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The $\delta^{18}O$ record along the Greenland Ice Core Project deep ice core and the problem of possible Eemian climatic instability

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Abstract. Over 70,000 samples from the 3029-m-long Greenland Ice Core Project (GRIP) ice core drilled on the top of the Greenland Ice Sheet (Summit) have been analyzed for $\delta^{18}O$. A highly detailed and continuous $\delta^{18}O$ profile has thus been obtained and is discussed in terms of past temperatures in Greenland. We also discuss a three-core stacked annual $\delta^{18}O$ profile for the past 917 years. The short-term (<50 years) variability of the annual $\delta^{18}O$ signal is found to be 1‰ in the Holocene, and estimates for the coldest parts of the last glacial are 3‰ or higher. These data also provide insights into possible disturbances of the stratigraphic layering in the core which seems to be sound down to the onset of the Eemian. Spectral analysis of highly detailed sequences of the profile helps determine the smoothing of the $\delta^{18}O$ signal, which for the Holocene ice is found to be considerably stronger than expected. We suggest this is due to a process involving diffusion of water molecules along crystal boundaries in the recrystallizing ice matrix. Deconvolution techniques were employed for restoring with great confidence the highly attenuated annual $\delta^{18}O$ signal in the Holocene. We confirm earlier findings of dramatic temperature changes in Greenland during the last glacial cycle. Abrupt and strong climatic shifts are also found within the Eem/Sangamon Interglaciation, which is normally recorded as a period of warm and stable climate in lower latitudes. The stratigraphic continuity of the Eemian layers is consequently discussed in section 3 of this paper in terms of all pertinent data which we are not able to reconcile.

1. Introduction

The main objective of ice-core drilling and analysis is to study past climates and other past environmental changes. Ice cores offer great potential for such studies, because each annual layer deposited on polar ice sheets is assumed to preserve all components that were deposited with the snow in the year the layer was formed. The isotopic composition of the ice yields information about the temperature [*Cuffey et al.*, 1995; *Dansgaard*, 1964; *Johnsen et al.*, 1995a, 1989; *Jouzel et al.*, this issue]; the dust about storminess and source efficiency [*Hammer*, 1977; *Steffensen*, this issue]; the air bubbles about the greenhouse gases in the atmosphere [*Bender et al.*, 1994; *Chappellaz et al.*, 1993; *Oeschger et al.*, 1984; *Raynaud et al.*, 1993; *Sowers and Bender*, 1995]; the acidity about volcanic eruptions in the northern hemisphere [*Clausen et al.*, 1995; *Hammer et al.*, 1980]; and chemical traces about various processes on land, in the sea, and in the atmosphere [*Clausen*, 1995; *Fuhrer et al.*, 1993; *Mayewski et al.*, 1993], at the time of formation of the ice. Together, they give a detailed picture of the environment and climate of the past. All these parameters can be examined in great temporal detail and thus throw light on various facets of the complicated mechanism of climate and the leads and lags between individual parameters.

A total of four ice cores, reaching bedrock, have been drilled through the Greenland Ice Sheet, and two more have been drilled to bedrock on marginal ice caps around the inland ice (Figure 1). The main object of this study is the deep ice core drilled in 1990–1992 by the joint European Greenland Ice Core Project (GRIP). The ice core was recovered at the top of the ice sheet, Summit (72°34′N; 37°37′W; elevation 3232 m above sea level (masl)) [*Dansgaard et al.*, 1993; *GRIP Project Members*, 1993; *Johnsen et al.*, 1992a], which is the only locality in central Greenland where at present no horizontal ice movement is thought to take place. The core is 3028.8 m long and reaches into the silty ice layers immediately above bedrock. It is of excellent quality with no known missing pieces and very few extra breaks except for the brittle zone from 700 to 1300 m.

Five GRIP shallow cores have also been drilled at Summit. Their stable isotope profiles are discussed in sections 2.5.2 and 2.6 and in another paper in this issue [*White et al.*, this issue].

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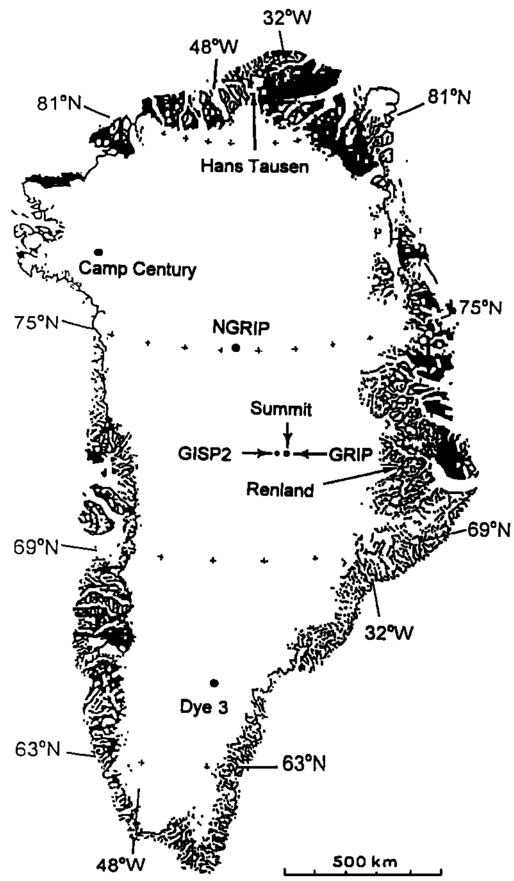

Figure 1. Six deep drilling sites in Greenland: Camp Century (U.S. Army Cold Regions Research and Engineering Laboratory, 1966), Dye 3 (GISP, 1981), Renland (Nordic Council of Ministers, 1988), Summit (GRIP, 1992), the U.S.A. camp (GISP2, 1993), Hans Tausen (Nordic Environmental Research Programme, 1995), and new deep drilling started at NGRIP in 1996.

2. The $\delta^{18}O$ Record

2.1. Conversion of $\delta^{18}O$ to Temperature

Various isotopic, chemical, and physical properties of the GRIP core have been measured at several European laboratories. The stable isotope ratio, $^{18}O/^{16}O$ (or D/H), is the main reference parameter, since its variability is determined mainly by the cloud temperature at the moment of snow formation and thus has direct climatic relevance, assuming unchanged temperature and humidity at the original moisture source areas [*Jouzel et al.*, this issue]. On the Greenland Ice Sheet, the present mean annual $\delta^{18}O$ (the per mil deviation of the $^{18}O/^{16}O$ ratio in a sample from the $^{18}O/^{16}O$ value in standard mean ocean water (SMOW), hereinafter denoted by δ) of the snow is related closely to the mean annual surface temperature, T, in degrees Celsius, by the formula [*Johnsen et al.*, 1989]

$$\delta = 0.67T - 13.7‰ \qquad (1)$$

The temporal relationship between the δ and the surface temperature (T) can be inferred by modeling the borehole temperature profile, based on a surface temperature history determined by a simple relationship to the well-dated δ profile. The parameters defining the relationship are determined by a least squares fit of the modeled temperature profile to the measured temperature profile. Calibration for decadal to century long temperature changes can only be done for the past several hundred years and is obtained from shallow core data [*Cuffey et al.*, 1994; *Johnsen*, 1977b]. Deep ice-core and borehole data provide information for calibrating millennial or longer-term temperature changes [*Cuffey and Clow*, this issue; *Cuffey et al.*, 1995; *Dahl-Jensen and Johnsen*, 1986; *Johnsen et al.*, 1995a]. In the modeling of the GRIP temperature profile, a second-order dependency of temperature on δ was allowed and corrections were made for the past δ variations of seawater. The Greenland Ice Sheet Project 2 (GISP2) calibration assumed a linear dependency of T on δ and no seawater correction was applied. Figure 2 shows the calculated cloud condensation temperature plotted against δ, based on a single-source precipitation model for Greenland [*Johnsen et al.*, 1989], the surface (spatial) temperature to δ relationship for (1), and the long-term borehole temperature calibrations obtained from the GISP2 [*Cuffey et al.*, 1995] and the GRIP [*Johnsen et al.*, 1995a] deep boreholes.

The GRIP calibration curve in Figure 2 is based on a slightly improved model for the temperature profile calculations by accounting fully for the thermal properties of the firn layer. This resulted in slight changes in the $\delta - T$ relationship published earlier [*Johnsen et al.*, 1995a], from 1.7 to 2.0°C/‰ at −35‰ and from 3.5 to 3.1°C/‰ at −42‰. Changes in other results were negligible. The difference in the two calibration curves for the more negative delta values is mainly due to the different δ scales used because the GISP2 δ values were not corrected for the δ seawater history.

Several points can be raised concerning the data in Figure 2. We observe that the modeled cloud temperatures are generally higher than the surface temperatures, which account for the transport of heat by storm tracks to the Greenland Ice Sheet [*Loewe*, 1936]; the slope of the cloud curve is only 1.04°C/‰ which indicates that precipitating clouds are not easily cooled off and their δ values are lowered, even during times of cold climate; and finally, the linear GISP2 calibration curve shows temperatures higher than the cloud temperatures at the upper

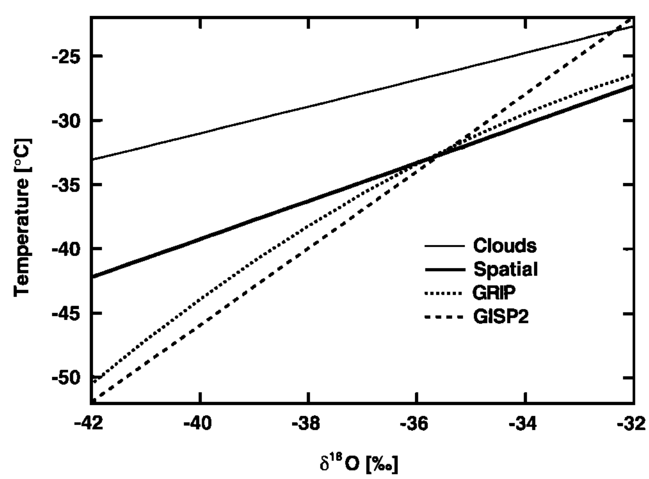

Figure 2. The relationship between temperatures and $\delta^{18}O$ in Greenland. The cloud condensation temperature is shown as the thin line, and the spatial relationship is shown as the thicker line. For long-term temporal changes at Summit, the dotted curve shows the GRIP temperature profile calibration [*Johnsen et al.*, 1995a] and the dashed line shows the GISP2 temperature profile calibration [*Cuffey et al.*, 1995].

end of the δ scale, which is probably not a realistic scenario. This questions the validity of assuming a strict linear relationship between δ and T for millennial timescales and warm climate.

The obvious difference in the $\delta - T$ relationship has generated a great deal of debate and is further discussed by other authors in this issue [*Jouzel et al.*, this issue].

2.2. Stable Isotope Sampling

Over 102,000 samples from the GRIP deep core and supplementary shallow cores have been analyzed for $\delta^{18}O$. The main objective of the sampling was to provide a continuous and detailed δ profile from surface to bedrock. The sampling resolution approaches or exceeds the inherent resolution of the δ profile, which is determined by diffusion processes in firn and solid ice [*Hammer et al.*, 1978; *Johnsen*, 1977a]. The entire core was cut in 55-cm continuous increments, so-called bag samples, and selected core sections were additionally sampled in much finer details, ranging from 2 to 53 samples per bag (Table 1). A maximum degree of data integrity was thus accomplished by independently sampling and measuring two continuous profiles from the core, except for a 230-m-long interval in the brittle zone, where it was only intermittently resampled (Table 1).

