Synchronized Terrestrial-Atmospheric Deglacial Records Around the North Atlantic

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On the basis of synchronization of three carbon-14 (^{14}C)–dated lacustrine sequences from Sweden with tree ring and ice core records, the absolute age of the Younger Dryas–Preboreal climatic shift was determined to be 11,450 to 11,390 ± 80 years before the present. A 150-year-long cooling in the early Preboreal, associated with rising $\Delta^{14}C$ values, is evident in all records and indicates an ocean ventilation change. This cooling is similar to earlier deglacial coolings, and box-model calculations suggest that they all may have been the result of increased freshwater forcing that inhibited the strength of the North Atlantic heat conveyor, although the Younger Dryas may have begun as an anomalous meltwater event.

The rapid melting of ice sheets during the earliest Holocene must have had an important effect on climate because of the suddenly increased impact of fresh water on the ocean. The major $\Delta^{14}C$ changes of the Last Termination (the deglaciation at the end of the last glaciation), which are reflected in offsets of the radiocarbon time scale, are related to changes in ocean ventilation and deep water formation (*1–4*) as well as to ^{14}C production changes resulting from changes in geomagnetic field strength and solar activity. The effect of production changes can be ignored, because the geomagnetic signal varies on a longer time scale and the solar modulation alone is inadequate for the oscillations of the last deglaciation (*2*). A year-to-year correlation between the $\Delta^{14}C$ record (from absolutely dated and ^{14}C-dated tree rings) and the temperature record (from ice cores) is needed to address the link between ocean ventilation and climate change. Here, we combined high-resolution tree ring, ice core, and lacustrine records of the Younger Dryas–Preboreal (YD-PB) transition to connect the tree rings with the ice cores through lacustrine sediments in order to attain a common chronology and integrate these different climatic signals. This integration greatly improves understanding of the mechanisms and rates of change that underlie the deglacial climatic oscillations and larger climatic changes in general.

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A synchronized chronology. The calibration of ^{14}C ages older than 9000 years to absolute years on the basis of tree rings has been uncertain (*5*), because the YD-PB pine chronology is not securely linked to the younger, absolutely dated German oak tree ring chronology (Fig. 1A). In addition, the calibration set must be corrected. First, in a recent comparison between the oak chronologies of the Hohenheim and Göttingen tree-ring laboratories (*6*), it was discovered that 41 rings are missing in the Hohenheim oak chronology at 7200 dendrochronological (dendro) years B.P. (reference year, 1950 A.D.). Second, after recent additions of new trees to both chronologies, the previous synchronization of the pine and oak chronologies (*7*) appears less plausible. Instead, the general sloping trend of the $\Delta^{14}C$ values (Fig. 2A) implies that the pine chronology is ~120 years older than previously thought (*5*). With these corrections the pine chronology becomes 161 ± 80 years older (Fig. 2A). A new tentative tree ring–based calibration is achieved with the 10,000–^{14}C year plateau ending ~11,200 dendro years B.P. (Fig. 2B).

A lithologic change (LC) in lacustrine sediments in southern Sweden marks the YD-PB transition. It is usually seen as a change from grayish to brownish sediments, reflecting increased organic content. For dating purposes, we define LC as the first lithologic change seen in lacustrine sediments at the YD-PB transition. Although it may be sharp or gradual depending on the sedimentation rate, it can usually be defined within 1 to 5 mm. We dated LC in three lakes along a north-south transect (Fig. 1, A and B) using dense accelerator mass spectrometry (AMS) ^{14}C measurements on plant remains (Table 1). The AMS dates show that LC in all the lakes occurred well before the end of the ^{14}C plateau at 10,000 to 9900 ^{14}C years B.P. (*5*), and pollen, macrofossils, and stable isotopes (*8*) also change abruptly at this level (Table 1 and Fig. 3). These records imply that air and water temperatures increased and aquatic production rose in shallow limnic systems.

To establish the calendar (dendro) age

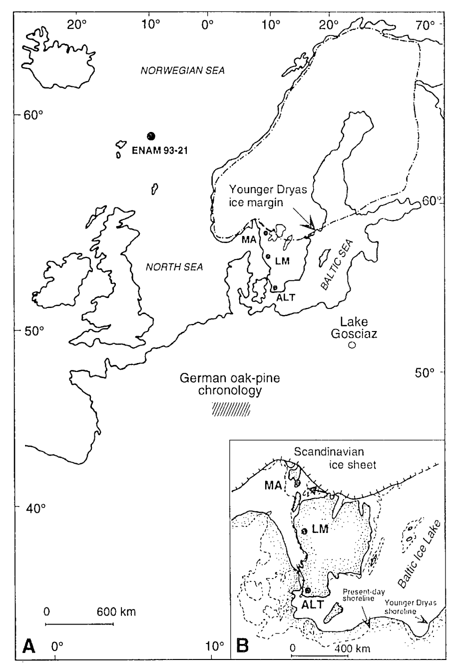
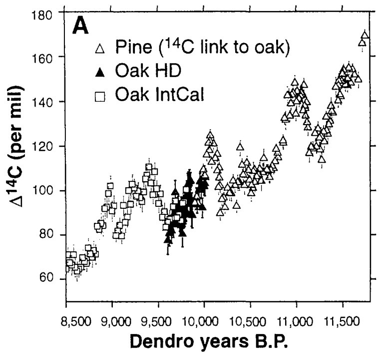


