Deuterium excess record from central Greenland over the last millennium: Hints of a North Atlantic signal during the Little Ice Age

Georg Hoffmann, Jean Jouzel, Sigfus Johnsen

JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 106, NO. D13, PAGES 14,265–14,274, JULY 16, 2001

Deuterium excess record from central Greenland over the last millennium: Hints of a North Atlantic signal during the Little Ice Age

Georg Hoffmann and Jean Jouzel

Laboratoire des Sciences du Climat et de l'Environnement, Unité Mixte de la Recherche CEA/CNRS, Gif-sur-Yvette, France

Sigfus Johnsen

Science Institute, University of Iceland, Reykjavik

Abstract. A stacked water isotope record for Summit, central Greenland, has been established covering approximately the last 900 years. Measurements of $\delta^{18}O$ and δD in the GRIP deep ice core and the 230-m core, S93, allow the reconstruction of a millennial record of the deuterium excess in a near-annual resolution. A short period of particularly high values of the Deuterium Excess at the beginning of the fourteenth century may be associated with the medieval warm period (MWP). The Little Ice Age (LIA) might be represented by a 100-year period of very low excess values in the sixteenth and seventeenth centuries. Using a simple isotope model, $\delta^{18}O$ and deuterium excess are interpreted in terms of surface temperature variations over central Greenland and over the subtropical North Atlantic, Greenland's principal vapor source region. An estimated cooling of –0.7°C of subtropical sea surface temperatures (SSTs) during the Little Ice Age and a warming of 0.6°C during the short warm period in the medieval is in agreement with previous studies. Over periods of about 100 years, an antiphasing between gradually decreasing $\delta^{18}O$ and increasing deuterium excess is observed. Interannual-to-decadal -scale variability associated to the North Atlantic Oscillation may be responsible for this anti-phase relationship. An alternative explication is a North-South oscillation in North Atlantic sea surface temperatures associated with short-term changes in the thermohaline circulation.

1. Introduction

The quantitative reconstruction of climate along the last few hundred years has become a focus of much ongoing research [*CLIVAR*, 1998]. Multiproxy data sets have been used to compare past climate variability with the global warming observed during the twentieth century [*Bradley and Jones*, 1993; *Mann et al.*, 1998; *Overpeck et al.*, 1997]. However, most of the high resolution records (i.e., at least decadal), such as tree ring widths or lake sediment records, originate from terrestrial sites and reflect climate conditions on land. Equivalent deep sea records are rare and on short timescales their exact dating and interpretation is not straightforward [*Keigwin*, 1996]. Some information from coral records is now available, too [*Boiseau et al.*, 1998; *Cole et al.*, 1993] and in the future will certainly become more important, although with a spatial coverage of coral archives basically restricted to the tropics.

As long as our knowledge of climate variability in the last millenium is mainly based on continental records, an important piece in the climate puzzle is missing. Whether internal variability of the ocean-atmosphere system or rather external factors such as solar luminosity or atmospheric aerosol concentration drive centennial scale climate, variability is difficult to determine without suitable marine information.

In the North Atlantic sector, two major climate variations in the last about thousand years are often discussed, the Little Ice Age (LIA) and the medieval warm period (MWP). Although uncertainties exist in both their exact timing and their spatial extent, these climatic episodes are seen as extremes in climate variability during the last 1000 years. Usually, the MWP is considered as a comparable warm period from approximately 900-1400 [*Hughes and Diaz*, 1994]. The LIA, originally associated with the observed advances of alpine glaciers, denotes cool conditions particularly during the seventeenth and nineteenth centuries. In western Europe, for instance, the more quantitative analysis of tree ring records show a transition to cooler summer conditions (~-1.5°C compared to modern conditions) at about 1550 lasting until 1750, a recovery of climate to former conditions in the second half of the eighteenth century and again a reversal to the former cooler state during the nineteenth century [*Bradley and Jones*, 1993].

For a long time these centennial temperature variations were not found in Greenland's ice cores. This is particularly surprising since the appearance of Viking settlements along Greenland's coast was often mentioned as suggesting milder conditions during the medieval. Consequently, the disappearance of this settlement at the beginning of the fiveteenth

Paper number 2000JD900585.
0148-0227/01/2000JD900585$09.00

century was interpreted as a hint for a change to colder conditions [*Dansgaard et al.*, 1975]. Solely, the isotope record of the deep drilling at Camp Century in the far northwest of Greenland [*Johnsen et al.*, 1970] documented local temperature variations in accordance with the described sequence of warmer and cooler periods during the last millennium. Recently, results from a transec of shallow ice cores drilled in northern Greenland confirmed these early findings. *Fischer et al.* [1998] found an isotope signal documenting the LIA climate variation with a continuously increasing amplitude when moving from central Greenland toward northwest Greenland. Furthermore, numerical analysis of borehole temperature measurements in the Greenland ice core project (GRIP) record give evidence for a cooling of −1.5°C during the sixteenth, seventeenth, and nineteenth centuries relative to the MWP even in central Greenland [*Dahl-Jensen et al.*, 1998]. This implies that there actually exists a LIA cooling also on the central Greenland ice sheet, but for some reason, it is not clearly recorded in most isotope records.

In this study we present the isotope records of two ice cores from Summit (37°37'W,72°34'N, 3232 m above sea level (a.s.l.) elevation), central Greenland. Because of the annual mean accumulation there, these records enable sufficiently high temporal resolution for studies of interdecadal variability over the last millennium. The isotopic composition ($\delta^{18}O$, δD) of polar ice cores is usually interpreted as reflecting regional temperatures over the ice sheet [*Dansgaard*, 1964; *Jouzel et al.*, 1987]. However, the second order parameter deuterium excess, defined by *Dansgaard* [1964] as $d=\delta D-8*\delta^{18}O$, offers an interesting possibility to work around the problem of missing quantitative information from the ocean. This second-order quantity depends less on condensation temperatures but reflects climate conditions at the sea surface when water vapor is formed [*Jouzel et al.*, 1982; *Vimeux et al.*, 1999]. During evaporation, kinetic nonequilibrium processes take place affecting the relationship between the two water isotopes ($H_2{}^{18}O$, HDO) and tagging the oceanic vapor with a deuterium excess signal, hereinafter referred to as *d*. At least in simple isotope models, designed for describing these fractionation processes, the deuterium excess in the final precipitation is a quite sensitive indicator of humidity and temperature at the sea surface.

The reason for this tagging mechanism by fractionation processes reads as thus: The colder and more humid the evaporative conditions are, the "slower" the evaporation occurs. Under such conditions the evaporation process is close to isotopic equilibrium, and the relation between the two isotopes (^{18}O, D) mainly reflects the ratio s of the equilibrium fractionation factors α ($s=\alpha^{18}O/\alpha D \sim 1/8$). Consequently, the influence of the kinetics is reduced and the deuterium excess of the corresponding water vapor, which reflects the deviation from the equilibrium slope *s*, is small. On the other hand, warm and dry conditions lead to a "faster" evaporation which is out of isotopic equilibrium and produces a larger deuterium excess [*Johnsen et al.*, 1989; *Merlivat and Jouzel*, 1979]. Also wind speed has a weak influence with d going down by app. 1% if wind speed doubles [*Merlivat and Jouzel*, 1979]. Assuming the distribution of Greenland's vapor source regions has not significantly changed over the last millennium and that the relative humidity at the ocean surface stayed approximately constant, a higher deuterium excess would then indicate warmer conditions at the ocean surface and, vice versa, a lower excess cooler ocean conditions.