2.3. Sample Handling

Before logging and sampling the core in the field, the cores from each drilling run were matched together with the preceding core to affect a perfect fit. It is important to process the core in this way to ensure continuity in the recovered cores and to permit accurate cross correlation of subsequent measurements. A specially designed band saw, built at the Physical Institute, University of Bern, was then used to cut long slices from the core without introducing further breaks. Samples, ranging from 10 to 30 mL in volume, were then taken from the long cut with a band saw and stored in premarked plastic bags. The ice samples were kept frozen until they were melted in hermetically sealed metal cans and then stored in plastic bottles. The integrity of the stable isotope data was further assured by keeping the samples frozen at all times, except for a few hours in the laboratory, when aliquots were taken for the δ measurements.

2.4. The $\delta^{18}O$ Measurements

Most of the stable isotope measurements were made at the Department of Geophysics, University of Copenhagen, at a rate of 115 double measurements per day. Selected samples, including sequences for the diffusion study, were measured at the Science Institute, University of Iceland, Reykjavik. The standard CO_2 equilibrium method was employed at both laboratories, using normally 5-mL water samples and two sets of secondary working standards for each batch of samples. The routine measuring accuracy is $\pm 0.07‰$ at the Copenhagen laboratory and $\pm 0.03‰$ at the Reykjavik laboratory. In order to prevent unwanted shifts in the stable isotope data sets, all the secondary standards were calibrated regularly against V-SMOW and standard low Antarctic precipitation (SLAP) at the Reykjavik laboratory.

Table 1. The $\delta^{18}O$ Sampling Sequence for the GRIP Deep Core

Depth Interval, m	Number of Samples per Bag	Sample Length, cm
0–3022	1	55.0
0–770	22	2.5
799–1640[a]	28–53	1.96–1.04
1000–1500	2	27.5
1500–1900	4	13.75
1900–2695	8	6.88
2695–3028.6	16	3.44

[a]Fifteen core increments from 5 to 9 m long were resampled.

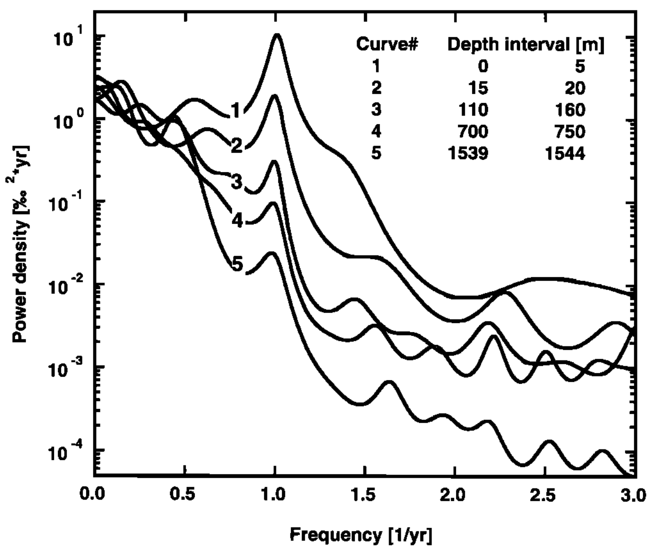

Figure 3. MEM power spectra for detailed $\delta^{18}O$ profiles demonstrating, for five different levels in the GRIP deep core; the presence of the weak annual $\delta^{18}O$ cycle by the spectral peaks is observed at 1 cycle/yr. The effect of smoothing by diffusion is manifested in the drastic lowering of the power densities at higher frequencies.

2.5. Some Inherent Properties of the δ Record

The climate signal is normally inferred from the mean values of the δ profile. Second-order parameters, like the standard deviation from the mean or RMS values, and power spectral densities [*Yiou et al.*, this issue (b)], are also interesting as an aspect of the climate signal. These properties are furthermore essential for studying deposition noise [*Fisher et al.*, 1985], disturbed stratigraphy, and in situ changes of the δ signal due to processes like firnification and diffusion in solid ice.

2.5.1. Spectral fingerprints. The maximum entropy spectral analysis method (MEM) [*Ulrych and Bishop*, 1975], used in this paper, has several properties that make it suitable for characterizing the spectral properties of our δ records: (1) the resolution of the calculated spectra can be controlled by selecting an appropriate autoregressive (AR) order of the prediction error filter employed by the method, (2) by increasing the AR order, the splitting of spectral peaks indicates quasi periodicity of the corresponding signal component, (3) the frequency resolution is not fixed but adapts to the structure of the data [*Lacoss*, 1971], and (4) spectral peaks can be integrated analytically [*Johnsen and Andersen*, 1978] in order to estimate the amplitude of the corresponding data oscillations.

MEM power spectra of detailed δ profiles at five depth levels in the core are shown in Figure 3. The damping of the δ signal due to diffusion lowers the spectral densities at higher frequencies down to the flat white noise level of the measuring errors. This is found to be consistent with a measuring accuracy of

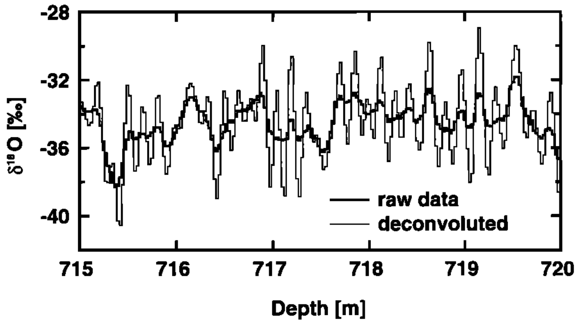

Figure 4. A detailed $\delta^{18}O$ profile from 715- to 720-m depth is shown as the thick curve. The corresponding deconvoluted profile (thin curve) shows the restored annual $\delta^{18}O$ cycle, which is used for counting annual layers.

0.07‰ for the higher four profiles and 0.03‰ for the lowest one, which was measured at the Reykjavik laboratory. The spectral peak of the annual cycle is clearly discernible on the spectral ramp at 1 cycle/yr and is obviously being increasingly attenuated by diffusion with depth. Analytical integration of the annual spectral peaks [*Johnsen and Andersen*, 1978] shows the mean amplitude of the annual δ cycle to be about 4.0‰ at the surface, 0.4‰ at 130 m, 0.24‰ at 725 m, and 0.13‰ at 1540 m.

In the main part of the Holocene period the annual δ cycle is sufficiently well preserved to be used for dating purpose after it has been restored by deconvolution [*Johnsen*, 1977a].

The progressive reduction of the annual δ cycle amplitude with depth was unexpected, due to the low temperatures at Summit, and indicates that processes other than self-diffusion in single crystals [*Ramseier*, 1967b] are responsible for a strong interstrata mixing. The processes responsible for the observed excess diffusion, are believed to be a form of crystal boundary ("CB") diffusion (S. J. Johnsen et al., manuscript in preparation, 1997) combined with ongoing recrystallization with crystal rotation and diagonalization in the ice [*Thorsteinsson et al.*, this issue]. The CB diffusion constant is found to be about 10 times higher in the −32°C cold Holocene ice than the normal single-crystal diffusion constant [*Ramseier*, 1967a]. Studies of the detailed δ profile, at 1650-m depth in the cold and dusty Younger Dryas period, show less smoothing, suggesting that CB diffusion has been much less active. This implies that the impurities sitting at the crystal boundaries [*Fisher*, 1987] inhibit the mobility of the water molecules along the crystal boundaries. Furthermore, we could expect that ions could also move along these boundaries and even take part in chemical reactions there. Further studies are needed, however, in order to investigate this unexpected behavior of the δ signal and its possible implications for the chemical signals.

An example of a deconvoluted high-resolution δ profile is presented in Figure 4. The thick line is the measured δ profile, and the thin line shows the reconstructed δ signal. The annual δ cycle is restored using the deconvolution technique and is used subsequently for dating the core by counting the high δ summer peaks. This procedure also reveals the annual accumulation rates after corrections have been made for ice densities and vertical strain since deposition [*Dahl-Jensen et al.*, 1993; *Reeh et al.*, 1978].

All six cores drilled at Summit have subsequently been dated by counting annual cycles observed in two continuous parameters, the deconvoluted δ profiles and the detailed electrical conductivity measurements (ECM) or acidity profiles, which also provide several volcanic markers for cross correlation [*Clausen et al.*, this issue]. We have, based on this counting effort, which assigns each individual year to a prescribed sequence of samples, constructed six pairs of accurately dated mean annual δ and accumulation rate records. The longest record, from the top 770 m of the deep core, reaches 3809 years back in time; three of the records have a 917-year overlap from A.D. 1063 to 1979 and are discussed in section 2.6. Five of the records overlap from A.D. 1772 to 1979 and are discussed in another paper in this issue [*White et al.*, this issue] together with their GISP2 counterpart.

2.5.2. The δ noise. The high number of long, overlapping δ records from Summit provides us with an excellent opportunity for studying the δ noise, often referred to as deposition noise [*Fisher and Koerner*, 1994; *Fisher et al.*, 1985], inherent in the Summit δ records for present-day climate conditions. One way of estimating the signal to noise (S/N) variance ratio is by comparing the variance of a stacked record (VAR_s) based on n overlapping records, with the mean of the variances (VAR_M) for the n individual records. The estimate of a single record S/N variance ratio then becomes

$$S/N = (VAR_s - VAR_M/n)/(VAR_s - VAR_M). \qquad (2)$$

For the five 208-year-long δ records we find VAR_s = 1.38‰², VAR_M = 0.82‰², and S/N = 0.97. For the three 917-year-long δ records we find VAR_s = 1.12‰², VAR_M = 0.72‰², and S/N = 0.87, in agreement with earlier estimates for other Greenland δ records [*Fisher et al.*, 1985]. The single-record variance of the deposition noise for the 917-year-long records thus becomes 0.6‰² and reduces to 0.2‰² for the stacked record, which contains an estimated 0.52‰² signal variance. We have thus increased the S/N ratio from 0.87 to 2.6, which allows firmer conclusions to be drawn about the mean annual δ signal at Summit back to A.D. 1063. It should be pointed out that the remaining 0.2‰² noise component does not include distortions to the δ signal common to all the cores due, for example, to changing seasonality of the precipitation on Summit. We have therefore to distinguish between the δ signal at Summit and its climatic component as discussed below.

