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Oxygen 18/16 variability in Greenland snow and ice with 10^{-3}- to 10^{5}-year time resolution

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Abstract. The 3-km-long Greenland Ice Sheet Project 2 (GISP2) ice core presents a $100{,}000^{+}$-year detailed oxygen isotope profile covering almost a full glacial-interglacial cycle. Measurements of isotopic fluctuations in snow, frost, and atmospheric water vapor samples collected during summer field seasons (up to 20‰) are compatible with the large and abrupt $^{18}O/^{16}O$ changes observed in accumulated firn. Snow pit $\delta^{18}O$ profiles from the GISP2 summit area, however, show rapid smoothing of the $^{18}O/^{16}O$ signal near the surface. Beyond about 2-m depth the smoothed $\delta^{18}O$ signal is fairly well preserved and can be interpreted in terms of average local weather conditions and climate. The longer climate fluctuations also have regional and often global significance. In the older part of the record, corresponding to marine isotope stages (MIS) 5a to 5d, the effect of orbital climate forcing via the 19- and 23-kyr precession cycles and the 41-kyr obliquity cycle is obvious. From the end of MIS 5a, at about 75,000 years B.P., till the end of the glacial at the Younger Dryas-Preboreal transition, at 11,650 years B.P., the $O^{18}O/^{16}O$ record shows frequent, rapid switches between intermediate interstadial and low stadial values. Fourier spectra of the oscillations that are superimposed on the orbitally induced changes contain a strong periodicity at 1.5 kyr, a broad peak at 4.0 kyr, and additional shorter periods. Detailed comparison of the GISP2 $^{18}O/^{16}O$ record with the Vostok, Antarctica, δD record; Pacific Ocean foraminiferal $^{18}O/^{16}O$; Grande Pile, France, tree pollen; and insolation indicates that a counterpart to many of the rapid $^{18}O/^{16}O$ fluctuations of GISP2 can be found in the other records, and that the GISP2 isotopic changes clearly are the local expression of climate changes of worldwide extent. Correlation of events on the independent GISP2 and SPECMAP time scales for the interval 10,000-50,000 years B.P. shows excellent chronometric agreement, except possibly for the event labeled 3.1. The glacial to interglacial transition evidently started simultaneously in the Arctic and the Antarctic, but its development and its expression in Greenland isotopes was later suppressed by the influence of meltwater, especially from the Barents Sea ice sheet, on deep water formation and ocean circulation. Meltwaters from different ice sheets bordering the North Atlantic also influenced ocean circulation during the Bølling-Allerød interstadial complex and the Younger Dryas and led to a distinct development of European climate and Greenland $^{18}O/^{16}O$ values. The Holocene interval with long-term stable mean isotopic values contains several fluctuations with periods from years to millennia. Dominant is a 6.3-year oscillation with amplitude up to 3 to 4‰. Periodicities of 11 and 210 years, also found in the solar-modulated records of the cosmogenic isotopes ^{10}Be and ^{14}C, suggest solar processes as the cause of these cycles. Depression of $^{18}O/^{16}O$ values (cooling) by volcanic eruptions is observed in stacked GISP2 $\delta^{18}O$ records, but the effect is small and not likely to trigger major climate changes.

Introduction

The conditions of the atmosphere such as the amount and distribution of sunshine, cloud cover, precipitation, and wind, as well as temperatures over day and night and over the seasons, are determined by local and global boundary conditions and by external forcing. The forcing is mainly due to the modulation of solar heating by the rotation of the Earth around its axis, leading to a diurnal cycle, and the parameters of its orbit around the Sun such as the seasonal cycle, the precession cycle (19 and 23 kyr), the obliquity cycle (41 kyr) and the eccentricity cycle(s) (95 and 124 kyr). Geologic processes changing the topography of the Earth provide long-time climate forcing by changing the global boundary conditions. On a daily basis our weather also shows strong stochastic changes. Realization of the natural weather and climate variability with timescales from hours ($<10^{-3}$ year) to 10^{5} year and concern over the effects of man-made increases in atmospheric greenhouse gas concentrations and aerosols have inspired major research into the prediction of the climate for the next decades to centuries. Successful predictive models need to incorporate all the relevant processes and forcings of the natural and anthropogenically influenced climate system. This requires detailed knowledge of as many diverse climate modes as can be identified. Important sources of detailed information on past climates are the cold ice sheets and high-elevation ice caps.

A close connection between the isotopic and chemical composition of snow accumulating at the Greenland summit and atmospheric conditions of the time is crucial for paleoclimatic re-

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constructions from the ice core record [*Jouzel et al.*, this issue]. Simultaneous observations of the isotopic composition of atmospheric water vapor, frost and snow, discussed below, show a simple direct relationship. Accumulating snow initially contains the imprint of the diurnal and seasonal weather cycles, as is discussed below (also see *Shuman et al.* [this issue]). Mass and isotope transport during the firnification process subsequently smooths the deposited signal and leads ultimately to elimination of the isotopic signature of individual depositional events. Continuing firnification also attenuates the seasonal cycle, leaving the deeper ice core record dominated by the composite of the longer climatic cycles with periods from 10 to >10^5 years. Interpretation of this environmental information poses the problem of deconvolving the different climatic cycles and reconstructing the associated environmental and atmospheric parameters.

The long climate cycles attributed to orbital obliquity and precession as well as the "100-kyr eccentricity" cycle (Milankovitch cycles) have extensively been studied and documented in ocean sediment cores [*Hays et al.*, 1976; *Imbrie et al.*, 1984; *Martinson et al.*, 1987] and in ice cores [*Jouzel et al.*, 1987; *Chappellaz et al.*, 1990; *Johnsen et al.*, 1995a]. The high-quality Greenland Ice Sheet Project 2 (GISP2) and Greenland Ice Core Project (GRIP) [*Johnsen et al*, this issue] cores from the Greenland summit area allow shorter cycles to be documented and, for the first time, provide a window for the study of decadal to millennium scale climate processes over an (almost) full glacial-interglacial cycle.

Vapor-Snow Isotopic Connection

We sampled accumulation (snow and frost) and, simultaneously, atmospheric water vapor in the clean-air sector near the GISP2 camp during successive field seasons as a first step to determine the isotope transfer function between atmospheric water vapor and firn. Snowfall and frost were collected on a vertical nylon mesh screen placed between about 1 and 2 m above the snow surface. Samples were collected separately for each significant accumulation. Water vapor from the atmosphere was collected by pumping air through a cold trap in an alcohol-filled dewar cooled to about -90°C by a Cryocool CC100-II. Twelve hours of pumping usually yielded about 10 g of water, and samples were taken, where possible, twice daily ("day" and "night"). The $^{18}O/^{16}O$ ratios of all water samples were measured in the laboratory with an automated Micromass 903 isotope ratio mass spectrometer on CO_2 equilibrated with the water at 25°C. The oxygen isotope composition of samples throughout this paper is, as usual, expressed as the relative difference in the abundance ratios $^{18}O/^{16}O$ of the sample and Vienna standard mean ocean water (V-SMOW), $\delta^{18}O$, given as per mil. The results of the 1990 vapor-snow-frost samples are discussed below as an example below.

The 1990 $\delta^{18}O$ values of snow and frost samples (Figure 1) vary during the field season by about 20‰, a variation about as large as the full seasonal cycle. Moreover, these values do not change gradually over the season, but the full range may be covered over as little as a week, and large changes occur several times in one season. This applies equally to fresh snowfall and to frost samples collected from the screen during episodes of ice fog at night (i.e. at low solar angle). Snow and frost $\delta^{18}O$ values were plotted together in Figure 1 because they do not differ significantly. The water vapor samples collected during the same period show the same changes in $\delta^{18}O$ as snow with an offset of about 13‰, corresponding to the expected liquid-gas fractionation at ~ -20°C [*Majoube*, 1971]. Evidently, precipitation faithfully represents the atmospheric water vapor from which it condensed. The change in $\delta^{18}O$ with mean air temperature at GISP2 during vapor sample collection is 0.83‰/°C, well above the modern spatial gradient of 0.63‰/°C derived from snow pits [*Dansgaard et al.*, 1973]. Moreover, comparing major peaks and valleys in $\delta^{18}O$ and temperature gives gradients as high as 1.4‰/°C, well above the thermodynamic gradient for water vapor of the Rayleigh distilllation process. This applies equally to the water vapor collected when no precipitation was formed. In the simple Rayleigh model $\delta^{18}O$ values of precipitation are derived from evaporation conditions (temperature, pressure, water vapor saturation) and temperature and pressure at the precipitation site [e.g., *Jouzel et al.*, this issue, Figure 5]. This is in principle insensitive to what happens in between. The large and rapid changes in $\delta^{18}O$ values of water vapor, frost, and snow, observed during a summer field season, cannot be explained this way but also reflect air mass history with exchange of water vapor and isotopes between the air mass and its environment. This underscores the need for atmospheric general circulation models (GCMs) including the water isotope cycles as discussed by *Jouzel et al.* [this issue]. The implied influence of air mass history means that $\delta^{18}O$ is sensitive to a wider range of environmental conditions and holds out the promise of richer paleoenvironmental information from the ice core isotope record.

Isotopic Changes During Firnification

In the firn at GISP2, little can be seen of the strong isotopic variability observed in precipitation, and the yearly range of $\delta^{18}O$ values, about 20‰ at the surface, decreases rapidly with increasing depth. Transport of water vapor in the firn pore spaces has been identified as the main cause of smoothing of isotope fluctuations [*Johnsen*, 1977, *Whillans and Grootes*, 1985]. It was assumed that a redistribution of the mass and the isotopes in the firn takes place without significant water vapor loss to the atmosphere. Were such a loss to take place, it would most likely

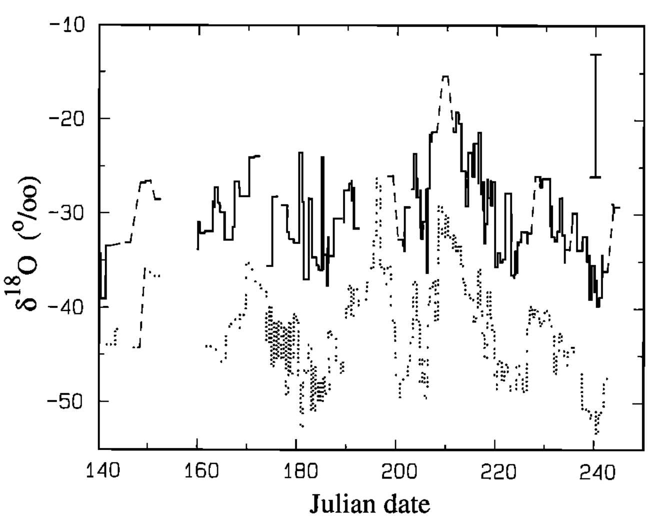

Figure 1. The $\delta^{18}O$ values of snow and frost (heavy line) and atmospheric water vapor (dotted line) collected simultaneously at the GISP2 camp during the 1990 summer field season as a function of time (Julian days). Snow and frost samples were collected whenever a meaningful sample could be obtained. The $\delta^{18}O$ values for snow and frost samples were nearly identical, so the results were combined in a single record. Vapor samples were collected regularly, once or twice daily, from two elevations above the snow surface. Dashed periods did not yield a sample. Bar gives predicted fractionation difference in $\delta^{18}O$ between water vapor and liquid relative to V-SMOW at -20°C.

lead to isotopic enrichment of the remaining firn. Studies of the process of summer depth hoar formation [*Alley et al.*, 1990; *Shuman et al.*, 1995] as well as studies of wind pumping [*Clarke and Waddington*, 1991; *Waddington and Morse*, 1994] and measured temperature variability in the snow indicate that post-depositional ^{18}O enrichment of the firn may be possible. To gain a better understanding of isotopic changes in near-surface firn during firnification, we carried out snow pit studies over the five GISP2 field seasons. An additional goal was to determine the spatial variability of the δ^{18}O signal accumulating in the GISP2 summit area.

Figure 2 depicts four GISP2 snow pit δ^{18}O profiles of the years 1989-1991. The depth scales of the pits are displaced so as to align the surface of one year's pit with the summer δ^{18}O maximum in the next year's pit. Several features indicate rapid isotopic smoothing, especially in the near-surface firn. (1) In each pit the seasonal range of δ^{18}O values near the surface is large, up to 28.5‰ in 1990, but decreases by about a factor 2 in the first 2 m (about 3 years at GISP2). (2) The 1991 δ^{18}O summer maximum decreases from -22 to -25‰ in two pits sampled about 3 km and 1 week apart in August 1991, although this difference could also mean that the heavy summer snow of pit 4 is missing in pit 5. (3) The 1990 profile, sampled at 1-cm resolution, shows smoothly changing δ^{18}O values without discontinuities corresponding to discrete snowfalls. The 3-cm profiles of the other pits appear to be more discontinuous, but this may be mainly due to stepwise sampling of steep gradients. (4) Detailed stratigraphy for 1991 pit 4 [*Duvall*, 1993] shows that δ^{18}O values change continuously across crusts of the profile. Such crusts form at the surface during periods without accumulation, mainly in summer. Since it is unlikely that the δ^{18}O of the first snow accumulating after such a period will exactly match the δ^{18}O of the snow last deposited, the continuity in δ^{18}O across crusts, even near the surface, is proof of rapid isotopic smoothing in the near-surface firn layers. The large δ^{18}O variability observed in vapor and precipitation studies on timescales of days (10^{-3} year) is not preserved in accumulating snow but is smoothed to yield a distinct seasonal cycle, often with subdued secondary peak [see *Shuman et al.*, 1995, this issue]. The deeper sections of the four pit profiles show a reduced seasonal δ^{18}O cycle with but little decrease in amplitude. There is no obvious increase in mean δ^{18}O with depth, so ^{18}O enrichment due to vapor loss from the firn must be small, except maybe in the very top layers of the firn. The results of our vapor, snow/frost, and firn studies thus indicate a direct link between atmospheric water vapor and weather and the seasonal and longer isotopic signals preserved after isotopic smoothing in the firn, and support the validity of climate reconstructions from the ice core record [*Jouzel et al.*, this issue]. Comparison of the four pit profiles, for example for the 1988-1989 "winter", however, also shows spatial variability, which limits the degree of climatic detail that can reliably be obtained from the isotopic record.

