate forcing in the early Preboreal.

From lake cores recently sampled from the Faeroe Islands (Bjorck *et al*., unpublished) there are strong indications of a fairly short, but distinct sediment change somewhere between the beginning of the Preboreal and the deposition of the Saksunarvatn Ash, dated to 8900–9000 ^{14}C yr BP (Bjorck *et al*., 1992; Birks *et al*., 1995). In total this period spans ca. 1300 calendar yr and the relative position of this more clastic-rich sediment horizon, overlain and underlain by more organic-rich deposits, makes its relationship to the PBO oscillation very likely. Apart from several ^{14}C dates we have, as yet, no detailed analyses from these Faeroe Island sediments, and hence our discussion of the proxy records from this region will focus on records from three Late Weichselian–early Holocene lake sediment sequences on the Skagi peninsula, northern Iceland (Fig. 1B).

All three lakes are situated close to the northern coast of the peninsula, and they had all been isolated from the sea by 9900 BP (Rundgren *et al*., in press). We directed our study in the core sections corresponding to early Preboreal regional pollen assemblage zone (RPAZ) Skagi-4, dated to 9900–9600 yr BP (Rundgren *et al*., in press). No sediments corresponding to RPAZ Skagi-3, deposited during the Younger Dryas (Rundgren, 1995; Rundgren *et al*., in press), have yet been recorded in Lake Geitakarlsvotn. This probably is due to the presence of coarse-grained sediments deposited in connection with basin isolation at 9900 BP (Rundgren *et al*., in press), making older sediments difficult to sample.

We performed detailed pollen analysis, including counting of coenobia of the green algal genus *Pediastrum*, on the cores from Lake Torfadalsvatn and Lake Geitakarlsvotn, and organic carbon content was determined in closely spaced samples from Lake Hraunsvatn and Lake Geitakarlsvotn (Fig. 5). A detailed carbon record could not be recovered from the Lake Torfadalsvatn sequence because most of this section of the core had been used for other purposes. In order to produce a representative record of lake productivity, one sample from the Lake Geitakarlsvotn sequence had to be omitted from the carbon analysis due to the presence of a tephra horizon at 5.00–5.01 m.

Only terrestrial taxa already present in all three lake sequences at the base of RPAZ Skagi-4, and continuously recorded throughout that zone, are shown in Fig. 5. We thereby exclude problems relating to the possible influence of climatic change on the appearance of new taxa. We also avoid making climatic inferences from pollen taxa with a more sporadic occurrence. The pollen data are presented as concentration values in order to clarify responses of individual pollen taxa.

All three lake sequences display a decrease in *Pediastrum* concentration values in the earlier part of RPAZ Skagi-4, and this is coincident with a lithological change in all sequences (Fig. 5). We regard the observed fall in *Pediastrum* concentration values to be a cooling response, reflecting lower water temperatures and prolonged winter ice cover. In addition, it may reflect an increase in suspended particles as a result of increased soil erosion.

Reduced limnic productivity is further indicated by a decrease in organic carbon content in all records except Lake Geitakarlsvotn. The minerogenic character of the sediments below 4.988 m in this sequence suggests significant allochtonous sediment input, which could be connected to threshold erosion following basin isolation (see above). This