The problem of the δ noise is further discussed by *Jouzel et al.* [this issue]. A comprehensive study involving measurements and modeling of the detailed δ signal in shallow and intermediate ice cores from several sites in Greenland has been performed [*Andersen*, 1993; also U. Andersen, manuscript in preparation, 1997). The approach chosen was to model the individual precipitation events (time and amount), and their uneven deposition due to sastrugi, in such a way that after a simulated firn diffusion, the MEM power spectra of the modeled series matched the power spectra of the real data, calculated with the same AR order around the annual peak. At lower frequencies an additional variance, corresponding to a climatic signal, had to be added in order to obtain an acceptable overall fit to the measured spectra. For Summit this hypothetical climatic δ component was found to have a variance of the order of 0.4‰². These preliminary results suggest that after the deposition noise of the δ records has been reduced by stacking the three records, resulting in a 0.52‰² signal variance stacked record, we still have a residual noise component, of nonclimatic origin, with a variance of ~0.3‰². Of this variance, 0.2‰² is from deposition effects, as discussed above, and 0.1‰² is from seasonality. In order to reduce this ultimate noise component further, a greater number of δ records, also from other Greenland sites, need to be included in the stacking [*Hibler and Johnsen*, 1979].

2.5.3. Short-term variability of the δ record. In the Holocene, the RMS variability of the mean annual δ values is generally found to be of the order of 1‰. For the cold glacial periods in the lower half of the core, much higher short-term variability is normally encountered in the data [*Ditlevsen et al.*, 1996]. We have estimated this variability by filtering the data with a low-pass 1-m Gaussian filter (σ = 0.221 m) and calculating the RMS variability of the raw data deviations from the filtered data set over 101-point running intervals. The result is shown in Figure 5b, along with the filtered δ profile in Figure 5a. A clear anticorrelation is observed between these two records, which together represent the main features of the highly detailed δ record. Another interesting feature is the steadily decreasing variability down to the onset of the Eemian at 2870 m, manifesting the increased diffusion and number of years in each sample. This trend does not, however, continue deeper down. This suggests that stratigraphic disturbances on the scale of 1 m or less are quite common below the Eemian strata.

We have estimated the variability of the mean annual δ values by multiplying the data variability, as described above, with the square root of the estimated number of years in each sample (Table 1), according to the ss08 age model developed for the core [*Johnsen et al.*, 1995b]. This procedure assumes a flat spectral structure for the original signal, disregarding the long-term components, which are not included in the calculations. These estimates are shown in Figure 5c, and we consider them to be minimum estimates due to diffusion damping, particularly below 2600 m, as expected from modeling of the diffusion lengths. The highest annual δ variability in the last glacial, around 3‰ or more, is found in the coldest periods and is unlikely to be due to deposition noise alone, which is expected to be of the order of 1.3‰ for the 3 times lower accumulation rates during these periods as compared to the present conditions [*Alley et al.*, 1993; *Dahl-Jensen et al.*, 1993], where we find 0.8‰. We thus conclude that the Greenland climate, during the cold glacial stades, was considerably more unstable than our present-day interglacial climate.

Another interesting feature of the annual δ variability data for the last glacial (Figure 5c) is the unexpected stable level encountered in the warm interstadials, which is generally truncated at 1.8‰ when their δ values are higher than −39‰. This implies that the anticorrelated annual δ variability does not mirror the sawtooth shape of the δ profiles of the interstadials. This stability, not observed in the uncorrected data (Figure 5b), renders some support to the validity of the age model used for making the corrections from sample variability to variability of the annual δ values. The lower variability observed in the warm interstadials 5a and 5c at around 2640 m and 2740 m, respectively, is due to the increasing smoothing by diffusion, but also shifts to higher values for δs below −39‰ are observed in this part of the core.

A clue to the significance of the −39‰ level of the GRIP δ record can perhaps be found in a comparison between the GRIP stage 5 δ record and the Deep Sea Drilling Project (DSDP) 609 and V29-191 ocean sediment records [*McManus et al.*, 1994]. It is observed that when the GRIP δ values are below a −38.5 to −39‰ trigger level, the *N. pachyderma* s. and ice-rafted debris (IRD) concentrations are generally high, pointing to colder sea surface temperatures (SSTs) and melting icebergs west of Ireland and a strong southward shift of the North Atlantic polar front. This suggests that general instability in the position of the polar front on decadal timescales is a primary cause of the high short-term variability observed in the Greenland climate during the last glacial as well as of the associated atmospheric circulation changes [*Ditlevsen et al.*, 1996; *Kapsner et al.*, 1995].

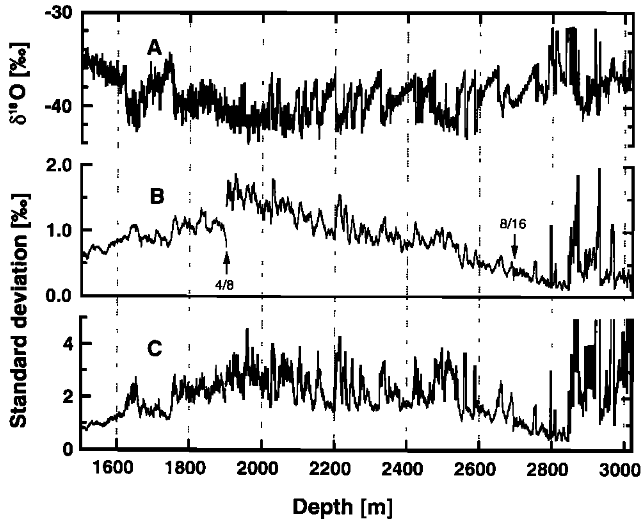

Figure 5. (a) The detailed GRIP $\delta^{18}O$ profile smoothed with a 1-m Gaussian low-pass filter. (b) The high-resolution RMS variability of the deviations of the unfiltered data from the filtered curve in Figure 5a, calculated for a 101-point running window. The two arrows show where the sampling lengths were changed from four to eight samples per bag and from eight to 16 samples per bag, resulting in a jump in the variability by a factor of $\sqrt{2}$. (c) Estimates of the short-term variability for the mean annual δ signal. The estimates are found by dividing the data in Figure 5b by the square root of the numbers of calculated years represented by each sample. The shifts at the arrows in Figure 5a are thus no more discernible. The anticorrelation with the δ record in Figure 5a is evident.

The data discussed above on the short-term variability of climate can be compared with predictions from general circulation models (GCMs) which deal with the water isotopes in the hydrological cycle [*Joussaume et al.*, 1984; *Jouzel et al.*, 1987] to help constrain the models in terms of the effect of possible ocean circulation changes and to improve our understanding of the data.

2.6. The Stacked Annual δ Record Since A.D. 1063

Three of the cores drilled at Summit, Eurocore, Sum89-3, and Sum93, overlap in the time period A.D. 1063 to 1979. In Figure 6b we present the stacked annual δ record for the three cores filtered by a 30-year Gaussian low-pass filter in order to emphasize the longer-term changes. We recognize the well-known warming around A.D. 1930 and the very cold conditions at the end of the seventeenth century. The most dramatic feature of the filtered δ record is the strong and rapid shift from very cold to very warm conditions at the end of the fourteenth century, which is also shown as annual δ values from A.D. 1320 to 1420 in Figure 6a. A second interesting feature of the record in Figure 6a is the warm sawtooth-like peak, leading up to the very cold period around A.D. 1380. Such cycles, which are often discernible in the stacked annual δ record on timescales from 10 to 200 years, probably represent

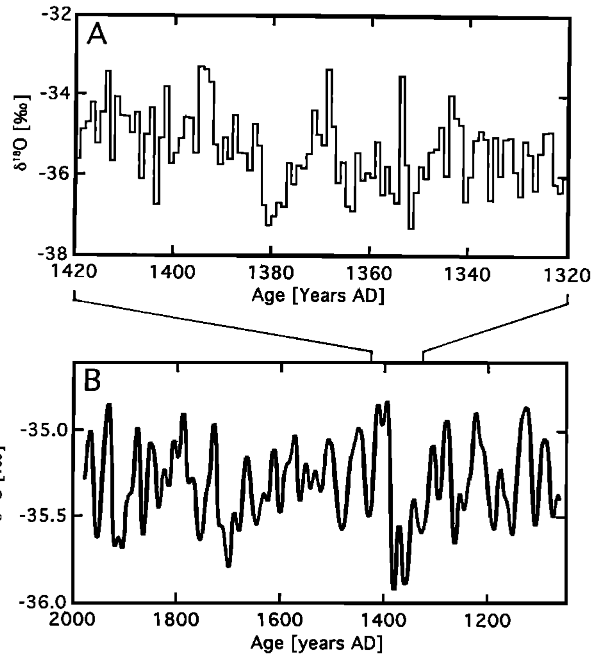

Figure 6. (a) The stacked mean annual $\delta^{18}O$ record for the period A.D. 1320 to 1420. (b) The 917-year-long stacked mean annual δ record, based on three Summit δ records, smoothed with a 30-year Gaussian low-pass filter.

a true feature of the Greenland climate: a rapid warming followed by a much slower cooling similar to that of the 1920s and 1930s. A similar sawtooth shape also characterizes the much longer-lasting warm interstadials of the last glacial period (compare Figure 8).

Due to the reduced noise in the stacked record, we should be able to detect possible periodic features in the data with greater confidence than normally done when using single-core records. The MEM power spectrum (AR = 30) for the unfiltered 2-year mean stacked record is shown in Figure 7. The three strongest spectral peaks seen represent enhanced variability of the δ record centered on periods of 61.6, 19.4, and 11.6 years. The data structures corresponding to the spectral region around the 61.6-year peak are presented in Figure 6a. That peak does not seem to represent any strictly periodic component, since it keeps splitting up with increased AR order. The second peak at 19.4 years is more stable and converges rapidly to a 19.9-year peak with increased AR order. A ~20-year spectral peak has also been discussed and found significant, at the 99% level [*Hibler and Johnsen*, 1979], in a similar study on stacked annual δ records from accurately dated cores, drilled at Dye 3, Milcent, and Crète on the Greenland Ice Sheet. The authors suggest a correlation with solar activity modulated by the tidal forces on the Sun, which display two sharp periodic components of 19.8 and 11.9 years, caused by the orbital periods and gravitational pull of Jupiter and Saturn.

Two spectral peaks at 19 and 10.5 years, significant at the 90% level, have been detected in the detailed ^{10}Be data from the Holocene part of the GRIP core [*Yiou et al.*, this issue (a)]. The authors find the 19-year periodicity more stable and assign an error of 1–2 years to the period lengths, which makes a correlation with our 19.4- and 11.6-year peaks possible.

The third distinct spectral peak, representing cyclic periods from about 11 to 12 years, shows some stability with respect to increasing AR order but does eventually split up into two stable peaks at 11.2 and 11.9 years. This behavior is confirmed by band-pass filtering (9.5 to 13 years), which reveals a long-term (100 to 200 years) beating signal with an amplitude up to 0.5‰. This component in the mean annual δ signal at Summit can generally not be traced back to the 11-year sunspot cycle. There are two reasons for this conclusion: (1) the maximum of the spectral peak is found at 11.6 years (it cannot be shifted to 11 years by invoking dating errors) and (2) the band-passed record mentioned above displays considerable amplitudes in the periods of minimum sunspot activity, the so-called Maunder and Sporer minima from A.D. 1640 to 1710 and from A.D. 1455 to 1545, respectively [*Eddy*, 1976, 1977]. In fact, the band-passed signal is dominated by the longer 11.9-year periodicity during these periods of low sunspot activity, which points to the solar tides as a possible cause for the 11.9-year component of the 11.6-year peak. The other component (11.2 years) of the peak could possibly be related to the sunspot cycle. Evidence for the influence of solar variability on the GISP2 δ record has also been discussed [*Stuiver et al.*, 1995]. They find, contrary to our results, correlation with the 11-year sunspot cycle over most of the past 250 years, but the correlation with the longer-term solar variability (60 to 550 years) as implied by the $\Delta^{14}C$ record is less clear.

