Anatomy of a Dansgaard-Oeschger warming transition: High-resolution analysis of the North Greenland Ice Core Project ice core

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[1] Large and abrupt temperature oscillations during the last glacial period, known as Dansgaard-Oeschger (DO) events, are clearly observed in the Greenland ice core record. Here we present a new high-resolution chemical (2 mm) and stable isotope (20 mm) record from the North Greenland Ice Core Project (NGRIP) ice core at the onset of one of the most prominent DO events of the last glacial, DO-8, observed ~38,000 years ago. The unique, subannual-resolution NGRIP record provides a true sequence of change during a DO warming with detailed annual layer counting of very high depth resolution geochemical measurements used to determine the exact duration of the transition. The continental ions, indicative of long-range atmospheric loading and dustiness from East Asia, are the first to change, followed by the snow accumulation, the moisture source conditions, and finally the atmospheric temperature in Greenland. The sequence of events shows that atmospheric and oceanic source and circulation changes preceded the DO warming by several years.

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1. Introduction

[2] The climate of the last glacial period was dominated, at least in the North Atlantic region, by large temperature variations, known as Dansgaard-Oeschger (DO) events [*Bond et al.*, 1993; *Dansgaard et al.*, 1993; *Grootes*, 1993; *Taylor et al.*, 1993; *Mayewski et al.*, 1997; *North Greenland Ice Core Project Members* (*NGRIP Members*), 2004]. These DO events, most clearly seen in the isotopic records from Greenland ice cores, were abrupt jumps between periods of extreme cold to relatively warm conditions, each lasting a few centuries to several thousand years. The magnitude of the warming in Greenland ranged from 9°C to 16°C [*Lang et al.*, 1999; *Severinghaus and Brook*, 1999; *Huber et al.*, 2006], with a general saw-toothed pattern of rapid warming followed by a gradual cooling. The numbering scheme used for DO events identifies 25 of them between 10 and 120 ka B.P. (before present, 1950) [*NGRIP Members*, 2004].

[3] Even in Greenland ice cores there is clear evidence that DO events have a footprint that is much larger than the region around Greenland itself. Analysis of atmospheric trace gases from the North Greenland Ice Core Project (NGRIP), the Greenland Ice Core Project (GRIP), and the Greenland Ice Sheet Project 2 (GISP2) ice cores has shown that methane (CH_4) and nitrous oxide (N_2O) concentrations are high in the warm phases of DO events and low during the cold phases [*Brook et al.*, 1996; *Flückiger et al.*, 2004; *Huber et al.*, 2006]. The CH_4 increase at the onset of each DO event is in phase with temperature, with a lag of between 0 and 30 years estimated for the most recent event [*Severinghaus et al.*, 1998; *Huber et al.*, 2006]. During cold phases of DO cycles, there are very large increases in deposition of dust and sea salt in Greenland ice [*De Angelis et al.*, 1997; *Legrand et al.*, 1997; *Fuhrer et al.*, 1999; *Mayewski et al.*, 1994; *NGRIP Members*, 2004]. Sea salt originates over the ocean, while dust arriving in Greenland originates from Asia [e.g., *Biscaye et al.*, 1997]. These large increases arise from changes in the source region and transport, and therefore suggest that atmospheric circulation was altered during DO events at a hemispheric scale.

[4] Although it can be difficult to confirm that events seen in other records are exactly synchronous with Greenland DO events, clear counterparts to the DO events are seen in numerous records across the northern hemisphere. Particularly clear examples include records of sea surface temperatures (SSTs) off Portugal [*Shackleton et al.*, 2000], the bioturbation of sediments in the Santa Barbara Basin on

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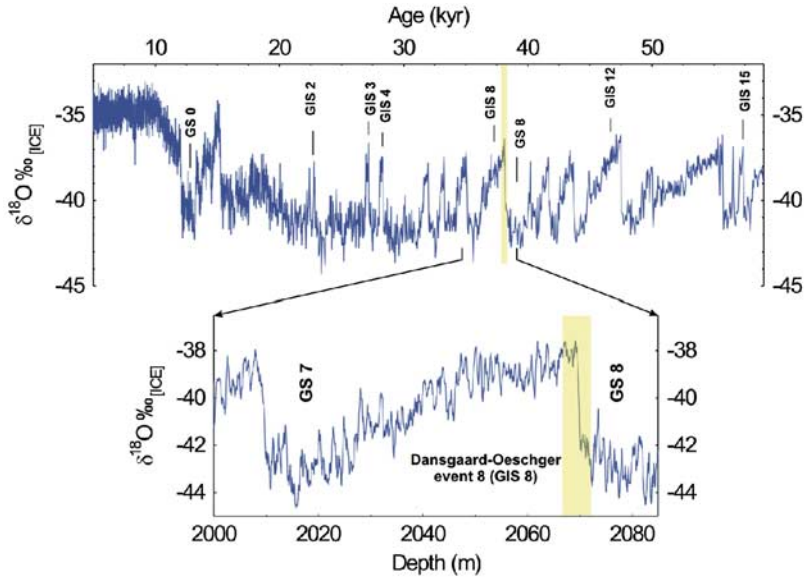

Figure 1. Age and depth of ice used in this study (highlighted yellow). (top) North Greenland Ice Core Project (NGRIP) δ^{18}O [*NGRIP Members*, 2004] on the Greenland Ice Core Chronology 2005 (GICC05) age scale [*Andersen et al.*, 2006] between 5 and 60 ka before present (B.P. 1950) and (bottom) Dansgaard-Oeschger event 8 (DO-8) on the NGRIP depth scale between 2000 and 2090 m. Selected Greenland stadials (GS) and interstadials (GIS) are marked for reference (numbered according to *Rousseau et al.* [2006].

the Pacific coast of North America [*Behl and Kennett*, 1996], and the oxygen isotope record of speleothems in Hulu Cave, China [*Wang et al.*, 2001]. Indeed, *Voelker et al.* [2002] identified 183 glaciological, marine and terrestrial sites across the globe where DO events are observed.

[5] For the most prominent DO events, it has been shown (by synchronizing Antarctic and Greenland ice core records through their common signal of methane variations) that there are counterparts to DO events in the south, but with an interesting phase relationship [*Blunier and Brook*, 2001; *Stocker and Johnsen*, 2003]. While Greenland is in a cold DO phase, Antarctic temperature shows a slow warming tendency; as soon as Greenland jumps into a warm DO period, the Antarctic trend is reversed, and Antarctica cools again. Recently, it has been shown that there is probably a counterpart in Antarctica to every DO event [*EPICA Community Members*, 2006], although it is not yet possible to confirm whether the phasing is similar in the less prominent events to that seen for the more prominent ones.