A recent modelling study [*Armengaud et al.*, 1998] compared the results of such simple isotope models with an atmospheric general circulation model fitted with water isotope diagnostics. Both types of models showed a similar response of the deuterium excess in central Greenland on the seasonally varying conditions at the sea surface (temperature and relative humidity). On the basis of this outcome we interpret in this paper quantitatively the deuterium excess signal of two stacked ice cores from central Greenland using a simple isotopic rainout model.

2. Data

In Figure 1 we show the isotope records ($\delta^{18}O$ and *d*) of the shallow core S93 and the upper part of the GRIP deep ice core [*Dansgaard et al.*, 1993] together with their stacked records. The sample length of 55 cm of both cores corresponds to a temporal resolution of about 2-3 years. Stacking several ice cores helps to reduce postdepositional effects such as snow drifting or wind scouring of winter snow which impose a high noise level on the original signal [*Fisher et al.*, 1985; *White et al.*, 1997]. We find a signal to noise ratio between the two cores of 0.95 for the $\delta^{18}O$ and 0.3 for the deuterium excess signal. The lower correlation between both cores for *d* is mainly due to larger uncertainties as *d* results of two independent isotope measurements.

For further reducing the high noise level existing even in the stacked records (Figure 1) we calculated the 20-year running mean presented in Figure 2. For inferring a climate-related signal from Greenland's ice cores usually low-pass filtering from between 30 and 50 years are applied [*Fisher et al.*, 1996]. Comparing the two composite signals once again demonstrates that the deuterium excess is a quantity independent of the actual isotope signal. In agreement with what has been stated before this central Greenland $\delta^{18}O$ signal indicate s only slightly cooler conditions during the LIA. During the eighteenth and seventeenth centuries century the $\delta^{18}O$ is, at least periodically, about −0.4‰ lower than the long-term mean value. However, if we would apply the classical spatial relationship for Greenland between $\delta^{18}O$ and temperature of 0.69‰/°C on the maximal temperature amplitude reconstructed from bore hole temperatures (-1.5°C), a 1‰ decrease of the $\delta^{18}O$ signal should be expected. Even with a smaller gradient of 0.33‰/°C, as it was predicted for the last glacial-interglacial change [*Cuffey et al.*, 1995; *Dahl-Jensen et al.*, 1998], the observed response of the water isotopes remains weak and points to some peculiarities in the physics of the water isotopes that we will discuss later.

The spectral properties of both isotope series are analyzed using both the maximum entropy method (MEM) and the multi taper method (MTM). The latter allows a significance test of the found periodicities against a supposed red noise process (see Figure 3). Using, in this paper, just a subset of the isotope series available from the GRIP site (the Eurocore or S89 core $\delta^{18}O$ data used in [*Johnsen et al.*, 1998] are not considered here since the corresponding excess was not measured), we find nevertheless approximately the same frequencies in the stacked isotope series as discussed by *Johnsen et al.* [1998]: There is weak spectral power in our stacked $\delta^{18}O$ record between 50 and 60 (significant at the

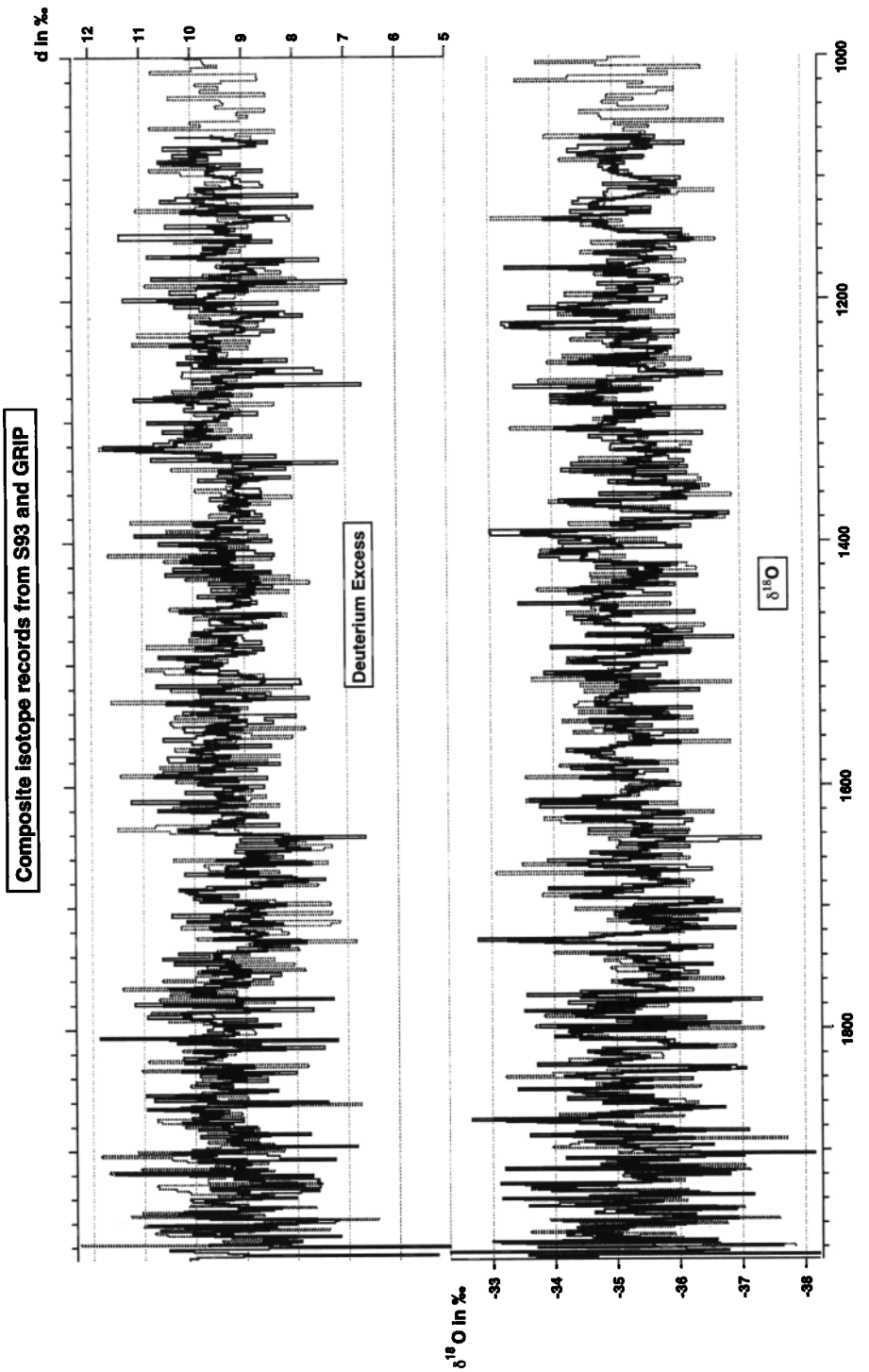

Figure 1. Isotope records of two ice cores from Summit (37°37'W, 72°34'N, 3232 /m a. s. l. elevation), central Greenland. Data are shown of the GRIP deep ice core (thin straight line), of the shallow core (230m) S93 (pointed line) together with their composite record (bold straight line). S93 was taken in 1993 at the Summit site roughly 30 meter from the location of the GRIP drilling site. The two cores were independently dated by counting the deconvoluted seasonal cycle of $\delta^{18}O$ and cross-checked with ECM signals of historically known volcanic eruptions. This guarantees an approximate precision of the age scale of 1 year. The 55cm bag samples were analyzed for both $\delta^{18}O$ and δD. The experimental precision for the deuterium excess is ± 0.8‰.