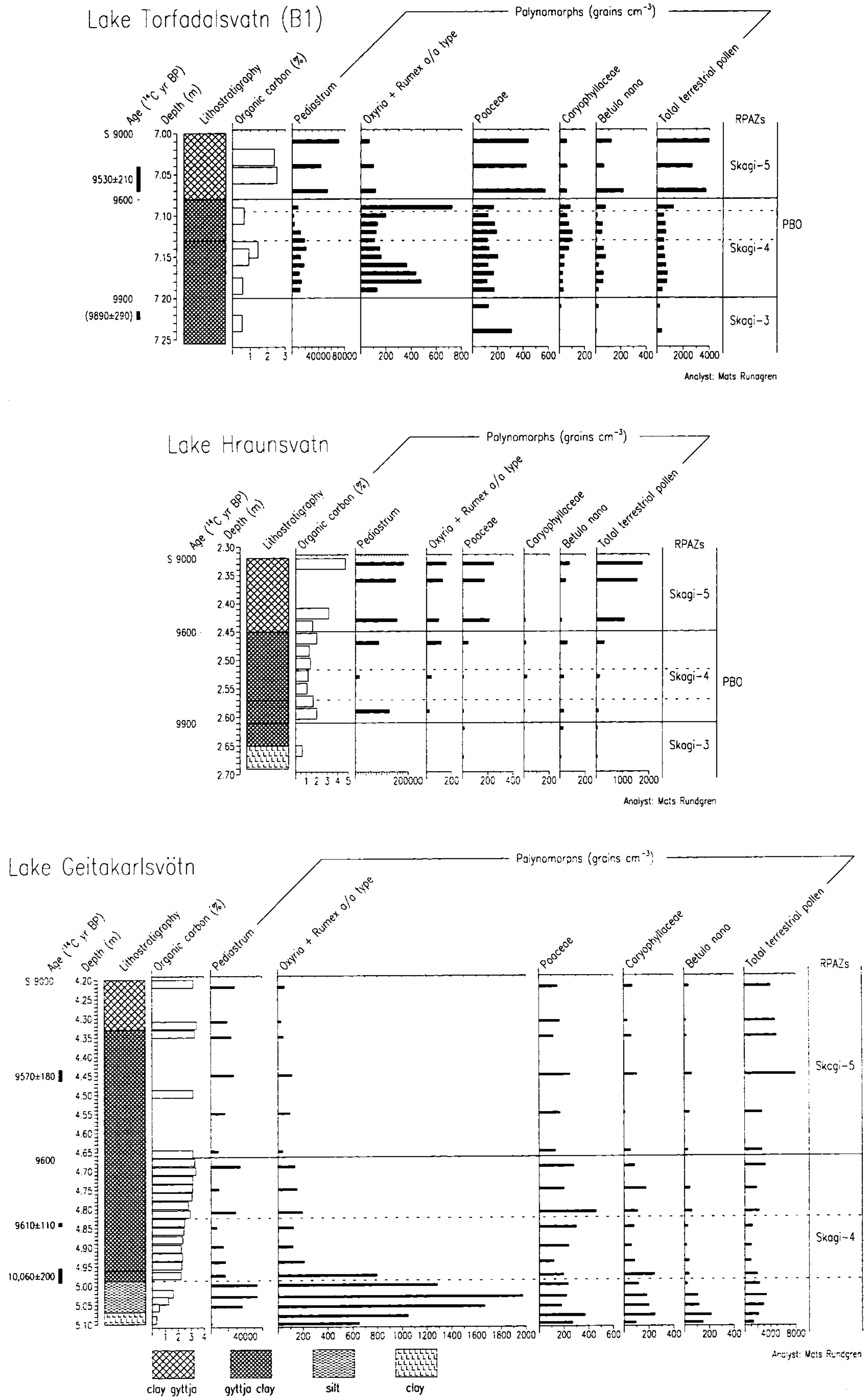

Figure 5 Late Younger Dryas–Preboreal pollen diagrams from the three Icelandic sites, showing concentration values of all terrestrial taxa that are continuously present in all three sequences throughout RPAZ Skagi-4 (9900–9600 yr BP; Rundgren *et al*., in press). In addition, lithostratigraphy, organic carbon content and concentration values for *Pediastrum* and total terrestrial pollen grains are shown. Radiocarbon dates, inferred ages for pollen zone boundaries, and the position of the Saksunarvatn ash (S) are displayed to the left. The lower ^{14}C date in Lake Torfadalsvatn is lithostratigraphically correlated from another core studied by Bjorck *et al*. (1992). Broken lines mark the inferred boundaries of the Preboreal oscillation (PBO).

may explain the absence of an organic carbon decline in this record.

This concordant limnic evidence suggests a regional cooling in northern Iceland at ca. 9800 yr BP that we attribute to the onset of the Preboreal oscillation. We further correlate the next rise in *Pediastrum* concentration values and associated increase in organic carbon content in all three lake sequences with the end of the PBO. Two samples, one of mosses and one of bulk sediment, were taken just below the inferred upper boundary of the PBO in Lake Geitakarlsvotn and submitted for AMS measurements. The moss sample (4.82–4.85 m) was <2 mg and yielded an age of 9400 ± 270 yr BP (LuA-4219). The high σ-value makes it useless for a precise dating of the PBO. However, the bulk date (4.83–4.84 m) gave a more precise age, 9610 ± 110 BP (LuA-4220), and is shown in Fig. 4. These dates, together with those situated above the Skagi-4–Skagi-5 boundary in Lake Torfadalsvatn and Lake Geitakarlsvotn, suggest that the PBO ended slightly before 9600 yr BP.

