A continuous record of temperature evolution over a sequence of Dansgaard-Oeschger events during Marine Isotopic Stage 4 (76 to 62 kyr BP)

A. Landais, J. M., Barnola, Valérie Masson-Delmotte, Jean Jouzel, J., Chappellaz, N., Caillon, C., Huber, M., Leuenberger, S. J., Johnsen

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A. Landais,[1] J. M. Barnola,[2] V. Masson-Delmotte,[1] J. Jouzel,[1] J. Chappellaz,[2] N. Caillon,[1] C. Huber,[3] M. Leuenberger,[3] and S. J. Johnsen[4]

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[1] Our knowledge of the temperature evolution over Greenland during Dansgaard-Oeschger events (DO) is currently qualitatively described through the water isotopic profile. Using two independent paleothermometry methods, one based on air isotopic measurements and the other on the combined measurements of water isotopes (δD and δ^{18}O), we show a complete and quantitative reconstruction of temperature at the NorthGRIP site over the period 76 to 62 kyr BP (DO 18, 19 and 20). We confirm that the associated warmings are larger than those conventionally depicted by the water isotopes (11°C, 16°C and 11°C for DO 18, 19 and 20). Secondly, we demonstrate that the relationship between temperature and δ^{18}O varies rapidly during the last glacial period, even over a DO. Finally, our temperature reconstruction over DO 19 agrees well with that predicted from simple climate models linking the DO to iceberg discharges. *INDEX TERMS:* 1040 Geochemistry: Isotopic composition/chemistry; 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology; 3354 Meteorology and Atmospheric Dynamics: Precipitation (1854). **Citation:** Landais, A., J. M. Barnola, V. Masson-Delmotte, J. Jouzel, J. Chappellaz, N. Caillon, C. Huber, M. Leuenberger, and S. J. Johnsen (2004), A continuous record of temperature evolution over a sequence of Dansgaard-Oeschger events during Marine Isotopic Stage 4 (76 to 62 kyr BP), *Geophys. Res. Lett.*, *31*, L22211, doi:10.1029/2004GL021193.

[2] The last glacial cycle was characterized by millenial scale climate fluctuations that have been documented in the North Atlantic region through numerous marine and continental records [*Bond et al.*, 1997; *Genty et al.*, 2003]. The GRIP and GISP2 Summit ice cores [*Dansgaard et al.*, 1993; *Grootes et al.*, 1993] and the newly NorthGRIP [*NorthGRIP Members*, 2004] ice core exhibit 25 DO during the last glacial period. These events are characterized by rapid, e.g., in less than 100 years, and large, up to 16°C [*Lang et al.*, 1999] warmings over Greenland.

[1]Institut Pierre-Simon Laplace/Laboratoire des Sciences du Climat et de l'Environnement, CEA Saclay, Gif-sur-Yvette, France.
[2]Laboratoire de Glaciologie Geophysique de l'Environnement, Centre National de la Recherche Scientifique, Saint Martin d'Heres, France.
[3]Physics Institute, University of Bern, Bern, Switzerland.
[4]Department of Geophysics, University of Copenhagen, Copenhagen, Denmark.

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[3] Ice cores have already provided a wealth of information on DO in Greenland through the water isotopes for temperature changes, chemical records for atmospheric circulation and analysis of air bubbles for changes in greenhouse gases concentration. However, temperature reconstruction from water isotopes is subject to large biases mainly due to (i) the seasonality of the precipitation (i.e., periods without precipitation will not have their temperature recorded) [*Fawcett et al.*, 1997] and (ii) changes in the oceanic source of Greenland snow [*Boyle*, 1997]. Whereas the latter can be estimated from the combined measurement of δD and δ^{18}O through the deuterium excess parameter ($d = \delta D - 8 * \delta^{18}O$), we need additional information to account for the influence of seasonality and possible variation of the vertical atmospheric temperature profile (the isotopic composition of the snow depends on the condensation temperature).

[4] One elegant way to overcome those difficulties is to use the isotopic composition of the air trapped in ice, a method based on the thermal diffusion of gases (nitrogen and argon), which allows estimates of the amplitude of rapid temperature increases [*Severinghaus and Brook*, 1999; *Lang et al.*, 1999; *Leuenberger et al.*, 1999]. This method gives directly access to the local mean surface temperature, and is therefore not affected by seasonality, the vertical temperature profile in the atmosphere, nor the source temperature of the precipitation. It confirms that the conventional temperature to water isotopes relationship, based on the spatial slope α_s (the $\Delta\delta^{18}O_{ice}$/temperature slope calculated from present-day surface data) underestimates Greenland temperature change [*Cuffey et al.*, 1995]. The temporal slope $\alpha_t = \Delta\delta^{18}O_{ice}/\Delta T$ (at a given site between two different climates) is lower than α_s by up to a factor of 2. We have recently refined this approach, using a sophisticated firnification and heat diffusion model [*Goujon et al.*, 2003] to quantify the temperature increase associated with large DO 12 [*Landais et al.*, 2004].