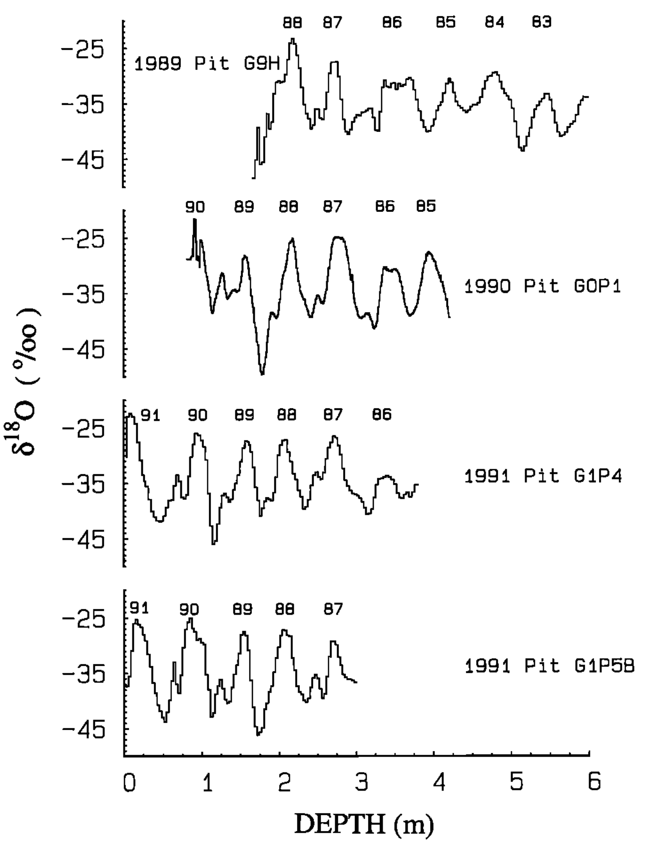

Figure 2. Snow pit δ^{18}O profiles from the GISP2 area during the summers 1989-1991 with depth scales aligned so the surface of one year's pit matches the δ^{18}O summer in the next year's pit. G9H - G0P1 offset is 0.79 m; G0P1 -G1P4 offset is 0.87 m.

The GISP2 Long Core and Its Reliability

The GISP2 1-m-resolution δ^{18}O profile (Figure 3) is a composite of the 196-m-long B core (dry drilled in 1989) and the D core, which ranges from a depth of 92 m (dry drilled in 1989) down to 3047.9 m (core length is 3056.6 m (G. Clow and N. Gundestrup, personal communication, 1996), matched at 150-m depth. Except for some small corrections, necessary because some of the isotope samples were compromised by vapor diffusion enrichment through the walls of a batch of faulty sample bottles [*Stuiver et al.*, 1995], Figure 3 is quite similar to the 2-m-resolution record used to demonstrate the excellent agreement between the GISP2 and the GRIP records down to about 2700 m [*Grootes et al.*, 1993]. The 1-m δ^{18}O values down to 300 m were calculated from high-resolution data. For this about 10 samples were cut per year (~ 5 cm/sample near the surface to ~2 cm at 300 m) based on a layer thickness calculated from present surface accumulation refined by visible stratigraphy observations. Below 300-m depth the 1-m values are based partly on sampling of the outside of the core at 1-m resolution and partly on detailed samples, collected for chemistry by melting the inner part of the core, that were split with stable isotopes, volcanics, and cosmogenic isotopes. Comparison of 1-m resolution results with those of the parallel 0.2-m sample set for the section 2793 to 2856 m, containing ice presumably from the last interglacial period, shows that the large δ^{18}O variations in this section are adequately represented by the 1-m record, while the 0.2 m results add only limited fine structure. The 1-m resolution of Figure 3 is for the interpretable part of the record down to about 2800 m, sufficient to reveal the full variability of the isotope record for century scale and longer fluctuations. For the 2931- to 3020-m section, containing "warm" ice from 2949 to 3005 m, 0.2-m samples reveal large δ^{18}O fluctuations not visible at 1-m resolution. However, increasing disturbance of the ice near the bottom makes a reliable climatic interpretation of the ice core record increasingly difficult [*Grootes et al.*, 1993; *Taylor et al.*, 1993; *Alley et al.*, 1995]. Unless it can be demonstrated that this part of the isotope record can reliably be given an environmental/climatic interpretation, there is little to be gained by higher-resolution measurements.

Three breakpoints at which the interpretability of the GISP2 record deteriorates can be identified. The first is at 2679-m depth.

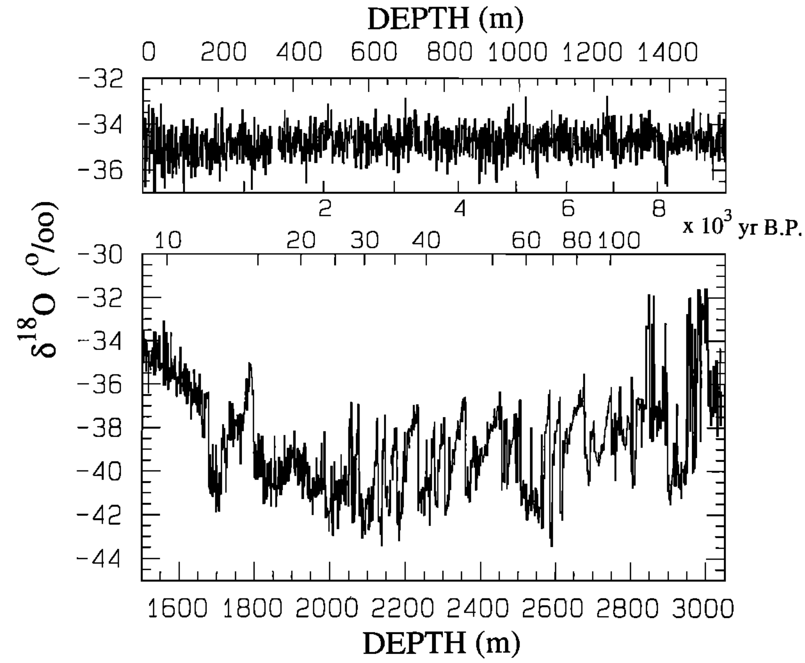

Figure 3. The $\delta^{18}O$ profile of the GISP2 ice core from the summit area of the Greenland ice sheet at 1-m resolution: depth interval 0 - 1500 m, Holocene ice; 1500 - 3040 m, mostly pre-Holocene, glacial ice. Depth is measured along the core. A timescale based on layer counting down to 50,000 years B.P. [*Meese et al.*, 1994] and on correlation with the Vostok ice core and the SPECMAP timescale via the $\delta^{18}O$ of O_2 in trapped air from 50,000 to about 110,000 years B.P. [*Bender et al.*, 1994] is also indicated. The age scale is highly nonlinear in the lower half as a result of ice flow thinning.

Here a 0.5-m-long region of ice shows inclined layering with dips up to 20° [*Alley et al.*, 1995]. The $\delta^{18}O$ value for the section 2678-2679 m depth deviates from the overlying and underlying sections by +1.45‰ and + 1.78‰, respectively. At this point we lose the close agreement, observed from the surface down, between the depth intervals covered by the same isotopic fluctuation in the GISP2 and the GRIP core. The GISP2 and GRIP $\delta^{18}O$ fluctuations, though on different depth scales, still agree closely down to the next break near 2760-m depth in both cores. Here a negative $\delta^{18}O$ spike in the section 2759-2760 m of the GISP2 core coincides isotopically with a region of inclined layers (up to 20°) at 2757-m depth in the GRIP core [*Alley et al.*, 1995]. Below this point, correlation between the GISP2 and GRIP records is lost. This indicates that flow deformation, of which the bottom sections of both cores contain more evidence [*Taylor et al.*,1993; *Grootes et al.*, 1993; *Alley et al.*, 1995; *Johnsen et al.*, 1995a], has destroyed the age-depth relationship in the bottom part of the cores. *Bender et al.* [1994] reported a third break around 2810 m depth in the GISP2 core based on a loss of correspondence between $\delta^{18}O$ of O_2 in air in the GISP2 and Vostok cores. The GISP2 $\delta^{18}O$ of ice shows a 6-m section with very low values from 2806 to 2812 m. Though the general agreement between air $\delta^{18}O$ in GISP2 and Vostok was taken as evidence that the GISP2 record could be interpreted down to 2800-m depth, it should be noted that no sign of the $\delta^{18}O$ changes in the ice between 2759 and 2806 m is seen in the $\delta^{18}O$ of air, which makes a climatic interpretation of $\delta^{18}O$ ice questionable. The deepest part of the record, below 2800 m, requires an absolute, flow-independent timescale for its interpretation.

Timescale

The GISP2 timescale (Figures 3 and 5) is based on annual layer counting down to about 50,000 years (2430 m [*Meese et al.*, 1994]). Beyond that, time was derived by *Bender et al.* [1994] by correlating the $\delta^{18}O$ of O_2 in the gas bubbles in the GISP2 core to that of the Vostok core, where the timescale is based on a combination of flow modeling [*Lorius et al.*, 1985] and a match to the Spectral Mapping Project (SPECMAP) deep-sea chronology of *Imbrie et al.* [1984] by *Sowers et al.* [1993]. The age of the ice at the 2759-m breakpoint is 103 ka with an uncertainty estimated at ± 10 ka [*Meese et al.*, this issue]. A layer recount of the GISP2 core below 2430 m based on microparticles has since given an age in agreement (± 10%) with this correlation timescale. The relative uncertainty for comparison with other long climate records is much smaller (< 3 ka [*Bender et al.*, 1994]). The timescale for the parallel GRIP core was established by layer counting and correlation from the surface back to 14,500 years B.P. (before present, where present means A.D. 1950) and farther back in time by ice flow modeling between the end of the Younger Dryas at 11,500 years B.P. and the $\delta^{18}O$ minimum at 2788 m equated with marine isotope stage (MIS) 5d at 110,000 years [*Johnsen et al.*, 1992a, *Dansgaard et al.*, 1993]. The GRIP age uncertainty is that of the flow model and the *Martinson et al.* [1987] timescale (generally up to ± 5 kyr; ± 6280 years at 110,790 years). Agreement between the GISP2 and GRIP timescales is good considering the age uncertainties connected with the counting and SPECMAP correlation of GISP2 and the flow modeling of GRIP. Agreement near 100 ka is, as expected, good as here both records were tied to the SPECMAP timescale. The first significant GRIP flow disturbance at 2757 m occurs close to the 110-kyr timescale anchor at 2788 m, so its influence on the timescale back to 110 ka should be small. For MIS 5, the interglacial and early glacial period, comparison of climate records has to be qualitative, as the available timescales are too imprecise to determine leads and lags. For the period back to about 50 ka the GISP2 direct-counting timescale offers the opportunity to verify and improve other

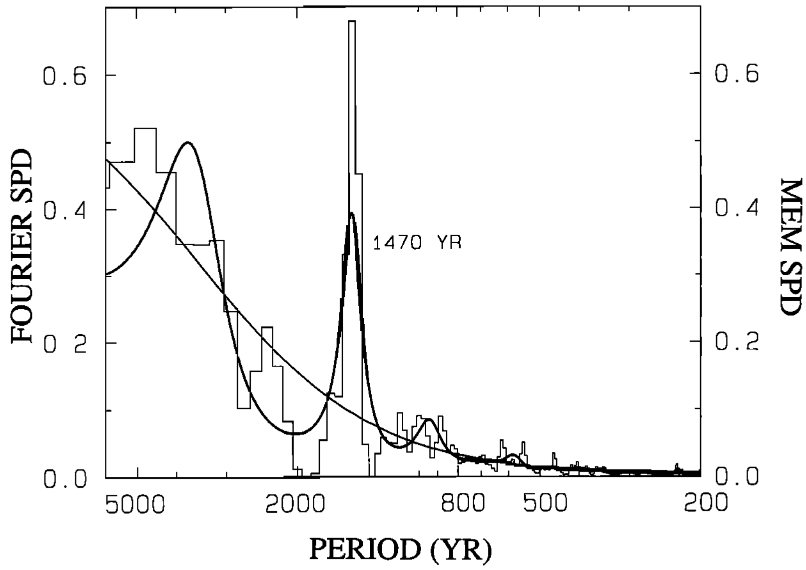

Figure 4. Relative spectral power density (SPD) normalized to the largest peak obtained for the layer-counted period 12,000 - 50,000 years B.P. of the GISP2 $\delta^{18}O$ record of the last glacial. Heavy solid line is maximum entropy method (MEM) with auto regressive (AR) order 27, light solid line is Fourier derived. Red noise is depicted by the smooth curve.

timescales that were based on interpolation, orbital tuning, and/or time information from radiometric and chemical dating techniques.