[6] The results from Antarctica lend weight to the commonly held paradigm that DO events are related to changes in the thermohaline circulation (THC) [*Broecker et al.*, 1985; *Broecker*, 1997; *Stocker and Johnsen*, 2003; *Knutti et al.*, 2004] and intensified formation of North Atlantic deep water (NADW) triggered by changes in surface water density [*Schmidt et al.*, 2006]. The abrupt atmospheric warmings over Greenland are replicated in North Atlantic sediment cores as increases in the sea surface temperatures (SSTs) at high and low latitudes [*Bond et al.*, 1993; *Curry and Oppo*, 1997; *Hendy and Kennett*, 1999]. In the Pacific, DO events are synchronous with fluctuations in surface and intermediate waters in the Santa Barbara Basin, on the Californian margin, and in the geochemical records of deepwater ventilation [*Keigwin and Jones*, 1994]. The sediment record from the Cariaco Basin in the western tropical Atlantic off the coast of Venezuela show shifts in the Atlantic Intertropical Convergence Zone (ITCZ) coincident with Greenland temperature fluctuations [*Hughen et al.*, 1996].

[7] While these and other studies shows that some of the oceanic changes one would expect to be associated with changes in THC did occur, there is still only limited evidence [*Gherardi et al.*, 2005] that changes in the overturning circulation were indeed responsible for the DO events. An alternative approach to seeking such evidence is therefore to map the spatial and temporal signature of the changes occurring at the sharp transitions into and out of DO events, to assess whether the spatial pattern and temporal phasing of the climate changes observed are consistent with model predictions based on a particular cause (such as a change in THC). Because Greenland ice cores contain signatures that are representative of different parts of the northern hemisphere climate system, and because the phasing of the different signals within a particular core (at least for the water isotope and aerosol components) can be determined very precisely, well-resolved records from Greenland ice cores are a highly suitable piece of evidence for such comparisons.

[8] In this paper we investigate Dansgaard-Oeschger event 8 (DO-8), one of the most prominent interstadials of the last glacial period observed during Marine Isotope Stage 3 (MIS 3), an interval characterized by various abrupt temperature jumps in the Greenland ice core record of up to 15°C [*Huber et al.*, 2006]. DO-8 is observed approximately 38 ka B.P. (Figure 1), according to the Greenland Ice Core Chronology 2005 (GICC05) dating [*Andersen et al.*, 2006], preceded by Heinrich event 4 (H4), observed in North Atlantic marine cores as a period of high concen-

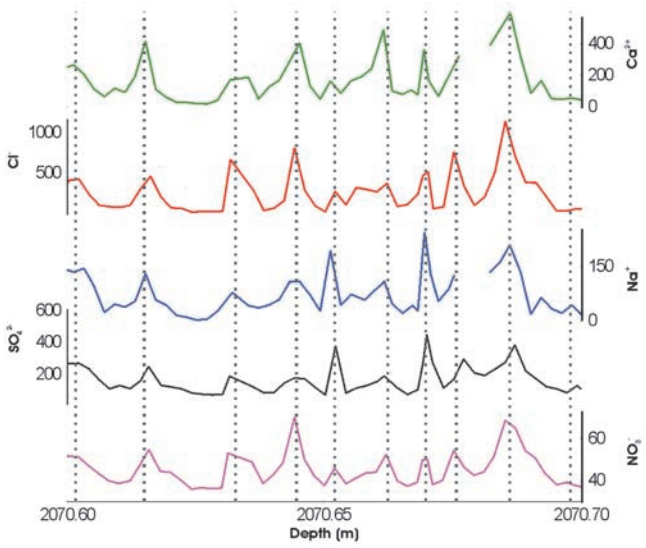

Figure 2. Annual layer counting from the concentrations of five ions: from top to bottom, calcium, chloride, sodium, sulphate, and nitrate, all in ppb at 2-mm resolution. Vertical dotted lines indicate years.

trations of ice rafted debris [*Heinrich*, 1988]. Antarctica is clearly slowly warming in the approximately 1.5 ka cold period that precedes it [*Blunier and Brook*, 2001], and cooling during the warm phase itself. CH_4 (not measured in this paper) jumps by over 100 ppbv at the onset of DO-8 [e.g., *Flückiger et al.*, 2004] indicating a significant change occurred in tropical wetlands. The onset of DO-8 therefore is a good target for testing possible causes of DO warmings, including a resumption of a vigorous THC.

[9] We present a new high-resolution (2 mm) chemical record from the North Greenland Ice Core Project (NGRIP) ice core [*NGRIP Members*, 2004] at the onset of interstadial 8 (DO-8). The new record is used to show the rate of change of chemical deposition over Greenland during the large and abrupt transition at the onset of DO-8, at a subannual resolution, and to assess the phasing between changes in different components. Annual layer counting is used to determine the duration of the transition and the true rate of change. In addition to the chemical measurements, a new high-resolution (20 mm) oxygen and deuterium isotope record is presented to determine the deuterium excess over the transition at high resolution.

2. Methods

[10] Sections of the NGRIP core, between 2068.5 and 2073 m (38,118–38,350 years B.P. on the GICC05 timescale; note that the dates in the official GICC05 timescale [*Andersen et al.*, 2006] were given as years before 2000 AD (b2k) and therefore have to be adjusted by 50 years to map them onto the B.P. dates given here), corresponding to the warming transition into DO-8, were cut into discrete samples of 2-mm thickness using a microtome device. A new cutting and small volume ion chromatography method was developed [*Thomas et al.*, 2008], and this is the first time that this resolution has been achieved at this ice depth.

[11] The section of ice used in this study was from the outermost part of the core; the outer surfaces were removed using a band saw to prevent contamination from drilling fluid and handling, leaving a horizontal area of just 20 mm $\times$ 30 mm. A sledge microtome was then used to cut 2-mm samples (using several passes of the blade to cut a succession of 40-μm shavings), with no ice loss, that were placed in clean sample vials (“accuvettes”) for analysis. Powder free gloves and protective clothing was worn throughout to minimize contamination. Tests of this method were made using frozen ultra pure water to show that the blanks were negligible (<5% of glacial concentration levels) for all ions [*Thomas*, 2006] except potassium (blank around 25% of typical concentrations). The measurements for this ion should be treated with caution.

[12] Ion chromatography was used to determine five anions; chloride (Cl^-), fluoride (F^-), methansulphonate (MSA^-), nitrate (NO_3^-) and sulphate (SO_4^{2-}), using a reagent-free Dionex ICS-2500 (2-mm column, 250-μL sample loop, 0.25 mL min^{-1} with an isocratic elution of 23 mM KOH), and four cations; calcium (Ca^{2+}), potassium (K^+), sodium (Na^+) and magnesium (Mg^+), using a reagent free Dionex IC-2000 (3-mm column, 250-μL sample loop, 0.5 mL min^{-1} with an isocratic elution of 20 mM MSA-). A sequential flow method was used so that both anions and cations could be analyzed from the same sample vial, using a Dionex AS50 autosampler, thus minimizing sample volume requisites [*Thomas*, 2006].

[13] Water isotopes were sampled at 20-mm resolution for oxygen ($\delta^{18}O$) and deuterium (δD) (Figure 3a), at the Natural Environment Research Council (NERC) Isotope Geosciences Laboratory (NIGL) using stable isotope mass spectrometer VG SIRA. Isotopic ratios ($^{18}O/^{16}O$ and $^2H/^1H$) are expressed in delta units, $\delta^{18}O$ and δD (‰, parts per mil), and defined in relation to the international standard, VSMOW (Vienna Standard Mean Ocean Water). Analytical precision is typically ±0.05‰ for $\delta^{18}O$ and ±1.0‰ for δD.