Fig. 1. (**A**) Map of northwest central Europe with the three Swedish lakes [Torreberga (ALT), Lake Madtjärn (MA), and Lake Mjällsjön (LM)], Lake Gosciaz (*3*) in central Poland, the ENAM 93-21 core (*50*), and the main find area for the YD-PB pine stumps (*5–7*) in southern Germany. The Scandinavian ice sheet margin in the late YD is also indicated. (**B**) The paleogeographic situation in south Scandinavia with the extent of the Baltic Ice Lake just before the YD-PB boundary (*44*), when the 25-m up-dammed lake still drained through the Danish-Swedish Strait. The pathway of the subsequent catastrophic drainage in south central Sweden is indicated by an arrow.

of LC, we smoothed our AMS dating series (Fig. 2B). The age is 11,450 to 11,400 years B.P. with the new dendrochronology. Ring widths in German pines increase markedly between 11,450 and 11,390 years B.P. (Fig. 4). This increase suggests a general change in pine type, resembling a shift from tree-ring types of modern high-altitude (cold-tolerant) to low-altitude pines (9), and constitutes the first response to the Preboreal warming in the dendrochronological record. The age of the YD-PB boundary can thus be established from Scandinavia to central and eastern (3) Europe (Fig. 1A). This dendro age is 75 years younger than the $\delta^{18}O$ rise in the Greenland Ice Core Project (GRIP) ice core (*10–12*) when the ice chronology is corrected for a reference date of 1950 A.D. If the ^{14}C age of 8900 years B.P. of the Saksunarvatn ash (*13*) is calibrated against our new dendrochronology, an age of ~10,050 years B.P. is obtained. The age for the same tephra in the Greenland ice is 10,065 years B.P. (*14*) if the GRIP chronology is reduced by 75 years.

We conclude that the rapid warming registered in shallow, temperature-sensitive lakes in northern Europe, independent of vicinity to the ice sheet (Fig. 1B), corresponds to the atmospheric warming in the GRIP core (*10–12*) at 11,525 to 11,465 years B.P. and to the abruptly increased tree-ring widths in the south German pines at 11,450 to 11,390 years B.P. We can thus synchronize the GRIP and dendrochronological records either by a tentative addition of 75 years to the tree ring chronology or by a corresponding subtraction from the GRIP chronology. We chose the latter and refer all absolute dates to this setting. This synchronization is well within quoted errors for the two time scales, and it produces a close fit between the two proxy records (Fig. 4).

A Preboreal oscillation. The synchronization shows that the first 150 years of the Preboreal were characterized by steadily ris-



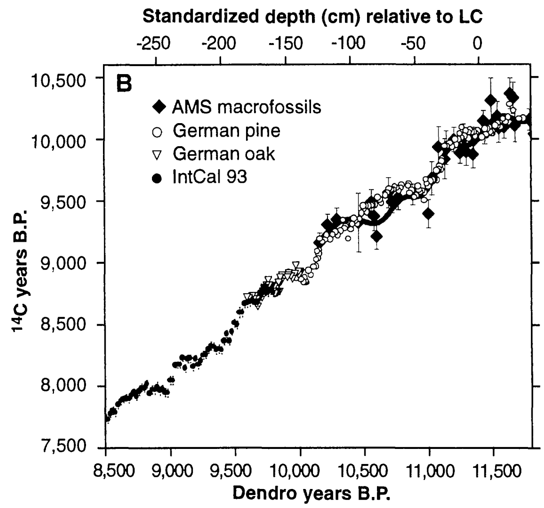


Fig. 2. (**A**) $\Delta^{14}C$ as derived from the IntCal 93 (*51*) German oak and pine chronology (*5*) and additional oak data (HD). The IntCal 93 and German oak series are corrected for a 41-year gap. The German pine is matched to the absolutely dated oak chronology by aligning the long-term $\Delta^{14}C$ trend. The uncertainty of this link is difficult to quantify, but the long-term slope of the ^{14}C concentration allows for a range of ±80 years. (**B**) Standardized radiocarbon dates (◆) from the three lakes smoothed against the new ^{14}C calibration set. Standardization was based on the known thickness of the YD in each lake, the calibrated age of the youngest ^{14}C-dated levels, and the following assumptions: LC corresponds to the YD-PB boundary; sedimentation rates are more or less continuous during the YD and change at LC, where they are continuous up to the youngest PB dates; the YD is 1150 years long (*3, 10–12*); and sedimentation rates are standardized to 1 mm year^{-1}. The solid line is the fast Fourier transform (FFT)–smoothed curve through the AMS dates versus standardized depth (cut-off length of 24 cm, equivalent to a low-pass filter of 240 years). Note the two ^{14}C age plateaus at 10,000 and 9600 ^{14}C years B.P. and the rapid transition between the two plateaus.