Frequency Analysis

The GISP2 timescale beyond 50 ka is based on correlation with Vostok and SPECMAP [*Bender et al.*, 1994]. This may have introduced Milankovitch periodicities, making a frequency analysis of our record for the 41- and 20 kyr Milankovitch cycles useless. Proof of Milankovitch cycles at the Greenland summit has already been obtained from the GRIP record [*Johnsen et al.*, 1995a]. Our analysis focuses on the many rapid stadial-interstadial climate fluctuations, labeled irregular expressions of sub-orbital climate variability by *Johnsen et al.* [1992a]. The layer-counted GISP2 timescale down to 51 ka offers a unique opportunity for spectral analysis independent of radiometric or orbital dating. Fourier analysis of the 12- to 50-ka $\delta^{18}O$ time series, binned at century resolution, shows a dominant peak at 1470 years with a further minor peak at about 4.0 kyr (Figure 4). The 1470-year signal is robust and persists in 0-50 ka, 0-100 ka, and 12-100 ka time series for different resolutions but is not found in the 0-11 ka Holocene $\delta^{18}O$ record. It is also clear in the ice chemistry, where it continues weakly during the Holocene [*Mayewski et al.*, this issue]. Cycles at 2.5 kyr, 1450 years and 1150 years were reported earlier for the Greenland Camp Century ice core by *Dansgaard et al.* [1971]. A new core from the subpolar North Atlantic also shows an approximately 1.5-kyr signal that is strong during the last glacial and persists weakly during interglacial times [*Bond et al.*, 1996]. A periodicity near 1500 years was further observed in sediments from Lake Tulane, Florida (1425 years [*Grimm et al.*, 1993]), the Santa Barbara Basin (1550 years [*Heusser and Sirocko*, 1997]), and core 74KL in the Arabian Sea (1450 years [*Sirocko et al.*, 1996]). Core 74KL from the upwelling area of the Arabian Sea records monsoonal climate fluctuations. Its 1450-year periodicity reflects the strength of northwesterlies carrying dust from north Arabia [*Sirocko et al.*, 1996] and persists through the Holocene. The combined ice chemistry and ocean cores identify the 1.5-kyr cycles as fluctuations in the strength of the North Atlantic atmospheric circulation advecting cold Arctic waters and drifting sea ice south [*Bond et al.*, 1996]. The Arabian Sea record shows that the 1.5-kyr changes in atmospheric circulation are not limited to the North Atlantic but are also present outside the Atlantic at low latitudes in the monsoonal zone. Though the effects of the 1.5-kyr cycle in and around the North Atlantic basin are particularly strong under glacial conditions, its cause remains to be determined and could be located elsewhere. This is of crucial importance for our understanding of both interglacial and glacial climate variability.

Interpretation of the Oxygen Isotope Record

The upper part of the GISP2 $\delta^{18}O$ record (down to 1678 m and 11,640 years B.P. [*Alley et al.*, 1993]) is characterized by long-term stability and considerable decadal-type variability. The striking features of the deeper, pre-Holocene GISP2 record are the frequent, large, and rapid fluctuations. The important question is the wider climatic significance of these fluctuations. We will address this via comparison with other paleoclimate records.

The GISP2 and GRIP $\delta^{18}O$ records agree closely down to 2750 m [*Grootes et al.*,1993], proving their regional significance. Recent calculations, based on detailed borehole temperature profiles, have recalibrated the glacial-interglacial $\delta^{18}O$ change in terms of local temperature [*Cuffey et al.*,1995; *Johnsen et al.*,1995b] and indicate an unexpectedly large glacial-interglacial temperature change of about 20°C. *Johnsen et al.* [1992a] noted that the GRIP ^{18}O fluctuations can also be seen in all previously measured long Greenland ice core records. A link with the North Atlantic was proposed by *Bond et al.* [1993], who correlated the GRIP stadial-interstadial $\delta^{18}O$ cycles to sea surface temperature (SST) fluctuations in the North Atlantic as indicated by the abundance of the foraminifera *Neogloboquadrina pachyderma* (s) in the sediment of cores VM23-81 and DSDP609. This would make the Greenland summit $\delta^{18}O$ fluctuations representative for climate changes over Greenland and at least part of the North Atlantic. In Figure 5 we explore their global significance by comparison with the Antarctic ice core record from Vostok, the

Pacific Ocean δ^{18}O record V19-29, the European pollen record from Grande Pile, and the insolation signal. The approach is similar to that of *Dansgaard et al.* [1993], who focussed on longer time periods and the European interstadial record as identified from pollen analyses. In the discussion we indicate interstadials (IS) with numbers as proposed for GRIP by *Dansgaard et al.* [1993] or with the numbers of the equivalent marine isotope stages [*Prell et al.*, 1986]. We first discuss the distinct pre-Holocene record with the older part equivalent to MIS 5; the middle, pleniglacial part, equivalent to MIS 2-4; and the glacial-interglacial transition; and then the Holocene.

Last Interglacial and Early Glacial Period (MIS 5)

Greenland. The ice with δ^{18}O values around -33‰ (i.e., heavier than Holocene ice) from 2839- to 2849-m depth at GISP2 and between about 2800 and 2860 m at GRIP almost certainly originated during the last interglacial (MIS 5e). A simple climatic interpretation of this part of both records is unfortunately not possible in view of (1) the large differences between GISP2 and GRIP, (2) the evidence of disturbed layering above this section in both cores (see above and *Alley et al.* [1995]), and (3) the evidence from the analyses of gases trapped in the ice [*Bender et al.*, 1994; *Peel*, 1995; *Chappellaz et al.*, this issue]. The interpretable early phase of the last glacial extends from the end of IS 19 at 2570-m depth in the GISP2 core down to the breakpoint at 2759 m at the base of IS 23. Agreement between the GISP2 and GRIP δ^{18}O records is quite good for this interval [*Grootes et al.*, 1993].

Antarctica. Correspondence between the major GISP2 interstadials IS 19-23 and isotope peaks in the Antarctic Vostok record (Figures 5d and 5e [*Jouzel et al.*, 1987, 1993; *Lorius et al.*, 1985]) has been proven by *Bender et al.* [1994] by using the δ^{18}O of oxygen in air extracted from both cores as a common marker. The uncertainties in the GISP2-Vostok correlation and in the age difference between ice and the gas it encloses in the Vostok core were, however, too large (up to 3 kyr) to decide on synchroneity or lead/lag relationships of less than a few thousand years for these climate fluctuations in Greenland and the Antarctic.

Deep sea. Deep-sea sediment cores have long provided the standard for past climatic developments because of their long, continuous sequences of relatively undisturbed sedimentation. These have been instrumental in establishing the climatic significance of the Milankovitch precession and obliquity cycles [*Hays et al.*, 1976; *Imbrie et al.*, 1984]. Deep-sea sediment δ^{18}O (of carbonate shells) and global ice volume are connected, because the isotopically light ice stored in ice sheets and shelves causes a direct oceanic ^{18}O enrichment. A detailed comparison between ice- and ocean core δ^{18}O records is difficult because (1) the δ^{18}O of foraminiferal carbonates is also influenced by ocean temperature, (2) oceanic conditions are spatially variable, (3) forams show a species-dependent offset from isotopic equilibrium (vital effects), (4) low sedimentation rate and bioturbation limit temporal resolution in the sediment, and (5) the mean isotopic composition of the stored ice, and thus the ocean δ^{18}O enrichment for a certain ice volume, will vary over glacial-interglacial and stadial-interstadial cycles. Moreover, the ice directly records changes in atmospheric conditions and has a temporal resolution limited only by gradual diffusional smoothing, while the ocean δ^{18}O represents changes in the global ice volume and its isotopic composition in response to long-term climate fluctuations. To remove local effects and bring out globally important features, individual deep-sea δ^{18}O records were "stacked" [*Imbrie et al.*, 1984; *Pisias et al.*,1984; *Prell et al.*, 1986]. Yet these local effects may have real climatic significance and correlate, for example, in the North Atlantic with Heinrich events and the Greenland ice core records [*Shackleton et al.*, 1996] (also see below). We chose for our comparison the V19-29 record from the eastern equatorial Pacific, used by *Pisias et al.* [1984] to identify marine isotope events (Figure 5b).

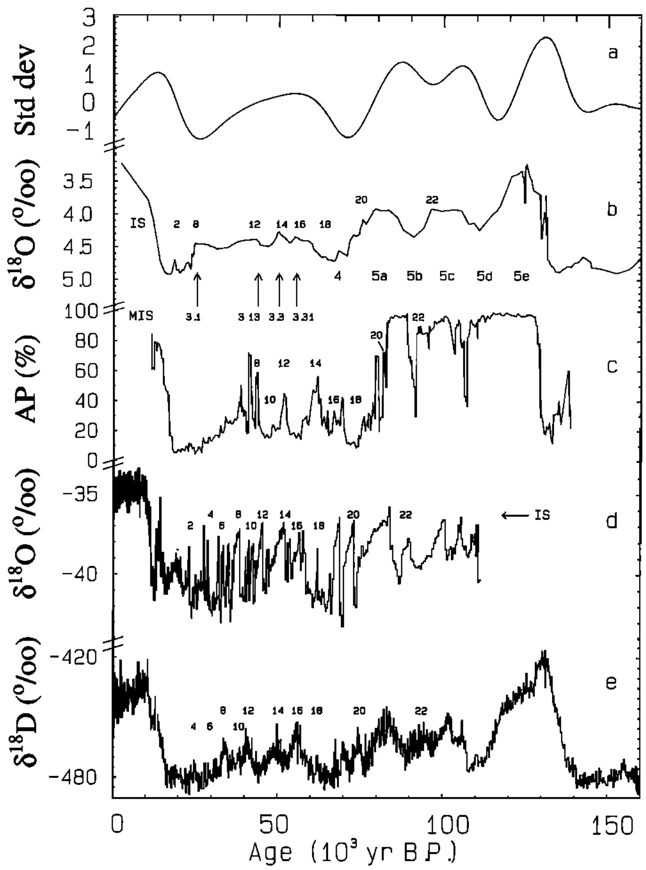

Figure 5. Climate data spanning the last 160,000 years B.P.: (a) Insolation variations due to the Earth's orbital parameters eccentricity, obliquity, and precession, normalized over 800,000 years and combined. The scale is in standard deviation units [*Imbrie et al.*, 1984]. (b) Benthic δ^{18}O record from Pacific Ocean core V19-29 [*Pisias et al.*, 1984] with ages linearly interpolated between isotopic events dated by *Martinson et al.* [1987]. Arrows mark marine isotope stages (MIS) 3.1, 3.13, 3.3, and 3.31 as tabulated by Martinson et al. (c) Grande Pile arboreal pollen (AP) [*Woillard and Mook*, 1982]. Ages were estimated by linear interpolation between the end of the Younger Dryas at 11,650 years B.P. [*Alley et al.*, 1993] and the marine isotope stage 5/6 transition at 129,840 years B.P. [*Martinson et al.*, 1987]. (d) GISP2 δ^{18}O isotope record [*Grootes et al.*, 1993; *Stuiver et al.*, 1995] with timescale developed by *Meese et al.* [1994] and *Bender et al.* [1994] and interstadials IS 2-22 as proposed for GRIP by *Dansgaard et al.* [1993]. (e) Vostok δD isotope record [*Jouzel et al.*, 1993].

The correspondence between GISP2 and V19-29 (Figures 5b and 5d) is not immediately obvious due to considerably more fine-structure of GISP2. The correspondence was established via the Vostok record (Figure 5e), of which the oceanic correlation, already discussed by *Lorius et al.* [1985], had been shown by *Sowers et al.* [1993] via the δ^{18}O of O_2 in air trapped in the ice which follows the δ^{18}O of seawater due to photosynthesis. The

GISP2-Vostok correlation by *Bender et al.* [1994] thus provided a firm link between the early phase of the last glacial sequence, IS 19-23, at GISP2 and MIS 5a to 5c (Figures 5b and 5d). Interesting is that the two cold periods separating the strong interstadials IS 19, 20, 21 were so short and/or so local that they are hardly significant in V19-29, where the $\delta^{18}O$ fluctuations over the MIS 5a to MIS 4 transition are unremarkable. Evidence of IS 19 and 20 is even totally lacking in the stacked benthic $\delta^{18}O$ records [*Imbrie et al.*, 1984; *Pisias et al.*, 1984; *Prell et al.*, 1986]. While the $\delta^{18}O$ of O_2 correlation with oceanic $\delta^{18}O$ confirms the presence of a 20-kyr precessional signal at GISP2 during MIS 5 (compare Figure 5a), climate at the Greenland summit was obviously also influenced by factors acting at shorter timescales, especially toward the end of MIS 5, that are not evident in the ocean $\delta^{18}O$ records of global ice volume.

Benthic $\delta^{18}O$ profiles for MIS 5 cores from the North Atlantic and the Nordic seas are similar to V19-29 [e.g., *Keigwin et al.*, 1994; *Jung*, 1996]. Evidence of climate variability similar to IS 19-24 was, however, found in proxies of (near)surface ocean conditions during MIS 5 [*Cortijo et al.*, 1995; *McManus et al.*, 1994; *Keigwin et al.*, 1994] in cores ranging from about 60°N to the Bahama Outer Ridge at 28°N. Bottom conditions at the Bahama Outer Ridge reflect the balance between North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW) [*Keigwin et al.*, 1994]. Significant fluctuations in carbonate content, a proxy for productivity and SST, and in the $\delta^{13}C$ of benthic forams during MIS 5 that is unremarkable in benthic $\delta^{18}O$ prove a link between changes in local SST and NADW formation to the north, possibly connected with IS 19-24 in Greenland.

Land. *Keigwin et al.* [1994] extended their comparison of the Bahama record to the tree-pollen-based landclimates of Grande Pile [*Woillard*, 1978] (Figure 5c) and Tenaghi Philippon [*Tzedakis*, 1993] in Europe and Clear Lake in California [*Adam et al.*, 1981]. At Grande Pile, total tree pollen percentages for the warm MIS 5e and the temperate MIS 5a and 5c are over 90% and lack climate sensitivity. The assemblage and succession of tree species, however, indicates differences between MIS 5a, 5c, and 5e as well as further climate oscillations, for example, at the end of the last interglacial (MIS 5e) and during St. Germain I (MIS 5c) [*de Beaulieu and Reille*, 1992] more in agreement with the Greenland and Bahama records. *Keigwin et al.* [1994] correlate their carbonate peaks (SST) during the MIS 5a to 4 transition with two sharp fluctuations in tree pollen percentages, Ognon I and II, that follow the warm St.Germain II. They conclude that although Ognon I and II (Figure 5c [*Woillard and Mook*, 1982]) occur in disturbed sediment and their significance has been questioned by *de Beaulieu and Reille*, [1992], the similarity of the pattern of the MIS 5a to 4 transition in the Greenland ice, the Bahama Ridge sediment, and the Grande Pile peat bog suggests that IS 19 and 20 represent real major climate fluctuations in the North Atlantic basin at the end of MIS 5a.