[5] Here, we go beyond the simple quantification of the temperature increase. A detailed set of δ^{15}N and δ^{40}Ar measurements over DO 18, 19 and 20 provides us with strong constraints on the complete temperature scenario between 76 and 62 kyr BP. This period roughly corresponds to Marine Isotope Stage 4 with rapid ice sheet growth [*Shackleton*, 1987]. We have also measured a continuous δD profile in the ice that, combined with the existing δ^{18}O record, allows us to account for the source temperature effects. This complete study makes possible the precise estimation of the temperature change over a sequence of DO.

[6] The δ^{15}N and δ^{40}Ar profiles (Figure 1) show a sharp peak corresponding to each warming. Rapid surface warm-

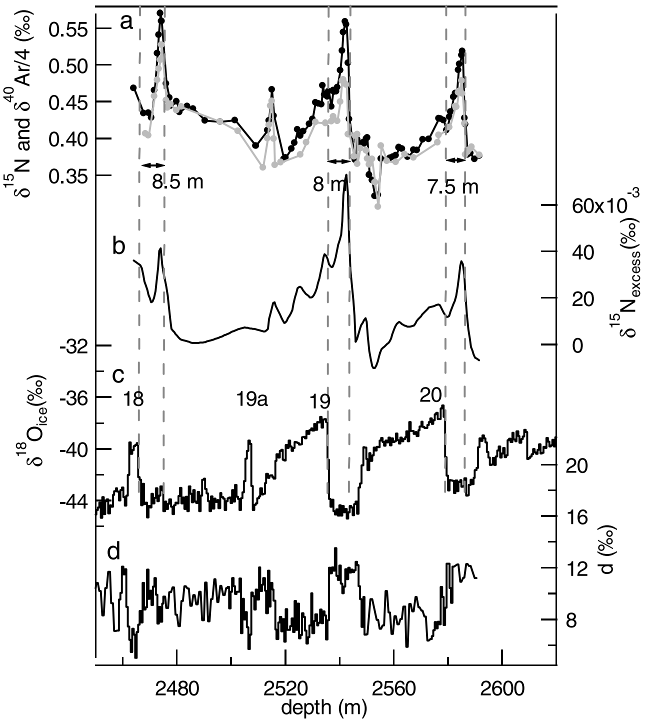

Figure 1. a: $\delta^{15}N$ (black) and $\delta^{40}Ar$ (grey) measured over DO 18, 19 and 20. The analytical uncertainty is 0.006‰ for $\delta^{15}N$ and 0.025‰ for $\delta^{40}Ar$. b: $\delta^{15}N_{excess}$ smoothed over 4 points. c: $\delta^{18}O_{ice}$ [*NorthGRIP Members*, 2004]. d: Deuterium excess. The Δdepth corresponding to each DO warming is indicated (both side arrows).

ing creates a temperature gradient in the firn, where thermal fractionation drives the heaviest isotopes towards its coldest end [*Severinghaus et al.*, 1998]. Part of the signal is also due to gravitational fractionation, which varies significantly during the warming phase of a DO because of the simultaneous changes in temperature and accumulation rate. By combining $\delta^{15}N$ and $\delta^{40}Ar$ data, we can separate the thermal and gravitational signals. The temperature gradient in the firn is then simply proportional to the $\delta^{15}N_{excess} = \delta^{15}N - \delta^{40}Ar/4$ [*Severinghaus and Brook*, 1999]. We estimate the surface temperature change using a firnification model accounting for the diffusion of heat in the firn and in the underlying ice sheet [e.g., *Goujon et al.*, 2003].

[7] The d profile shows a strong antiphase with $\delta^{18}O$. We follow *Johnsen et al.* [1989] in interpreting this feature as reflecting a significant warming of the oceanic source region for NorthGRIP precipitation during the cold phase of a DO, probably induced by a southward displacement of this evaporative region. To reconstruct a temperature scenario over Greenland from δD and $\delta^{18}O$ of ice, we have inverted a Rayleigh-type isotopic model [*Ciais and Jouzel*, 1994] as currently done for interpreting co-isotopic profiles both from Greenland [*Hoffmann et al.*, 2001] and Antarctica [*Stenni et al.*, 2001] (Appendix A).