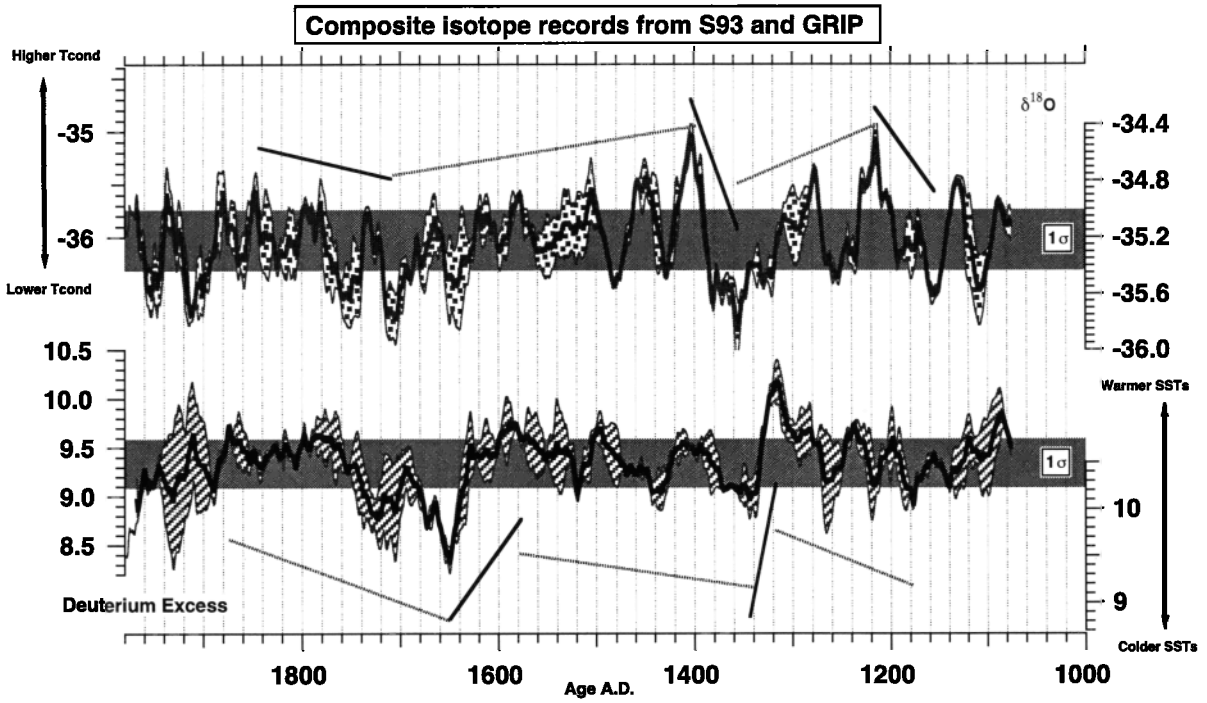

Figure 2. Twenty years running mean of the stacked isotope records ($\delta^{18}O$, top panel, and deuterium excess d, bottom panel) of the GRIP and the S93 ice cores. We show the single records as an envelope curve around the composite records in order to give an estimate of the measurement precision. Linear regressions were calculated over certain periods defined by, alternately, increasing maxima and decreasing minima. The regressions are shown in the figure slightly displaced relative to the original data (axes for the regressions are on the right for the excess and on the left for the $\delta^{18}O$).

90% level only), 19 and 20 (99%), 11 and 12 (99%), and between 8 and 9 years (99%). Some of these frequencies (19 and 11 years) have also been found analyzing ^{10}Be in the GRIP and ^{14}C in the Greenland Ice sheet program (GISP) core [*Stuiver et al.*, 1995], respectively. Since the distribution of these radioisotopes is primarily controlled by solar activity, both $\delta^{18}O$ peaks were tentatively associated to variations of solar insulation. In the low-frequency range, periodicities between 50 to 60 years, however, might rather be due to internal oscillations of the coupled ocean/atmosphere/sea ice climate system. Analyzing the deuterium excess signal in the same manner reveals a similar sequence of periodicities; however, most of them significantly shifted relative to the $\delta^{18}O$ signal: Among 30-40 (99%), 16 - 17 (95%), 9 - 9.5 years (95%), and 6-7 years, there is significant spectral power. The most striking differences between both spectra certainly is the different low-frequency behavior. The $\delta^{18}O$ isotope record shows much less low frequency power than the excess record.

Furthermore the ^{18}O record shows two gradual cooling cycles (from 1220 A.D. to 1360 A.D. and from 1400 A.D. to 1720 A.D.) both interrupted by quite abrupt warming events. In Figure 2 the sawtooth-like cooling cycles are sketched out by calculating the linear regression of both isotope series over certain periods (see Figure 2 caption for details). This sawtooth-like form has been noted previously [*Johnsen et al.*, 1998]. This structure of gradual cooling followed by an abrupt warming can be found on a comparably short timescale (10-300 years) and on the long timescale (some 1000 years) of the warm interstadials during the last glacial period as well. We have no conclusive explanation for the apparent existence of such a climate pattern on different timescales. However, it might point to a common mechanism affecting the circum-Greenland climate such as a general asymmetry of the thermohaline circulation (THC) in the North Atlantic (this is a rather gradual slow down of the THC, but an abrupt intensification).

Interestingly, a similar but inverse pattern can be seen in the deuterium excess record. The long periods of gradual cooling over Greenland are accompanied by a slight gradual increase of the excess and thus following our interpretation based on simple isotope models, by a gradual warming of the source regions of Greenland's precipitation. These warming cycles are interrupted by quite rapid coolings that are leading the corresponding shifts to warmer conditions over Greenland by about 50 years. Two extremes in the stacked excess record can be identified both clearly exceeding the 1σ band: one rather short positive deviation (+0.5‰) from about 1290-1330 A.D. and a rather longer lasting negative one from 1630 to 1740 A.D. (maximal amplitude -0.9‰). We associate these variations of *d* with the MWP and the LIA. However, in the nineteenth century, often mentioned as the second phase of the LIA, there is no pronounced cool period in our records, and in the most modern part of the isotope records, there is no particular tendency possibly linked to a general warming.