The qualitative agreement during MIS 5 between the Greenland Summit ice cores, Grande Pile, and sea surface proxies in ocean cores reflects their common sensitivity to North Atlantic sea surface conditions, especially temperature. The atmospheric response to SST changes, recorded in Greenland ice and west European pollen is immediate, while the response of NADW formation is of the order of years to decades. The sharp pattern of IS19/20 in the transition MIS 5-4, recognizable in the ocean and on land, confirms the predicted close temporal agreement. As changes in NADW formation are felt in the oceans worldwide, and the existence of atmospheric teleconnections has been established, it is likely, though not yet proven, that not only the long-lasting GISP2 climate episodes like IS 21, but also the shorter ones like IS 19 and 20, have global significance. The character of the equivalent episodes elsewhere (e.g. warm/cold, wet/dry, strong/weak) will vary with latitude and longitude and also depend on local conditions.

Pleniglacial (MIS 2-4)

Greenland. The ice between 2570- and 1798-m depth represents the second, pleniglacial period between the MIS 5a-4 transition at about 74 ka and the start of IS 1, the Bølling/Allerød interstadial complex (14,820 years B.P.) (Figure 5d). It includes two cold phases (equivalent to MIS 4 and MIS 2, the "last glacial maximum") and the warmer MIS 3 complex. This complex (about 58 to 27 ka) is characterized by frequent $\delta^{18}O$ fluctuations between interstadial values of -37 to -39‰ (IS 3-17), about 3‰ less than the Holocene interglacial mean value, and cold stadial $\delta^{18}O$ values of -41 to -43‰. The transition from stadial to interstadial conditions occurs very rapidly and leads initially to the least negative $\delta^{18}O$ values. After a slow, gradual drop in $\delta^{18}O$ values (down to -39‰ for some of the longer interstadials), the return to stadial values again is very rapid. This suggests a climate switching between two "stable" modes. Excellent agreement between the GISP2 and GRIP records was observed for this interval [*Grootes et al.*, 1993]. The wider climatic significance of the Greenland summit ice core records will again be discussed via the comparison with other records (Figure 5). The Bølling/Allerød interstadial, IS 1, and the following Younger Dryas stadial, though treated in the "glacial-interglacial transition", have similar amplitude, rapidity of change, and duration as the climate oscillations of MIS 3.

Antarctica. Correspondence between the major interstadials of GISP2 (IS 8, 12, 14, 16-17) and the Vostok record (Figure 5e) was established by *Bender et al.* [1994]. The shorter GISP2 interstadials, however, lacked statistically significant counterparts. A new $\delta^{18}O$ record from Taylor Dome, Antarctica (P.M.Grootes et al., manuscript in preparation, 1997) resembles that of Vostok, but has larger and, especially, sharper stadial-interstadial fluctuations. It suggests that climate fluctuations in parts of Antarctica were more similar to those in the northern hemisphere than is indicated by comparison of the GISP2 and Vostok records and also that some of the shorter interstadials are found in both Greenland and Antarctica.

Deep sea. As was discussed above, rapid climate fluctuations are represented poorly, or not at all, in the V19-29 benthic $\delta^{18}O$ record of global ice volume (Figure 5b) and in the benthic "standard" stacks [*Imbrie et al.*, 1984; *Pisias et al.*, 1984; *Prell et al.*, 1986]. The large climate variability shown by the Greenland ice cores, supported by increasing evidence from the North Atlantic (below), puts into question the use of events as defined in the deep-sea isotopic taxonomy [*Prell et al.*, 1986] for global climate correlation [*Shackleton et al.*, 1996]. For MIS 3 this is especially clear, as the "5 minor peaks" identified by *Pisias et al.* [1984] in V19-29 cover the same time span as at least 15 interstadials (IS 3-17) of the GISP2 record. The standard isotope taxonomy, based on low accumulation cores, identifies only two events for stage 3 [*Prell et al.*, 1986], while the standard stack used by *Martinson et al.* [1987] retains four events for the five MIS 3 peaks of *Pisias et al.* [1984] (3.01 is no longer recognized, and 3.1 has moved up to near the top of MIS 3(Figure 5b)). With ice sheet volume slowly responding to climate change, a series of short interstadials separated by likewise short cold episodes could generate a fairly constant ice volume. Only major interstadials like IS 8, 12, 14, and 16/17 and prolonged cold periods would then produce significant changes in global ocean $\delta^{18}O$. Correlations in Figure 5b are based on these inferences. The ages of ice core and deep-sea events agree surprisingly well (Table 1). This may indicate that the estimated age uncertainties

Table 1. Timing of Glacial Climate Fluctuations in the Greenland GISP2 Ice Core, Benthic Ocean Record, and Huon Peninsula Coral Reefs

	GISP2 Ice Core	Ocean Sediment		Huon Peninsula Corals	
IS	Age[a], ka	MIS	Age[b], ka	Terrace	Age[c], ka
3 / 4	27.7 / 28.8 ± 1.4	3.1[d]	25.4 ± 5.9[d]	-	-
6	33.5 ± 1.7			IIc (B-U21/24)	33.4 ± 0.6/ 33.0 ± 0.5
7	35.0 ± 1.7			IIa (K-U14)	34.8 ± 0.3
8	37.3 ± 1.9	3.1	25.4 ± 5.9	IIa (B-U10)	37.8 ± 0.3
11	42.2 ± 2.1			IIa (K-U9)	41.8 ± 0.6/ 42.2 ± 0.3
12	44.3 ± 2.2	3.13	43.9 ± 4.7	IIIc,b (K-U10/34)	43.9 ± 0.7/ 44.5 ±0 7
14	50.4 ± 2.5	3.3	50.2 ± 3.9	IIIa$_{m,u}$ (K-4/3)	51.2 ± 0.8/ 51.8± 0.8
16/17	56.4 ± 5.6	3.31	55.5 ± 5.0	IIIa$_l$ (K-9)	54.6 ± 0.7
19	67.8 ± 6.8			IV (K)	64.9 ± 1.7
20	71.5 ± 7.2			IV (B /K)	72.8 ± 2.2/ 71.9 ± 3.6
21	79.0 ± 7.9	5.1	79.3 ± 3.6	-	-

[a] - *Meese et al.* [1994] and *Bender et al.* [1994]; age uncertainty is ± 5 % for the counted scale up to 50 ka and (absolute) 10% beyond 50 ka.
[b] - *Martinson et al.* [1987] with tabulated age uncertainty.
[c] - *Chappell et al.* [1996] with age uncertainty as tabulated, B indicates transect Bobo, and K is transect Kanz.
[d] - Proposed correlation MIS 3.1 based on AMS ^{14}C-dating.

are too large. For event 3.1, dated at 25,420 years B.P. by *Martinson et al.* [1987], the correlation with IS 8 dated at 36.3 to 38.4 ka (± 1.9 ka) on the GISP2 timescale is obviously wrong. Yet a weak δ^{18}O minimum near the end of MIS 3, labeled 3.1, has consistently been dated by accelerator mass spectrometer (AMS) ^{14}C between 25.5 and 26.9 ka (around 29 ka in calibrated years [*Sarnthein et al.*,1995]), in agreement with the Martinson age, and with ages of 27.7 ka and 28.8 ka (± 1.5 ka) for IS 3 and 4 in GISP2. A possible solution to the apparent MIS 3.1 controversy is that the fifth peak of V19-29, which disappeared from the stacked record, is real. Event 3.01 of *Pisias et al.* [1984] then corresponds to IS 3 and 4 and the isotope event currently labeled 3.1 Their event 3.1 of V19-29 would be new in the stacked record where it corresponds to IS 8 (age from 36.8 to 38.4 ka on the GISP2 layer-counting scale), thus filling the presently large time gap between MIS 3.13 and 3.1.

Sea level derived from uplifted coral reefs generally fluctuates in step with δ^{18}O in deep-ocean cores [*Chappell and Shackleton*, 1986], i.e., with global ice volume. Thus the extensively dated coral terraces of the Huon Peninsula, New Guinea [*Chappell and Shackleton*, 1986; *Chappell et al.*, 1996] offer an independent timescale for sea level highstands to verify the SPECMAP timescale and the proposed correlation between Greenland interstadials and global ice volume (Table 1). The ages of the episodes of high sea level in the coral terraces agree in all cases very well with GISP2 interstadials. A complete and detailed stratigraphy of the coral reefs is needed, however, to decide whether this is due to the large number of interstadials available for correlation in GISP2 or whether all the individually correlated terraces indeed represent different climate and sea level episodes. In the latter case we have the first proof that the GISP2 interstadials are the expression of a globally warmer climate with less global ice volume.

High-sedimentation-rate deep-sea cores provide the best records to verify the climatic significance of the Greenland ice core records. In the North Atlantic, progressively colder SST cycles terminate with a so-called Heinrich layer of ice-rafted detritus (IRD) and look similar to the GRIP and GISP2 interstadials [*Bond et al.*, 1993]. *Bond and Lotti* [1995] later identified smaller IRD peaks corresponding to most of the other MIS 3 stadials. Millennial-scale carbonate fluctuations on the Bahama Outer Ridge and Bermuda Rise [*Keigwin and Jones*, 1994] occur during MIS 3 with a frequency similar to the interstadials in the Renland δ^{18}O record [*Johnsen et al.*,1992b]. Accompanying δ^{13}C changes indicate changes in either NADW formation or productivity. *Cortijo et al.* [1995] observed grey scale fluctuations in a suite of North Atlantic cores along a north-south transect and proposed a detailed correlation with the Greenland interstadials IS 3-24. A general correlation between the Greenland ice core records and surface ocean conditions in various parts of the North Atlantic is thus well established.

A new, high-accumulation-rate core from the Faeroe-Shetland Channel [*Rasmussen et al.*,1996] reveals this correlation in high detail. A multiproperty paleoceanographic reconstruction for the past 58 kyr shows a pattern of changes in NADW formation and SST consistent with the atmospheric climate fluctuations IS 3-15 at the Greenland summit. Detailed measurements produced a magnetic record very similar to the δ^{18}O record at the Greenland summit which allows a detailed correlation and firmly establishes the ocean-atmosphere-ice connection. Warm interstadial conditions in Greenland, warm SST, and vigorous NADW formation in the Nordic Seas were followed by a cooling phase with decreasing δ^{18}O in Greenland, weakening and irregular NADW formation and SST. The cycle ends with no NADW being formed in the Nordic Seas, low SST and Greenland δ^{18}O, and evidence for icebergs moving south from IRD and negative meltwater δ^{18}O anomalies in planktonic forams. This covariance of Greenland δ^{18}O, SST, and NADW formation provides a first glimpse of the workings of the coupled climate system under glacial conditions. Independent evidence for the detailed climate correlation between ocean sediment and ice cores is provided by detailed AMS ^{14}C dating of meltwater spikes in a new high-resolution core off Iceland. The varying offsets between the AMS ^{14}C ages of the $\delta^{18}O$ meltwater spikes and the GISP2 ages of the correlated $\delta^{18}O$ stadials are in good agreement with a proposed magnetic calibration of the radiocarbon timescale of *Laj et al.* [1996 and personal communication, 1997] [*Sarnthein et al.*, 1996]. The identification of groups of interstadials [*Bond et al.*, 1993] allows the correlation of closely spaced interstadials and stadials despite the considerable uncertainties in their dating (GISP2 timescale: ± 5% or ± 2 ka at 40 ka [*Meese et al.*, this issue]; ^{14}C dates ± 0.5 - 1 ka). The correlation is further supported by the observation of a 1.5-kyr periodicity, similar to that of the glacial GISP2 $\delta^{18}O$ record, in a North Atlantic core [*Bond et al.*, 1996].

The link between Greenland ice core climate fluctuations and variations in NADW formation [*Rasmussen et al.*,1996] provides a mechanism for oceanic teleconnections for the large Greenland/North Atlantic climate fluctuations. Changes in NADW are also recorded in the $\delta^{13}C$ of benthic foraminifera in a core from the South Atlantic Ocean. *Charles et al.* [1996] argue from an offset between changes in benthic $\delta^{13}C$ and planktonic $\delta^{18}O$ values (SST) in this core that climate change in the southern hemisphere led North Atlantic changes by 1500 years. This lead corresponds to a full 1.5-kyr cycle, suggesting the possibility of a miscorrelation between the still partly ununderstood records [*Charles et al.*, 1996]. A reported presence of the 1.5-kyr periodicity in the Vostok record [*Yiou et al.*, 1991], strengthens the Arctic-Antarctic connection. Together, the ice core and ocean evidence indicate that the large glacial climatic variability was not restricted to Greenland but extended throughout the Atlantic. Observations of a strong 1.5-kyr periodicity outside Greenland and the Atlantic in the Arabian Sea and the Santa Barbara Basin and the link between Greenland climate and NADW formation support the likelihood of climatic teleconnections between Greenland and the Atlantic and the rest of the world. Local differences in ice sheet behavior and/or landmasses may have resulted temporarily in locally different climate changes.