[14] The accumulation rate was determined from the thickness of the annual layers of the chemical species using a Dansgaard-Johnsen-type model, which assumes that ice thinning, from the vertical compression and longitudinal stress in the ice sheet, is proportional to burial [*Johnsen et al.*, 1992, 1995]. Although this method may introduce some error for the absolute accumulation rate, the uncertainty in relative accumulation rate between adjacent layers will be determined only by the accuracy of determination of the layer thickness.

[15] Records plotted on an age scale were determined from the annual cycles in concentration observed in ions measured in this study. Years were assigned if a near-synchronous (within the same determined layer) peak was observed in calcium, chloride, nitrate, sodium, and sulphate, considered to be the most robust and reliable indicators of seasonal deposition (Figure 2). The resolution was such that annual layers contained an average of 7 data points during the stadial and 11 during the interstadial leading to an estimated error, either as a result of damaged ice or insufficient liquid sample for analysis, of just ±3 years (between depths 2068.60 m and 2071.30 m) [*Thomas et al.*, 2008]. A reference date of 38,167 years B.P. (1950) from the GICC05 age scale [*Andersen et al.*, 2006], coinciding

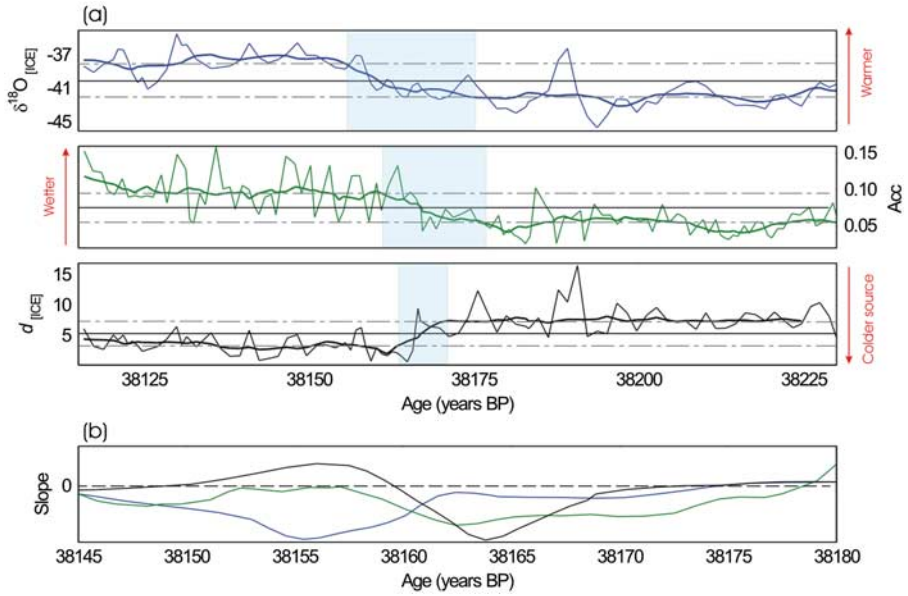

Figure 3. (a) δ^{18}O (blue curve), accumulation rate (Acc, green curve), and deuterium excess (d, black curve), plotted on an age scale (determined from annual layer counting with a reference point from GICC05 [*Andersen et al.*, 2006]), during the warming transition into DO-8. The δ^{18}O and d were measured at 20-mm resolution, and the accumulation rate was plotted at an annual resolution. Running decadal averages are plotted in thick curves on all graphs with the mean shown by a solid horizontal line and ±1 standard deviation above and below this plotted with dash-dotted horizontal lines. Blue shading indicates the transition. (b) Slopes of the decadal averaged records for δ^{18}O (blue curve), accumulation rate (green curve), and deuterium excess (black curve) for an expanded section across the transition. Note that increases in a quantity with time (decreasing age) appear as negative slopes; deuterium excess is plotted in reverse to allow comparison.

with the deuterium excess transition at 2069.96 m, was used to center the dating.

3. Results

[16] The warming transition at the onset of DO-8 is observed as a shift to less negative values in δ^{18}O and δD, an increase in accumulation, and a negative shift in the deuterium excess (d) (Figure 3). Despite the clarity of the transition in the context of the long-term signal, it is difficult to determine the precise timing of the transition because at the very short timescales considered here, the noise in the signal on either side of the transition is still significant compared to the size of the transition itself. We used two methods to estimate the onset and termination of the warming transition in each measured parameter.

[17] The transition is observed as an abrupt gradient change as the isotopic ratios and accumulation increase (decrease in d). In the first method therefore the onset and termination of the transitions were determined from the first derivatives (slope) of the decadal averaged records, shown in Figure 3b. Moving forward in time (decreasing age), increased δ^{18}O and accumulation appear as negative slopes (deuterium excess is plotted in reverse to allow comparison) and thus the onset of the transition is the point at which the slope falls below zero (above zero for d), while the end of the transition is the point at which the slope reaches a minimum (maximum for d) before returning to zero.

[18] As an alternative approach, the mean and standard deviation of the record was determined (Figure 3a). The point at which the decadal average exceeded one standard deviation above the record mean was determined as the termination and vice versa for the onset. This method is somewhat arbitrary, because the exact mean and standard deviation depend on the length of the record used before or after the transition, but nonetheless it succeeds in identifying the transition. Both methods produced comparable ages with the difference in the age of the onset of ±1.5, ±1.1 and ±0.3 years for δ^{18}O (δD), accumulation rate and d compared to ±0.6, ±1.2 and ±0.6 years at the termination. The annual and decadal records of δ^{18}O, accumulation rate, and d during the warming transition are plotted in Figure 3a with the duration of the transition (from the average of the two methods) highlighted in blue. It should be noted that the use of the decadal averages to determine the transition period is robust in distinguishing the transition from noise but tends to underplay the speed of extremely abrupt, subdecadal changes such as those observed in d.