Before discussing possible climate mechanisms that might link Greenland's climate and the conditions in its

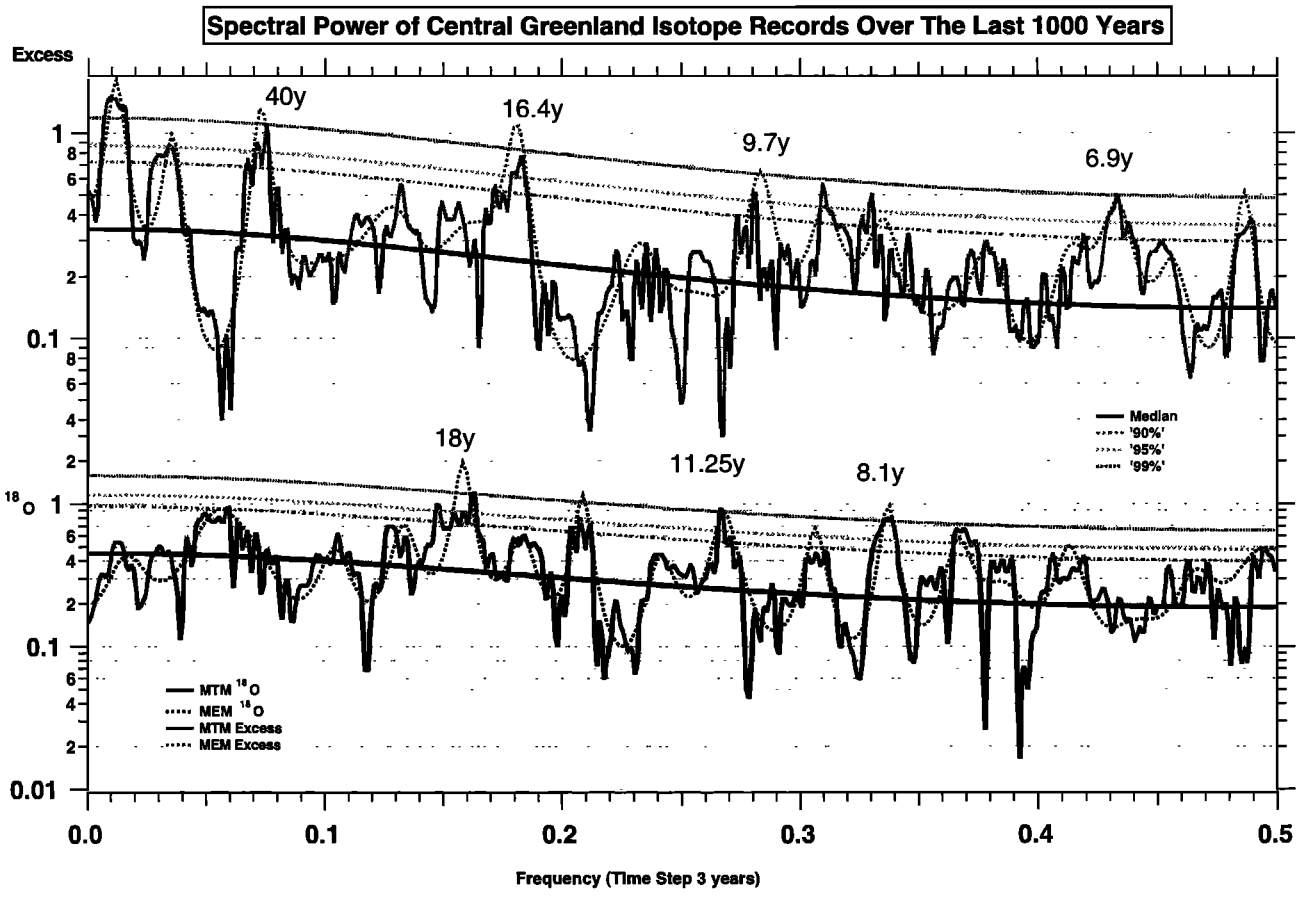

Figure 3. Spectral power of the composite deuterium excess and ^{18}O records versus frequency (the highest frequency corresponds to 1/(2*3 years). The original data were resampled with an equidistant interval of 3 years. The spectra were calculated using the maximum entropy method (MEM of the order of 40 corresponding to sample frequency of 256 data points) and the multi-taper method (number of tapers is 3). Statistical significance has been tested against a red noise spectrum using the MTM method. We used the spectral analysis SSA-MTM tool kit version 4.

water vapor source regions, we will try to better quantify both isotope series in terms of temperature variations by using a simple Rayleigh-type distillation model.

3. Modeling

The mixed cloud isotope model (MCIM) model [*Ciais and Jouzel*, 1994] describes in detail the fractionation processes during evaporation at the ocean surface and during phase changes among vapor, liquid, and solid cloud water. As with all of these types of plume models, it is limited by the basic assumption of a single source region from which vapor is transported to the site of precipitation without any further mixing with other air masses. Though this is a serious limitation, these models enable us quantifying both isotope series ($\delta^{18}O$, d) simultaneously. The simultaneity is an important point since the isotopic composition, even in the most simple and conceptual Rayleigh model [*Dansgaard*, 1964], actually is sensitive to the temperature difference between the evaporation source and the site of condensation, both possibly varying independently. Therefore applying a constant isotope/condensation-temperature relation implicitly assumes not only a constant (seasonally and regionally) contribution of the vapor source regions to the total precipitation but also that the climate conditions in the source regions have not changed significantly. However, using the deuterium excess we have access to information on the climate conditions at the sea surface and are able to correct the interpretation of the $\delta^{18}O$ signal in terms of local temperature variations by the simultaneous SST signal. Vice versa, the deuterium excess depends, though weakly, on the condensation temperatures too, mainly via the temperature dependence of both ($\delta^{18}O$ and δD) equilibrium fractionation coefficients α. The objective of this section is therefore a temperature interpretation of the $\delta^{18}O$ record taking into account the changes in the source region as documented by the deuterium excess and, vice versa, a simultaneous interpretation of the excess in terms of climate variations at the sea surface corrected by the condensation temperatures, i.e., the $\delta^{18}O$ record.

The MCIM model calculates $\delta^{18}O$ and d in polar precipitation depending on the relative humidity RH_{Sea}, the wind velocity w, the temperature at the ocean surface T_{Sea}, and the condensation temperature T_{Cond}. At the beginning of the hydrological cycle the vapor formed at the ocean surface is calculated under the assumption of a global closure condition of the water isotopes [*Merlivat and Jouzel*, 1979]: $\delta_{vap}=(1-k)/\alpha(1-kh)-1$.