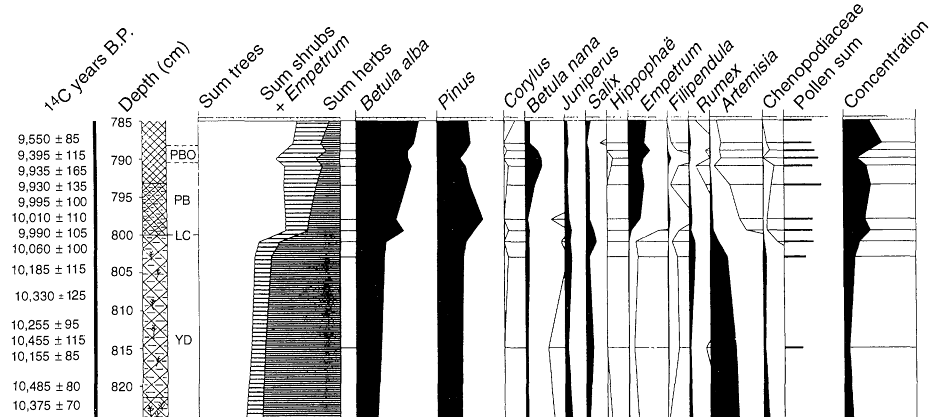


Fig. 3. Pollen diagram from Lake Madtjärn with the most diagnostic AL-PB pollen types, ^{14}C dates (Table 1), a sediment log, and the positions of LC, Allerød (AL), the YD, and the PBO. The outer pollen curve is a 10× exaggeration of the inner solid curve. The values at the bottom are percentages of total pollen from terriphytic spermatophytes (that is, the pollen sum at the right). Concentration represents the amount of terriphytic spermatophyte pollen in grains per cubic centimeter. The sediments are as follows: 900 to 835 cm, blackish blue-gray silty clay (marine) with some FeS coloring, upper boundary (UB) is very gradual; 835 to 832.5 cm, gray to beige gyttja clay, UB is rather gradual; 832.5 to 829.5 cm, greenish-gray clay gyttja with moss remains, UB is rather sharp; 829.5 to 821.5 cm, brown clay gyttja, rich in mosses, UB is rather sharp; 821.5 to 800 cm, brownish-gray clay gyttja with occasional mosses, UB is very gradual; 800 to 793 cm, brown clayey fine detritus gyttja, UB is very gradual; 793 to 737 cm, dark brown fine detritus gyttja.

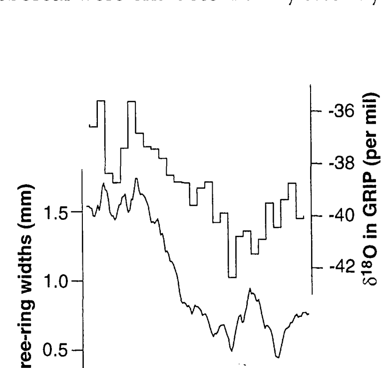


Fig. 4. By a tentative 75-year reduction of the GRIP chronology (1950 A.D.) and the addition of 161 years to the dendrochronology, a close fit is achieved between the 5-year mean $\delta^{18}O$ record from GRIP (upper curve) and the 5-year running average of the ring-width record of the German pines (lower curve) (*6*) at the YD-PB boundary in both records.

ing $\delta^{18}O$ values and decreasing Ca values (15) in the ice. This is in phase with declining $\Delta^{14}C$ values (Fig. 2A) and suggests good ocean ventilation (1). At 11,300 years B.P., air temperatures (inferred from decreasing $\delta^{18}O$ values in the GRIP record) slowly began to decrease when the $\Delta^{14}C$ reduction phased out after the long decline, which implies that ventilation was not as efficient as before. About 300 years after the YD-PB boundary, $\Delta^{14}C$ values rise abruptly (24 per mil between 11,170 and 11,050 years B.P.; Fig. 2A), indicating an ocean ventilation minimum (1). At the corresponding time, the GRIP record shows a distinct $\delta^{18}O$ and temperature minimum (Fig. 5) together with high Ca (15) and low CH_4 (16) values. This cooling is also well displayed (or even more pronounced) in the Dye 3, Camp Century, Renland, and Greenland Ice Sheet Project 2 (GISP2) records (17). Thereafter, the GRIP record shows a slow return to higher air temperatures while the $\Delta^{14}C$ rise begins to phase out, suggesting a transition into slowly better mixing rates. At 11,000 years B.P., the GRIP values attain normal Holocene levels, and this temperature recovery was thus completed ~50 years after the $\Delta^{14}C$ rise had ceased (Fig. 5) and good ventilation had resumed. In addition, the early Preboreal cooling deduced from foraminiferal and diatom changes in the Norwegian Sea (18, 19) lends further support for the ocean ventilation-cooling relation that we infer from our comparisons.