[19] The onset of the warming transition in δ^{18}O and δD occurs at an age of 38,176 years B.P. (2070.14 m) and terminates at 38,155 years B.P. (2069.64 m), considered the start of DO-8. The warming lasts just 21 years, and during this period the values of δ^{18}O (δD) increase from a stadial average of 42.1 ± 1.9‰ (329.4 ± 15.9‰ in δD) to an interstadial average of 38.3 ± 1.5‰ (297.6 ± 12.9‰ in δD), increasing at a rate of 0.15‰ a^{-1}. This would equate to a temperature increase of 11.4°C, if a glacial calibration

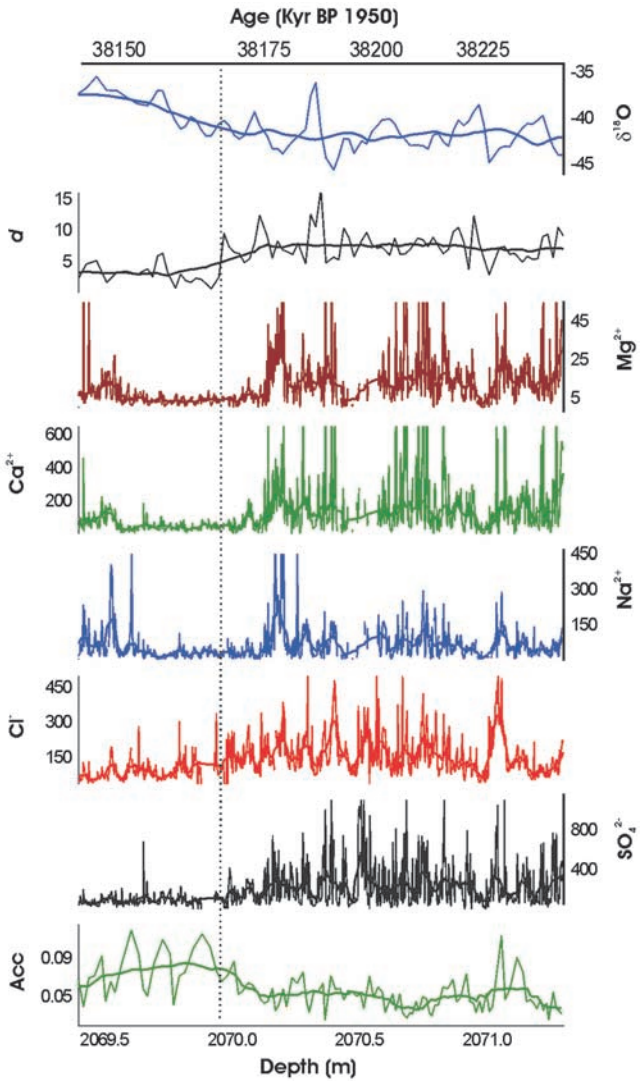

Figure 4. The warming transition. Shown, from top to bottom, are $\delta^{18}O$, deuterium excess, magnesium, calcium, sodium, chloride, sulphate, all in ppb at 2-mm resolution (thin curves) and approximate annual resolution (thick curves, based on an average annual layer thickness of 18 mm from annual layer counting), and accumulation rate. Vertical dashed line indicates the abrupt transition in deuterium excess (d) at 38,167 years B.P.

(0.33‰/°C) [*Cuffey et al.*, 1995]) is used, consistent with isotope-calibrated rates for other DO events [*Huber et al.*, 2006].

3.1. Deuterium Excess

[20] The deuterium excess (d) is a measure of the relative proportions of ^{18}O and ^{2}H contained in water ($d = \delta D - 8\ \delta^{18}O$ [*Dansgaard*, 1964]), regarded as a residual that relates the slight differences between D and ^{18}O to SSTs and sea surface humidity at the initial evaporation site [*Jouzel et al.*, 1997, 2007; *Petit et al.*, 1991]. The transition in d is more rapid than $\delta^{18}O$ and δD, observed in the decadal record as an 8-year transition starting at 38,171 years B.P. The transition in the 20-mm record, however, occurs in a single step lasting just 1 year at 38,167 years B.P. (2069.96 m), where values decrease by 4.4‰. The drop is in antiphase to $\delta^{18}O$ and δD, indicating a shift to colder source waters with reduced evaporation. This antiphase behavior has been observed to be common in all Greenland interstadials [*Steffensen et al.*, 2008; *Jouzel et al.*, 2007; *Masson-Delmotte et al.*, 2005a].

[21] Despite the antiphase behavior across the transition itself, the d record in the period after the warming transition (GIS 8), is positively correlated to $\delta^{18}O$ and δD ($r^2 = 0.57$, $n = 44 > 98\%$ confidence) but only weakly so in the period prior to it (GS8) ($r^2 = 0.2$, $n = 115$, >98% confidence), where large and abrupt changes are observed indicating instability in both atmospheric and oceanic source conditions at this time.

3.2. Accumulation Rate

[22] By using the seasonal signal of each ion, it was possible to count (independently of the GICC05 dating) the number of annual layers in and around the transition and to determine the layer thickness of each annual layer (shown in Figure 2). The annual layer count obtained in this way showed good agreement with the elapsed time in the new GICC05 age scale [*Andersen et al.*, 2006; *Thomas et al.*, 2008]. The annual layer thickness increased from an average of 15 mm during the stadial to 22 mm during the interstadial, representing an increase in accumulation rate (m ice a^{-1}) of 47%. The start of the accumulation transition (the end of GS8) was determined to be at 38,178 years B.P. (2070.17 m), 2 years earlier than the $\delta^{18}O$ transition. The duration of the accumulation transition seems to be shorter than that of $\delta^{18}O$ with rates stabilized 7 years earlier (Figure 3); inspection of the 20-mm data in Figure 3 confirms that the accumulation rate does appear to settle at its new higher value before the values of $\delta^{18}O$ reach their interstadial level.

3.3. Chemistry

[23] The subannual resolution chemical record during the rapid warming at the onset of DO-8 is compared with the change in $\delta^{18}O$, deuterium excess, and accumulation in Figure 4, plotted on a depth and age scale for the section between 2069.4 and 2071.3 m. There is a large degree of annual and decadal variability during the stadial making it hard to determine a robust transition point. The stadial record culminates in a decade-long (2070.25–2070.12 m) period of elevated deposition in Ca^{2+}, Mg^{2+}, SO_4^{2-}, and Na^{+}, after which concentrations drop to interstadial levels. A brief (2 year) increase is observed in all ions between 2070.08 and 2070.05 m, and it is only after this time that the transition can be considered complete. However, some ions that have a major component of marine origin (Cl^{-} and SO_4^{2-}) do not drop consistently below stadial levels until 2069.98 m, approximately coincident with the final shift in d (in the 20-mm record). The missing data in the later part of the chemical transition for the anions (Cl^{-}, Figure 4) makes a precise definition of the transition timing more difficult.

[24] All species show a significant decrease during the transition from cold conditions to relatively warm conditions. The largest changes are observed in Ca^{2+}, Mg^{2+} and SO_4^{2-}, decreasing by 70%, 70% and 57%, respectively, with a 59% decrease in sulphate defined as non–sea salt in

origin. The marine species show a less dramatic decrease with sodium dropping by 47%, while chloride was just 34% lower during the interstadial. Note that these values reflect only the changes estimated across the short and noisy sections shown in Figure 4 and are slightly less than the long-term changes deduced from longer records [*Mayewski et al.*, 1997; *Fuhrer et al.*, 1999]. Deposition changes of the continental ions occur in rapid steps, as observed previously for calcium in the GRIP record for this and other DO events [*Fuhrer et al.*, 1999]. The point at which the continental transition ends, and relatively constant interstadial concentrations are achieved, is coincident with the onset of the δ^{18}O transition and 21 years before temperatures stabilize.