Here δ_{vap} denotes the isotopic composition of the formed vapor, α the equilibrium fractionation between vapor and liquid, h is the relative humidity above the sea surface and k is a kinetic fractionation that depends smoothly on the wind regime (k is about 6‰ for $H_2{}^{18}O$ in a wide range of wind speed). Subsequently, the vapor is cooled down to its condensation point from which on the coexistence of vapor, liquid and solid is taken into account by the MCIM model. An additional kinetic fractionation process becomes important at low temperatures during formation of ice crystals. In our model, this hardly investigated fractionation mechanism is parameterized by a temperature-dependent supersaturation function S according to *Jouzel and Merlivat* [1984] and *Petit et al.* [1991]. For polar sites this function describing the supersaturation of liquid in the appearance of ice is the most important tuning parameter. This mechanism is responsible for the approximately linear relation between the isotopes and the condensation temperatures even at very low temperatures (-20° - -60°C) [*Petit et al.*, 1991]. Since the supersaturation is hardly known from direct observations a simple linear parameterization by cloud temperatures, T_{Cond}, has been chosen. In our case the model is tuned to simulate satisfyingly the seasonal cycle of $\delta^{18}O$ and d at the Summit site for the period from 1970 to 1980 by playing solely on the supersaturation function S which was tuned to S=1.003-0.007*T_{Cond}. For further details of the tuning process see *Hoffmann et al.* [1998]. In agreement with other similar studies [*Johnsen et al.*, 1989] we identified the subtropical North Atlantic as the most important source region with a mean relative humidity of 79% and a mean temperature of 26.1°C.

After this tuning process we perform a simple "inversion" of the MCIM model (see Figure 4). The model is modified to find by a simple trial and error algorithm for each pair of observed $\delta^{18}O$ and d the corresponding T_{Sea} and T_{Cond} under the assumption that the relative humidity (constant 79%), and the wind speed (4 m/s) stayed unchanged in the source region. The step-like form of the temperature reconstruction (see Figure 5) is due to the measuring errors of 0.1‰ for $\delta^{18}O$ and 0.8‰ for d which define when a simulated pair of $\delta^{18}O$ and d is close enough to the observations. This calculation gives a maximum cooling of -0.7°C in the estimated vapor source regions between about 1650 until 1750. As expected from our discussion above, calculated condensation temperatures appear to be determined by the $\delta^{18}O$ signal, whereas sea surface temperatures are determined by the deuterium excess. However, this is not strictly correct, as both isotopic signals also influence one another. For instance, a short "warm" $\delta^{18}O$ isotope signal at around 1400 amplifies the synchronous T_{Sea} signal by about 0.2°C, thereby reducing the temperature gradient between T_{Cond} and T_{Sea} (see Example 1 in Figure 5). Otherwise, a very large gradient between the source and the precipitation site could not fit with both isotope records.

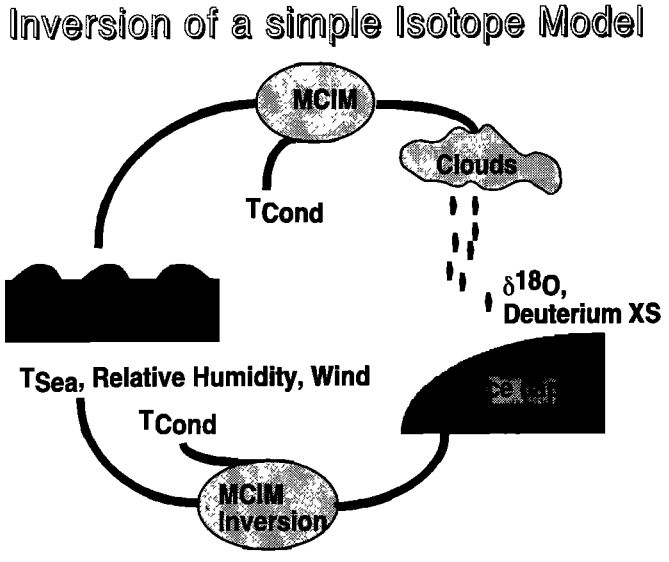

Figure 4. Scheme of our methodology to reconstruct sea surface temperatures and cloud condensation temperatures by means of a simple Rayleigh isotope model. Instead of using this model in the "forward" mode it is here used to find for each pair of $\delta^{18}O$ and d a corresponding solution for the SSTs and T_{cloud} given that no other changes affect the isotopic composition.

Another example (Example 2) for the nonlinearity of the temperature reconstruction by the MCIM model is the period of cold sea surface temperatures in the eighteenth and seventeenth centuries. In fact, after being corrected for these anomalous low T_{Sea} the corresponding T_{Cond} are comparably cooler than by just assuming a constant $\delta^{18}O/T$ gradient. A part of the problem that the LIA is not clearly seen in many Greenland ice core records might therefore be directly due to synchronous lower sea surface temperatures masking somewhat the isotope signal ($\delta^{18}O$).

With the parameters chosen here for the MCIM model we yield a mean temporal $\delta^{18}O/T$ gradient of 0.7‰/°C over the last 900 years, which is in good correspondence to the spatial gradient over Greenland (0.67‰/°C) [see *Johnsen et al.*,1989] However, this result depends crucially on the validity of our basic assumptions: an unchanged vapor source region for central Greenland and a constant relative humidity over the subtropical North Atlantic. The isotopic signal we found here could alternatively be explained by variations of the relative humidity near the sea surface of the order of 5%. For the climatic conditions (central Greenland) and the time period (the last millennium) considered here this gives us roughly a sensitivity for the deuterium excess of -0.2‰/% for the relative humidity and 1.3‰/°C for the sea surface temperature. Our crude estimate of T_{Sea} variations of about -0.8°C during the LIA is, however, consistent with SST reconstruction for the northern Saragossa Sea being about -1°C cooler than today *[Keigwin, 1969]*.

4. Discussion

In the discussion we will mainly focus on two question: (1) What is the reason for the weak LIA signal in isotope records of central and southern Greenland? (2) Is the sawtooth-like structure of $\delta^{18}O$ and d real and are they indeed anticorrelated? What could possibly be the mechanism linking the two records?

There might be several reasons for the astonishing fact that isotope records of central and south Greenland do not show a clear LIA (or MWP), whereas this signal becomes increasingly clear in ice cores from northwest Greenland. First, the most simple explanation is, of course, that central southern Greenland was indeed consistently warmer during