[8] Following the approach described by *Landais et al.* [2004], the temperature changes are estimated to be of 11°C, both for DO 18 and 20, and of 16°C for DO 19 ($\sigma = \pm 2.5$°C). This latter estimate is consistent with that derived for the same event at GRIP [*Lang et al.*, 1999]. The corresponding α_t (expressed in ‰/°C) are of 0.3, 0.4 and 0.5 for DO 18, 19 and 20 respectively. Again, these values are well below α_s (present-day value of 0.67 [*Dansgaard*, 1964]) which is likely due to changes in precipitation seasonality. Interestingly, the decrease of α_t with time parallels the observed sea-level decrease during Marine Isotope Stage 4 (74 to 59 kyr BP), suggesting a link between precipitation seasonality and the build-up of the Laurentide and Fennoscandian ice sheets with an associated southward deviation of the winter storm tracks and a drastic decrease of winter snowfall in central Greenland.

[9] Going beyond the estimation of the temperature amplitude of DO warmings, we show next that the complete temperature scenario can be reconstructed through the combined use of $\delta^{15}N_{excess}$, Δdepth (the depth difference between the warming recorded in the $\delta^{18}O_{ice}$ and in the $\delta^{15}N$) and $\delta^{15}N$ (Appendix B). The use of a constant α_t for a given DO is indeed not consistent with the $\delta^{15}N$ profile for DO 19 and 20 as calculated with a firnification and heat diffusion model (Figure 2). A second argument against a constant α_t comes from our d profile that suggests large reorganisations of the hydrological cycle between the cold and warm phases. Hence, in order to better reproduce the measured $\delta^{15}N$ profile, we assume that α_t varies freely with time, i.e., that the temperature is no longer linearly related to $\delta^{18}O_{ice}$. Sensitivity experiments were performed step by step over 100 year windows by modifying the initial temperature scenario deduced from $\delta^{18}O_{ice}$ for input to the model for each window. We constrained the temperature scenario by minimizing, for each time window, the area between modelled and measured $\delta^{15}N$ profiles.

[10] The shape of the gas derived temperature record differs from the one inferred from $\delta^{18}O_{ice}$ assuming a constant α_t. We illustrate this point on DO 19 where more detailed $\delta^{15}N$ and $\delta^{40}Ar$ measurements were performed. Whereas the cold and long phase (1500 yr) preceding the rapid DO warming is flat for $\delta^{18}O_{ice}$ (Figure 3), the reconstructed gas temperature depicts a cold dip lasting ~300 yr (I), followed by a ~5°C temperature increase in ~1000 yr (II), and a last cooling (~1°C in ~200 years, III). The main rapid 16°C warming (~100 yr) is followed by a short maximum (IV lasting only 200 yr) and then by a cooling much more rapid than suggested by $\delta^{18}O_{ice}$ (V). This shape agrees well with that independently derived from combined δD and $\delta^{18}O$, including the temperature increase in two main steps, I and II (but not the small intervening cooling, III) and the shortness of the temperature maximum (IV). A similar relatively small amplitude is also obtained for DO 19a.

[11] Our water isotope interpretation, which allows the shape of the gas reconstructed temperature scenario to be reproduced fails however to reproduce the amplitudes of the temperature variations inferred from the $\delta^{15}N_{excess}$. This underestimation of the temperature changes using the combined analysis of water isotopes is probably due to the seasonality bias that is neglected with this approach (we consider that $\delta^{18}O_{ice}$ and d represent annual mean, whereas they are probably shifted toward summer values during glacial period). We consequently show that the temperature shape of the DO as inferred by $\delta^{18}O_{ice}$ is biased by T_{source} while the amplitude is biased by the seasonality of the precipitation. As others [*Fawcett et al.*, 1997; *Boyle*, 1997;

Hoffmann et al., 2001], we conclude that both changes in precipitation origin and seasonality should be taken into account when interpreting Greenland water isotope data.

[12] Our reconstructed temperature scenario provides a quantitative picture of a DO sequence over Greenland. The DO cold phase is viewed as a decrease in northward heat transport by the ocean, associated with a weakened thermohaline circulation (THC) driven by the discharge of icebergs from the Laurentide or the Fennoscandian ice sheet [*Bond et al.*, 1997; *Ganopolski and Rahmstorf*, 2001]. The short cold maximum (I) lasting ~300 years should thus correspond to the duration of the associated discharge, which is in agreement with recent model estimate [*Roche et al.*, 2004]. The following slow warming (~1000 years, II) suggests that no large changes in the THC and in the atmospheric circulation occurred at that time. The fresh water input is possibly reduced during this period due to the landward retreat of the ice sheet after the iceberg discharge. This leads to a restart of the THC, after 1000 years, with the associated atmospheric feedbacks that bring the accumulated heat in low latitudes toward the northern Atlantic and are at the origin of the large and rapid warming (16°C in ~100 years). After this heat input (the temperature maximum lasts ~200 years), this transient state slowly relaxes back to cooler values.