If neither the sunspot record nor the $\Delta^{14}C$ record [*Stuiver et al.*, 1995] correlates significantly with the central Greenland δ records, we have to seek other explanations for the quasi-periodic 61.6-year component. We could appeal to internal oscillations in the atmosphere/ocean system, or the North Atlantic Oscillation [*Barlow et al.*, 1993], as found in a fully coupled ocean/atmosphere circulation model that shows high variability of the thermohaline circulation at periodicities between 40 and 100 years [*Delworth et al.*, 1993]. Other explanations include volcanic activity, which has been shown to correlate with Greenland δ records, both for long-term [*Hammer*, 1980] and short-term [*Stuiver et al.*, 1995] changes.

2.7. The Long-Term δ Profile

The continuous δ record along the upper 2900 m of the GRIP core is plotted in Figure 8 on a calculated timescale, the so-called ss08 chronology [*Dansgaard et al.*, 1993], with a 100-year resolution over most of the record. The correlation to the marine isotope stages (MIS) is marked to the right.

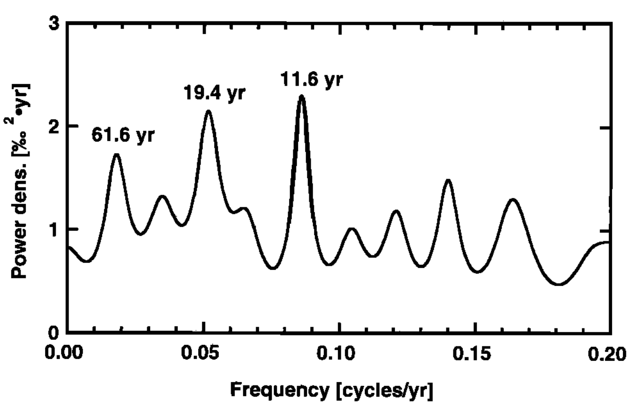

Figure 7. The MEM power spectrum (AR = 30) for the unfiltered 917-year-long Summit annual δ record shown in Figure 6.

The timescale was derived by identifying and counting annual layers downward from the surface back to 14,500 years B.P. and further back in time by calculations based on ice flow modeling [*Dansgaard et al.*, 1993; *Johnsen and Dansgaard*, 1992]. Two nonobservable parameters of the model were chosen to assign well-established ages to two characteristic features in the δ record: (1) 11,500 years for the end of the Younger Dryas event [*Bard et al.*, 1993; *Johnsen et al.*, 1992a] and (2) 110,000 years for the marine isotope stage 5d [*Martinson et al.*, 1987], shown by the arrows marked a and b in Figure 8. A new chronology for the GRIP core, based on counting annual layers in a continuous dust profile, has recently been established.

Apart from the δ minimum at 8210 years B.P. (1334-m depth), the Holocene record in Figure 8 depicts a remarkably stable δ signal at Summit. Modeling of the GRIP temperature profile with elevation changes due to varying accumulation rates and shifts of the ice margins suggests, however, a cooling trend in Greenland since 8 kyr B.P. of some 3°C [*Cuffey and Clow*, this issue; *Johnsen et al.*, 1995a]. This is in good agreement with melt feature observations on cores drilled in the Agassiz Ice Cap, Canada [*Fisher et al.*, 1995]. In contrast, the older part of the record shown in Figure 8 is dominated by large and abrupt δ shifts, which suggest that climatic stability in Greenland is rare. The warm glacial interstadials, the so-called Dansgaard-Oeschger cycles, have also been observed in ice cores from northwest, south, and east Greenland [*Dansgaard et al.*, 1982, 1971; *Johnsen et al.*, 1992b, 1972], as shown in Figure 9 and to be discussed in the next section, in the nearby GISP2 core [*Grootes et al.*, 1993], and in North Atlantic sediment cores [*Bond et al.*, 1993; *Fronval et al.*, 1995; *McManus et al.*, 1994].

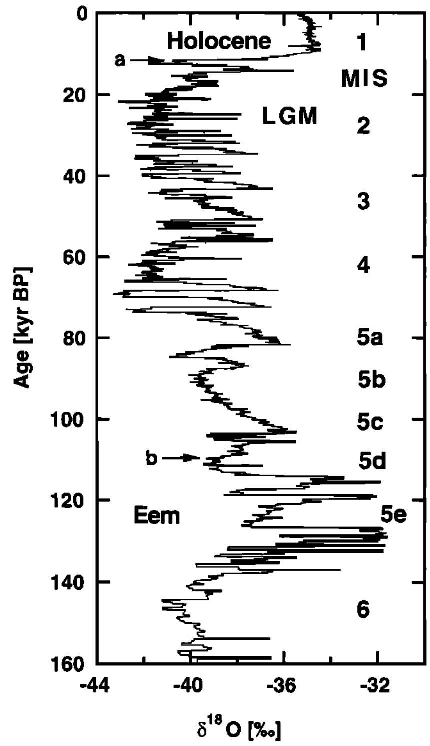

Figure 8. The GRIP $\delta^{18}O$ profile plotted on a timescale, which is established by counting annual layers back to 14,500 years B.P. and beyond that by ice flow modeling using the points a and b as fix points for the timescale. The marine isotope stages (MIS) are indicated on the figure together with the last glacial maximum (LGM).

2.7.1. Comparison with other Greenland δ profiles. Continuous $\delta^{18}O$ profiles along sections of four Greenland ice cores (located in Figure 1) from Camp Century (northwest), Dye 3 (southeast), Summit (central), and Renland (east), spanning nearly the same time interval including the last glacial period, are shown in Figure 9. The GISP2 profile, which is almost identical with the Summit profile except for the lowest 100 m shown, is not included in this comparison. The four records are all plotted on linear depth scales and extend down to bedrock except for the GRIP record. The silty ice layers sitting immediately above bedrock are indicated for both Camp Century and Dye 3 profiles.

Some of the warm interstadials (IS), as defined from the GRIP δ record [*Dansgaard et al.*, 1993], are shown by the numbers close to the δ profiles and serve to highlight the correlation among all four records. Judging from the correlation with the GRIP record, the Dye 3 record seems to deteriorate below IS 8 even though a correlation with the GRIP record can be reestablished by the Z2 volcanic ash layer [*Grönvold et al.*, 1995; *Ram and Gayley*, 1991] shown by arrows marked Z2 in the Figure 9. The Camp Century record shows the highest and the Renland record shows the lowest degree of glacial instability, suggesting different source regions for the precipitation falling at these two sites. A similar conclusion would also hold for the Summit site since the GRIP record shows less variability during IS 21 and 23 than is found in the Camp Century counterparts.

The Renland record [*Hansson*, 1994; *Johnsen et al.*, 1992b], which spans only 20 m of ice, shows a surprisingly accurate correlation with the more than 1200-m-long GRIP record when the 4-m low δ interval above IS 19 is excluded. The low amplitude and blocky shape of the Renland interstadials are still a matter of debate [*Johnsen et al.*, 1992b].

The high δ sections of the profiles situated below IS 23 in Figure 9 are assumed to represent parts of the last interglacial period, Eem. Evidently, no consensus is found in the data about the shape of the Eemian. The Dye 3 profile is clearly disturbed and can be ruled out in this discussion. The Camp Century and the Renland profiles could represent the latest stages of the Eemian period provided the ice at these sites was melted away in earlier stages of the Eemian. This is consistent with heavy melt features and a associated high dust content observed in the lowest meter of the Renland core pointing to a rebuilding stage of the Renland Ice Cap. The relatively great thickness of the warm Eemian ice in the Camp Century and the Renland ice cores seems to preclude any correlation with the GRIP Eemian record. If we assume, however, that the Camp Century and Renland Eemian ice was deposited during a buildup phase with low initial strain rates for thinning the Eemian layers, a correlation with the last warm phase (5e1) of the GRIP Eemian seems possible. The occurrence of a very cold and sharp δ spike in the Camp Century Eemian ice as well as in the GRIP 5e1 ice shown in Figure 9 and discussed in section 3.3 provides further support for this correlation.

2.7.2. Correlation with Antarctic profiles. Pronounced δ variability on the timescales of the Dansgaard-Oeschger cycles is also found in the Antarctic δ records of the Byrd core [*Johnsen et al.*, 1972; *Sowers and Bender*, 1995] and the Vostok core [*Jouzel et al.*, 1992; *Yiou et al.*, this issue (b)]. This emphasizes the global nature of these cycles [*Bender et al.*, 1994]. The

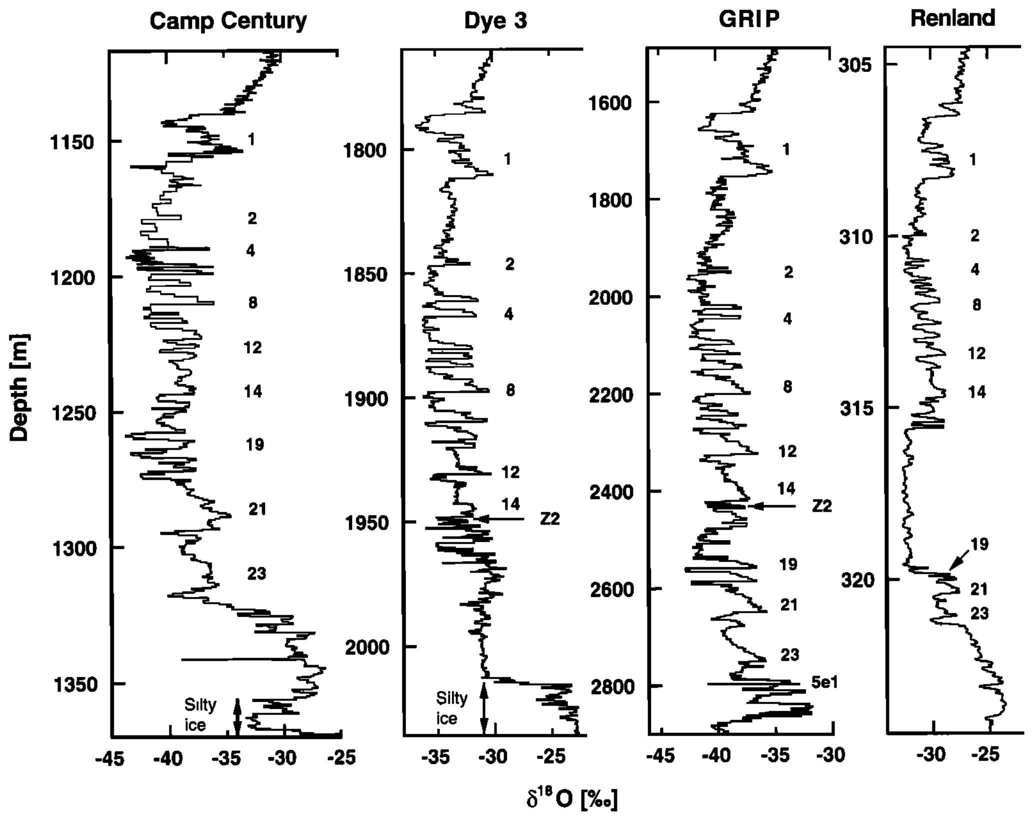

Figure 9. Continuous $\delta^{18}O$ profiles along sections of four Greenland ice cores (located in Figure 1) from Camp Century (northwest), Dye 3 (southeast), Summit (central), and Renland (east), spanning nearly the same time interval including the last glacial period. The GISP2 profile, which is almost identical with the Summit profile except for the lowest 100 m shown, is not included. The four records are all plotted on linear depth scales and extend down to bedrock except for the GRIP record. The silty ice layers are indicated for both the Camp Century and the Dye 3 core. Some of the warm interstadials (IS), as defined from the GRIP δ record, are shown by the numbers close to the δ profiles and serve to highlight the correlation between all four records.

exact phase between these cycles is still being debated; correlations using deep ocean sedimentary cores indicate that the south is leading the north by some 1500 years [*Charles et al.*, 1996]. The Antarctic counterpart of the Younger Dryas-Allerød-Bølling oscillation found in Greenland seems, however, to be of opposite phase based on a recent study of the global Younger Dryas methane drop in the GRIP and the Byrd ice cores [*Blunier et al.*, 1997].