4. Discussion

[25] Although the large uncertainties in determining the exact timing of the transition lead to some uncertainties that should not be overlooked, there appears to be clear evidence of a transition in the chemical record (continental and marine) before there is any obvious sign of a shift in other parameters. The start of the transition in accumulation rate and temperature (inferred from δ^{18}O and δD) is very close, and the abrupt deuterium excess drop occurs in the middle of their transition. Deuterium excess and accumulation rate stabilize at new interstadial values a few years before the δ^{18}O and δD (temperature). The total period from the apparent chemistry change to the final stabilization of the δ^{18}O and δD signal is 26 years.

4.1. Deuterium Excess

[26] The δ^{18}O and δD can be considered primarily to reflect the local temperature at Summit, Greenland, but with some influence from the conditions at the source (on these timescales, changes in seawater composition should be minimal). The deuterium excess, related to SSTs and sea surface humidity at the initial evaporation site [*Jouzel et al.*, 1982; *Petit et al.*, 1991], has been used from the GRIP record to reconstruct millennial-scale changes in source temperatures. The observed antiphase relationship between source (*d*) and site (δ^{18}O) temperature has been attributed to drastic changes in the location of the polar front [*Masson-Delmotte et al.*, 2005b; *Jouzel et al.*, 2007].

[27] The interstadials were characterized by North Atlantic circulation patterns that appear to resemble present conditions [*Rasmussen and Thomsen*, 2004; *Schmidt et al.*, 2006], with relatively warm surface waters flowing north to the Nordic seas where it sinks and returns to the North Atlantic as deep water. However, during the cold stadials, the northward flow of warm, saline surface waters into the subpolar North Atlantic was reduced, and the sites of deepwater formation migrated south [*Schmidt et al.*, 2006]. The associated expansion of sea ice cover and the more southerly location of the polar front, resulted in a southward shift in the NGRIP moisture source, with evaporation maintained in the warmer tropical and subtropical oceans [*Masson-Delmotte et al.*, 2005b; *Jouzel et al.*, 2007]. Therefore the decrease in *d* values from our NGRIP record, indicating a shift to colder source conditions, was probably the result of a northward shift in the polar front thus returning the NGRIP moisture source location to the cold but ice-free Greenland-Norwegian Sea. This mechanism is consistent with temperature and salinity profiles of marine sediments [*Cortijo et al.*, 1997; *Vidal et al.*, 1997] and SST reconstructions from GRIP *d*, which indicate a northward shift in the moisture source of 5° latitude during rapid changes [*Masson-Delmotte et al.*, 2005b].

[28] The warming in Greenland itself appears to be largely a result of these changes because the signals of this migration are observed near the start of the warming, at a time when temperatures in Greenland are still very cold. Indeed it is only after completion of the *d* transition, when the moisture source location is believed to have shifted north, that the warming appears to accelerate. The decadal rate of increase in δ^{18}O more than doubles from 0.13‰ a^{-1} before 38,167 years B.P. (the date of the *d* transition in the 20-mm record) to 0.28‰ a^{-1} after it. This enhanced warming is consistent with a transfer of oceanic heat to the atmosphere after the insulating sea ice has been removed.

4.2. Chemistry

[29] The evidence that the deposition of continental ions to Greenland has reached interstadial levels 21 years earlier than the δ^{18}O and δD indicates atmospheric source or circulation alterations also preceded the temperature increase, and were not caused by it. The decline in continental ions (Ca^{2+} and Mg^{2+}), indicative of long-range atmospheric loading and dustiness from East Asia [*Biscaye et al.*, 1997; *Svensson et al.*, 2000; *Bory et al.*, 2003], suggest a shift in the circulation pattern that transport dust to Greenland or alterations at the source location.

[30] Model studies suggest there were strong northwesterly winds over Greenland during stadials, with a split jet stream primarily due to a blocking anticyclone over the Laurentide ice sheet [*Bromwich et al.*, 2004]. The reduction in continental ions at the start of the transition could indicate a change in the jet stream location or a reduction in strength thus affecting the wind stress, shown to have an important impact on water masses during their passage in the North Atlantic current [*Orvik and Skagseth*, 2003].

[31] Studies of ice core particle size [*Ruth et al.*, 2003], however, suggest that changes in transport strength are probably not the main cause of the change in chemical concentrations. We also note that the change in chemistry precedes the shift in moisture source in our record, represented by the *d* transition that occurs some years later.

[32] Another explanation is a reduction in residence time and dust mobilization at the Asian source [*Fuhrer et al.*, 1999], as a result of lower local wind speeds or reduced uplift. Paleoclimate records from the Pacific and Indian oceans have been used to show an increase in the East Asian [*Wang et al.*, 2001] and the Indian [*Schulz et al.*, 1998; *Altabet et al.*, 2002] summer monsoon intensity, synchronous with interstadials in Greenland. The rapid decrease in continental ions could therefore be a result of increased strength of the Asian monsoon thus reducing dust mobility at the source as conditions move from dry to wet, the speed of which is consistent with speleothem records from the Indian ocean during DO-12 (similar amplitude to DO-8) [*Burns et al.*, 2003].

[33] The fluctuations in marine species (Cl^- and SO_4^{2-}) could be important with regard to the ocean circulation and sea ice retreat during the warming transition. The drop to interstadial levels appears to occur some time after the

continental ions (supporting the proposed mechanism that it is source rather than circulation changes causing the continental decrease), close to coincident with the shift in *d*. However, the degree of annual and decadal variation in our records makes it difficult to draw conclusions as to the exact transition point. Similar moderate changes in sodium have also been reported during the Bølling and Allerød periods from the NGRIP ice core suggesting a gradual change in sea ice conditions during abrupt glacial warmings [*Steffensen et al.*, 2008]

[34] Our NGRIP record provides a range of proxies that are free of relative dating uncertainty and therefore provides a true sequence of change during a Dansgaard-Oeschger warming. It is evident in our record that the warming in Greenland was preceded by both atmospheric and oceanic source and circulation changes. The simplest narrative would be that changes in atmospheric circulation (continental ions) and a reduction in the northwesterly winds, led to a major reduction in sea ice (*d* and marine-derived ions (Cl^- and SO_4^{2-})); this in turn allowed Greenland to warm to its new state. This sequence of events might even suggest a remote trigger from outside of the North Atlantic.

[35] The combined records from the tropical Atlantic [*Hughen et al.*, 1996; *Peterson et al.*, 2000], the tropical Pacific [*Bard et al.*, 1997; *Stott et al.*, 2002], the Indian Ocean [*Burns et al.*, 2003], and Asia [*Wang et al.*, 2001] support the theory that DO events coincided with a warming of the tropical warm pool and have been proposed not as a result of high-latitude warming but as a tropical precursor to DO events [*Hendy et al.*, 2002; *Broecker*, 2003; *Denton et al.*, 2005]. Indeed, model results have shown that SST patterns in the tropical Pacific can influence the behavior of the Atlantic THC through their impact on the freshwater budget [*Thorpe et al.*, 2001]. The evidence that a reduction in Asian dust mobilization and uplift, related to local wind and monsoon conditions, also occurred when temperatures in Greenland were cold add weight to the argument that large alterations were occurring at lower latitudes before the warming in Greenland.