The lake sequences we analyzed (8) show that LC is synchronous with the onset of the YD-PB pollen transition zone (20, 21) in southern Scandinavia, characterized by plants like *Filipendula*, *Juniperus*, and *Empetrum*, which were already present and were instantly favored by warmer temperatures (Fig. 3). At the end of the 10,000–^{14}C year plateau (Fig. 3 and Table 1), which coincides with the onset of the Preboreal oscillation (PBO) in GRIP, lake sediments display changes (Fig. 6) that are most likely related to a harsher climate. The pioneer vegetation in Sweden was fairly insensitive to this cooling, but in some of our pollen records, decreases in tree pollen and total pollen abundance and increases in herb and shrub pollen abundance are evident (Fig. 3). This event is most likely the equivalent of an early Preboreal climatic cooling, for which there is evidence from at least 31 European pollen sites starting ~300 years after the termination of the YD (22). In Lake Gosciaz (3), a clear minimum in elm pollen and $\delta^{18}O$ of authigenic carbonates is also seen 300 to 350 years after the end of the YD. In addition, tree ring widths (Fig. 5), isotopes in tree rings (23), and lake level studies (24) imply that the PBO was char-

Table 1. AMS dates (excluding nine dates with <1 mg C) and other records from the three lake sites. Abbreviations for macrofossils: B, *Betula* sp.; D, *Dryas octopetala*; Di, *Distichium* sp.; E, *Empetrum nigrum*; M, *Menyanthes trifoliata*; Ny, *Nymphaea* sp.; P, *Polygonum viviparum*; Pi, *Pinus sylvestris*; Po, *Populus tremula*; Pol, *Polytrichum* sp.; S, *Salix* sp.; Sc, *Scirpus lacustris*; So, *Saxifraga oppositifolia*; and W, undetermined wood. Calibrated ages [(51) with additional 161 years] are shown [without the often high σ values on the 10,000– to 9900–^{14}C year plateau (5)] for dates around LC when the age is covered by the dendrochronology-based calibration. The AL-YD boundary in the Lake Madtjärn record is at 832.5 cm. The sedimentation rate in Torreberga is three times that in the other lakes. Climate proxy changes at LC are as follows: For Lake Madtjärn, relative increase of organic C, 28%; mean temperature rise in July and January [according to mutual climatic range analysis of beetles (53)], 9.3° and 18.2°C, respectively. For Lake Mjällsjön, relative increase of organic C, 25%. For Torreberga, relative increase of organic C and carbonate C, 108% and 116%, respectively; absolute changes in $\delta^{13}C$ and $\delta^{18}O$ of *Candona* ostracodes, −2.3 and +6.1 per mil, respectively.

Depth (cm) for AMS date	Dated macrofossils	^{14}C age B.P. ($\pm\sigma$)	Calibrated age B.P.	^{13}C (per mil) of dated material	AMS sample number
Lake Madtjärn (LC depth, 800 cm)					
773 to 777	B + S	9,670 ± 195		−29.30	Ua-10258
777 to 781	B + S	9,490 ± 100		−28.50	Ua-10259
781 to 785	B + S + D	9,490 ± 90		−28.70	Ua-10260
785 to 789	B + D + E + S	9,550 ± 85		−28.44	Ua-10223
789 to 791	B + D	9,395 ± 115		−28.21	Ua-10261
791 to 793	B + D + S	9,935 ± 165	11,190	−28.02	Ua-10262
793 to 794	B	9,930 ± 135	11,160	−28.12	Ua-10263
794 to 796	B + D + S	9,995 ± 100	11,330	−29.22	Ua-10264
796 to 798	B + D + S	10,010 ± 110	11,365	−28.31	Ua-10265
798 to 800	D + S	9,990 ± 105	11,330	−27.81	Ua-10266
800 to 802.5	B + D + S	10,060 ± 100	11,490	−27.90	Ua-10267
802.5 to 806	D + S	10,185 ± 115	12,080	−28.02	Ua-10268
806 to 810	D + S	10,330 ± 125		−28.38	Ua-10269
810 to 813.5	D + S	10,155 ± 85	12,000	−28.29	Ua-10222
813.5 to 814.5	S	10,455 ± 115		−28.48	Ua-10270
814.5 to 818	D + P + S	10,375 ± 70		−28.35	Ua-10221
818 to 821.5	D + P + S	10,425 ± 80		−28.07	Ua-10220
821.5 to 824	S	10,340 ± 70		−27.82	Ua-10219
824 to 827	D + S	10,440 ± 65		−27.95	Ua-10218
827 to 829.5	S	10,400 ± 80		−28.59	Ua-10217
829.5 to 831	D + S + So	10,625 ± 75		−28.49	Ua-10216
831 to 832.5	D + E + O + P + S + So	10,625 ± 70		−29.68	Ua-10215
832.5 to 835	D + E + S + So	10,995 ± 75		−28.96	Ua-10214
835 to 845	S	10,820 ± 85		−27.85	Ua-10272
835 to 845	D + E + P	11,065 ± 150		−27.00	Ua-10271
845 to 855	E + S	10,750 ± 100		−26.85	Ua-10274
875 to 885	S	11,630 ± 190		−28.56	Ua-10281
885 to 895	S	11,470 ± 135		−29.53	Ua-10352
Lake Mjällsjön (LC depth, 435 cm)					
413 to 416	Pi	9,375 ± 120		−27.24	Ua-4643
416 to 418.5	B + Pi	9,565 ± 125		−28.22	Ua-4644
421.5 to 424.5	B	9,910 ± 165	11,160	−27.99	Ua-4646
424.5 to 427	B + Pi	9,680 ± 135	11,090	−28.92	Ua-4647
427 to 429.5	B + Po	9,840 ± 165	11,150	−27.44	Ua-4648
431.5 to 433.5	B + E + Po + S	9,900 ± 115	11,150	−28.09	Ua-4650
437 to 439	B + Pol + S	10,085 ± 115	11,640	−28.70	Ua-4654
439 to 442	Pol + S	10,110 ± 135	11,860	−27.78	Ua-4656
442 to 445	B + D + Di + Pol	10,040 ± 160	11,465	−26.10	Ua-4657
Torreberga (LC depth, 341 cm)					
215 to 220	M	9,160 ± 80		−26.58	Ua-4468
220 to 223	M	9,305 ± 85		−26.70	Ua-4467
223 to 235	M	9,350 ± 90		−26.96	Ua-4466
255 to 265	B + M + Po	9,210 ± 105		−27.13	Ua-4465
274.5 to 280	Ny	9,525 ± 95		−25.64	Ua-4464
290 to 300	Sc	9,805 ± 125	11,150	−27.78	Ua-4463
320 to 330.5	B + Po + Sc	9,890 ± 85	11,150	−29.24	Ua-4462
330.5 to 341	B + S	9,875 ± 110	11,150	−31.32	Ua-4461
341 to 360	W	10,145 ± 85	11,970	−29.77	Ua-4460
360 to 380	B + D	10,310 ± 180		−30.26	Ua-4458
395 to 410	B + D + S	10,095 ± 90	11,820	−30.63	Ua-4457
410 to 425	B + D	10,365 ± 125		−30.27	Ua-4456
425 to 450	B + D + S	10,510 ± 235		−31.29	Ua-4455