The immediate cause of the Dansgaard-Oeschger cycles has been assigned to circulation changes in the North Atlantic [*Broecker et al.*, 1985; *Dansgaard et al.*, 1984; *Oeschger et al.*, 1984], which are controlled by the delicate salinity balance in North Atlantic Ocean. The salinity is modified by variable freshwater fluxes from pecipation, icebergs, surging glaciers, and glacial lakes, which are subsequently manifested in the variable strength of the North Atlantic heat conveyor [*Bond and Lotti*, 1995; *Broecker et al.*, 1992]. The immediate effect of these North Atlantic circulation changes on the ocean circulation in the southern oceans is not fully understood, but some clues can certainly be found by careful correlation of Greenland and Antarctic deep ice-core records.

2.7.3. The last interglacial. Series of δ values higher than the Holocene mean value of $-35‰$ occur between 115 and 130 kyr B.P. in the record of Figure 8. This clearly represents a period with temperatures exceeding those of today, and we interpret this core increment as being deposited during the last interglaciation, or Eem, which is usually recorded in other climate proxies as warmer than the Holocene (e.g., European pollen records [*Zagwijn*, 1996]) and warmer than the other most recent interglaciations in ocean sediment records [*Kellogg*, 1980].

The most spectacular feature of the δ record is the continuation of abrupt δ shifts within the Eem Interglaciation (Figures 8 and 10), which seems dominated by three different climate stades: Periods of high δ, corresponding to Greenland temperatures 4°–5°C higher than today, shift abruptly to periods of Holocene-like δ or to periods of interstadial-type δ corresponding to Greenland temperatures 5°–6°C colder than now. The longest of the low δ intervals seems to have lasted approximately 4000 years. These shifts in the GRIP Eemian have no close parallel in climate records published earlier, in other Greenland ice cores (including the GISP2 deep core), in midlatitude North Atlantic sediments [*Cortijo et al.*, 1994], or in western and central European pollen records [*Zagwijn*, 1996]. The latter two studies reveal no indications for climate instability in the Eemian period at midlatitudes. There is, however, some recently obtained evidence pointing to Eemian instability in records from the Nordic Seas sediments [*Cortijo et al.*, 1994; *Fronval and Jansen*, 1996; *Sejrup et al.*, 1995] and pointing to rapid changes in the ocean circulation and the heat flux.

The δ values for the cold Eemian stages are similar to the δ values found in the relatively warm interstadials 5a and 5c (Figure 8) that range from -36 to -38.5‰. During these interstadials the central European climate is found, as a contrast, to be quite stable and warm [*Guiot et al.*, 1989]. This suggests that the warm and stable Eemian climate in central Europe could coexist with relatively cold climate in Greenland as found during the colder stages of the GRIP Eemian, possibly due to eastward shift of the North Atlantic Current and reduced ocean heat flux to the Nordic Seas not affecting the central European climate [*Johnsen et al.*, 1995b].

In order to investigate other possible explanations the ice-core stratigraphy, as defined by several continuously observed parameters, has been scrutinized for disturbances, which may have invalidated the orderly layer sequence in the Eemian part of the GRIP core [*Alley et al.*, 1995; *GRIP Project Members*, 1993; *Johnsen et al.*, 1995b]. Below we will reevaluate the evidence already published on this topic and discuss the implications of the new evidence based on total gas, CH_4, $\delta^{18}O_{atm}$, and ^{10}Be measurements through the GRIP Eemian [*Chappellaz et al.*, this issue; *Fuchs and Leuenberger*, 1996; *Peel*, 1995].

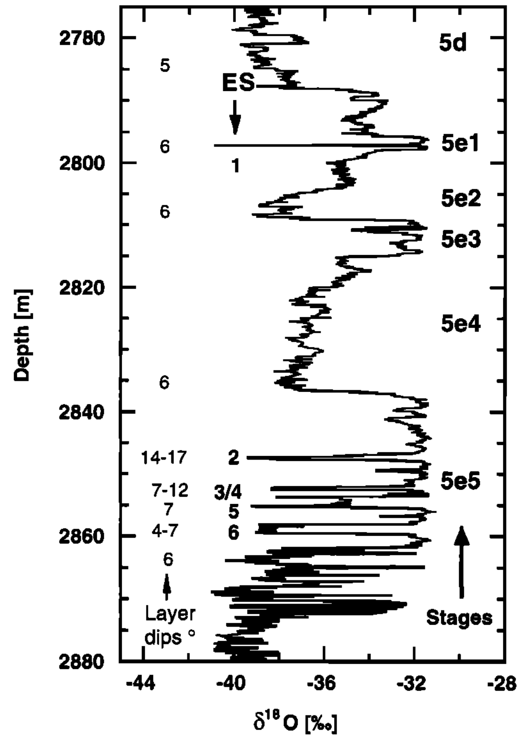

Figure 10. High-resolution δ profile spanning the Eemian ice. The numbers farthest to the left indicate the tilt (in degrees) of cloudy bands in low $\delta^{18}O$ ice. The numbers immediately to the left of the curve indicate the suggested numbering of the sharp Eemian $\delta^{18}O$ spikes (ES1 to ES6), and the numbers to the right of the curve indicate the Eemian isotopic substages.

3. The Eemian Problem

3.1. The GRIP Eemian

The GRIP Eemian sequence is best described by the detailed δ record in Figure 10. Five main stages have been identified and are labeled as 5el to 5e5 [*Dansgaard et al.*, 1993; *GRIP Project Members*, 1993]. Stages 5e2 and 5e4 represent the cool stages interrupting the Eemian warmth for relatively long periods. A total of six sharp and very cold spikes are also identified in the sequence and are here labeled ES1 through ES6 for convenience. The first and the sixth spikes have previously been discussed as event 1 and event 2 and were found to possess unique characteristics compared to other strata in the GRIP core [*GRIP Project Members*, 1993].

3.2. Gas Evidence

Samples from the atmosphere are continuously being trapped in glacier ice as air bubbles in the firn become closed off from overburden pressure. Several important characteristics of the Earth's past atmosphere can thus be studied in ice cores, including its composition and the isotopic ratios of its constituents [*Chappellaz et al.*, this issue].

The gas records can be used to check the integrity of the GRIP Eemian or assist in its interpretation. This is because the gas records should be of global nature due to the well-mixed atmosphere, given a sufficiently large residence time for the gas in the atmosphere. They should consequently match similar records through the Eemian sequence of the Vostok core, which is not believed to be affected by flow disturbances [*Bender et al.*, 1994]. The whole exercise relies on the assumption of the stability of the gases during and after their trapping in the ice, first as air bubbles and later as clathrates in a recrystallizing ice matrix loaded with impurities. Unfortunately, this stability requirement does not hold for the CO_2 gas trapped in Greenland glacier ice [*Anklin et al.*, 1995; *Barnola et al.*, 1995], which consequently cannot be used in this comparison.

The two new gas records, $\delta^{18}O_{atm}$ [*Fuchs and Leuenberger*, 1996] and CH_4 [*Chappellaz et al.*, this issue] have been compared with the Vostok gas records. Both studies conclude that the GRIP Eemian sequence is disturbed by foldings, intrusions, or both, due to the very poor correlation with the Vostok counterparts. The study by Chappellaz et al., which discusses both types of gas data, suggests that the two cold substages, 5e2 and 5e4, carry a MIS 5d or MIS 7 fingerprint and should therefore originate from layers originally sitting outside of the Eemian sequence. A 10-kyr-long part of MIS 6 strata seems also to be missing in the core. The CH_4 data for samples taken from the impure and cold ES2 spike further suggest disturbed stratigraphy, since its signature does not match that found in the isotopically warm ambient ice. This conclusion would, however, be challenged if the CH_4 concentrations and the $\delta^{18}O_{atm}$ values were somehow affected by the varying impurity content in the core, a possibility which seems to be contradicted by the otherwise close resemblance of the Vostok gas records with the independent GRIP and GISP2 counterparts [*Brook et al.*, 1996] from higher up in the cores where their respective δ profiles show a close match [*Grootes et al.*, 1993].

We go on to examine other available evidence pertinent to the question of folding and intrusions in the GRIP Eemian sequence in order to better understand the implications of such events taking place within an ice sheet.

3.3. Evidence for Intrusions

In geology, intrusions are mainly discussed in connection with intrusive igneous rock. In glaciology, we may not have an equivalent process, but if we for a moment assume that the sharp δ spikes in the Eemian sequence (Figure 10) are due to alien ice from outside the isotopically warm δ interval, then

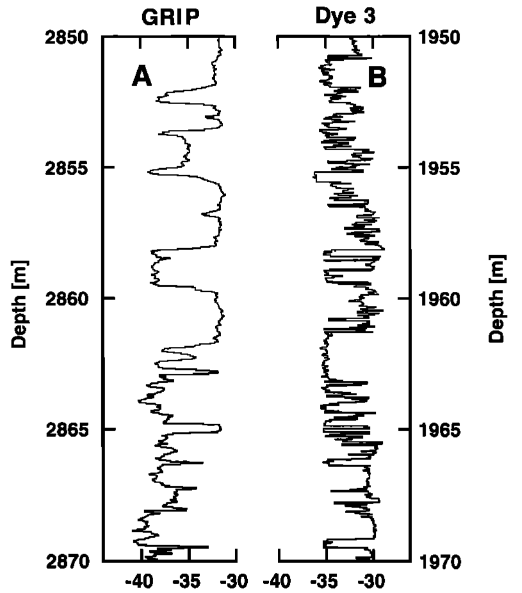

Figure 11. (a) The detailed $\delta^{18}O$ profile showing the onset of the GRIP Eemian. (b) A disturbed, highly detailed $\delta^{18}O$ profile from the Dye 3 core, which depicts extremely steep gradients and large shifts due to intrusions and folding. No similarly sharp features are seen in the GRIP profile.

they would be of intrusive origin. We thus propose to use the term intrusion for such narrow and apparent alien layers if they are truly out of stratigraphic order.