With the boundaries of the Preboreal oscillation defined, we can now evaluate the response of terrestrial vegetation to this cooling event. Due to the low resolution in the Lake Hraunsvatn pollen record, we limit our discussion to the other two records. *Oxyria* + *Rumex* a/a type, which is the dominant pollen taxon of RPAZ Skagi-4 (Rundgren, 1995; Rundgren *et al*., in press), peaks shortly after the Younger Dryas–Preboreal transition, and is already in decline before the onset of the PBO. *Oxyria digyna*, the species which we believe has contributed the majority of pollen grains referred to this pollen type, is a common plant in Icelandic ravines, on cliffs, gravelly slopes and rocky ground (Kristinsson, 1987). It has also been found to be an early coloniser on disturbed and recently deposited materials in the high Arctic (Edlund and Alt, 1989) and in front of receding Icelandic glaciers (Persson, 1964). *Oxyria digyna* is, moreover, often dominant on moist soils below snowbeds (Nordiska ministerrådet, 1984). The low concentration values of *Oxyria* + *Rumex* a/a type, both during the Younger Dryas and the PBO, indicate that climatic conditions were unfavourable for *Oxyria digyna* during these two events. The increased concentration values recorded immediately after these coolings possibly reflect phases of rapid colonisation during periods of limited competition by this opportunistic pioneer plant.

There is no clear common trend in the Poaceae pollen records, probably reflecting the fact that this pollen taxon includes a great variety of species with widely differing habitat requirements. Accordingly, no detailed climatic interpretations can be made from Poaceae pollen concentration data.

Both records show decreasing Caryophyllaceae pollen concentration values during the Preboreal oscillation, and the end of this event is followed by a short-lived expansion in both sequences. Similar responses are seen in the *Betula nana* records, which may reflect cooling and destabilisation of soils during the PBO. As this cooling event was soon followed by a regional expansion of dwarf-shrub vegetation on Skagi (Rundgren *et al*., in press), it is likely that *Betula nana* was growing close to its physiological limit during the PBO. Total terrestrial pollen concentration values also rise abruptly at the end of the PBO, although there is no clear common trend in pollen taxa during the earlier part of this event. This implies a gradual initial cooling, permitting individualistic behaviour of terrestrial taxa, followed by a more abrupt warming at the end of the PBO.

Pollen data show that northernmost Skagi was characterised by tundra vegetation during the whole of the Preboreal (Rundgren, 1995). As this is a vegetation type with species well adapted to arctic conditions, one should not expect dramatic vegetational changes to be associated with the PBO. Nevertheless, we can identify some responses to this event, which implies that it was associated with significant cooling.

To estimate the severity of the Preboreal oscillation, it may be appropriate to compare the responses of limnic and terrestrial parameters with those recorded during the Younger Dryas (Rundgren, 1995). Most of the terrestrial taxa discussed above show lower concentration values during the Younger Dryas compared with the PBO, and also the limnic parameters suggest that the Preboreal oscillation was less severe compared with the Younger Dryas. Organic carbon content in Lake Hraunsvatn sediments is greater during the PBO compared with the Younger Dryas (Rundgren *et al*., in press), and *Pediastrum* concentration values in Lake Torfadalsvatn are also higher. The similarity in responses of limnic and terrestrial proxy records to the PBO and the Younger Dryas suggests that these two climatic events were similar in character, but the PBO was clearly a less severe event. Therefore, it is possible that the polar front moved southwards during the PBO, but it probably did not reach south of Iceland as was the case during the Younger Dryas (e.g. Ruddiman and McIntyre, 1981; Koç and Jansen, 1994).

Early Preboreal climate in Greenland

The evidence for Late Weichselian–early Holocene climatic change in Greenland has recently been reviewed by Funder and Hansen (1996), who suggested that the initial retreat of the Greenland ice-sheet after the Last Glacial Maximum was primarily a result of rising global sea-level, which destabilised the marine portions of the ice-sheet. There is no evidence from Greenland of a Younger Dryas glacial advance, probably because the climate was too cold and dry for ice growth. Instead, readvance and unstable ice margins characterised the Early Preboreal until ca. 9500 yr BP in many parts of the country.

Hjort (1981) suggested that the Milne Land stade began at ca. 10 300 yr BP, and in north Greenland, Kelly and Bennike (1992) found that the correlative Warming Land stade began after 10 500 yr BP. This seems to indicate that the readvance began after the YD–PB transition, and may reflect the establishment of the post-glacial atmospheric and oceanic circulation pattern around Greenland, with increased precipitation but low temperatures (Funder and Hansen, 1996). More significantly, in all parts of the country this period ended in a phase of rapid ice-margin retreat, showing that temperatures finally became high enough to promote melting of land-based ice. A time-scale for this transition, the end of the PBO, has been established by dating the marine invasion in areas that were ice covered. In East Greenland this was dated to ca. 9800 yr BP (Funder, 1978), in north Greenland to shortly before 9500 yr BP (Kelly and Bennike, 1992), and in areas of west Greenland the similar Taserqat Stage was dated to ca. 9500 yr BP (Ten Brink and Weidick, 1974).