acterized by a humid, cool climate, followed by arid, warm conditions slightly before 11,000 years B.P., when ring widths decreased, lake levels fell (*24*), and isotopes in tree rings increased (*23*).

The timing of the PBO coincides with the short and enigmatic brackish phase of the Yoldia Sea stage of the Baltic Sea. It may reflect decreased meltwater flux from the Scandinavian ice sheet into the Baltic basin that resulted in a temporary entry of saline bottom water. Our dendro ages of LC and the PBO are up to 700 years older than corresponding varve ages of the recently revised Swedish clay varve chronology (*25*). Until the missing varves are identified, parts of this time scale should be regarded as uncertain.

There are several other records of a short, early Preboreal cooling around the North Atlantic and even the tropical Atlantic (*26*). These studies have yielded ages between 9900 and 9500 ^{14}C years B.P., which have been difficult to correlate to one single event. Because Δ^{14}C values rise rapidly through the 150-year PBO, it is almost impossible to assign it a well-defined ^{14}C age. Thus, we suggest that most, if not all, of these early Preboreal cooling records belong to the same event.

The PBO followed the onset of rapid melting of the Scandinavian–Laurentide–Barents Sea ice sheets. Together with the final drainage of the Baltic Ice Lake (Fig. 1B), which delivered 0.15 to 0.3 sverdrups of fresh water in 1 to 2 years to the North Sea, this melting led to a major influx of fresh water to the North Atlantic. This meltwater input, possibly corresponding to meltwater peak MWP IB (*27*), may have been the trigger for the PBO.

Earlier deglacial oscillations. Detailed pollen and ^{14}C records across the Allerød–Younger Dryas (AL-YD) boundary from Lake Madtjärn (Fig. 3 and Table 1) show that Δ^{14}C values increase abruptly (Fig. 5) by 30 to 60 per mil in <2 cm (<70 years) over this boundary. Thus, ^{14}C ages become suddenly younger after dates of 11,000 to 10,800 ^{14}C years B.P. (Table 1). Two high-precision data sets of floating tree ring series also show a decrease in Δ^{14}C over at least 120 dendro years in the interval 11,000 to 10,900 ^{14}C years B.P. and an abrupt jump to AMS dates of ~10,600 years B.P. for the onset of the YD (*28, 29*). These records imply that a period of good ocean ventilation suddenly came to an end (*1*) at the AL-YD boundary.

With a normal production of 1% ^{14}C per 80 years and with 90% of the total C reservoir in the intermediate-deep ocean (*30*), a rise of 30 to 60 per mil atmospheric Δ^{14}C in 80 years would mean an approximately 35 to 70% reduction in the exchange between the intermediate-deep ocean and the surface ocean–atmosphere reservoirs. We made several runs with a box-diffusion model (*31*) to test whether a rapid Δ^{14}C rise followed by a slower rise and a subsequent gradual decrease could be caused by a 200-year-long reduction in deep-ocean ventilation at the onset of the YD, followed by a gradual return to the Holocene mode (Fig. 5). The Δ^{14}C variations were modeled by varying the diffusional mixing of the deep ocean and the exchange rates between the atmosphere and the mixed layer (Fig. 7). The resulting change in atmospheric and mixed-layer Δ^{14}C is consistent with our data and also explains the apparent rise in the atmosphere–mixed layer Δ^{14}C difference ("marine reservoir age") during the YD (*32*). Such a gradual increase in ocean ventilation over many centuries may seem unrealistic. However, because of the damping of the atmospheric ^{14}C signal, an intermittent operation of the ocean conveyor with a gradually increasing duration of exchange would leave an equivalent imprint in the ^{14}C record. The reduction in the exchange between the atmosphere and the ocean mixed layer can be explained by the