Land. The pleniglacial at Grande Pile (Figure 5c) consists of long cold sections with short warming spikes. This timing was mostly derived by linear translation of depth to time. In the GISP2 core we see generally short cold episodes. This suggests that sedimentation rates at Grande Pile under cold climate with reduced vegetation cover have been higher, distorting the timescale, and/or that the response of tree pollen to climate change may have been asymmetric with a rapid response to cooling and a slow response, determined by ecosystem recovery, to warming. Tree pollen occurs during the pleniglacial generally in low percentages, but reaches 50% during several intervals. One can equate the tree pollen peaks to the major ice core interstadials of the series IS 3-17 and find signs of most of the others in the fine structure between peaks, similar to the ocean $\delta^{18}O$ record. Similar records, showing generally cool and variable glacial climate conditions, have been obtained from sediments in the volcanic lakes of Les Echets [*de Beaulieu and Reille*, 1989], Le Bouchet [*Reille and de Beaulieu*, 1990], Vico [*Leroy et al.*, 1996] and Monticchio [*Watts et al.*, 1996]. Though this shows that the climate fluctuations of Greenland and the North Atlantic may, as expected, have influenced climate in a large part of southwestern and southern Europe, it does not yet prove a direct correlation. This requires a detailed record of fast responding, climatically sensitive pollen or other climate proxies with an independent, detailed timescale. For northwest Europe, *Behre and van der Plicht* [1992] identified only seven interstadials. These were correlated with the major interstadials of the GRIP IS 1-24 between 10 and 110 ka by *Dansgaard et al.* [1993]. Since the generally westerly atmospheric circulation over Europe makes it likely that the equivalents of the GISP2 interstadials existed in Europe, and since the more continuous sedimentary records from volcanic lakes seem to bear this out, we attribute their apparent partial absence in northwestern Europe to an incomplete sedimentation record and insufficiently precise dating. This is further supported by signs of frequent climate fluctuations in sediment sequences reported by *Zagwijn and Paepe* [1968], *Paepe et al.* [1990)], and *Mania* [1967]. At Grande Pile and other European sites of continuous sedimentation we may expect a record of climate variability equivalent to the Greenland ice cores and North Atlantic ocean cores. Much further work on land is still needed to refine the terrestrial record of climate change.

Comparing Early Glacial (MIS 5a-d) and Pleniglacial (MIS 2-4) Climate

During the early phase of the last glacial (MIS 5a-d), insolation was mostly above average (Figure 5a) and land ice volume was less than half of its full glacial value [*Chappell et al.*, 1996; *Chappell and Shackleton*, 1986]. Climate was "intermediate" with limited variability shown by both Greenland and Antarctic ice-core isotopes and by European pollen (Figures 5c, 5d, and 5e). For the pleniglacial period (IS 2-18), insolation and $\delta^{18}O$ were at a minimum during MIS 4 and MIS 2, while the MIS 3 interval had an insolation slightly below average and highly variable $\delta^{18}O$ (Fig. 5a, 5b, and 5d). The absence of significant interstadial interruptions (except maybe for ~100 years at 23.5 ka) during MIS 2 (about 14.8 - 28 ka) suggests that for this period the large global ice volume and low sea level combined with low levels of insolation and greenhouse gases to give a relatively stable cold climate. MIS 3 is characterized by large climate variability with abrupt transitions between periods with mild and cold conditions containing a strong 1500-year periodicity. For the period between about 70 ka and the glacial-interglacial transition, the Atlantic Ocean was in a full glacial mode with deep water temperatures in the 0°C range [*Labeyrie et al.*, 1987], concentrations of the greenhouse gases CO_2 and CH_4 were low [*Chappellaz et al.*, 1990,1993], and ice cover of the northern continents was large. The climate system had apparently two favored modes and strong positive feedbacks giving fast climate transitions. *Rasmussen et al.* [1996] showed that the Greenland interstadial/stadial cycles were accompanied by formation of large amounts of NADW and no NADW, respectively. We speculate that the varying NADW formation modulates heat transport north in the North Atlantic, which during MIS 3 reduced/increased snow cover and thereby albedo of the European continent and provided a strong positive feedback leading to rapid switches between stadial and interstadial climates. The transitional cooling [*Rasmussen et al.*, 1996], which mirrors the gradual drop in Greenland $\delta^{18}O$ after the interstadial maximum, documents irregular and weakening NADW formation. Ice sheet surges, producing the ice rafted detritus associated with most of the cold stadial episodes [*Bond et al.*, 1993; *Bond and Lotti*, 1995], did therefore not initiate cooling as proposed by *Cortijo* [1995] but may have cut short the transitional cooling with a drop to full glacial conditions. A precise determination of the relative timing of IRD and temperature change is needed to determine whether ice sheet surges, giving rise to IRD, are needed to develop full stadial climate, or whether a longer survival of icebergs under cold conditions may explain some of the smaller IRD deposits.

Glacial-Interglacial Transition

The exact start and the duration of the transition from the last glacial maximum to the current Holocene interglacial seems to

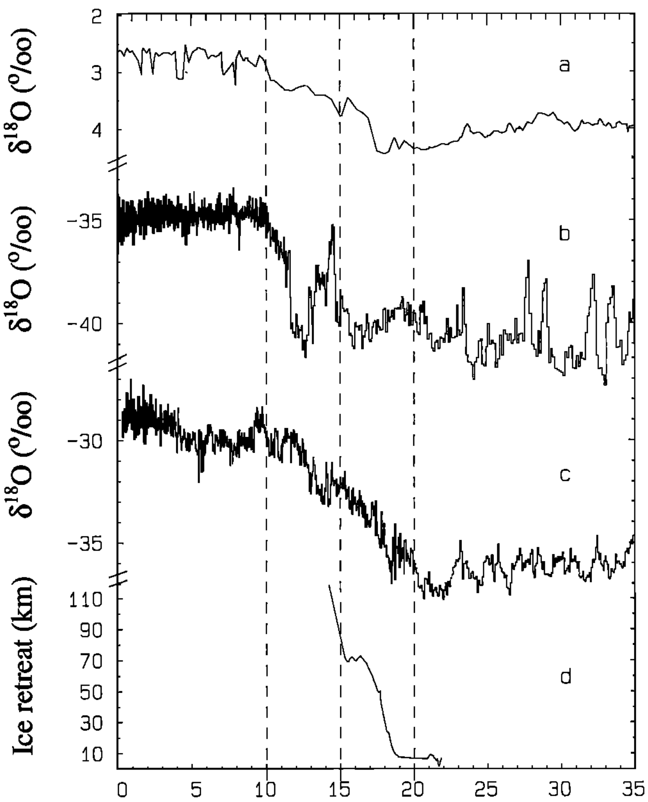

Figure 6. Climate data spanning the last glacial to interglacial transition 0 - 35,000 yr. B.P. (a) Benthic foram *C. wuellerstorfi* $\delta^{18}O$ from Ocean Drilling Program site 658, core C, off northwest Africa (M. Sarnthein, personal communication, 1996). (b) GISP2 $\delta^{18}O$ isotope record [*Grootes et al.*, 1993; *Stuiver et al.*, 1995], (c) Byrd Station, Antarctica, $\delta^{18}O$ isotope record [*Beer et al.*, 1992; S.J.Johnsen, personal communication, 1995]. GISP2 and Byrd Station records were correlated by *Sowers and Bender* [1995] on the basis of the $\delta^{18}O$ of O_2 in trapped air. (d) Retreat of the Scandinavian ice sheet from the continental shelf near Andøya, northern Norway [*Vorren et al.*, 1988].

differ between Greenland and Antarctica, yet this timing is of considerable importance for an evaluation of the connection between insolation and climate. *Sowers and Bender* [1995] tied the $\delta^{18}O$-ice record from Byrd Station, Antarctica, to GISP2 using $\delta^{18}O$ of O_2 in trapped air. At Byrd (Figure 6d) an increase in $\delta^{18}O$ of the ice, presumably indicating rising temperatures, began at about 21 ka and continued with minor interruptions until Holocene values are encountered around 12 ka. In Greenland, GISP2 $\delta^{18}O$ (Figure 6c) increases in a similar way from a minimum around 21 ka until 18.7 ka. However, while Byrd $\delta^{18}O$ continues to rise after a brief fall back at 18.5 to 18 ka, the GISP2 $\delta^{18}O$ decreases for the 18-16 ka interval. This signal applies to all of Greenland [*Johnsen et al.*, 1992a] yet is not global, as is evidenced by (1) a rise in CH_4 and CO_2 concentrations beginning at 17 ka [*Sowers and Bender*, 1995], (2) a decrease in benthic $\delta^{18}O$ indicating global ice sheet reduction around the same time (Figure 6a), and (3) the onset of sea level rise at Barbados [*Fairbanks*, 1989]. Retreat of the Scandinavian ice sheet from Andøya, an island at 69°15'N off the coast of Norway, started about 18 ka (Figure 6e [*Vorren et al.*, 1988]), while $\delta^{18}O$ of planktonic *N. pachyderma* (s) in eastern central Arctic Ocean sediment drops rapidly after 18.6 ka [*Stein et al.*,1994]. *Sarnthein et al.* [1995] find in this period meltwater $\delta^{18}O$ minima in front of the Barents Shelf in the northern Norwegian Sea and east of Greenland, extending from the Denmark Strait toward Jan Mayen. They conclude that major meltwater pulses from icebergs from the Barents shelf and the Laurentide Ice Sheet changed ocean circulation and facilitated cold water transport along Norway to latitudes as far south as Faeroes. A cooler Europe (Oldest Dryas?), and increased "paleo-Irminger" inflow of relatively warm and less saline water north through the Denmark Strait resulted. The meltwater also reduced deep water formation and its associated northward heat transport, further cooling the eastern North Atlantic and Europe. These combined changes must have increased the temperature difference between the ocean surface and the Greenland ice sheet and led to continued low GISP2 $\delta^{18}O$ values, while values in Antarctica increased. The evidence suggests a global onset of deglaciation with meltwater effects in the North Atlantic and the Nordic Seas leading to a local cold phase with persistent low $\delta^{18}O$ values in the Greenland ice till the start of the Bølling interstadial at 14.8 ka, while deglaciation proceeded around the Barents Sea (Figure 6e).

Sarnthein et al. [1995] observe in the Nordic Seas during the Bølling/Allerød period meltwater from the North Sea and from middle Norway and a circulation pattern resembling that of the Holocene, including significant deep water formation. This allowed rapid warming in Europe and concomitant $\delta^{18}O$ increase in Greenland. At Barbados, *Guilderson et al.* [1994] report during this period a rapid sea level rise and a drop in SST compatible with a large influx of meltwater into the ocean and a western ocean surface cooling. The Younger Dryas cold period may have been mainly a North Atlantic phenomenon. Continued sea level rise at Barbados shows continued melting, albeit at a reduced rate. The melting may have been largely in the Antarctic where an Antarctic cold reversal (ACR) [*Jouzel et al.*, 1995], which possibly preceded the Greenland Younger Dryas between about 14 and 12.5 ka [*Sowers and Bender*, 1995], was followed by renewed warming (Figure 6d). Also at Andøya there is no sign of significant glacier expansion like that found in southern Norway. This suggests that the Younger Dryas was meltwater induced and centered in the North Atlantic basin like the cold phase preceding the Bølling. Varved sediment cores from the Cariaco basin in the western tropical Atlantic show changes in relative reflectivity (grey scale) that can be correlated in great detail with the GRIP $\delta^{18}O$ record. This suggests that a coupled tropical/high-latitude North Atlantic climate system operated during the last deglaciation, including the Younger Dryas [*Hughen et al.*, 1996]. The Greenland ice cores clearly reflect a glacial to interglacial climate transition modified by meltwater forcing of Nordic Seas/North Atlantic Ocean thermohaline circulation. The increase in the greenhouse gases CO_2 and CH_4 during the "aborted" part of the Greenland deglaciation (18.5-15 ka), the continued ice retreat in the northern part of Scandinavia [*Vorren et al.*, 1988] and the Barents Sea, and the ocean circulation changes [*Sarnthein et al.*, 1995] all played a role in the very rapid and large Oldest Dryas-Bølling climate change recorded at the GISP2 site.

The well known terrestrial ^{14}C ages of the very distinct climate transitions of the classical sequence Oldest Dryas-Bølling-Older Dryas-Allerød-Younger Dryas-Holocene in northwest Europe [*Mangerud et al.*, 1974; *Nilsson*, 1983] allow a check on the ice core timescale. Calibrated ^{14}C-ages of these boundaries agree within 100 years with the GISP2 ice layer counts of these climate transitions [*Stuiver et al.*, 1995]. The summit ice cores [*Grootes et al.*, 1993; *Johnsen et al.*, 1992a; *Stuiver et al.*, 1995] reveal additional detail that has also been found in ocean sediment cores. *Karpuz and Jansen* [1992] identify four additional

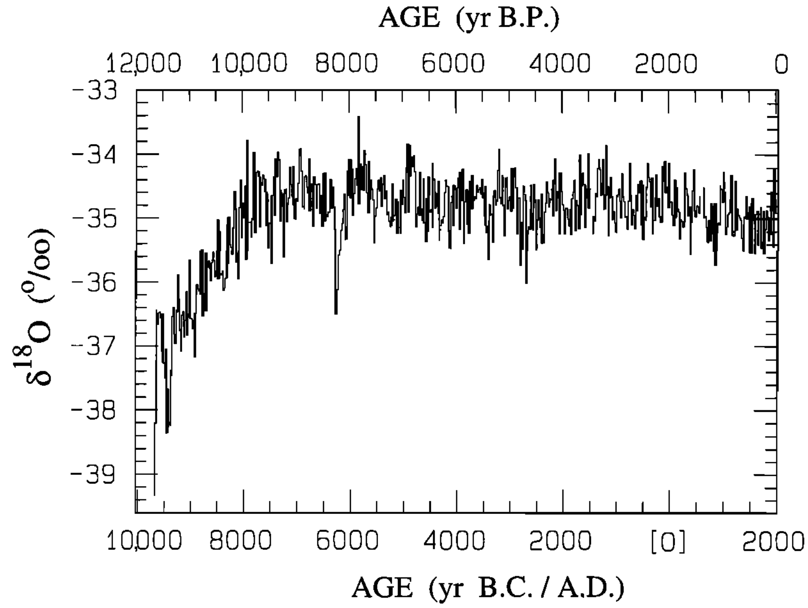

Figure 7. GISP2 bidecadal $\delta^{18}O$ record for the Holocene on the layer-counted timescale of *Meese et al.* [1994].

cold *N. pachyderma* (s) episodes in their AMS ^{14}C-dated cores from the Norwegian Sea, namely the Bølling cold period I, Bølling cold period II (probably equivalent with the Older Dryas), Older Dryas I, Older Dryas II (equivalent with the intra-Allerød cold period of *Lehman and Keigwin* [1992]), and Younger Dryas II (YDII). The calibrated radiocarbon ages of the first four climate transitions are fully compatible with their GISP2 layer counts [*Meese et al.*, 1994]. This supports the detailed correlation between the Greenland summit isotopes and Norwegian Sea SST as well as the dating of both records. Similarly good agreement is found between GRIP ice core ages and the calibrated ^{14}C ages of the Saksunarvatn and Vedde ash layers (10,180 years and 11,980 years [*Grönvold et al.*, 1995]). An even tighter time control to decades is provided by the varved sediments of the Cariaco basin [*Hughen et al.*, 1996].