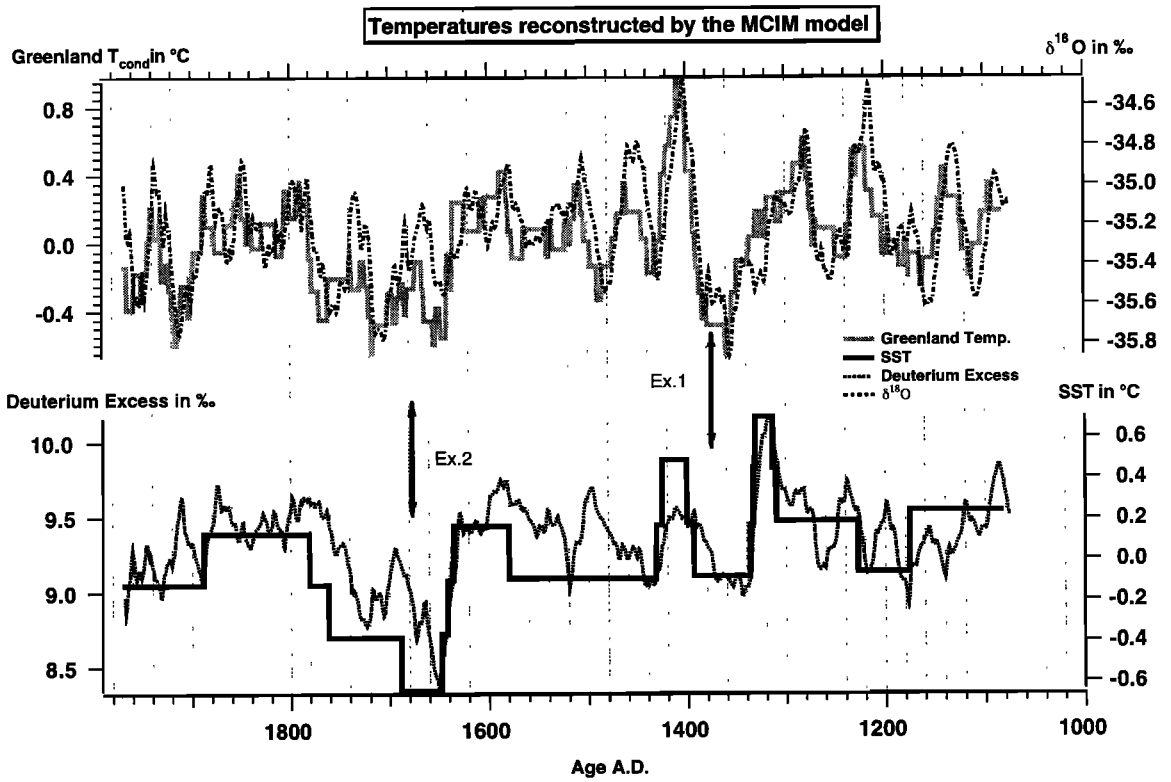

Figure 5. Twenty years running mean of $\delta^{18}O$ (top panel) and deuterium excess d (bottom) from the stacked central Greenland ice cores shown with the reconstructed anomalies of the condensation temperature (top) and the corresponding anomalies of the sea surface temperature (bottom). The square shape form of the temperature reconstruction originates from the experimental uncertainties of $\delta^{18}O$ (0.1‰) and d (0.8‰). The inversion procedure of the MCIM model calculates solely a new pair of temperatures, T_{cond} and T_{sea}, when the corresponding isotope values exit this interval of uncertainties. Example 1 (Ex1) and Example 2 (Ex2) demonstrate the nonlinear temperature reconstruction of the MCIM discussed in more detail in the text.

the LIA than the north and northwest. For the moment we have no observation or theory that might explain such a partial climatic decoupling of both regions. Second, since central and south Greenland are more exposed to cyclonic activity in the North Atlantic, the signal (in this case a possible cooling during the LIA) to noise ratio might be more favorable in the North. Third, northwest Greenland is located at the end of the distillation column which, typically starting in the North Atlantic source regions produces successively more and more depleted precipitation. The temperature sensitivity of this process, thus the $\delta^{18}O/T$ gradient, becomes increasingly larger at the very end of the vapor rainout process [*Dansgaard*, 1964]. This holds though the kinetic process during ice formation mentioned above largely dampens this steepening of the isotope/temperature relation [*Jouzel and Merlivat*, 1984]. In the work of *Rozanski et al.* [1993] the authors gave for all nonpolar stations (with a mean annual temperature below 15°C) of the global network of isotopes in precipitation (GNIP) a mean gradient of 0.58‰/°C, for south and west Greenland a gradient of 0.67‰/°C and for East Antarctica 0.79‰/°C. In north and northwest Greenland therefore a temperature variation of the same amplitude might produce a larger isotope signal due to a stronger isotope/temperature gradient at a later stage of the distillation process. Fourth, simultaneous changes in SSTs might slightly shadow the changes in temperature, as was shown in our model calculation above. For the argument being valid it requires that the vapor sources of north and northwest Greenland are different from those of the rest of Greenland and that they did not undergo the same SST variations. Indeed, hints from modeling studies exist confirming that northwest Greenland is under stronger influence of water vapor originating from the Pacific ocean than central and south Greenland [*Charles et al.*, 1994; *Werner et al.*, 2000a].

However, a completely independent argument tells us that we should take the GRIP isotope record as a reliable climate indicator even for the high-frequency oscillations (that means at least 100 years) during the Holocene. In the work of *Willemse and Toernquist* [1999] the authors showed a very good agreement between the GRIP $\delta^{18}O$ signal and the paleo-productivity records reconstructed from a near coastal lake in the southwest of Greenland throughout the entire Holocene. This important observation excludes the possibility that we are dealing here just with artefacts of the water isotope physics, though a careful analysis might be needed to calculate from the isotope record the actual temperature signal.

This discussion about the missing LIA signal in the water isotopes concerns equally the lack of any cool signal in the nineteenth century and of a warming signal over the last 100 years. Partly, this might be due to the extreme noisiness of the excess record during the most recent period (see Figure 2). Furthermore, two effects might at least partly cancel out each

other. The ongoing global warming is regionally covered by circulation-driven temperature anomalies. Decadal-to-interdecadal variability associated to the North Atlantic Oscillation (NAO) lead over certain periods to comparably cool conditions around Greenland (i.e., empirical orthogonal function (EOF) 1 for the global warming signal and EOF 3 for the NAO associated pattern in Figure 2 in the work of *Mann et al.* [1998]. However, even if such a cancelling effect is true for the twentieth century, it should not play any important role for the nineteenth century, which is documented as particularly cool [*Mann et al.*, 1998].

Obviously, the definition of the cooling cycles in Figure 2 is not straightforward. We need certainly longer time series of both isotopes to verify more quantitatively the discussed $\delta^{18}O$-*d* pattern. What mechanisms might possibly be responsible for that kind of anti-correlation between the excess (i.e., temperatures in the source regions of Greenland's precipitation) and the isotopes (i.e., local temperatures in central Greenland)?