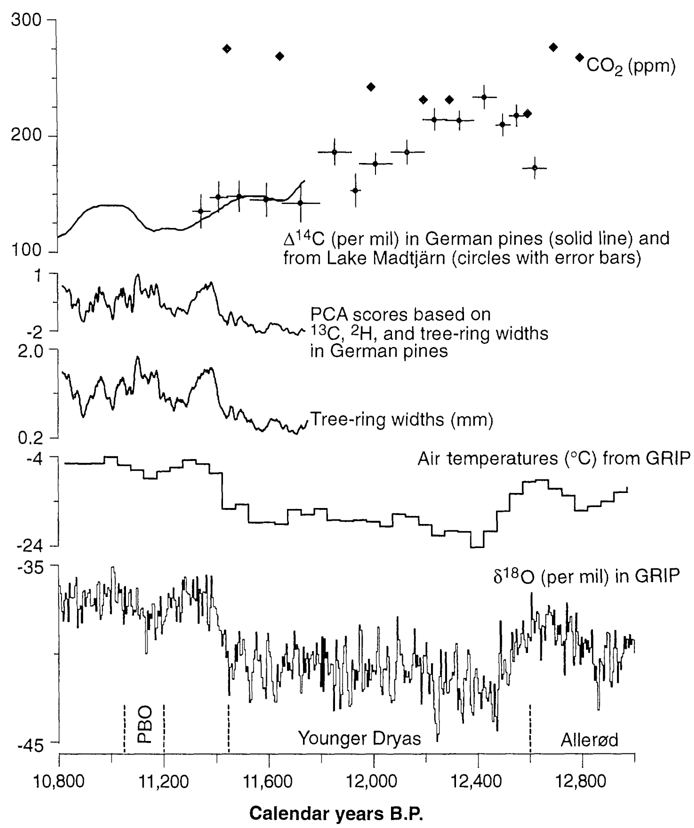


Fig. 5. A synchronized graph of different records between 13,000 and 10,800 years B.P. The δ^{18}O values (per mil relative to SMOW) are 5-year mean values. Air temperatures are estimated from 50-year periods (*12*). Ring widths are the mean of all pines with a 15-year running average. Principal components analysis (PCA) scores represent a 15-year running average of the results from the first PCA axis (covering 97.3% of total variance). The analyzed data set consists of the measured ^{13}C and ^{2}H values together with the mean ring widths of the chemically analyzed tree rings without weighting or data transformation. Centering was done by variables on a covariance matrix. The Δ^{14}C curve is based on the ^{14}C-dated dendrochronology with the additional 161 years (Fig. 5); the data were low-pass filtered by FFT smoothing. The Δ^{14}C values from Lake Madtjärn were calculated by aligning calendar ages for the AMS-dated levels 796 to 835 cm (Fig. 3 and Table 1); the age of LC was fixed at 11,450 calendar years B.P., and two constant sedimentation rates [0.28 mm year^{-1} (late AL-YD) and 0.3 mm year^{-1} (PB)] were used on the basis of the 1150–calendar year length of the YD (*3, 10–12*) and the calibrated calendar ages of the youngest ^{14}C dates. The CO_2 values [from (*33*)] were related to our chronology by assuming a constant sedimentation rate through the YD (*33*). The older values illustrate the general drop in the late Allerød.

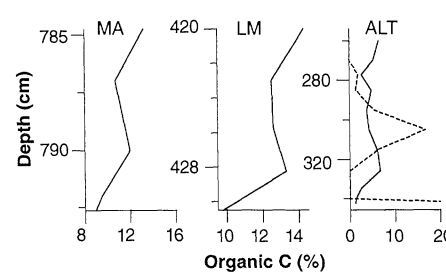


Fig. 6. Organic C content (reflecting aquatic production) in the early Preboreal lacustrine sediments (MA, Lake Madtjärn; LM, Lake Mjällsjön; ALT, Torreberga), as well as the content of detrital carbonates from Torreberga (dashed line), indicating soil erosion and allochthonous input related to depth below lake surface. Detrital carbonate content (DC) was obtained by a mass balance calculation (*52*). All values are related to percentage of dry weight. DC exhibits values of 70 to 80% in the YD before it rapidly decreases to 0% at LC (341 cm). The stratigraphic position of the PBO is marked by shading.

abundant cover of sea ice and icebergs during the YD (*18, 19*) [see also the ice-rafted debris (IRD) record in Fig. 8].

It is thus plausible that a short period of greatly reduced ventilation and bottom-water production may have led the climate into the YD. Atmospheric CO_2 concentrations decreased distinctly at the end of the Allerød warming (*33*); this trend is most likely related to our $\Delta^{14}C$ rise (Fig. 5). These possibly coupled signals can be explained in terms of decreased ventilation and sea-surface temperatures: Meltwater fluxes lead to a decrease in atmospheric CO_2 as ocean uptake of CO_2 increases with sea-surface cooling. This process thus acts as a positive feedback for more cooling. A decrease in atmospheric CO_2 concentration immediately after the post-YD CO_2 peak in peat records and possibly in ice-core records (*34, 35*) may correspond to the $\Delta^{14}C$ rise at the PBO.