Similar layers of suspected intrusions or foldings have also been found in the Devon Island ice core [*Paterson et al.*, 1977] and are clearly seen in the highly disturbed part of the Dye 3 core (Figure 11b) discussed below.

Reordering and splitting up of a well-ordered sequence of ice layers will most likely take place in ice under intensive shear especially when the shear direction is different from that of the original layering. Any change in flow line directions due to growth or decay of the ice sheet would furthermore expedite such processes, particularly close to any bedrock undulations. An example of such restacking processes is found in the detailed δ profile from the lower strata of the Dye 3 core, where the otherwise perfect correlation with the GRIP δ profile (Figure 9) encountered at a distance higher than 120 m from bedrock has broken down. Part of the mechanically disturbed Dye 3 δ profile, probably originating from MIS 3 and MIS 4, is shown in Figure 11b, along with the detailed δ profile from the onset of the GRIP Eemian for comparison, where most of the cold spikes are located (Figure 11a). The sample increments are 2.5 cm for the Dye 3 δ profile and 3.44 cm for the GRIP δ profile. Both profiles shown represent a 20-m-long depth interval in the cores.

The highly disturbed Dye 3 profile (Figure 11b) displays numerous large and exceptionally sharp δ shifts that occur over core increments of less than 1 to 2 cm. The detailed dust record [*Hammer et al.*, 1985] from the core (not shown) also contains similarly sharp shifts. The effect of the restacking process in the 1950-m-deep Dye 3 ice is furthermore demonstrated by the separation of the Z2 ash layer (shown in Figure 9) into several bands of ash aggregates over more than 50-cm core increments in the Dye 3 core [*Ram and Gayley*, 1991], which is found as a 2.5-cm-thick dark and banded layer in the undisturbed part of GRIP core [*Grönvold et al.*, 1995].

We infer that any intrusions in the GRIP Eemian sequence should have similar characteristics. No such sharp shifts are seen, however, in the GRIP δ profile (Figure 11a). It could be argued that the cold Eemian spikes have been formed in a similar manner and that the initially sharp gradients have been smoothed out by diffusion over long time. This possibility was rejected [*Johnsen et al.*, 1995b], based on back diffusion, with a diffusion length of 3.5 cm, performed on the detailed δ profile in Figure 10. This conclusion seems no longer to hold, however, due to complications with the diffusion lengths in the Eemian strata, to be further discussed below.

One interesting feature, common to all the Eemian spikes, is the gradual lowering of the high δ values, starting at a distance of some 20 cm from the steepest parts of the δ spikes (Figures 10 and 11b). If diffusion is responsible for this feature, a diffusion length of some 6 to 7 cm, or twice that found by modeling the diffusion length using the single-crystal diffusion constant [*Ramseier*, 1967a], is needed. When the 10 times more effective CB diffusion, as observed in the clean Holocene ice, is included in the modeling, we end up with a diffusion length for the warm Eemian strata of about 6 cm. The ice in the coldest parts of the spikes contains a high level of impurities and we believe this inhibits crystal boundary diffusion, as observed in the similarly impure Younger Dryas strata. We thus expect to have a diffusion length of some 3.5 cm for the impure ice in the spikes. It should be noted that if the delta spikes at some stage had sharp edges, it would take only about 20 kyr for the diffusion to smooth them to the shape they have today. In other words, the present shape of the Eemian δ spikes cannot be used to determine their possible origin, except that they must have formed more than 20 kyr ago.

The detailed dust profile through the Eemian (C. U. Hammer, manuscript in preparation, 1997), which could not have been smoothed by diffusion, records all the Eemian spikes as high (2 to 6 mg/kg) levels of dust with very sharp gradients at the boundaries of the spikes. The dust evidence is thus consistent with the intrusion hypothesis, but it does not rule out possible dramatic changes in the Greenland climate as being their primary cause.

The Eemian spikes can also be characterized by their chemistry [*Steffensen et al.*, this issue], which finds his data difficult to reconcile with mechanical origin of these sharp events. Detailed measurements of ammonium, calcium [*Fuhrer et al.*, 1996], and the isotopes presented here show that for each of the spikes, the chemical fingerprints are unique and found nowhere else in the core. This implies that the intrusion hypothesis must be rejected unless (1) the originally undisturbed layers were totally absorbed into the Eemian sequence or (2) the chemical fingerprints of these layers have been drastically changed in the course of time.

The detailed δ and dust profiles for ES1 are shown in Figure 12. This spike stands out as being different from the others in several ways: (1) its δ values plunge down some 9‰ to a value of −41‰, which is lower than the lowest δ value found anywhere near the Eemian, (2) the dust concentration shows a double peak and is generally higher (6 mg/kg) and trails off gradually at the upper transition over a core length of 4 cm, while the other spikes show a lower dust concentration from 2 to 4 mg/kg with abrupt transitions, and (3) the dip and shape of the visible layering show no clear signs of stratigraphic disturbances, while the others are characterized by generally higher and variable dip and small open folds [*Alley et al.*, 1995; *Johnsen et al.*, 1995b], as indicated in Figure 10. These observations seem to support a nonmechanical origin of ES1 [*GRIP Project Members*, 1993], keeping in mind that a similarly deep δ spike has been found in the warm Eemian part of the Camp Century core [*Dansgaard et al.*, 1972]. It has been subjected to much less diffusion smoothing than ES1 and seems to have some features in common with the ES1 high-resolution dust profile: a double peak and a gradual recovery of the δ values at the upper transition. We thus consider it highly probable that this is a real event and suggest that it corresponds to the cold and sharp Melesey I event found in the Grande Pile pollen profile [*de Beaulieu and Reille*, 1992] and to the sharp IRD double spike found at 642-cm-depth in the DSDP 609 marine sedimentary core from the early part of MIS 5d [*McManus et al.*, 1994].

The δ gradients at the transitions of the longer-lasting cold substages 5e2 and 5e4, the supposedly alien parts of the GRIP Eemian sequence of possible MIS 5d or MIS 7b origin [*Chappellaz et al.*, this issue], are generally much smoother than the steep gradients in the δ spikes (Figure 10). Since the dust concentrations over the transitions are strictly anticorrelated with the smooth δ and show no evidence of abrupt breaks, as would be expected at the boundaries of intrusions, that mechanism has to be ruled out as being the cause of these cold strata of the Eemian sequence. If we still assume, as suggested by the study of CH_4 and $\delta^{18}O_{atm}$ [*Chappellaz et al.*, this issue], that the Eemian stages 5e2 and 5e4 have been introduced mechanically into the Eemian sequence, we will have to think of folding as the only other obvious possibility.

3.4. Evidence for Folding

Several mechanisms for folding in the lowest strata of ice sheets have been proposed, (1) ice flowing over and around bedrock obstacles, (2) changes in the directions of ice flow over undulating bedrock, and (3) ice divides migrating back and forth over bedrock that might even be flat and horizontal.

A core drilled through overfolded ice should generally encounter the same strata two to three times, and some irregularities in the stratification or layer dipping should be encountered. This has not been observed in the visible layering in stages 5d, 5e1, 5e2, and 5e4, which were all found to be gently dipping 5° to 6° (Figure 10) toward the same relative azimuth [*Johnsen et al.*, 1995b]. This should probably not be taken as strong evidence against possible folding since subsequent progressive layer thinning and stretching would eventually erase all differences in the dip, provided enough time has elapsed. In that case, the much higher and variable dips observed in ES2 through ES6 have to be caused by other processes like active boudinage [*Staffelbach et al.*, 1988]. A more direct way to detect folding would be to check for layers that repeat themselves in all measured constituents. This has been done for layers inside and close to the Eemian sequence, using detailed continuous measurements of δ, ECM, dust, nitrate, Ca^{++}, and NH_4^+. No evidence for repeated layers was found [*Johnsen et al.*, 1995b].

The stability of both the ^{10}Be concentrations [*Yiou et al.*, this issue (a)] and the total gas content [*Raynaud et al.*, this issue] through stages 5e2 and 5e4 of the Eemian sequence seem furthermore to speak against disturbed layering.

Since the core must be strictly continuous, due to the perfect fits observed during logging between the drilled core pieces, and furthermore by using the evidence presented above that rejects the intrusion hypothesis, we are not able to reconcile the apparently convincing evidence from the methane and $\delta^{18}O_{atm}$ measurements with the evidence based on other important ice-core data regarding the mechanical origin of the cold Eemian substages 5e2 and 5e4.

We hope to shed new light on this important issue when the recently started North Greenland Ice Core Project (NGRIP) deep drilling project, organized by the Department of Geophysics, University of Copenhagen, has recovered a new deep ice core in an area 320 km NNW of Summit, where conditions for uniform layering at great depths are believed to be much more favorable than at Summit [*Dahl-Jensen et al.*, 1997].

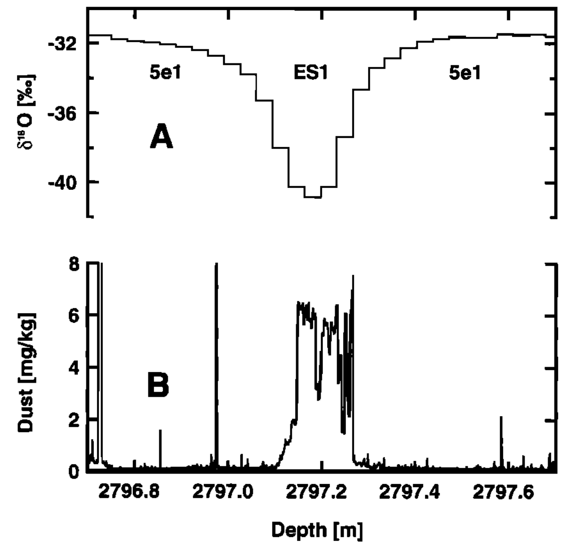

Figure 12. (a) The detailed $\delta^{18}O$ profile for ES1. The gradual lowering of the $\delta^{18}O$ values on both sides of the spike is due to enhanced diffusion in the high $\delta^{18}O$ Eemian ice. (b) The detailed dust profile for the same depth interval. The isolated peaks are most probably data errors. Note the sharp transitions in the dust concentrations, which are typical for all the Eemian $\delta^{18}O$ spikes except for this particular one, which shows a gradual trailing off at the upper transition, suggesting that the peak might be of climatic origin.

4. Summary and Conclusions

On the basis of the data and discussion presented above, we emphasize the following main observations:

1. The variability of the mean annual δ signal at Summit, for the past 917 years, is of the order of 1.05‰. Of the 1.1‰^{2} total variance, we estimate 0.6‰^{2} to be depositional noise and 0.5‰^{2} to be the real δ signal. Stacking of three independent δ records reduces the noise variance to 0.2‰^{2}, which allows much firmer conclusions to be drawn from the data.

2. Spectral analysis of the stacked 917-year-long δ records indicates two persistent periodicities at 19.8 and 11.9 years. The 19.8-year periodicity has also been found in a similar study of other stacked Greenland δ records [*Hibler and Johnsen*, 1979]. Both periods are also found in the tidal forces of the sun. No convincing evidence is found for the 11-year sunspot period over this time interval.