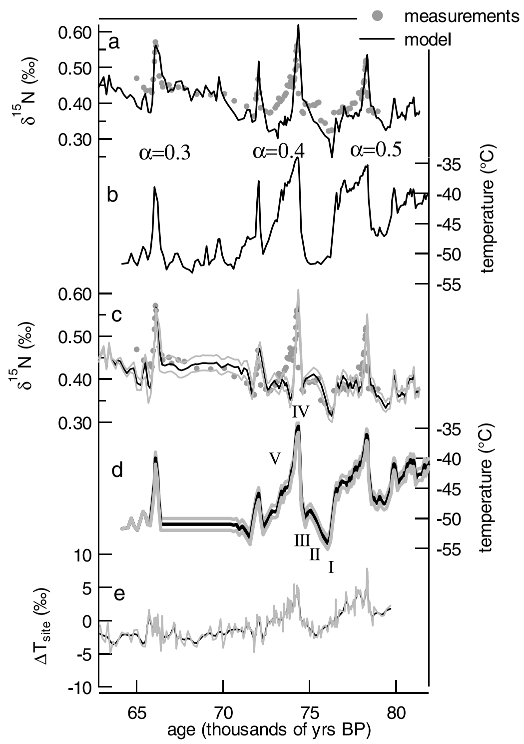

Figure 2. Temperature reconstructions respecting the 11, 16 and 11°C for DO 18, 19 and 20 warmings as inferred from $\delta^{15}N_{excess}$ variations. a: measured $\delta^{15}N$ compared to $\delta^{15}N$ modelled with the temperature scenario shown below. b: Surface temperature scenario used as input to the model by *Goujon et al.* [2003]. α ($\Delta\delta^{18}O_{ice}/\Delta T$) was taken as a constant over each DO event. c: measured $\delta^{15}N$ compared to $\delta^{15}N$ modelled with the temperature scenarii shown below. The agreement is rather poor over phase V even if it minimizes the area between model and measurements. Better agreement could be reached after improvement of the firn model (e.g., addition of gas diffusion) as shown by preliminary intercomparisons of firn models. d: Surface temperature scenarii used as input to the model by *Goujon et al.* [2003]. α is allowed to change during a DO. e: ΔT_{site} obtained by inversion of the $\delta^{18}O_{ice}$, d and $\delta^{18}O_{sw}$ profiles. The $\delta^{18}O_{sw}$ correction does not influence the reconstructed temperature scenario over a DO because of the low resolution of the profile (~1500 yrs).

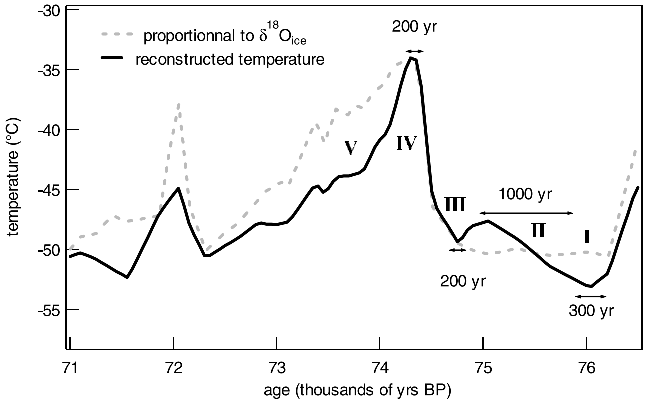

Figure 3. Gas temperature reconstruction over DO 19 (solid line). The temperature scenario proportional to $\delta^{18}O_{ice}$ is in dotted grey line.

[13] This reconstructed Greenland temperature scenario essentially follows the high latitude temperature evolution predicted by simple models [*Paillard and Labeyrie*, 1994]. However, the sharp cooling of ~1°C in 200 years preceding the main temperature increase is missing from the different modelling approaches. If the cooling indicated by the gas isotopic measurements on DO 19 did indeed occur (in agreement with a previous study on DO 12 [*Landais et al.*, 2004]), it would reflect an instability in the THC "off mode" or a triggering of the restart through sea-ice formation.