Several records (Fig. 8) indicate substantial climatic oscillations after the extensive retreat of ice sheets 14,200 to 13,200 ^{14}C years B.P. (*36*). This retreat resulted in a high meltwater input to the North Atlantic and the long Oldest Dryas cold spell. The transition into the Bølling warming is shown by increasing temperatures in the GRIP core (*12*) followed by decreasing $\Delta^{14}C$ values until at least 12,500 ^{14}C years B.P. (*29, 37*), when another huge meltwater peak (MWP IA) occurred (*27*). This peak preceded the 100- to 150-year-long Older Dryas (*10, 38, 39*) at 12,200 ^{14}C years B.P. During the Older Dryas, $\Delta^{14}C$ values rose, as indicated by rapidly younger ^{14}C ages (*29, 37*), and atmospheric CO_2 concentrations seem to have reached a minimum (*35*). The 200-year-long Gerzensee/Killarney event, originally detected in European and North American lake sediments (*40*), began 400 years before the onset of the YD and appears as a distinct $\delta^{18}O$ depletion of 2.5 per mil in the GRIP core (Fig. 5). Large jumps in ^{14}C age from ~11,400 to ~10,900 years B.P. are seen in several records (*3, 29, 37*) as well as in our data (Table 1). This cooling was thus also accompanied by rising $\Delta^{14}C$ values.

The 1150-year-long YD (*3, 10–12*) is the exception to a pattern of longer warming periods interrupted by short cool events. Although the pattern at the onset of the YD appears consistent with the other coolings, one discernible difference is that the rate and extent of $\Delta^{14}C$ rise at the AL-YD boundary (30 to 60 per mil in <70 years) is considerably higher than at the onset of the PBO (15 per mil in 60 years). The same difference is evident when the temperature lowerings in the GRIP ice core (*12*) are compared with the lowerings at the other oscillations: 15°C in 150 years at the onset of the YD, compared with 4°C in 100 years (PBO), 7.5°C in 200 years (Gerzensee oscillation), and 11°C in 200 years (Older Dryas).

The $\Delta^{14}C$ record (Fig. 5) suggests that the strength of the North Atlantic conveyor decreased substantially at the AL-YD boundary and that cooling was rapid. This signature could have been caused by a sudden freshwater flux compared with more gradual freshwater build-ups at the other oscillations, and it implies that the cause of the YD may be related to the draining of huge glacial lakes (*41, 42*). At the AL-YD transition, both Lake Agassiz and the Baltic Ice Lake evidently drained, although the actual ^{14}C ages of these events vary between 11,000 and 10,500 years B.P. (*43, 44*). This range of dates likely reflects the change from declining to suddenly rising $\Delta^{14}C$ values at the AL-YD boundary (Fig. 5). We thus imply that these drainages may have created

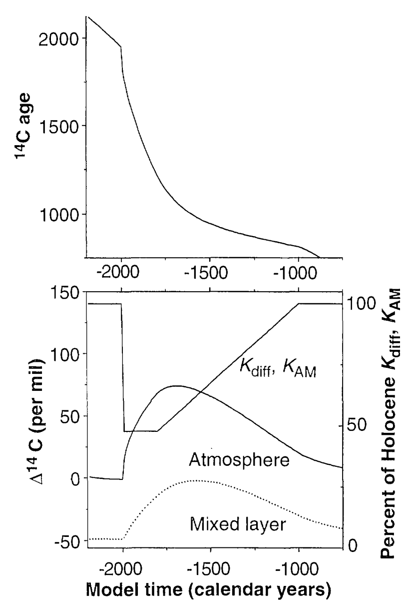


Fig. 7. A box-diffusion model (*31*) on $\Delta^{14}C$ variations (lower graph) and resulting ^{14}C ages (upper graph) caused by changes in deep-ocean mixing and in the atmosphere–mixed layer exchange rate (upper trace), in the atmosphere (middle trace), and in the mixed layer (lower trace). The maximum reduction in the model parameter is 50% (from 4200 m^2 $year^{-1}$ and 1/6.9 years, respectively). K_{diff} denotes the coefficient of vertical diffusive exchange, and K_{AM} is the coefficient of exchange between the atmosphere and the mixed layer. The long-term trend in $\Delta^{14}C$, mainly attributed to an increase in the geomagnetic field strength, was ignored, and we thus only modeled the residual over the long-term trend.