1. *Barlow et al.* [1993] analyzed the isotope signal ($\delta^{18}O$ and excess) in the upper part of the GISP2 core (central Greenland as well) in a subannual resolution. Temporally, the core covers roughly the period for which direct meteorological observations are still abundant in the North Atlantic region, i.e. from 1840 to 1970. At least seasonally, the authors found an anticorrelation between the excess and the isotopes as in our record here though on a much shorter timescale. They explained such behaviour by analyzing meteorological data over the last 40 years during high and low phases of the NAO. During high phases of the NAO, i.e., during conditions of a particularly strong pressure gradient between Island and the Azores, Greenland's temperature are anomalously low, whereas Europe is warmer than normal. Simultaneously, stronger zonal winds amplify the wind-driven circulation in the North Atlantic and intensify the transport of comparably warm surface waters northward. This leads to a stronger temperature gradient between Greenland (colder than normal) and its source regions (warmer than normal). Conversely, during low phases of the NAO the temperature seesaw turns to the other side (Greenland warmer than normal and Europe colder), the zonal circulation weakens, thereby producing anomalously low SSTs between 40° and 30°N in the North Atlantic. Obviously, such a scenario could very well explain an anticorrelation between the excess linked to the SSTs and the isotopes controlled by local temperatures. Further arguments for this concept applied to the centennial timescale considered here comes from a paper by *Keigwin and Pickart* [1999]. Analyzing high-resolution sediment cores from the Laurentian fan (43°-44°N, 56°-54°W) and comparing these results with the former ones from the Bermuda rise, the authors state that at least during the last 2000 years there is a North Atlantic temperature seesaw into action. According to *Keigwin and Pickart* [1999] this seesaw turns around approximately 44°N and leads to temperatures out of phase north and south of this apparently critical latitude. Colder temperatures during the LIA, for example, are accompanied by warm anomalies (1°-2°C) north of it. To explain these results, the authors refer to the known interannual and decadal SST variability which was discussed by *Barlow et al.* [1993] as well and suggest an enhanced NAO variability during the LIA. However, as the authors mention, this is in contradiction to results from Greenland, indicating no exceptionally high phase of the NAO during the LIA [*Appenzeller et al.*, 1998]. Moreover, such a mechanism affects the temperatures (and the water isotopes) over the ocean and over Greenland nearly instantaneously. It therefore seems to us not clear how it possibly has lead to such phase lags as were documented in our combined $\delta^{18}O$-excess record.

2. Another reasonable mechanism linking both isotope signals, however, might be found in analyzing the variability of the THC. In its present mode of functioning the THC transports warm saline water toward high latitudes, where it cools and forms Atlantic bottom water. This mechanism represents the most effective way to redistribute heat in the ocean. A possible weakening of the THC during the LIA surely would shift the polar front southward and produce cooler conditions not only in Greenland but also in parts of Europe. The simultaneous blocking of warm water masses in the subtropics, under such conditions less effectively advected to the North, would lead to a compensating warming in low latitudes. These processes fit well to the described $\delta^{18}O$-*d* pattern: Periods of gradually descending $\delta^{18}O$ isotope values, this means local cooling over Greenland, are accompanied by a correspondingly increasing deuterium excess, indicating a gradual warming of lowlatitudinal parts of the Atlantic. A sudden reintensification of the thermohaline circulation might explain the abrupt jump to cooler conditions in the source regions first, followed by a corresponding abrupt warming in high latitudes. During these small rapid events the excess signal leads the Greenland isotope record by approximately 40 years, possibly due to a certain inertia in the coupled ocean/atmosphere/sea ice system at the northern end of the THC.

External forcing factors, such as reconstructed solar activity [*Lean et al.*, 1995] or the atmospheric dust content [*Overpeck et al.*, 1997], which is largely controlled by volcanic eruptions, do not show any resemblance with the deuterium excess record. Even if these factors might have contributed to centennial scale climate variations (for example, the minimum of the excess record roughly coincides with the Maunder sunspot minimum), they seem not to have played a dominating role for the temperatures in the subtropical North Atlantic as given here by our reconstruction from the deuterium excess.

More evidence for a possible role of the THC even on the comparably short centennial timescale came up recently in a paper of *Broecker et al.* [1999]. Analyzing various tracers in the deep Southern Ocean, the authors claim a mismatch between modern, comparable low rates of deep water formation and the actual state of the deep ocean appearing relatively "fresh" and well ventilated. In fact, the observations suggest a 3 times more intense average production of deep water in the Southern Ocean during the LIA. In their interpretation the North Atlantic has varied inversely, forming therefore less deep water masses and transporting less heat from low to high latitudes during the same time period. Our tentative interpretation of the $\delta^{18}O$-*d* signal from Greenland seems in general agreement with this hypothesis. However, we propose a compensating effect for the diminished heat transport to high latitudes already in low latitudes of the North Atlantic, whereas *Broecker et al.* [1999] suggest such a compensation rather in the Southern Hemisphere.

Some last remarks to the credibility of the isotopes as a climate indicator which might be biased by atmospheric

circulation changes: An increasing number of available proxy records from different regions and a more rigorous interpretation of these records gave evidence that the simplifying idea of a globally uniform warm climate at the beginning of this millennium, thus the MWP, followed by a global cooling in the sixteenth and the seventeenth centuries, the LIA, is misleading [*Hughes and Diaz*, 1994; *Mann et al.*, 1998]. Rather, we are faced with a temporally and spatially heterogeneous pattern of warming and cooling, both potentially influencing the Greenland isotope record: First, a strengthening or weakening of the seasonal cycle of temperature or precipitation amount strongly affects the water isotopes. Recently, it was demonstrated in numerical simulations of the last glacial maximum with atmospheric general circulation models (AGCM) that a shift of the polar front south of Greenland reduces considerably the amount of winter precipitation at Summit. This mechanism shifts the water isotopes toward higher summer $\delta^{18}O$ values and leads to a mismatch between real surface temperatures and reconstructed "isotope temperatures *[Krinner et al., 1997; Werner et al. 2000b]*. Even on the shorter millennial timescale of our study here such seasonal climate variations might bias our temperature reconstruction. Second, a further risk of our model calculation is a considerable change of the regional vapor sources of Greenland's precipitation. However, tagging experiments with AGCMs rather argue for a remarkable constancy of the contribution of the different source areas *[Werner et al., 2000a]*. The reconstructed sea surface temperature variations must be seen in the light of these considerable uncertainties. Only a long-term simulation of a fully coupled ocean-atmosphere, model which is undergoing LIA like centennial climate variations, and moreover is fitted with water isotope diagnostics is able to address directly all these questions for the interpretation of the water isotopes.

Acknowledgments. This study has been funded by EEC within the MilEClim project, contract number 260504. We like to thank Michel Stievenard, Olivier Cattani, and Sonja Falourd for their assistance at performing the isotope measurements. Early versions of the manuscript were improved a lot by the careful reading of Ulrich von Grafenstein, Valerie Masson, and three anonymous reviewers. We further like to thank the team of the SSA-MTM spectral analysis tool kit, in particular Pascal Yiou, for help and for making available this very helpful software.

References

Appenzeller, C., T. F. Stocker, and M. Anklin, North Atlantic oscillation dynamics recorded in Greenland ice cores, *Nature*, *282*, 446-449, 1998.

Armengaud, A., R.D. Koster, J. Jouzel, and J. Ciais, Deuterium Excess in Greenland snow: Analysis with simple and complex models, *J. Geophys. Res.*, *103*, 8947-8953, 1998.

Barlow, L.K., J.W.C. White, R.G. Barry, and P.M. Grootes, The North Atlantic oscillation signature in deuterium and deuterium excess in the Greenland ice sheet project 2 ice core, 1840-1970, *Geophys. Res. Lett.*, *20*(24), 2901-2904, 1993.

Boiseau, M , A. Juillet-Leclerc, P. Yiou, B. Salvat, P. Isdale, and M. Guillaume, Atmospheric and oceanic evidences of ENSO events in the south central Pacific Ocean from coral stable isotopic records over the past 137 years, *Paleoceanography*, *13*, 671-685, 1998.