[36] A recent study by *Brown et al.* [2007] of sediment cores from Lake Malawi, East Africa, reveal millennial-scale events which strongly resemble, but precede, corresponding DO events in Greenland. The authors conclude that the tropical component of the bipolar seesaw [*Stocker and Johnsen*, 2003] plays a major role in setting the abruptness and amplitude of DO events. In addition, *Knorr and Lohmann* [2003] proposed that warmer southern oceans triggered the interstadials in the northern hemisphere. A three-dimensional ocean circulation model was used to show that an abrupt resumption of the interstadial mode of the THC is observed when the southern ocean is warmed gradually. The southern ocean warming and reduction in sea ice extent enhances mass transport into the Atlantic, via the warm Indian Ocean and the cold Pacific route, resulting in increased northern hemisphere temperatures and reduced southern hemisphere temperatures [*Knorr and Lohmann*, 2003]. This is observed in the Antarctic warmings that precede at least the longest DO events (including DO-8) [*Blunier and Brook*, 2001].

[37] This mechanism would fit with the interpretation of the NGRIP data that the reduction in continental deposition prior to the warming in Greenland is connected with alterations at lower latitudes leading to large changes in atmospheric and oceanic circulation.

4.3. Can We Generalize Our Findings to Other DO Events?

[38] It is important to remember that this study is based on just one DO event. The magnitude and direction of changes in all the proxies we have discussed is similar in other DO events, but information on the phasing between proxies at decadal and shorter timescales is sparse. A study of the phasing between Na^+, Ca^{2+} and $\delta^{18}O$ in the DO events of the last 50 ka did suggest that, on average, during DO warmings, the chemistry showed changes before Greenland temperature, consistent with our finding for DO-8 [*Mogensen and Johnsen*, 2002]. However, the resolution of that study was probably inadequate to reliably determine such phasing. The phasing between Ca^{2+} and $\delta^{18}O$ at the start of DO-3 and DO-20 in the GRIP core has been shown before at sufficiently high resolution [*Fuhrer et al.*, 1999]; in both cases, significant changes in Ca^{2+} appear to occur right at the start of the $\delta^{18}O$ transition, but it is less clear that the change in Ca^{2+} is completed before Greenland temperature starts to change. A recent study of the Bølling and Allerød warm periods, at the end of the last glacial period, reveals a similar sequence of events with dust deposition preceding the temperature increase in Greenland [*Steffensen et al.*, 2008]. However, preliminary indications from a study of phasing between chemical measures and $\delta^{18}O$ at NGRIP suggest that the apparent phasing may be variable between different events, so the conclusions of this study of one event should be treated with caution.

5. Conclusions

[39] The most striking finding in our high-depth-resolution NGRIP record is the order of events. The continental ions appear to be the first to change, followed by the snow accumulation and the moisture source conditions and finally the atmospheric temperature in Greenland. The proposed mechanism for this sequence of events is an atmospheric and oceanic circulation change that originates outside of the North Atlantic.

[40] Whether such a sequence of events and consequences is plausible could now be tested in a model environment, where the sequence expected from other hypothetical triggers might also be tested. While it will be difficult to bring other proxy data into the same precise relative timing alignment, at least the magnitude and style of changes in other paleorecords can also be part of the model testing procedure.

[41] Finally, we note the caution that, given the noise on the records at DO-8, making it hard to be certain of the timing of different events, it would be desirable to carry out similar studies on other Greenland interstadials, in order to establish a more general climatology of such transitions.

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References

Altabet, M. A., M. J. Higginson, and D. W. Murray (2002), The effect of millennial-scale changes in Arabian Sea denitrification on atmospheric CO2, *Nature*, *415*, 159–162, doi:10.1038/415159a.

Andersen, K. K., et al. (2006), The Greenland Ice Core Chronology 2005, 15–42 kyr: part 1. Constructing the time scale, *Quat. Sci. Rev.*, *25*, 3246–3257, doi:10.1016/j.quascirev.2006.08.002.

Bard, E., F. Rostek, and C. Sonzogni (1997), Interhemispheric synchrony of the last deglaciation inferred from alkenone palaeothermometry, *Nature*, *385*, 707–710, doi:10.1038/385707a0.

Behl, R. J., and J. P. Kennett (1996), Brief interstadial events in the Santa Barbara basin, NE Pacific, during the past 60 kyr, *Nature*, *379*, 243–246, doi:10.1038/379243a0.

Biscaye, P. E., F. E. Grousset, M. Revel, S. Van der Gaast, G. A. Zielinski, A. Vaars, and G. Kukla (1997), Asian provenance of glacial dust (stage 2) in the Greenland Ice Sheet Project 2 Ice Core, Summit, Greenland, *J. Geophys. Res.*, *102*, 26,765–26,781, doi:10.1029/97JC01249.

Blunier, T., and E. J. Brook (2001), Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period, *Science*, *291*, 109–112.

Bond, G., W. Broecker, S. Johnsen, J. McManus, L. Labeyrie, J. Jouzel, and G. Bonani (1993), Correlations between climate records from North Atlantic sediments and Greenland ice, *Nature*, *365*, 143–147, doi:10.1038/365143a0.

Bory, A. J.-M., P. E. Biscaye, and F. E. Grousset (2003), Two distinct seasonal Asian source regions for mineral dust deposited in Greenland (NorthGRIP), *Geophys. Res. Lett.*, *30*(4), 1167, doi:10.1029/2002GL016446.

Broecker, W. S. (1997), Thermohaline circulation, the Achilles heel of our climate system: Will man-made CO_2 upset the current balance?, *Science*, *278*, 1582–1588, doi:10.1126/science.278.5343.1582.

Broecker, W. S. (2003), Does the trigger for abrupt climate change reside in the ocean or in the atmosphere?, *Science*, *300*(5625), 1519–1522, doi:10.1126/science.1083797.

Broecker, W. S., D. M. Peteet, and D. Rind (1985), Does the ocean-atmosphere system have more than one stable mode of operation?, *Nature*, *315*, 21–26, doi:10.1038/315021a0.

Bromwich, D. H., E. R. Toracinta, H. Wei, R. J. Oglesby, J. L. Fastook, and T. J. Hughes (2004), Polar MM5 simulations of the winter climate of the Laurentide Ice Sheet at the LGM, *J. Clim.*, *17*, 3415–3433, doi:10.1175/1520-0442(2004)017<3415:PMSOTW>2.0.CO;2.

Brook, E. J., T. Sowers, and J. Orchardo (1996), Rapid variations in atmospheric methane concentrations during the past 110,000 years, *Science*, *273*, 1087–1093, doi:10.1126/science.273.5278.1087.

Brown, E. T., T. C. Johnson, C. A. Scholz, A. S. Cohen, and J. W. King (2007), Abrupt change in tropical African climate linked to the bipolar seesaw over the past 55,000 years, *Geophys. Res. Lett.*, *34*, L20702, doi:10.1029/2007GL031240.

Burns, S. J., D. Fleitmann, A. Matter, J. Kramers, and A. A. Al-Subbary (2003), Indian Ocean climate and an absolute chronology over Dansgaard/Oeschger events 9 to 13, *Science*, *301*, 1365–1367, doi:10.1126/science.1086227.