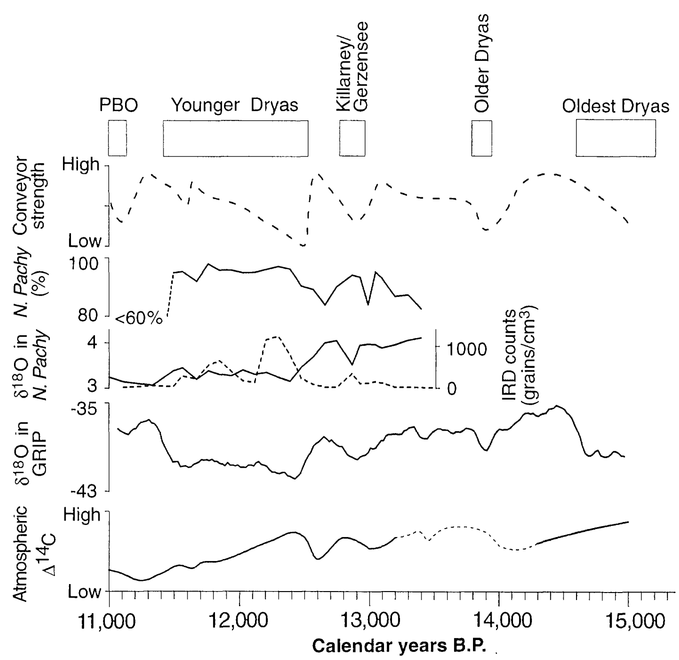


Fig. 8. Five different curves showing the oscillating development between 15,000 and 11,000 calendar years B.P. The chronology was fixed by setting the YD-PB boundary to 11,450 years B.P. in all records. The $\Delta^{14}C$ record only represents relative changes; the dashed part of the line indicates periods of uncertain changes. The $\delta^{18}O$ record (per mil relative to SMOW) from the GRIP ice core (*10–11*) is based on 100-year running mean values. The planktonic $\delta^{18}O$ (per mil) record (solid line) and percentages of the polar *Neogloboquadrina pachyderma* sinistral (s.) foraminifera and IRD counts (dashed line) derive from core ENAM 93-21 (*50*) in the Faeroe-Shetland Channel (Fig. 1A) at a water depth of 1020 m in the gateway to the Nordic Seas. The Gerzensee/Killarney oscillation is seen in all the marine records and the AL-YD transition is characterized by a distinct $\delta^{18}O$ depletion, interpreted as a freshwater spike, together with increased occurrence of *N. pachyderma* (s.) and a distinct IRD peak followed by relatively high IRD values throughout the YD. The onset of the PB is shown by abruptly decreasing *N. pachyderma* (s.) frequencies and another $\delta^{18}O$ depletion, perhaps caused by a meltwater signal. The marine chronology is based on two ^{14}C dates and the positions and ages of the Vedde and Saksunarvatn tephras (*13, 32*). The top curve depicts the strength of the North Atlantic conveyor on the basis of the combined terrestrial, ice-core, and marine records presented in this article. Across the top, the five most prominent deglacial cooling periods in the North Atlantic region between 15,000 and 11,000 years B.P. are shown.

brief but huge meltwater pulses on both sides of the North Atlantic, as indicated by a clear sea-surface salinity minimum in the Gulf of St. Lawrence and a meltwater spike on the Swedish west coast at the AL-YD transition (*45*). These meltwater inputs caused a temporary but substantial waning of the conveyor (*46*) and abruptly rising $\Delta^{14}C$ values, and may explain why only one YD occurred after the onset of deglacial warming.

An oscillating system. The inevitable effect of large influxes of fresh water and icebergs to the North Atlantic seems to be repeated periods of a gradual waning of the conveyor, resulting in brief coolings (Fig. 8). Thereafter, the conveyor accelerates again as freshwater input decreases and salinity builds up; northward heat advection and deep bottom-water production resume, resulting in declining $\Delta^{14}C$ values as ventilation increases. The onset of the YD was an extreme event that resulted in a long but oscillating recovery (*19, 21, 39, 47*) (Fig. 5) with reduced bottom-water production and low atmosphere–mixed layer exchange rates. The gradually better ventilation in the upper ocean during the YD is shown by our data and model (Figs. 5 and 7) as well as by other $\Delta^{14}C$ and paleoceanographic records (*42, 48*).

The abrupt YD-PB warming shows that the onset of northward heat advection was almost instantaneous but was not accompanied by a markedly increased ventilation rate. The deglacial North Atlantic may thus have been characterized by gradual changes in freshwater forcing combined with sudden instability during massive freshwater fluxes (*49*). An oscillating behavior of the ocean and climate thus seems to have prevailed during the time of the establishment of Holocene-style circulation. Such a conclusion also implies that catastrophic freshwater discharges are necessary to create YD-type situations during the stable modes of dominantly meridional cyclonic current patterns in the northeastern Atlantic. We also note that the distinct $\Delta^{14}C$ anomaly at 10,100 years B.P. (Fig. 2A) may signal one of the last deglacial oscillations, indicated by a simultaneous 1.5° to 2.0°C temperature drop in the GRIP core (*12*).

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$$DC(\%) = 1 - \left[\frac{(\delta^{18}O_{sed} - \delta^{18}O_{detr})}{(\delta^{18}O_{Chara} - \delta^{18}O_{detr})}\right]$$
because $\delta^{18}O_{sed}$ is a mixture of two isotopic end members in Torreberga: *Chara* encrustations and detrital carbonates ($\delta^{18}O_{detr}$). Values of $\delta^{18}O_{Chara}$ vary between −9.5 per mil (pre-LC) and −6.5 per mil (post-LC), whereas a constant value of −4 per mil for $\delta^{18}O_{detr}$ is assumed because of almost constant values of $\delta^{18}O_{sed}$ in the down-core (AL-YD) record [D. Hammarlund *et al.*, unpublished data; D. Hammarlund and D. H. Keen, *Geol. Foeren. Stockholm Foerh.* **116**, 235 (1994)].
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