Stuiver et al. [1995] used the GISP2 $\delta^{18}O$ record to evaluate the oceanic influence, via deep water formation, on global ^{14}C levels. After removing the effect of geomagnetic dipole intensity changes from the ^{14}C-Th/U coral calibration curve [*Bard et al.*, 1993; *Edwards et al.*, 1993] a correlation coefficient of -0.83 between the residual ^{14}C and GISP2 $\delta^{18}O$ was obtained for the 16,500 - 10,000 calendar year B.P. interval. Such a correlation agrees with changes in the rate of NADW formation, where an increase lowers mixed layer and atmospheric ^{14}C levels, while it also leads to a larger heat transport north in the North Atlantic and high-latitude warming.The changes in ^{14}C at the onset of the Bølling and the Younger Dryas seem to lead GISP2 $\delta^{18}O$ by about 400 years. The correlation between ^{14}C and NADW formation indicates that the many glacial cycles of NADW formation intensity, documented by *Rasmussen et al.* [1996], will result in a ^{14}C calibration curve for the glacial period that varies with climate.

Holocene Record

The upper half of the core (above 1676 m depth) consists of ice from the Holocene. It is characterized by a rather stable long-term-mean $\delta^{18}O$ value around which many short-term fluctuations occur (Figure 7). With the exception of the $\delta^{18}O$ valley at 8.2 ka, these fluctuations are not distinctive, and unlike the bottom, glacial half of the core, the $\delta^{18}O$ record lacks clearly identifiable features that can easily be correlated with other records.

The last millennium was sampled with seasonal resolution (about 10 samples per year) in the upper 300 m of cores B and D. The $\delta^{18}O$ record of the last 200 years is discussed by *White et al.* [this issue]. Fourier analysis of the full annual record (Figure 8a) and of the bidecadal Holocene record (Figure 8b) was discussed by *Stuiver et al.* [1995]. They identified in the bidecadal GISP2 $\delta^{18}O$ record spectral periodicities (Figure 8) similar to those encountered in the bidecadal record of atmospheric tree ring ^{14}C concentrations near 2500-3300, 830-1050, 520, 210, and 150 years, and for annual data near 70, 61, and 46-44 years. Since the 210-year ^{14}C periodicity relates fairly unequivocally to solar forcing [*Stuiver and Braziunas*, 1993; *Damon and Sonett*, 1991], the same may hold true for the 210-year periodicity in the GISP2 $\delta^{18}O$ record. The 3300-year periodicity (Figure 8b) may correspond to the 4 kyr found in the glacial part of the record and the 3.5-kyr orbital harmonic [*Yiou et al.*, 1991]. *Stuiver et al.* [1995] conclude that although solar modulation could be responsible for the $\delta^{18}O$ cycles near 61, 210, and 530 years, this cannot be proven statistically given the uncertainties in the estimate of solar forcing and the neglect of possible positive climate feedbacks.

The amplitudes of the longer Holocene cycles are much smaller than that of the decadal-scale $\delta^{18}O$ cycle (Figure 9). *Stuiver et al.* [1995] discussed a possible connection of $\delta^{18}O$ with the 11-year sunspot cycle and found a probability of no correlation of only 0.4% for the A.D. 1700-1924 interval using a 5- to 20-year filter. Similar correlations were found between sunspots *S* and the cosmogenic ^{10}Be levels and ^{14}C production. These correlations account, however, for just 8.5% (^{10}Be-*S*) to 13% ($\delta^{18}O$ -*S*) of the total variance. Figure 9 shows that the 11-year solar (sunspot) signal modulates a stronger signal of about 7 years with peak-to-trough values of up to 3‰ at this resolution corresponding to the strong 6.3-year spectral peak of Figure 8a, also observed by *White et al.* [this issue]. Climate variability in the 4- to 8-year band has been associated with El Niño/Southern Oscillation (ENSO) and the North Atlantic Oscillation (NAO) [see also *White et al*, this issue], both large-scale atmosphere-ocean interactions. The similarity between the climatic $\delta^{18}O$ sig-

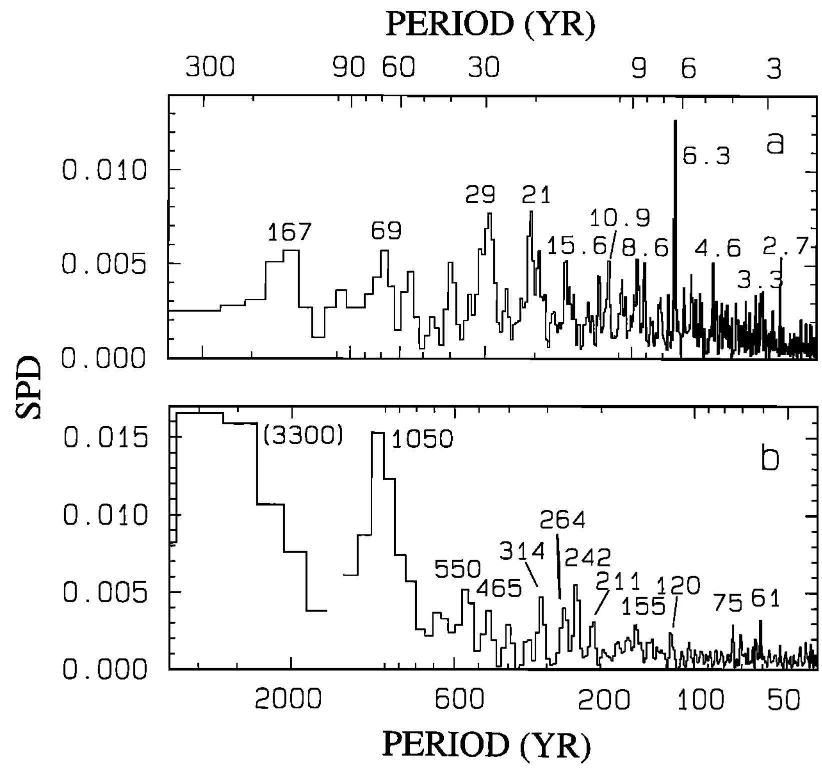

Figure 8. Relative spectral power density (SPD), normalized on the largest peak, obtained by Fourier power spectrum analysis from (a) The 1168-year-long annual GISP2 $\delta^{18}O$ record [*Stuiver et al.*, 1995] and (b) The bidecadal Holocene GISP2 $\delta^{18}O$ record (11,600- 120 calendar years B.P. [*Stuiver et al.*, 1995]. Periods of major peaks are given in years.

nal and the production/concentration variations of the cosmogenic isotopes ^{14}C and ^{10}Be seems surprising at first. Yet ^{10}Be concentrations in snow at the Greenland summit and ^{10}Be fluxes to the summit are the result of production, snow accumulation rate, and long-distance transport. The latter two are tied to atmospheric conditions just like $\delta^{18}O$. The ENSO phenomenon involves an oceanic component with significant changes in Pacific upwelling as well as large changes in biosphere-atmosphere carbon exchange that affect atmospheric CO_2 levels [*Keeling et al.*, 1989]. Since upwelling CO_2 is depleted in ^{14}C, it may also influ-

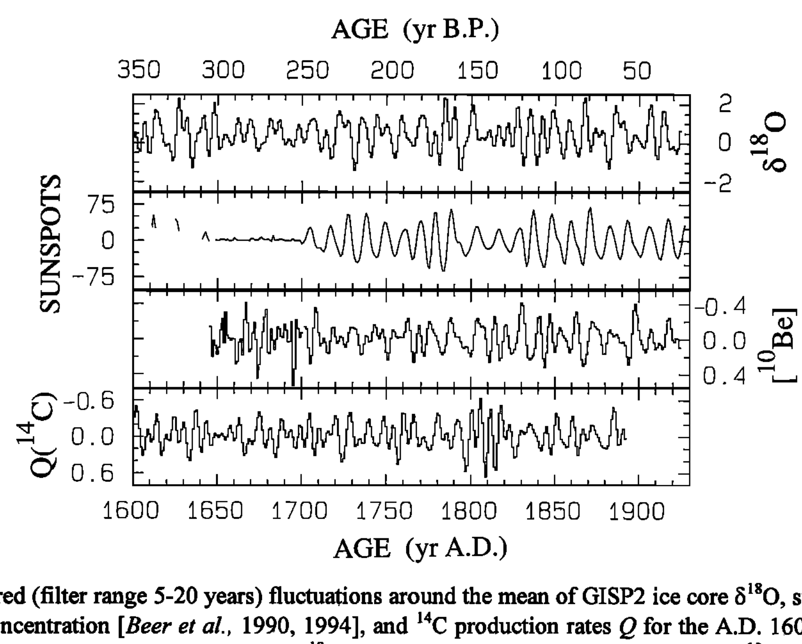

Figure 9. Filtered (filter range 5-20 years) fluctuations around the mean of GISP2 ice core $\delta^{18}O$, sunspot numbers, ice core ^{10}Be concentration [*Beer et al.*, 1990, 1994], and ^{14}C production rates *Q* for the A.D. 1600-1930 interval. The ^{10}Be and *Q* scales are inverted. Units are $\delta^{18}O$, per mil; sunspots, actual numbers; and ^{10}Be and *Q*, fractional deviation from the mean [*Stuiver et al.*, 1995].

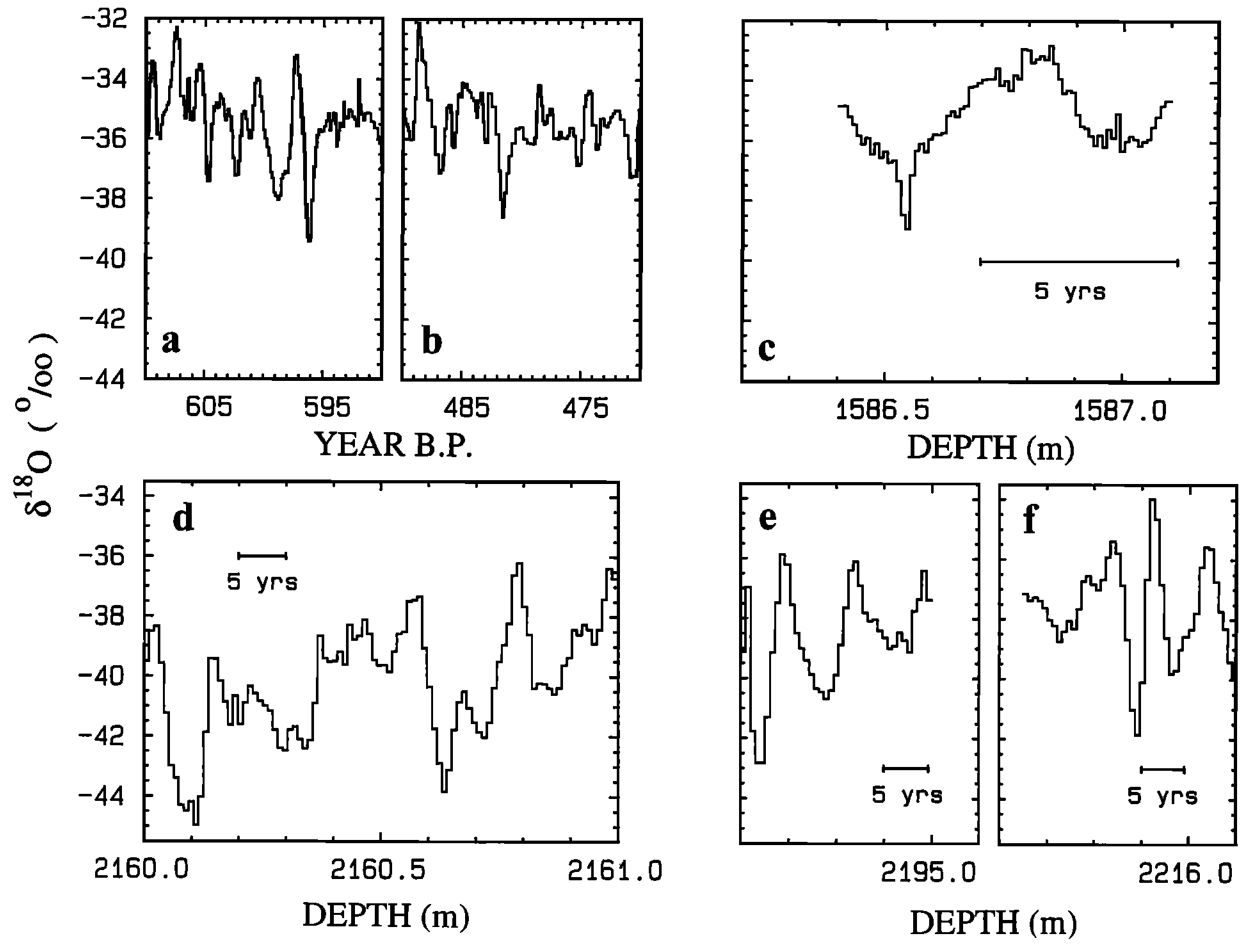

Figure 10. Annual resolution GISP2 $\delta^{18}O$ records showing the dominant decadal-type fluctuations for (a,b,c) Holocene, (d), stadial, (e) IS 8 to stadial transition, and (f) interstadial climates.. Because the general timescale is not reliable at this depth at this resolution, $\delta^{18}O$ is plotted as function of depth The bar indicates 5 years calculated from *Meese et al.* [1994].

ence atmospheric ^{14}C levels. The NAO involves the North Atlantic and Nordic Seas, presently the main area of deep water formation. Changes in the rate of deep water formation influence the ocean-atmosphere exchange of $^{14}CO_2$ and may thus contribute a climatic component to the cosmogenic ^{14}C record.