Bradley, R.S., and P.D. Jones, "Little Ice Age" summer temperature variations: Their nature and relevance to recent global warming trends, *The Holocene*, *3*, 367-376, 1993.

Broecker, W. S., S. Sutherland, and T.-H. Peng, A Possible 20th-Century Slowdown of southern Ocean deep water formation, *Science*, *286*, 1132-1135, 1999.

Charles, C.D., D. Rind, J. Jouzel, R.D. Koster, and R.G. Fairbanks, Glacial-interglacial changes in moisture sources for Greenland: Influence on the ice core record of climate, *Science*, *263*, 508-511, 1994.

Ciais, P., and J. Jouzel, Deuterium and oxygen 18 in precipitation: A new isotopic Model including Mixed Cloud Processes, *J. Geophys. Res.*, *99*, 16,793-16,803, 1994.

Climate Variability (CLIVAR), Initial Implementation Plan, *WCRP 103*, 1998.

Cole, J.E., R.G Fairbanks, and G.T. Shen, Recent variability in the Southern Oscillation: Isotopic results from a Tarawa atoll coral, *Science*, *260*, 17,690-17,693, 1993.

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

Dahl-Jensen, D., K. Mosegaard, N. Gundestrup, G.D. Clow, S.J. Johnsen, A.W. Hansen, and N. Balling, Past temperatures directly from Greenland ice sheet, *Science*, *282*, 271-281, 1998.

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

Dansgaard, W., S.J. Johnsen, N. Reeh, N. Gundestrup, H.B. Clausen, and C.U. Hammer, Climatic changes, Norsemen and modern man, *Nature*, *255*, 24-28, 1975.

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

Fischer, H., M. Werner, D. Wagenbach, M. Schwager, T. Torsteinsson, F. Wilhelms, J. Kipfstuhl, and S. Sommer, "Little Ice Age" clearly recorded in Northern Greenland Ice Cores, *Geophys. Res. Lett.*, *25*, 11749-11752, 1998.

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, K. Kuivinen, H.B. Clausen, S.J. Johnsen, J.P. Steffensen, N. Gundestrup, and C.U. Hammer, Intercomparison of ice core ^{18}O and precipitation records from sites in Canada and Greenland over the last 3500 years and over the last few centuries in detail using EOF Techniques, in *Climatic Variations and Forcing Mechanisms of the Last 2000 Years*, edited by P.D. Jones, R.S. Bradley, and J. Jouzel, pp. 297-328, Springer-Verlag, New York, 1996.

Hoffmann, G., M. Stievenaard, J. Jouzel, J.W.C. White, and S.J. Johnsen, The deuterium excess record from central Greenland: Modelling and observations, in *International Symposium on Isotope Techniques in the Study of Past and Current Environmental Changes in the Hydrosphere and the Atmosphere*, pp. 591-602, Int. At. Energ. Agency, Vienna, Austria, 1998.

Hughes, M.K., and H.F. Diaz, Was there a 'Medieval Warm Period', and if so, where and when?, *Clim. Change*, *26*, 109-142, 1994.

Johnsen, S.J., W. Dansgaard, H.B. Clausen, and C.C.J. Langway, Climatic oscillations 1200-2000 A.D., *Nature*, *227*, 482-483, 1970.

Johnsen, S.J., W. 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., et al., The $\delta^{18}O$ record along the GRIP deep ice core and the problem of possible Eemian climatic instability, *J. Geophys. Res.*, *102*, 26,397-26,410, 1998.

Jouzel, J., and L. Merlivat, Deuterium and oxygen 18 in precipitation, modeling of the isotopic effects during snow formation, *J. Geophys. Res.*, *89*, 11,749-11,757, 1984.

Jouzel, J., L. Merlivat, and C. Lorius, Deuterium excess in an East Antarctic ice core suggests higher relative humidity at the oceanic surface during the last glacial maximum, *Nature*, *299*, 688-691, 1982.

Jouzel, J., C. Lorius, J. 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*, *239*, 403-407, 1987.

Keigwin, L.D., The little ice age and medieval warm period in the Saragossa Sea, *Science*, *274*, 1504-1508, 1996.

Keigwin, L.D., and R.S. Pickart, slope water current over the Laurentian fan on interannual to millenial time scales, *Science*, *286*, 520-523, 1999.

Krinner, G., C. Genthon, and J. Jouzel, GCM analysis of local influences on ice core δ signals, *Geophys. Res. Lett.*, *24*(22), 2825-2828, 1997.

Lean, J., A. Skumanich, and O.R. White, Estimating the sun's radiative output during the Maunder minimum, *Geophys. Res. Lett.*, *22*, 3195-3198, 1995.

Mann, M.E., R.S. Bradley, and M.K. Hughes, Global-scale temperature patterns and climate forcing over the past six centuries, *Science*, *392*, 779-787, 1998.

Merlivat, L., and J. Jouzel, Global climatic interpretation of the deuterium-oxygen 18 relationship for precipitation, *J. Geophys. Res.*, *84*, 5029-5033, 1979.

Overpeck, J., et al., Arctic environmental change of the last centuries, *Science*, *278*, 1251-1256, 1997.

Petit, J.R., J.W.C. White, N.W. Young, J. Jouzel, and Y.S. Korotkevitsch, Deuterium excess in recent Antarctic snow, *J. Geophys. Res.*, *96*, 5113-5122, 1991.

Rozanski, K., L. Araguas-Araguas, and R. Gonfiantini, Isotopic patterns in modern global precipitation, in *Climate Change in Continental Isotopic Records*, *Geophys. Monogr. Ser.*, vol. 78, edited by P.K. Swart, K.C. Lohmann, J. MacKenzie, and S. Savin, pp. 1-37, AGU, Washington, D. C., 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.

Vimeux, F., V. Masson, J. Jouzel, M. Stievenaard, and J.-R. Petit, Glacial-Interglacial changes in ocean surface conditions in the Southern Hemisphere, *Nature*, *398*, 410-413, 1999.

Werner, M., M. Heimann, and G. Hoffmann, Isotopic composition and origin of polar precipitation in present and glacial climate simulations, *Tellus, Ser. A*, in press, 2000a.

Werner, M., U. Mikolayewicz, M. Heimann, and G. Hoffmann, Borehole versus Isotope Temperatures on Greenland: Seasonality does matter, *Geophys. Res. Lett.*, *27*(5), 723-726, 2000b.

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 Summit, Greenland Snow: Results of Comparisons with Modern Climate Observations, *J. Geophys. Res.*, *102* (C12), 26,425-26,439, 1997.

Willemse, N.W., and T.E. Toernquist, Holocene century-scale temperature variability from West Greenland lake records, *Geology*, *27*(7), 580-584, 1999.

G. Hoffmann and J. Jouzel, IPSL/LSCE Laboratoire des Sciences du Climat et de l'Environnement, UMR CEA/CNRS No. 1572, Orme des Merisiers, 91191 Gif-sur-Yvette, France (hoffmann@lsce.saclay.cea.fr).

S. Johnsen, Science Institute, University of Iceland, Reykjavik, Iceland.

(Received March 1, 2000; revised September 18, 2000; accepted September 19, 2000.)