3. For the coldest parts of the last glacial period the short-term (<50 years) δ variability is ~3% or much higher than for present-day interglacial conditions. A significant part of the glacial short-term δ variability is believed to be of climatic

origin. For the warmest parts of the glacial interstadials, when δ exceeds $-39‰$, the variability is quite stable and close to $1.8‰$.

4. The smoothing of the detailed δ signal in the Holocene is much stronger than anticipated. The diffusion constant for this excess diffusion is about 1 order of magnitude higher than the single-crystal diffusion constant for a water molecule in ice at $-32°C$. We suggest that diffusion of loosely bound water molecules along the crystal boundaries (CB diffusion) in the recrystallizing ice matrix is responsible for the excess diffusion observed. In the more impure glacial-type ice the impurities seem to inhibit the excess diffusion process.

5. The long-term δ record for the GRIP core correlates well with other proxy climate records from ice cores, high-resolution ocean sediment cores, and terrestrial pollen records back to MIS 5d or 115 kyr BP. The 5e or Eemian sequence in the core is questionable, since it correlates neither with other records from ice cores nor with most published records from deep-sea sediment cores. The substages 5e1 to 5e5 seem, however, to correlate with ocean sediment data from the Nordic Seas [*Fronval and Jansen*, 1996], which suggest rapid changes in the ocean circulation and heat flux. We consider, however, based on correlation with the Camp Century core that the cold ES1 event is a real climatic event.

6. Several high-quality data sets are available from the GRIP ice core that can be used to investigate signs of layer disturbances in the GRIP Eemian sequence. By correlating the GRIP gas data (CH_4 and $\delta^{18}O_{atm}$) with corresponding data from the Vostok and the GISP2 cores [*Chappellaz et al.*, this issue] and assuming the gases to be well mixed in the atmosphere and chemically stable in the ice, evidence for serious disturbances in the Eemian sequence is found [*Peel*, 1995]. The other data sets, including $\delta^{18}O$, dust, chemistry, and ECM, have been scrutinized for evidence in support of the above conclusion from the gas study. No direct support has been found in the data, which appear to contradict the notion of a GRIP Eemian disturbed by large-scale folding and intrusions.

Since we are not able to reconcile the evidence, pertinent to the question of a disturbed GRIP Eemian, we have to admit to serious flaws in our understanding related to the basic behavior of some of the ice-core parameters discussed above.

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References

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, J. Meese, and T. Thorsteinsson, Comparison of deep ice cores, *Nature*, *373*, 395–394, 1995.

Andersen, U., Spectralundessøgelse of $\delta^{18}O$ profiler fra Grønlandske iskerner, Cand. Sci. thesis, Univ. of Copenhagen, Copenhagen, 1993.

Anklin, M., J. M. Barnola, J. Schwander, B. Stauffer, and D. Raynaud, Processes affecting the CO^2 concentration measured in Greenland ice, *Tellus*, *Ser. B*, *47*, 461–470, 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.

Barlow, L. K., J. W. C. White, R. G. Barry, J. C. Rogers, and P. M. Grootes, The North Atlantic Oscillation signature in deuterium and deuterium excess signals in the Greenland Ice Sheet Project 2 ice core, 1840–1970, *Geophys. Res. Lett.*, *20*(24), 2901–2904, 1993.

Barnola, J. M., M. Anklin, J. Porcheron, D. Raynaud, J. Schwander, and B. Stauffer, CO^2 evolution during the last millenium as recorded by Antarctic and Greenland ice, *Tellus*, *Ser. B*, *47*, 264–272, 1995.

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

Blunier, T., J. Schwander, B. Stauffer, T. Stocker, A. Dällenbach, A. Indermühle, J. Tschumi, J. Chappellaz, D. Raynaud, and J.-M. Barnola, Timing of the Antarctic cold reversal and the atmospheric CO_2 increase with respect to the Younger Dryas event, *Geophys. Res. Lett.*, *24*, 2683–2686, 1997.

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

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

Broecker, W. S., D. M. Peteet, and D. Rind, Does the ocean-atmosphere system have more than one stable mode of operation?, *Nature*, *315*, 21–26, 1985.

Broecker, W., G. Bond, M. Klas, E. Clark, and J. McManus, Origin of the northern Atlantic's Heinrich events, *Clim. Dyn.*, *6*, 265–273, 1992.

Brook, E. J., T. Sowera, and J. Orchardo, Rapid variations in atmospheric methane concentration during the past 110,000 years, *Science*, *273*, 1087–1091, 1996.

Chappellaz, J., T. Blünier, 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. Blünier, and B. Malaizé, CH_4 and $\partial^{18}O$ 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 Greenland Ice Sheet Project 2 ice cores, *J. Geophys. Res.*, this issue.

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

Clausen, H. B., Group meeting on nitrate sources in Antarctica and Greenland, in *Ice Core Studies of Global Biogeochemical Cycles*, edited by R. J. Delmas, *NATO ASI Ser. I*, *30*, 247–248, 1995.

Clausen, H. B., C. U. Hammer, J. Christensen, C. S. Hvidberg, D. Dahl-Jensen, M. Legrand, and J. P. Steffensen, 1250 years of global volcanism as revealed by central Greenland ice cores, in *Ice Core Studies of Global Biogeochemical Cycles*, edited by R. J. Delmas, *NATO ASI Ser. I*, *30*, 175–194, 1995.

Clausen, H. B., C. U. Hammer, C. S. Hvidberg, D. Dahl-Jensen, J. P. Steffensen, J. Kipfstuhl, and M. Legrand, A comparison of the volcanic records over the past 4000 years from the Greenland Ice Core Project and Dye 3 Greenland Ice Cores, *J. Geophys. Res.*, this issue.

Cortijo, E., J. C. Duplessy, L. Labeyrie, H. Leclaire, J. Duprat, and T. C. E. van Weering, Eemian cooling in the Norwegian Sea and North Atlantic ocean preceding continental ice-sheet growth, *Nature*, *372*, 446–449, 1994.

Cuffey, K. M., and G. D. Clow, Temperature, accumulation, and ice sheet elevation in central Greenland through the last deglaciation deglacial transition, *J. Geophys. Res.*, this issue.

Cuffey, K. M., R. B. Alley, P. M. Grootes, J. M. Bolzan, and

S. Anandakrishnan, Calibration of the $\delta^{18}O$ isotopic paleothermometer for central Greenland, using borehole temperatures, *J. Glaciol.*, *40*, 341–349, 1994.

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.

Dahl-Jensen, D., and S. J. Johnsen, Palaeotemperatures still exist in the Greenland ice sheet, *Nature*, *320*, 250–252, 1986.

Dahl-Jensen, D., S. J. Johnsen, C. U. Hammer, H. B. Clausen, and J. Jouzel, Past accumulation rates derived from observed annual layers in the GRIP ice core from Summit, Central Greenland, in *Ice in the Climate System*, *NATO ASI Ser. 1*, *12*, 517–532, 1993.

Dahl-Jensen, D., N. S. Gundestrup, K. R. Keller, S. J. Johnsen, S. P. Gogineni, H. Miller, J. Kipfstuhl, and E. D. Waddington, A search in north Greenland for a drill site, *J. Glaciol.*, in press, 1997.

Dansgaard, W., Stable isotopes in precipitation, *Tellus*, *16*, 436–468, 1964.

Dansgaard, W., S. J. Johnsen, H. B. Clausen, and C. C. Langway Jr., Climatic record revealed by the Camp Century ice core, in *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 C. C. Langway Jr., Speculations about the next glaciation, *Quat. Res.*, *2*, 396–398, 1972.

Dansgaard, W., H. B. Clausen, N. Gundestrup, C. U. Hammer, S. J. Johnsen, P. M. Kristinsdóttir, and N. Reeh, A new Greenland deep ice core, *Science*, *218*, 1273–1277, 1982.

Dansgaard, W., S. J. Johnsen, H. B. Clausen, D. Dahl-Jensen, N. Gundestrup, C. U. Hammer, and H. Oeschger, North Atlantic climatic oscillations revealed by deep Greenland ice cores, in *Climatic Processes and Climate Sensitivity*, *Geophys. Monogr. Ser.*, Vol. *29*, edited by J. E. Hansen and T. Takahashi, pp. 288–298, AGU, Washington, D. C., 1984.

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 last climatic cycle at La Grande Pile (Vosges, France): A new pollen profile, *Quat. Sci. Rev.*, *11*, 431–438, 1992.

Delworth, T., S. Manabe, and R. J. Stouffer, Interdecadal variations of the thermohaline circulation in a coupled ocean-atmosphere model, *J. Clim.*, *6*(11), 1993–2011, 1993.

Ditlevsen, P. D., H. Svensmark, and S. J. Johnsen, Contrasting atmospheric and climate dynamics of the last-glacial and Holocene periods, *Nature*, *379*, 810–812, 1996.

Eddy, J. A., The Maunder Minimum, *Science*, *192*, 1189–1202, 1976.

Eddy, J. A., Climate and the changing Sun, *Clim. Change*, *1*, 173–190, 1977.

Fisher, D. A., Enhanced flow of Wisconsin ice related to solid conductivity through strain history and recrystallization, in *The Physical Basis of Ice Sheet Modelling* (*Proceedings of the Vancouver symposium, Canada, August 1987*), edited by E. Waddington, *IAHS Publ.*, 170, 45–51, 1987.

Fisher, D. A., and R. M. Koerner, Signal and noise in four ice-core records from the Agassiz Ice Cap, Ellesmere Island, Canada: Details of the last millennium for stable isotopes, melt and solid conductivity, *Holocene*, *4*(2), 113–120, 1994.

Fisher, D. A., N. Reeh, and H. B. Clausen, Stratigraphic noise in time series derived from ice cores, *Ann. Glaciol.*, *7*, 76–83, 1985.

Fisher, D. A., R. M. Koerner, and N. Reeh, Holocene climatic records from Agassiz Ice Cap, Ellesmere Island, NWT, Canada, *Holocene*, *5*(1), 19–24, 1995.

Fronval, T., and E. Jansen, Rapid changes in ocean circulation and heat flux in the Nordic seas during the last interglacial period, *Nature*, *383*, 806–810, 1996.

Fronval, T., E. Jansen, J. Bloemendal, and S. J. Johnsen, Oceanic evidence for coherent fluctuations in Fennoscandian and Laurentide ice sheets on millennium timescales, *Nature*, *374*, 443–446, 1995.

Fuchs, A., and M. Leuenberger, $\delta^{18}O$ of atmospheric oxygen measured on GRIP ice core shows unexpected stratigraphic features well above bedrock, *Geophys. Res. Lett.*, *23*(9), 1049–1052, 1996.

Fuhrer, K., A. Neftel, M. Anklin, and V. Maggi, Continuous measurements of hydrogen peroxide, formaldehyde, calcium and ammonium concentrations along the new GRIP ice core from Summit, Central Greenland, *Atmos. Environ.*, *Part A*, *27*, 1873–1880, 1993.

Fuhrer, K., A. Neftel, M. Anklin, T. Staffelbach, and M. Legrand, High-resolution ammonium ice core record covering a complete glacial-interglacial cycle, *J. Geophys. Res.*, *101*(D2), 4147–4164, 1996.

Greenland Ice Core Project (GRIP) Project Members, Climate instability during the last interglacial period recorded in the GRIP ice core, *Nature*, *364*, 203–208, 1993.