There is insufficient resolution in the biannual $\delta^{18}O$ record below 300 m depth to follow the strong "decadal" cycle through time down the core. Records from 1-m sections sampled with high resolution intermittently down the core demonstrate decadal-type variability throughout the Holocene and well into the last glacial (Figure 10). At 1580-m depth (10.2 ka), 1-cm sample resolution reveals the weak remnants of what once was the strong seasonal cycle modulating a 4 to 5‰ decadal cycle. With increased thinning in the deeper sections and reduced accumulation during the last glacial the annual layer thickness approaches the 1-cm sample resolution. Yet we see in the deeper sections decadal-type cycles of surprising sharpness and amplitude. These dominate the short-term variability and persist through glacial-interglacial, and stadial-interstadial climate changes (Figure 10), and may even have been stronger during the last glacial. *White et al.* [this issue] link a 7.6-year cycle in the NAO to the GISP2 record, which exhibits periodicity of 7.5-years for the last 200 years and 6.3 years for the last 1168 years (Figure 8a). For periodicities in the range 1 to 10 years, stochastic weather variability may, however, produce comparable cycles. Large decadal $\delta^{18}O$ variability, when linked to climate variability, has serious societal implications for lower, inhabited latitudes. Various brief but severe cold spells in the GISP2 $\delta^{18}O$ record during the Little Ice Age may have been responsible for the demise of the Norse colonies in Greenland [*Stuiver et al.*, 1995]. Further study of the 6.3-year cycle and its possible causes thus is important.

A final question is whether volcanic eruptions can influence climate and whether these aperiodic events can trigger major climate changes. By stacking the $\delta^{18}O$ records of periods containing a major volcanic eruption, *Stuiver et al.* [1995] for the first time demonstrated unambiguously the influence of these eruptions on climate as recorded at the Greenland summit. The acidity signal in the ice lags the first sign of a decrease in $\delta^{18}O$ (cooling) by as much as 1 to 2 years, and recovery of $\delta^{18}O$ (and temperature) takes 3 to 4 years depending on the size of the eruption. The changes in $\delta^{18}O$, from 0.90 ± 0.42‰ for eruptions with volcanic explosivity index (VEI) 6 or 7 to 0.45 ± 0.16‰ for all volcanic events identified by *Zielinski et al.* [1994] for the last millennium in the GISP2 core, are significantly smaller than the decadal type fluctuations just discussed. This and the speedy recovery, commensurate with aerosol atmospheric residence times, make it unlikely that volcanic eruptions play a major role in triggering climate change.

Conclusions

The isotopic composition of snow and frost parallels that of the water vapor in the air from which it is formed. The offset can be modified by local precipitation temperature. Short-term variability is large because of the diurnal temperature cycle and changing air masses. Strong isotopic smoothing in the near-surface firn layers transforms discrete $\delta^{18}O$ depositions into a smooth isotope-depth profile with a dominant seasonal cycle and subannual structure. Continued gas phase isotope transport eliminates most of the seasonal cycle before the firn-ice transition is reached and leaves a dominant decadal-type cycle (6.3-years) with weaker, longer cycles, some of which are attributable to the Sun (11 years, 210 years).

The relatively stable $\delta^{18}O$ values of the Holocene period contrast strongly with the large and rapid switches between stadial and interstadial climate for the interval 11,600 to about 70,000 years B.P. The transition from glacial to interglacial conditions lasted from about 21,000 to 11,600 years ago and started roughly simultaneously in the Arctic and Antarctic. Meltwater fluxes in the North Atlantic and the Nordic Seas occasionally changed ocean circulation and reduced the rate of deep water formation. This led to cold episodes in Europe and negative $\delta^{18}O$ excursions in GISP2 that are not seen with this severity outside the North Atlantic basin. The many stadial-interstadial cycles of the pleniglacial are found in both GISP2 and GRIP and are dominated by a 1470-year periodicity. They correlate closely with North Atlantic sea surface temperature and NADW formation. The global nature of the climate fluctuations can be recognized in Antarctic ice core isotope records like Vostok, in deep-sea carbonate content, planktonic forams, and sediment composition (e.g., grey-scale, magnetic susceptibility), and in European pollen. Benthic $\delta^{18}O$ records, reflecting global ice volume, are, however, insensitive to many of the shorter climate fluctuations. The GISP2 layer-counted timescale, the *Martinson et al.* [1987] orbitally tuned ocean timescale, and dated Huon Peninsula coral terraces show a surprisingly good agreement. The oldest part of the interpretable $\delta^{18}O$ record, corresponding to MIS 5a to 5d, has significantly less $\delta^{18}O$ variability than the full glacial part, with cycles dominated by the familiar precessional forcing modulated by obliquity.

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References

Adam, D.P., J.D.Sims, and C K.Throckmorton, 130,000-yr continuous pollen record from Clear Lake, Lake County, California, *Geology, 9*, 373-377, 1981.

Alley, R.B., E.S.Saltzman, K.M.Cuffey, and J.J.Fitzpatrick, Summer time formation of depth hoar in central Greenland, *Geophys. Res. Lett., 17*, 2393-2396, 1990.

Alley, R.B., et al., Abrupt increase in Greenland snow accumulation at the end of the Younger Dryas event, *Nature, 362*, 527-529, 1993.

Alley, R.B., A.J Gow, S.J.Johnsen, J.Kipfstuhl, D.A.Meese, andT.Thorsteinsson, Comparison of deep ice cores, *Nature, 373*, 393-394, 1995.

Bard, E., M.Arnold, R.G.Fairbanks, and B.Hamelin, ^{230}Th-^{234}U and ^{14}C ages obtained by mass spectrometry on corals, *Radiocarbon, 35*, 191-199, 1993.

Beer, J., et al., Use of ^{10}Be in polar ice to trace the 11-year cycle of solar activity, *Nature, 347*, 164-166, 1990.

Beer, J., S.J Johnsen, G.Bonani, R.C.Finkel, C.C.Langway, H.Oeschger, B.Stauffer, M.Suter, and W.Woelfli, in *The Last Deglaciation: Absolute and Radiocarbon Chronologies, NATO ASI Ser., Ser. I*, vol 2, edited by E.Bard and W.S.Broecker, pp 141-153, Springer-Verlag, New York, 1992.

Beer, J., F.Joos, C.Lukasczyk, W Mende, J.Rodriguez, U.Siegenthaler,and R.Stellmacher, ^{10}Be as an indicator of solar variability and climate. in *The Solar Engine and Its Influence on Terrestrial Atmosphere and Climate*, ed., *NATO ASI Ser., Ser. I*, vol. 25, edited by E. Nesme-Ribes, pp. 221-233, Springer-Verlag, New York, 1994.

Behre, K.E., and J. van der Plicht, Towards an absolute chronology for the last glacial period in Europe: Radiocarbon dates from Oerel, northern Germany, *Veg. Hist. Archaeobot., 1*, 111-117, 1992

Bender, M., T.Sowers, M.-L.Dickson, J.Orchardo, P Grootes, P A.Mayewski, and D.A.Meese, Climate correlations between Greenland and Antarctica during the past 100,000 years, *Nature, 372*, 663-666, 1994.

Bond, G.C., and R.Lotti, Iceberg discharges into the North Atlantic on millennial timescales during the last glaciation, *Science, 267*, 1005-1010, 1995.

Bond, G., W.Broecker, S.Johnsen, J.McManus, L.Labeyrie, J.Jouzel, and G.Bonani, Correlations between climate records from North Atlantic sediments and Greenland ice, *Nature, 365*, 143-147, 1993.

Bond, G., P. deMenocal, and W.Showers, Abrupt climate shifts on sub-Milankovitch time scales in the North Atlantic during the Holocene, and the last glacial-interglacial cycle, *EOS Trans. AGU, 77* (46), Fall Meet. Suppl., F15, 1996.

Chappell, J.M , and N.J Shackleton, Oxygen isotopes and sea level, *Nature, 324*, 137-140, 1986.

Chappell, J , A.Omura, T Esat, M.McCulloch, J.Pandolfi, Y Ota, and B Pillans, Reconciliation of late Quaternary sea levels derived from coral terraces at Huon Peninsula with deep sea oxygen isotope records, *Earth Planet. Sci. Lett., 141*, 227-236, 1996.

Chappellaz, J., J.M.Barnola, D.Raynaud, Y.S.Korotkevitch, and C.Lorius, Ice-core record of atmospheric methane over the past 160,000 years, *Nature, 345*, 127-131, 1990.

Chappellaz, J , T.Blunier, D.Raynaud, J M.Barnola, J.Schwander, and B Stauffer, Synchronous changes in atmospheric CH_4 and Greenland climate between 40 and 8 kyr BP, *Nature, 366*, 443-445, 1993

Chappellaz, J , E.Brook, T Blunier, and B Malaizé, CH_4 and $\delta^{18}O$ of O_2 records from Antarctic and Greenland ice. A clue for stratigraphic disturbance in the bottom part of the Greenland Ice Core Project and the Greenland Ice Sheet Project ice cores, *J. Geophys. Res.*, this issue.

Charles, C.D., J.Lynch-Stieglitz, U.S.Ninnemann, and R.G.Fairbanks, Climate connections between the hemisphere revealed by deep sea sediment core/ice core correlations. *Earth Planet. Sci. Lett., 142*, 19-27, 1996.

Clarke, G.K C., and E.D.Waddington, A three-dimensional theory of wind pumping, *J. Glaciol., 37*, 89-96, 1991.

Cortijo, E., La variabilité climatique rapide dans l'Atlantique Nord depuis 128 000 ans: Relations entre les calottes de glace et l'océan de surface, Ph.D. thesis, Univ. de Paris-Sud, Paris, 1995.

Cortijo, E., P.Yiou, L.Labeyrie, and M.Cremer, Sedimentary record of rapid climatic variability in the North Atlantic Ocean during the last glacial cycle, *Paleoceanography, 10*, 911-926, 1995.

Cuffey, K.M , G.D.Clow, R.B.Alley, M.Stuiver, E.D.Waddington, and R.W.Saltus, Large Arctic temperature change at the Wisconsin-Holocene glacial transition, *Science, 270*, 455-458, 1995

Damon, P.E , and C.P.Sonett, Solar and terrestrial components of the atmospheric ^{14}C variation spectrum, in *The Sun in Time*, edited by C.P.Sonett, M.S.Giampa, and M.S Matthews, pp. 360-388, Univ. of Ariz. Press, Tucson, 1991.

Dansgaard, W., S.J.Johnsen, H.B.Clausen, and C C.Langway Jr, Climatic record revealed by the Camp Century ice core, in *The Late Cenozoic Glacial Ages*, edited by K.K.Turekian, pp 37-56, Yale Univ Press, New Haven, Conn , 1971

Dansgaard, W., S.J.Johnsen, H.B.Clausen, and N.Gundestrup, Stable isotope glaciology, *Medd. Groenl., 197* (2), 1-53, 1973.

Dansgaard, W., et al., Evidence for general instability of past climate from a 250-kyr ice-core record, *Nature, 364*, 218-220, 1993.

de Beaulieu, J.-L., and M.Reille, The transition from temperate phases to stadials in the long Upper Pleistocene sequence from Les Echets (France), *Palaeogeogr. Palaeoclimatol, Palaeoecol., 72*, 147-159, 1989.

de Beaulieu, J.-L., and M.Reille, The last climatic cycle at La Grande Pile (Vosges, France): A new pollen profile, *Quat. Sci. Rev., 11*, 431-438, 1992.

Duvall, M.D., Climate implications in the near surface snow and firn at the summit of the Greenland ice sheet, undergraduate thesis, Univ. of Wash., Seattle, 1993.

Edwards, R.L., J.W.Beck, G.S Burr, D.J.Donahue, J.M.A Chappell, A.L.Bloom, E.R.M.Druffel, and F.W.Taylor, A large drop in atmospheric $^{14}C/^{12}C$ and reduced melting in the Younger Dryas, documented with ^{230}Th ages of corals, *Science, 260*, 962-968, 1993

Fairbanks, R.G., A 17,000-year glacio-eustatic sea level record Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation, *Nature, 342*, 637-642, 1989.

Grimm, E.C., G.L.Jacobson Jr., W.A.Watts, B.C.S.Hansen, and K.A.Maasch, A 50,000-year record of climate oscillations from Florida and its temporal correlation with the Heinrich events, *Science, 261*, 198-200, 1993.

Grönvold, K., N.Oskarsson, S.J.Johnsen, H.B.Clausen, C.U.Hammer, G.Bond, and E.Bard, Ash layers from Iceland in the Greenland GRIP ice core correlated with oceanic and land sediments, *Earth Planet. Sci. Lett., 135*, 149-155, 1995

Grootes, P.M., M.Stuiver, J.W C White, S.Johnsen, and J.Jouzel, Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores, *Nature, 366*, 552-554, 1993.

Guilderson, T.P., R.G.Fairbanks, and J.L.Rubenstone, Tropical temperature variations since 20,000 years ago: Modulating interhemispheric climate changes, *Science, 263*, 663-665, 1994.

Hays, J.D., J.Imbrie, and N.J.Shackleton, Variations in the Earth's orbit: Pacemaker of the ice ages, *Science, 194*, 1121-1132, 1976.

Heusser, L.E., and F Sirocko, Millennial pulsing of environmental change in southern California from the past 24 k.y.: A record of Indo-Pacific ENSO events?, *Geology, 25*, 243-246, 1997

Hughen, K A., J.T.Overpeck, L.C.Peterson, and S Trumbore, Rapid climate changes in the tropical Atlantic region during the last deglaciation, *Nature, 380*, 51-54, 1996

Imbrie, J , J.D.Hays, D.G.Martinson, A.McIntyre, A.C.Mix, J.J.Morley, N.G.Pisias, W.L.Prell, and N.J.Shackleton, The orbital theory of Pleistocene climate: Support from a revised chronology of the marine $\delta^{18}O$ record, in: *Milankovitch and Climate*, Part I, edited by A.L.Berger et al., pp. 269-305, D. Reidel, Norwell, Mass., 1984.