Cortijo, E., L. Labeyrie, L. Vidal, M. Vautravers, M. Chapman, J.-C. Duplessy, M. Elliot, M. Arnold, J.-C. Turon, and G. Auffret (1997), Changes in sea surface hydrology associated with Heinrich event 4 in the North Atlantic Ocean between 40° and 60°N, *Earth Planet. Sci. Lett.*, *146*, 29–45, doi:10.1016/S0012-821X(96)00217-8.

Cuffey, K. M., G. D. Clow, R. B. Alley, M. Stuiver, E. D. Waddington, and R. W. Saltus (1995), Large Arctic temperature-change at the Wisconsin-Holocene glacial transition, *Science*, *270*, 455–458, doi:10.1126/science.270.5235.455.

Curry, W. B., and D. W. Oppo (1997), Synchronous, high-frequency oscillations in tropical sea surface temperatures and North Atlantic Deep Water production during the last glacial cycle, *Paleoceanography*, *12*, 1–14, doi:10.1029/96PA02413.

Dansgaard, W. (1964), Stable isotopes in precipitation, *Tellus*, *16*, 436–467.

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

De Angelis, M., J. P. Steffensen, M. Legrand, H. Clausen, and C. Hammer (1997), Primary aerosol (sea salt and soil dust) deposited in Greenland ice during the last climatic cycle: Comparison with east Antarctic records, *J. Geophys. Res.*, *102*, 26,681–26,698, doi:10.1029/97JC01298.

Denton, G. H., R. B. Alley, G. C. Comer, and W. S. Broecker (2005), The role of seasonality in abrupt climate change, *Quat. Sci. Rev.*, *24*, 1159–1182, doi:10.1016/j.quascirev.2004.12.002.

EPICA Community Members (2006), One-to-one coupling of glacial climate variability in Greenland and Antarctica, *Nature*, *444*, 195–198, doi:10.1038/nature05301.

Flückiger, J., T. Blunier, B. Stauffer, J. Chappellaz, R. Spahni, K. Kawamura, J. Schwander, T. F. Stocker, and D. Dahl-Jensen (2004), N2O and CH4 variations during the last glacial epoch: Insight into global processes, *Global Biogeochem. Cycles*, *18*, GB1020, doi:10.1029/2003GB002122.

Fuhrer, K., E. W. Wolff, and S. J. Johnsen (1999), Timescales for dust variability in the Greenland Ice Core Project (GRIP) ice core in the 100,000 years, *J. Geophys. Res.*, *104*, 31,043–31,052, doi:10.1029/1999JD900929.

Gherardi, J. M., L. Labeyrie, J. F. McManus, R. Francois, L. C. Skinner, and E. Cortijo (2005), Evidence from the Northeastern Atlantic basin for variability in the rate of the meridional overturning circulation through the last deglaciation, *Earth Planet. Sci. Lett.*, *240*, 710–723, doi:10.1016/j.epsl.2005.09.061.

Grootes, P. M. (1993), Comparison of the oxygen isotope records from GISP2 and GRIP Greenland ice cores, *Nature*, *366*, 552–554, doi:10.1038/366552a0.

Heinrich, H. (1988), Origin and consequences of cyclic ice rafting in the northeast Atlantic ocean during the past 130,000 years, *Quat. Res.*, *29*, 142–152, doi:10.1016/0033-5894(88)90057-9.

Hendy, I. L., and J. P. Kennett (1999), Latest Quaternary North Pacific surface-water responses imply atmosphere-driven climate instability, *Geology*, *27*, 291–294, doi:10.1130/0091-7613(1999)027<0291:LQNPSW>2.3.CO;2.

Hendy, I. L., J. P. Kennett, E. B. Roark, and B. L. Ingram (2002), Apparent synchronicity of submillennial scale climate events between Greenland and Santa Barbara Basin, California from 30–10 ka, *Quat. Sci. Rev.*, *21*, 1167–1184, doi:10.1016/S0277-3791(01)00138-X.

Huber, C., M. Leuenberger, R. Spahni, J. Flückiger, J. Schwander, T. F. Stocker, S. Johnsen, A. Landais, and J. Jouzel (2006), Isotope calibrated Greenland temperature record over Marine Isotope Stage 3 and its relation to CH4, *Earth Planet. Sci. Lett.*, *243*, 504–519, doi:10.1016/j.epsl.2006.01.002.

Hughen, K. A., J. T. Overpeck, L. C. Peterson, and S. Trumbore (1996), Rapid climate changes in the tropical Atlantic region during the last deglaciation, *Nature*, *380*, 51–54, doi:10.1038/380051a0.

Johnsen, S. J., H. B. Clausen, W. Dansgaard, K. Fuhrer, N. Gundestrup, C. U. Hammer, P. Iversen, J. Jouzel, B. Stauffer, and J. P. Steffensen (1992), Irregular glacial interstadials recorded in a new Greenland ice core, *Nature*, *359*, 311–313, doi:10.1038/359311a0.

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

Jouzel, J., L. Merlivat, and C. Lorius (1982), 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.

Jouzel, J., et al. (1997), Validity of the temperature reconstruction from water isotopes in ice cores, *J. Geophys. Res.*, *102*, 26,471–26,487, doi:10.1029/97JC01283.

Jouzel, J., M. Stievenard, A. Landais, V. Masson-Delmotte, A. Sveinbjornsdottir, F. Vimeux, U. von Grafenstein, and J. W. C. White (2007), The GRIP deuterium-excess record, *Quat. Sci. Rev.*, *26*, 1–17, doi:10.1016/j.quascirev.2006.07.015.

Keigwin, L. D., and G. A. Jones (1994), Western North Atlantic evidence for millennial-scale changes in ocean circulation and climate, *J. Geophys. Res.*, *99*, 12,397–12,410, doi:10.1029/94JC00525.

Knorr, G., and G. Lohmann (2003), Southern Ocean origin for the resumption of Atlantic thermohaline circulation during deglaciation, *Nature*, *424*, 532–536, doi:10.1038/nature01855.

Knutti, R., J. Fluckiger, T. F. Stocker, and A. Timmermann (2004), Strong hemispheric coupling of glacial climate through freshwater discharge and ocean circulation, *Nature*, *430*, 851–856, doi:10.1038/nature02786.

Lang, C., M. Leuenberger, J. Schwander, and S. Johnsen (1999), 16°C rapid temperature variation in central Greenland 70,000 years ago, *Science*, *286*, 934–937, doi:10.1126/science.286.5441.934.

Legrand, M., C. Hammer, M. DeAngelis, J. Savarino, R. Delmas, H. Clausen, and S. J. Johnsen (1997), Sulfur-containing species (methanesulfonate and SO4) over the last climatic cycle in the Greenland Ice Core Project (central Greenland) ice core, *J. Geophys. Res.*, *102*, 26,663–26,679, doi:10.1029/97JC01436.

Masson-Delmotte, V., et al. (2005a), Holocene climatic changes in Greenland: Different deuterium excess signals at Greenland Ice Core Project (GRIP) and NorthGRIP, *J. Geophys. Res.*, *110*, D14102, doi:10.1029/2004JD005575.