The environmental change at the end of the PBO is also seen in lake sequences, where the beginning of organic sedimentation have been dated at 10 000–9500 yr BP in both the south, west and east (Funder and Fredskild, 1989; Bocher and Fredskild, 1993). It is also apparent in the temporal distribution of more than 1000 ^{14}C dates from all parts of Greenland. This shows a steep increase in the number of dates from only five dates in the 10 250–9750 yr BP interval, rising to 30 in the 9500–9250 yr BP interval (Funder and Hansen, 1996). The dates come from both coastal and some inner fjord localities, and suggest that these areas were rapidly deglaciated at that time. The spread of dates also implies that in lake basins and shallow-marine environments, from where most of the ^{14}C dates originate, there was a dramatic and climatically conditioned increase in organic productivity at that time. The implication of this is that in Greenland the PBO and its termination seem to have assumed the climatic role of the Younger Dryas and the YD–PB transition in more southerly areas.

Discussion and conclusions

There is strong and independent evidence for a short Preboreal event in the eastern and northern parts of the North Atlantic region (Fig. 1), dated to c. 11 300–11 150 calendar yr BP (Fig. 2). A comparable event may also have been recorded at the mouth of the Hudson Strait in Arctic Canada, where a major ice-stream advance (Kaufman *et al*., 1993) has been dated to 9900–9600 yr BP. The various proxy data also demonstrate that it occurred in the first half of the Preboreal, some hundreds of years after the termination of the Younger Dryas. Although different dating methods have been used, the miscellaneous records suggest that it was a broadly synchronous event. However, the signature of this event may vary between different areas and regions so we cannot, without exploring leads and lags of the various systems, state with certainty that it represents a larger regional cooling event.

In the case of ice advances, one can argue that maximum summer insolation together with greatly increased precipitation in the early Preboreal (Alley *et al*., 1993) may be the trigger for dynamically caused readvances. The other proxy data we have presented here do, however, suggest that it was a climatically induced oscillation. This is very clear in Norway and Iceland, where ice readvances (or stillstands) seem to occur simultaneously with records of cooling in lake sediments. The Swedish pollen data indicate that more open vegetation was favoured for a short time at the expense of the immigrating shading trees and shrubs. Pollen production also appears to have decreased and the lower *Pediastrum* and carbon values suggest that lake productivity fell in both Icelandic and Swedish aquatic systems. This may have been caused by a combination of colder water during the growing season, longer ice-cover seasons and an increase in suspended sediments in lake water, due to temporarily increased soil erosion. The well-dated Swedish lacustrine sediments clearly indicate that the PBO occurred between the two early Preboreal radiocarbon plateaux (Figs 2 and 4). By placing it between these two periods we can also conclude that it was characterised by a significant rise in atmospheric $\Delta^{14}C$ (Bjorck *et al*., 1996; Fig. 2). Such a distinct rise can be explained in terms of decreased ocean ventilation in the North Atlantic, caused by a significant flux of meltwater from the rapidly disintegrating continental ice sheets, including the huge amounts of fresh water from the 25 m lowering of the Baltic Ice Lake (Bjorck, 1995). If increased meltwater flux and decreased thermohaline circulation in the Nordic Seas were the triggers for the PBO, we would also expect a southerly displacement of the Polar Front and the westerly anticyclonic tracks in northwest Europe. This is also broadly supported by the Icelandic data, which suggest colder conditions, and the European records, which imply a cooler and more humid climate.

The study by Haflidason *et al*. (1995) suggests that the GRIP and Norwegian Sea records are in phase with each other at both the YD–PB boundary and during the PBO. This is confirmed by our data, which imply that the response in the atmospheric and terrestrial systems to this possibly marine forced oscillation was more or less immediate. In spite of what may be a common triggering mechanism, however, namely meltwater peak IB, the brief but distinct PBO should not be correlated to more regionally restricted, less well-dated coolings, such as the melt-water induced cooling in the Great Lakes region between 10 100–8200 yr BP (Anderson *et al*., 1997) and the 10 000–8800 yr BP cold event in northern Alaska (Epstein, 1995).

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