Johnsen, S.J., Stable isotope homogenisation of polar firn and ice, in *Isotopes and Impurities in Snow and Ice, IAHS Publ., 118*, 210-219, 1977.

Johnsen, S J., H.B.Clausen, W Dansgaard, K Fuhrer, N.Gundestrup, C.U.Hammer, P.Iversen, J.Jouzel, B.Stauffer, and A.P.Steffensen, Irregular glacial interstadials recorded in a new Greenland ice core, *Nature 359*, 311-313, 1992a

Johnsen, S.J., H.B.Clausen, W.Dansgaard, N.S.Gundestrup, M.Hansson, P.Jonsson, J.P.Steffensen, and A.E.Sveinbjørnsdottir, A "deep" ice core from East Greenland, *Medd. Groenl., Geosci., 29*, 3-22, 1992b.

Johnsen, S.J., H.B.Clausen, W.Dansgaard, N.S.Gundestrup, C.U.Hammer, and H Tauber, The Eem stable isotope record along the GRIP ice core and its interpretation, *Quat. Res., 43*, 117-124, 1995a.

Johnsen, S.J., D.Dahl-Jensen, W.Dansgaard, and N.S.Gundestrup, Greenland paleotemperatures derived from GRIP bore hole temperature and ice core isotope profiles, *Tellus, Ser. B.,47*, 624-629, 1995b.

Johnsen S.J., et al., The $\delta^{18}O$ record along the Greenland Ice Core Project deep ice core and the problem of possible Eemian climatic instability, *J. Geophys. Res.*, this issue.

Jouzel, J., C.Lorius, J.R.Petit, C.Genthon, N.I.Barkov, V.M Kotlyakov, and V.M.Petrov, Vostok ice core: A continuous isotope temperature record over the last climatic cycle (160,000 years), *Nature, 329*, 403-407, 1987.

Jouzel, J., et al., Vostok ice cores: Extending the climatic signal over the penultimate glacial period, *Nature, 364*, 407-412, 1993.

Jouzel, J , et al , The two-step shape and timing of the last deglaciation in Antarctica, *Clim. Dyn., 11*, 151-161, 1995.

Jouzel, J., et al., On the validity of the temperature reconstruction from water isotopes in ice cores, *J.Geophys. Res.*, this issue.

Jung, S.J.A., Wassermassenaustausch zwischen NE-Atlantik und Nordmeer während der letzten 300.000/80.000 Jahre im Abbild stabiler O- und C-Isotope, Ph.D. dissertation, 104 pp., Christian-Albrechts-Univ , Kiel, Germany, 1996.

Karpuz, N.K., and E.Jansen, A high-resolution diatom record of the last deglaciation from the SE Norwegian Sea: Documentation of rapid climatic changes, *Paleoceanography, 7*, 499-520, 1992.

Keeling, C D., R.B.Bacastow, A.F.Carter, S.C.Piper, T.P.Whorf, M.Heimann, W G.Mook, and H Roeloffzen, A three-dimensional model of atmospheric CO_2 transport based on observed winds, 1, Analysis of observational data, in *Aspects of Climate Variability in the Pacific and the Western Americas, Geophys. Monogr. Ser.*, vol. 55, edited by D.H.Peterson, pp.165-236, AGU, Washington, D.C., , 1989.

Keigwin, L.D., and G.A.Jones, Western North Atlantic evidence for millennial-scale changes in ocean circulation and climate, *J. Geophys. Res., 99*, 12,397-12,410, 1994.

Keigwin, L.D., W.B.Curry, S.J.Lehman, and S.Johnsen, The role of the deep ocean in North Atlantic climate change between 70 and 130 kyr ago, *Nature, 371*, 323-326, 1994

Labeyrie, L.D , J.C.Duplessy, and P.L.Blanc, Variations in mode of formation and temperature of oceanic deep waters over the past 125,000 years, *Nature, 327*, 477-482, 1987.

Laj, C., A.Mazaud, and J -C.Duplessy, Geomagnetic intensity and ^{14}C abundance in the atmosphere and ocean during the past 50 kyr, *Geophys. Res. Lett., 23*, 2045-2048, 1996.

Lehman, S.J., and L.D.Keigwin, Sudden changes in North Atlantic circulation during the last deglaciation, *Nature, 356*, 757-762, 1992.

Leroy, S.A.G., S.Giralt, P.Francus, and G.Seret, The high sensitivity of the palynological record in the Vico maar lacustrine sequence (Latium, Italy) highlights the climatic gradient through Europe for the last 90 ka, *Quat. Sci. Rev. 15*, 189-201, 1996.

Lorius, C., J.Jouzel, C.Ritz, L.Merlivat, N.I.Barkov, Y S.Korotkevich, and V.M.Kotlyakov, A 150,000-year climatic record from Antarctic ice, *Nature, 316*, 591-596, 1985.

Majoube, M., Fractionnement en oxygène-18 et en deutérium entre l'eau et sa vapeur, *J.Chim. Phys., 68*, 1423-1436, 1971.

Mangerud, J., S.T.Andersen, B.E.Berglund, and J J.Donner, Quaternary stratigraphy of Norden, A proposal for terminology and classification, *Boreas, 3*, 109-128, 1974.

Mania, D., Das Jungquartär aus dem ehemaligen Ascherslebener See im Nordharzvorland, *Petermanns Geogr. Mitt., 111*, 257- , 1967.

Martinson, D.G., N G.Pisias, J.D.Hays, J.Imbrie, T.C.Moore Jr., and N.J Shackleton, Age dating and the orbital theory of the ice ages: Development of a high-resolution 0 to 300,000-year chronostratigraphy, *Quat. Res., 27*, 1-29, 1987.

Mayewski, P.A., L.D.Meeker, M.S.Twickler, S.Whitlow, Q.Yang, and M.Prentice, Major features and forcing of a high-latitude northern hemisphere atmospheric circulation using a 110,000-year-long glaciochemical series, *J.Geophys. Res.*, this issue.

McManus, J.F., G.C.Bond, W.S.Broecker, S.Johnsen, L.Labeyrie, and S.Higgins, High-resolution climate records from the North Atlantic during the last interglacial, *Nature, 371*, 326-329, 1994.

Meese, D.A., R.B.Alley, A.J.Gow, P.Grootes, P.A.Mayewski, M.Ram, K.C.Taylor, E.D.Waddington, and G.Zielinski, Preliminary depth-age scale of the GISP2 ice core, *CRREL Spec. Rep. 94-1*, U.S. Army Cold Reg. Res. and Eng. Lab., 66 pp., Hanover, N.H., 1994.

Meese, D.A., A.J.Gow, R.B.Alley, G.A.Zielinski, P.M.Grootes, M.Ram, K.C.Taylor, P.A.Mayewski, and J F.Bolzan, The Greenland Ice Sheet Project 2 depth-age scale: Methods and results, *J.Geophys.Res.*, this issue.

Nilsson, T., *The Pleistocene, Geology and Life in the Quaternary Ice Age*, 651 pp., D.Reidel, Norwell, Mass., 1983.

Paepe, R., I. Mariolakos, E van Overloop, and E. Keppens, Last interglacial-glacial north-south geosoil traverse (from stratotypes in the North Sea basin and in the eastern Mediterranean). *Quat. Int., 5*, 57-70, 1990.

Peel, D.A., Profiles of the past, *Nature, 378*, 234-235, 1995

Pisias, N.G., D.G.Martinson, T.C.Moore Jr , N.J.Shackleton, W.Prell, J.Hays, and G.Boden, High resolution stratigraphic correlation of benthic oxygen isotopic records spanning the last 300,000 years, *Mar. Geol., 56*, 119-136, 1984.

Prell, W.L., J.Imbrie, D.G.Martinson, J.J.Morley, N.G.Pisias, N.J Shackleton, and H.F Streeter, Graphic correlation of oxygen isotope stratigraphy application to the late Quaternary, *Paleoceanography, 1*, 137-162, 1986.

Rasmussen, T.L., E.Thomsen, T C.E. van Weering, and L.Labeyrie, Rapid changes in surface and deep water conditions at the Faeroe margin during the last 58,000 years, *Paleoceanography, 11*, 757-771, 1996.

Reille, M., and J.-L de Beaulieu, Pollen analysis of a long upper Pleistocene continental sequence in a Velay maar (Massif Central, France), *Palaeogeogr. Palaeoclimatol. Palaeoecol., 80*, 35-48, 1990.

Sarnthein, M., et al., Variations in Atlantic surface ocean paleoceanography, 50°-80°N: A time-slice record of the last 30,000 years, *Paleoceanography, 10*, 1063-1094, 1995.

Sarnthein, M., D.Dreger, H.Erlenkeuser, P.Grootes, A.Voelker, and M.Weinelt, Climate oscillations during Stage 2-4 and changes in Atlantic water at 50-70N on bidecadal to centennial timescales, *EOS, Trans. AGU, 77* (46), Fall Meet. Suppl., F14, 1996.

Shackleton, N.J., L. de Abreu, M.A.Hall, E.A.Boyle, J-C Duplessy, and E Vincent, Oxygen isotope stratigraphy of Stage 3. *EOS, Trans. AGU, 77* (46), Fall Meet. Suppl., F15, 1996.

Shuman, C.A., R.B Alley, S.Anandakrishnan, J.W.C.White, P.M.Grootes, and C.R.Stearns, Temperature and accumulation at the Greenland summit, Comparison of high-resolution isotope profiles and passive microwave brightness temperature trend,. *J.Geophys. Res., 100*, 9165-9177, 1995

Shuman, C.A., R.B.Alley, M.A.Fahnestock, P J.Fawcett, R.A.Bindschadler, J.W.C.White, P.M.Grootes, S.Anandakrishnan, and C.R.Stearns, Detection and monitoring of strategic markers and temperature trends at the Greenland Ice Sheet Project 2 using passive microwave remote-sensing data, *J. Geophys. Res.*, this issue

Sirocko, F., D.Garbe-Schönberg, A.McIntyre, and B.Molfino, Teleconnections between the subtropical monsoons and high-latitude climates during the last deglaciation, *Science, 272*, 526-529, 1996.

Sowers, T., and M.Bender, Climate records covering the last deglaciation, *Science, 269*, 210-214, 1995.

Sowers, T., M.Bender, L.Labeyrie, D.Martinson, J.Jouzel, D.Raynaud, J.J.Pichon, and Y.S.Korotkevich, A 135,000-year Vostok-Specmap common temporal framework, *Paleoceanography, 8*, 737-766, 1993.

Stein, R., S.-I.Nam, C.Schubert, C.Vogt, D.Fütterer, and J.Heinemeier, The last deglaciation event in the eastern central Arctic Ocean, *Science, 264*, 692-696, 1994

Stuiver, M., and T.F.Braziunas, Sun, ocean, climate and atmospheric $^{14}CO_2$: An evaluation of causal and spectral relationships, *Holocene, 3*, 289-305, 1993.

Stuiver, M., P.M.Grootes, and T.F.Braziunas, The GISP2 $\delta^{18}O$ climate record of the past 16,500 years and the role of the Sun, ocean, and volcanoes, *Quat. Res., 44*, 341-354, 1995.

Taylor, K.C., C.U.Hammer, R B.Alley, H.B.Clausen, D.Dahl-Jensen, A.J.Gow, N.S.Gundestrup, J.Kipfstuhl, J.C.Moore, and E.D.Waddington, Electrical conductivity measurements from the GISP2 and GRIP Greenland ice cores, *Nature, 366*, 549-552, 1993.

Tzedakis, P.C., Long-term tree populations in northwest Greece through multiple Quaternary climatic cycles, Nature, 364, 437-440, 1993.

Vorren, T.O., K.-D.Vorren, T.Alm, S.Gulliksen, and R.Løvlie, The last deglaciation (20,000 to 11,000 B.P.) on Andøya, northern Norway, Boreas, 17, 41-77, 1988.

Waddington, E.D., and D.L.Morse, Spatial variations of local climate at Taylor Dome, Antarctica: Implications for paleoclimate from ice cores, Ann. Glaciol., 20, 219-225, 1994.

Watts, W.A., J.R.M.Allen, and B.Huntley, Vegetation history and climate of the last glacial period (73,000 to 12,000 years before present) at Laghi di Monticchio, southern Italy, Quat. Sci. Rev., 15, 113-132, 1996

Whillans, I.M., and P.M.Grootes, Isotopic diffusion in cold snow and firn, J.Geophys.Res., 90, 3910-3918, 1985.

White, J.W.C., L.K.Barlow, D.A.Fisher, P.M.Grootes, J.Jouzel, S.J.Johnsen, M.Stuiver, and H.Clausen. The climate signal in the stable isotopes of snow from Summit, Greenland: Results of comparisons with modern climate observations, J.Geophys. Res., this issue

Woillard, G.,Grande Pile peat bog: A continuous pollen record for the last 140,000 years, Quat. Res., 9, 1-21, 1978.

Woillard, G.M, and W.G.Mook, Carbon-14 dates at Grande Pile. Correlation of land and sea chronologies, Science, 215, 159-161, 1982.

Yiou, P., C. Genthon, M. Ghil, J. Jouzel, H. Le Treut, J.M.Barnola, C.Lorius, and Y.N.Korotkevitch, High-frequency paleovariability in climate and CO2 levels from Vostok ice core records, J. Geophys. Res., 96, 20,365-20,378, 1991.

Zagwijn, W., and R.Paepe, Die Stratigraphie der weichselzeitlichen Ablagerungen der Niederlande und Belgiens, *Eiszeitalter Ggw. 19*, 129-146, 1968.

Zielinski, G.A, P.A.Mayewski, L.D.Meeker, S.Whitlow, M.S.Twickler, M.Morrison, D.A.Meese, A.J.Gow, and R.B.Alley, Record of volcanism since 7000 B.C. from the GISP2 Greenland ice core and implications for the volcano-climate system, *Science, 264*, 948-952, 1994 (Technical commentary, *Science, 267*, 256-258, 1995.)

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