Masson-Delmotte, V., J. Jouzel, A. Landais, M. Stievenard, S. J. Johnsen, J. W. C. White, M. Werner, A. Sveinbjornsdottir, and K. Fuhrer (2005b), GRIP deuterium excess reveals rapid and orbital-scale changes in Greenland moisture origin, *Science*, *309*, 118–121, doi:10.1126/science.1108575.

Mayewski, P. A., et al. (1994), Changes in atmospheric circulation and ocean ice cover over the North Atlantic during the last 41,000 years, *Science*, *263*, 1747–1751, doi:10.1126/science.263.5154.1747.

Mayewski, P. A., L. D. Meeker, M. S. Twickler, S. I. Whitlow, Q. Yang, W. B. Lyons, and M. Prentice (1997), Major features and forcing of high-latitude northern hemisphere atmospheric circulation using a 110,000-year-long glaciochemical series, *J. Geophys. Res.*, *102*, 26,345–26,366, doi:10.1029/96JC03365.

Mogensen, I. A., and S. J. Johnsen (2002), Rapid climate changes during the last glacial, *Ann. Glaciol.*, *35*, 398–402, doi:10.3189/172756402781816654.

North Greenland Ice Core Project Members (NGRIP Members) (2004), High-resolution record of Northern Hemisphere climate extending into the last interglacial period, *Nature*, *431*, 147–151, doi:10.1038/nature02805.

Orvik, K. A., and Ø. Skagseth (2003), The impact of the wind stress curl in the North Atlantic on the Atlantic inflow to the Norwegian Sea toward the Arctic, *Geophys. Res. Lett.*, *30*(17), 1884, doi:10.1029/2003GL017932.

Peterson, L. C., G. H. Haug, K. A. Hughen, and U. Rohl (2000), Rapid changes in the hydrologic cycle of the tropical Atlantic during the last glacial, *Science*, *290*, 1947–1951, doi:10.1126/science.290.5498.1947.

Petit, J. R., J. W. C. White, N. W. Young, J. Jouzel, and Y. S. Korotkevich (1991), Deuterium excess in recent Antarctic snow, *J. Geophys. Res.*, *96*, 5113–5122, doi:10.1029/90JD02232.

Rasmussen, T. L., and E. Thomsen (2004), The role of the North Atlantic Drift in the millennial timescale glacial climate fluctuations, *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, *210*, 101–116, doi:10.1016/j.palaeo.2004.04.005.

Rousseau, D.-D., G. Kukla, and J. McManus (2006), What is what in the ice and the ocean?, *Quat. Sci. Rev.*, *25*, 2025–2030, doi:10.1016/j.quascirev.2006.03.009.

Ruth, U. D., J. P. Wagnbach, J. P. Steffensen, and M. Bigler (2003), Continuous record of microparticle concentration and size distribution in the central Greenland NGRIP ice core during the last glacial period, *J. Geophys. Res.*, *108*(D3), 4098, doi:10.1029/2002JD002376.

Schmidt, M. W., M. J. Vautravers, and H. J. Spero (2006), Rapid subtropical North Atlantic salinity oscillations across Dansgaard-Oeschger cycles, *Nature*, *443*, 561–564, doi:10.1038/nature05121.

Schulz, H., U. von Rad, and H. Erlenkeuser (1998), Correlation between Arabian Sea and Greenland climate oscillations of the past 110,000 years, *Nature*, *393*, 54–57.

Severinghaus, J. P., and E. J. Brook (1999), Abrupt climate change at the end of the last glacial period inferred from trapped gasses in polar air, *Science*, *286*, 930–934, doi:10.1126/science.286.5441.930.

Severinghaus, J. P., T. Sowers, E. J. Brook, R. B. Alley, and M. L. Bender (1998), Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice, *Nature*, *391*, 141–146, doi:10.1038/34346.

Shackleton, N. J., M. A. Hall, and E. Vincent (2000), Phase relationships between millennial-scale events 64,000–24,000 years ago, *Paleoceanography*, *15*, 565–569, doi:10.1029/2000PA000513.

Steffensen, J. P., et al. (2008), High-resolution Greenland ice core data show abrupt climate change happens in few years, *Science*, *321*, 680–684, doi:10.1126/science.1157707.

Stocker, T. F., and S. J. Johnsen (2003), A minimum thermodynamic model for the bipolar seesaw, *Paleoceanography*, *18*(4), 1087, doi:10.1029/2003PA000920.

Stott, L., C. Poulsen, S. Lund, and R. Thunell (2002), Super ENSO and global climate oscillations at millennial time scales, *Science*, *297*, 222–226, doi:10.1126/science.1071627.

Svensson, A., P. E. Biscaye, and F. E. Grousset (2000), Characterization of late glacial dust in the GRIP, *J. Geophys. Res.*, *105*, 4637–4656, doi:10.1029/1999JD901093.

Taylor, K. C., C. U. Hammer, R. B. Alley, H. B. Clausen, D. Dahl-Jensen, A. J. Gow, N. S. Gundestrup, J. Kipfstuhl, J. C. Moore, and E. D. Waddington (1993), Electrical-conductivity measurements from the GISP2 and GRIP Greenland ice cores, *Nature*, *366*, 549–552, doi:10.1038/366549a0.

Thomas, E. R. (2006), High resolution analysis of rapid climate change from Greenland ice cores, Ph.D., Open Univ., Milton Keynes, UK.

Thomas, E. R., R. Mulvaney, and E. W. W. Wolff (2008), A change in seasonality in Greenland during a Dansgaard-Oeschger warming, *Ann. Glaciol.*, *48*, 19–24, doi:10.3189/172756408784700590.

Thorpe, R. B., J. M. Gregory, T. C. Johns, R. A. Wood, and J. F. B. Mitchell (2001), Mechanisms determining the Atlantic thermohaline circulation response to greenhouse gas forcing in a non-flux-adjusted coupled climate model, *J. Clim.*, *14*, 3102–3116, doi:10.1175/1520-0442(2001)014<3102:MDTATC>2.0.CO;2.

Vidal, L., L. Labeyrie, E. Cortijo, M. Arnold, J. C. Duplessy, E. Michel, S. Becqué, and T. C. E. van Weering (1997), Evidence for changes in the North Atlantic Deep Water linked to meltwater surges during the Heinrich events, *Earth Planet. Sci. Lett.*, *146*, 13–27, doi:10.1016/S0012-821X(96)00192-6.

Voelker, A. H. L., et al. (2002), Global distribution of centennial-scale records from Marine Isotope Stage (MIS) 3: A database, *Quat. Sci. Rev.*, *21*, 1185–1212, doi:10.1016/S0277-3791(01)00139-1.

Wang, Y. J., H. Cheng, R. L. Edwards, Z. S. An, J. Y. Wu, C. C. Shen, and J. A. Dorale (2001), A high-resolution absolute-dated Late Pleistocene monsoon record from Hulu Cave, China, *Science*, *294*, 2345–2348, doi:10.1126/science.1064618.

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