Grönvold, K., N. Óskarsson, S. J. Johnsen, H. Clausen, B., 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. J. Johnsen, and J. Jouzel, Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores, *Nature*, *366*, 552–554, 1993.

Guiot, J., A. Pons, J. L. de Beaulieu, and M. Reille, A 140,000-year continental climate reconstruction from two European pollen records, *Nature*, *338*, 309–313, 1989.

Hammer, C. U., Dust studies on Greenland ice cores, in *Proceedings of Symposium on Isotopes and Impurities in Snow and Ice*, *Int. Assoc. of Hydrol. Sci.*, *Commission of Snow and Ice*, *IAHS-AISH Publ.*, *118*, 365–370, 1977.

Hammer, C. U., Acidity of polar ice cores in relation to absolute dating, past volcanism, and radio-echoes, *J. Glaciol.*, *25*, 359–372, 1980.

Hammer, C. U., H. B. Clausen, W. Dansgaard, N. Gundestrup, S. J. Johnsen, and N. Reeh, Dating of Greenland ice cores by flow models, isotopes, volcanic debris, and continental dust, *J. Glaciol.*, *20*, 3–26, 1978.

Hammer, C. U., H. B. Clausen, and W. Dansgaard, Greenland ice sheet evidence of post-glacial volcanism and its climatic impact, *Nature*, *288*, 230–235, 1980.

Hammer, C. U., H. B. Clausen, W. Dansgaard, A. Neftel, P. Kristinsdóttir, and E. Johnson, Continuous impurity analysis along the Dye-3 deep core, in *Greenland Ice Cores: Geophysics, Geochemistry and Environment*, *Geophys. Monogr. Ser.*, vol. 33, edited by C. C. Langway Jr., H. Oeshger, and W. Dansgaard, pp. 90–94, AGU, Washington, D. C., 1985.

Hansson, M. E., The Renland ice core: A northern hemisphere record of aerosol composition over 120,000 years, *Tellus*, *Ser. B*, *46*, 390–418, 1994.

Hibler, W. D., and S. J. Johnsen, The 20-yr cycle in Greenland ice core records, *Nature*, *280*, 481–483, 1979.

Johnsen, S. J., Stable isotope homogenization of polar firn and ice, in *Proceedings of Symposium on Isotopes and Impurities in Snow and Ice*, *IAHS-AISH Publ.*, *118*, 210–219, 1977a.

Johnsen, S. J., Stable isotope profiles compared with temperature profiles in firn and with historical temperature records, in *Proceedings of Symposium on Isotopes and Impurities in Snow and Ice*, *IAHS–AISH Publ.*, *118*, 388–392, 1977b.

Johnsen, S. J., and N. Andersen, On power estimation in maximum entropy spectral analysis, *Geophysics*, *43*, 681–690, 1978.

Johnsen, S. J., and W. Dansgaard, On flow model dating of stable isotope records from Greenland ice cores, in *The Last Deglaciation: Absolute and Radiocarbon Chronologies*, edited by E. Bard and W. S. Broecker, *NATO ASI Ser.*, *Ser. I*, *2*, 13–24, 1992.

Johnsen, S. J., W. Dansgaard, H. B. Clausen, and C. C. Langway Jr, Oxygen isotope profiles through the Antarctic and Greenland ice sheets, *Nature*, *235*, 429–434, 1972. (Errata, *Nature*, *236*, 249, 1972.)

Johnsen, S. J., W. Dansgaard, and J. W. C. White, The origin of Arctic precipitation under present and glacial conditions, *Tellus*, *Ser. B*, *41*, 452–468, 1989.

Johnsen, S. J., H. B. Clausen, W. Dansgaard, K. Fuhrer, N. Gundestrup, C. U. Hammer, P. Iversen, J. P. Steffensen, J. Jouzel, and B. Stauffer, 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. Hanson, P. Jonson, J. P. Steffensen, and A. E. Sveinbjörnsdóttir, A "deep" ice core from East Greenland, *Medd. Grøn. Geosci.*, *29*, 1–22, 1992b.

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

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, 1995b.

Joussaume, S., J. Jouzel, and R. Sadourny, A general circulation model

of water isotope cycles in the atmosphere, *Nature*, *311*(5981), 24–29, 1984.
Jouzel, J., G. L. Russell, R. J. Suozzo, R. D. Koster, J. W. C. White, and W. S. Broecker, Simulations of the HDO and H_2 ^{18}O atmospheric cycles using the NASA GISS general circulation model: The seasonal cycle for present-day conditions, *J. Geophys. Res.*, *92*(D12), 14,739–14,760, 1987.
Jouzel, J., et al., The last deglaciation in Antarctica: Further evidence of a "Younger Dryas" type climatic event, in *The Last Deglaciation: Absolute and Radiocarbon Chronologies*, edited by E. Bard and W. S. Broecker, *NATO ASI Ser., Ser. I, 2*, 229–266, 1992.
Jouzel, J., et al., On the validity of the temperature reconstruction from water isotopes in ice cores, *J. Geophys. Res.*, this issue.
Kapsner, W. R., R. B. Alley, C. A. Shuman, S. Anandakrishnan, and P. M. Grootes, Dominant influence of atmospheric circulation on snow accumulation in Greenland over the past 18,000 years, *Nature*, *373*, 52–54, 1995.
Kellogg, T. B., Paleoclimatology and paleo-oceanography of the Norwegian and Greenland seas: Glacial-interglacial contrasts, *Boreas*, *9*, 115–137, 1980.
Lacoss, R. T., Data adaptive spectral analysis methods, *Geophysics*, *36*, 661–675, 1971.
Loewe, F., The Greenland ice cap as seen by a meteorologist, *Q. J. R. Metorol. Soc.*, *62*, 359–377, 1936.
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. C. Morrison, M. S. Twickler, S. I. Whitlow, K. K. Ferland, D. A. Meese, M. Legrand, and J. P. Steffensen, Greenland ice core "signal" characteristics: An expanded view of climate change, *J. Geophys. Res.*, *98*(D7), 12,839–12,847, 1993.
McManus, J. F., G. C. Bond, W. S. Broecker, S. J. Johnsen, L. Labeyrie, and S. Higgins, High-resolution climate records from the North Atlantic during the last interglacial, *Nature*, *371*, 326–329, 1994.
Oeschger, H., J. Beer, U. Siegenthaler, B. Stauffer, W. Dansgaard, and C. C. Langway Jr., Late glacial climate history from ice cores, in *Climate Processes and Climate Sensitivity*, *Geophys. Monogr. Ser.*, vol. *29*, edited by J. E. Hansen and T. Takahashi, pp. 299–306, AGU, Washington, D. C., 1984.
Paterson, W. S. B., R. M. Koerner, D. Fisher, S. J. Johnsen, H. B. Clausen, W. Dansgaard, P. Bucher, and H. Oeschger, An oxygen-isotope climatic record from the Devon Island ice cap, arctic Canada, *Nature*, *266*, 508–511, 1977.
Peel, D. A., Profiles of the past, *Nature*, *378*, 234–235, 1995.
Ram, M., and R. I. Gayley, Long-range transport of volcanic ash to the Greenland ice sheet, *Nature*, *349*, 401–404, 1991.
Ramseier, R. O., Self-diffusion of tritium in natural and synthetic ice monocrystals, *J. Appl. Phys.*, *38*, 2553–2556, 1967a.
Ramseier, R. O., Self-diffusion in ice monocrystals, *CRREL Res. Rep. 232*, Cold Reg. Res. and Eng. Lab., Hanover, N. H., 1967b.
Raynaud, D., J. Jouzel, J. M. Barnola, J. Chappellaz, R. J. Delmas, and C. Lorius, The ice record of greenhouse gases, *Science*, *259*, 926–934, 1993.
Raynaud, D., J. Chappellaz, C. Ritz, and P. Martinerie, Air content along the Greenland Ice Core Project core and past surface elevations in central Greenland, *J. Geophys. Res.*, this issue.
Reeh, N., H. B. Clausen, W. Dansgaard, N. Gundestrup, C. U. Hammer, and S. J. Johnsen, Secular trends of accumulation rates at three Greenland stations, *J. Glaciol.*, *20*, 27–30, 1978.
Sejrup, H. P., H. Haflidason, D. Klintgaard Kristensen, and S. J. Johnsen, Last interglacial and Holocene climatic development in the Norwegian Sea region: Oceanic front movements and ice core data, *J. Quat. Sci.*, *10*(4), 385–390, 1995.
Sowers, T., and M. Bender, Climate records covering the last deglaciation, *Science*, *269*, 210–214, 1995.
Staffelbach, T., B. Stauffer, and H. Oeschger, A detailed analysis of the rapid changes in ice-core parameters during the last ice age, *Ann. Glaciol.*, *10*, 167–170, 1988.
Steffensen, J. P., The size distribution of microparticles from selected segments of the Greenland Ice Core Project ice core representing different climatic periods, *J. Geophys. Res.*, this issue.
Steffensen, J. P., H. B. Clausen, C. U. Hammer, M. Legrand, and M. De Angelis, The chemical composition of cold events within the Eemian section of the Greenland Ice Core Project core from Summit, Greenland, *J. Geophys. Res.*, this issue.
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.
Thorsteinsson, T., J. Kipfstuhl, and H. Miller, Textures and fabrics in the GRIP ice core, *J. Geophys. Res.*, this issue.
Ulrych, T. J., and T. N. Bishop, Maximum entropy spectral analysis and autoregressive decomposition, *Rev. Geophys.*, *13*, 183–200, 1975.
White, J. W. C., L. K. Barlow, D. A. Fisher, P. M. Grootes, J. Jouzel, S. J. Johnsen, M. Stuiver, and H. B. Clausen, The climate signal in the stable isotopes of Summit, Greenland Snow: Results of comparison with modern climate observations, *J. Geophys. Res.*, this issue.
Yiou, F., et al., Beryllium 10 in the Greenland Ice Core Project ice core at Summit, Greenland, *J. Geophys. Res.*, this issue (a).
Yiou, P., K. Fuhrer, L. D. Meeker, J. Jouzel, S. Johnsen, and P. A. Mayewski, Paleoclimatic variability inferred from the spectral analysis of Greenland and Antarctic ice-core data, *J. Geophys. Res.*, this issue (b).
Zagwijn, W. H., An analysis of Eemian climate in western and central Europe, *Quat. Sci. Rev.*, *15*, 451–469, 1996.

K. K. Andersen, U. Andersen, H. B. Clausen, D. Dahl-Jensen, W. Dansgaard, N. S. Gundestrup, C. U. Hammer, C. S. Hvidberg, S. J. Johnsen, and J. P. Steffensen, Niels Bohr Institute of Astronomy, Physics and Geophysics, Department of Geophysics, University of Copenhagen, Juliane Maries Vej 30, DK-2100, Copenhagen, Denmark. (e-mail: kku@gfy.ku.dk; ua@gfy.ku.dk; hbc@gfy.ku.dk; ddj@gfy.ku.dk; ng@gfy.ku.dk; cuh@gfy.ku.dk; ch@gfy.ku.dk; sigfus@gfy.ku.dk; jps@gfy.ku.dk)

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