[14] The THC switching between "off" and "on" modes satisfyingly explains the shape of our inferred Greenland temperature changes. It is, however, not sufficient to account for its large amplitude (16°C). Studies performed with a coupled climate model of intermediate complexity with a crude representation of the atmosphere [*Ganopolski and Rahmstorf*, 2001] suggest that the switching on of the THC explains no more than a 8°C warming over Greenland. We therefore conclude that strong atmospheric feedbacks such as storm tracks and cyclogenesis extent displacements must have occurred in association with these THC changes.

Appendix A

[15] $\delta^{18}O_{ice}$ and δD are mainly influenced by the temperature at which snow precipitates. They also bear the imprint of the isotopic composition of the global ocean, $\delta^{18}O_{sw}$ (regional variability in oceanic $\delta^{18}O$ is neglected according to isotopic Rayleigh distillation modelling or Mixed Cloud Isotopic Model, MCIM [*Ciais and Jouzel*, 1994]) as well as surface conditions in the evaporative regions. d is linked to such source conditions, mainly temperature (MCIM analyses show that the influence of the relative humidity and wind speed can be neglected). To quantify the sensitivities

of $\delta^{18}O_{ice}$ and d to T_{site} and T_{source}, we first correct them for the long term $\delta^{18}O_{sw}$ evolution [*Waelbroeck et al.*, 2002] following the procedure described by *Jouzel et al.* [2003] derived from simple Rayleigh distillation ($\Delta\delta^{18}O_{corr} = \Delta\delta^{18}O - 0.95\ \Delta\delta^{18}O_{sw}$ and $\Delta d_{corr} = \Delta d - 2\ \Delta\delta^{18}O_{sw}$, ($\Delta$ designs anomalies from modern conditions). The MCIM model is then run for varying T_{site} (−60°C to −30°C) and T_{source} (10° to 25°C) to infer their influences on $\delta^{18}O_{corr}$ and d_{corr}. After linear inversion, T_{site} and T_{source} can be expressed as functions of $\Delta\delta^{18}O_{corr}$ and Δd_{corr}. We used here the T_{site} reconstruction: $\Delta T_{site} = 1.32\ \Delta\delta^{18}O_{corr} + 1.04\ \Delta d_{corr}$.

Appendix B

[16] Rapid temperature changes affect the $\delta^{15}N$ as well as the $\delta^{15}N_{excess}$. The $\delta^{15}N_{excess}$ variations give the amplitude of the biggest rapid temperature changes. When the $\delta^{15}N_{excess}$ variations are smooth, the temperature scenario should preferentially be inferred through the $\delta^{15}N$ since (i) the analytical uncertainty is poorer for $\delta^{15}N_{excess}$ and (ii) the $\delta^{15}N$ gravitational signal [*Severinghaus et al.*, 1998] provides additional information on slow temperature change due to its dependence on firn depth, which depends on accumulation and temperature. Modelling of $\delta^{15}N$ together with Δdepth can hence yield information on accumulation and temperature.

[17] Of the 2 factors influencing the $\delta^{15}N$ gravitational signal, the mean temperature is the more important one: a 0.05‰ change in $\delta^{15}N$ can be induced by a 3.5°C change in temperature or a 40% change in accumulation rate. The accumulation rate has a larger influence on Δdepth: a 40% change in accumulation rate results in a 80% change in Δdepth while the Δdepth is modified by 8% for a 3.5°C change.

[18] The GRIP ss09sea age scale has been adopted as a preliminary time scale for the NorthGRIP core [*NorthGRIP Members*, 2004]. Such dating enables us to reproduce the measured Δdepth with a convective zone of 2 m (in agreement with modern firn measurements). Using this dating, we modify step by step the surface temperature scenario initially proportional to $\delta^{18}O_{ice}$ to reproduce the $\delta^{15}N$ profile (note that we also manage to reproduce the $\delta^{15}N_{excess}$). The surface temperature scenario must still conform with the 11, 16 and 11°C rapid warmings for DO 18, 19 and 20.

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J. M. Barnola and J. Chappellaz, LGGE, CNRS, 53 Rue Molière, F-38402 St Martin d'Heres, France.

N. Caillon, J. Jouzel, A. Landais, and V. Masson-Delmotte, IPSL/LSCE, CEA Saclay, F-91191 Gif-sur-Yvette, France. (landais@lsce.saclay.cea.fr)

C. Huber and M. Leuenberger, Physics Institute, University of Bern, CH-3012 Bern, Switzerland.

S. J. Johnsen, Department of Geophysics, University of Copenhagen, DK-2100 Copenhagen, Denmark.