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Precise timing and characterization of abrupt climate change 8200 years ago from air trapped in polar ice

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Abstract

How fast and how much climate can change has significant implications for concerns about future climate changes and their potential impacts on society. An abrupt climate change 8200 years ago (8.2 ka event) provides a test case to understand possible future climatic variability. Here, methane concentration (taken as an indicator for terrestrial hydrology) and nitrogen isotopes (Greenland temperature) in trapped air in a Greenland ice core (GISP2) are employed to scrutinize the evolution of the 8.2 ka event. The synchronous change in methane and nitrogen implies that the 8.2 ka event was a synchronous event (within ± 4 years) at a hemispheric scale, as indicated by recent climate model results [Legrande, A. N., Schmidt, G. A., Shindell, D. T., Field, C. V., Miller, R. L., Koch, D. M., Faluvegi, G., Hoffmann, G., 2006. Consistent simulations of multiple proxy responses to an abrupt climate change event. Proceedings of the National Academy of Sciences 103, 837–842]. The event began with a large-scale general cooling and drying around $\sim 8175 \pm 30$ years BP (Before Present, where Present is 1950 AD). Greenland temperature cooled by $3.3 \pm 1.1\,^{\circ}C$ (decadal average) in less than ~20 years, and atmospheric methane concentration decreased by $\sim 80 \pm 25$ ppb over ~40 years, corresponding to a $15 \pm 5\%$ emission reduction. Hemispheric scale cooling and drying, inferred from many paleoclimate proxies, likely contributed to this emission reduction. In central Greenland, the coldest period lasted for ~60 years, interrupted by a milder interval of a few decades, and temperature subsequently warmed in several steps over ~70 years. The total duration of the 8.2 ka event was roughly 150 years.

1. Introduction

Future climatic change associated with anthropogenic greenhouse gases is a primary concern for human society. Climate models and paleoclimatic records suggest that future climate change may involve large and abrupt changes (Manabe and Stouffer, 1993; Stocker and Schmittner, 1997; IPCC, 2001; National Research Council (US). Committee on Abrupt Climate Change, 2002). One way to quantify possible impacts of future abrupt climate change amid model uncertainties is to look back to the past, and find the closest possible analog (National Research Council (US). Committee on Abrupt Climate Change, 2002; Alley and Agustsdottir, 2005). The abrupt climate change ~8200 years ago, the largest abrupt climatic event in the past 10,000 years (Fig. 1), occurred at a time when the background climate was much like the present (the Holocene, a relatively warm period after the last glacial period, Fig. 1). The event was characterized by generally cool, dry, and windy climate, which affected ecosystems and early human societies (Alley et al., 1997a; Weiss, 2000). The widespread evidence for the 8.2 ka event, combined with a rapid decrease of atmospheric methane concentration (Spahni et al., 2003; this study) suggests that the event was at least hemispheric or "near-global' in its geographical extent (Alley and Agustsdottir, 2005; Morrill and Jacobsen, 2005; Wiersma and Renssen, 2006). The cause of the event may have been the largest proglacial lake (Lake Agassiz) outburst of the last deglaciation (Clarke et al., 2004), which has been dated to 8470 ± 300 (1σ) BP (Barber et al., 1999; Rohling and Palike, 2005).

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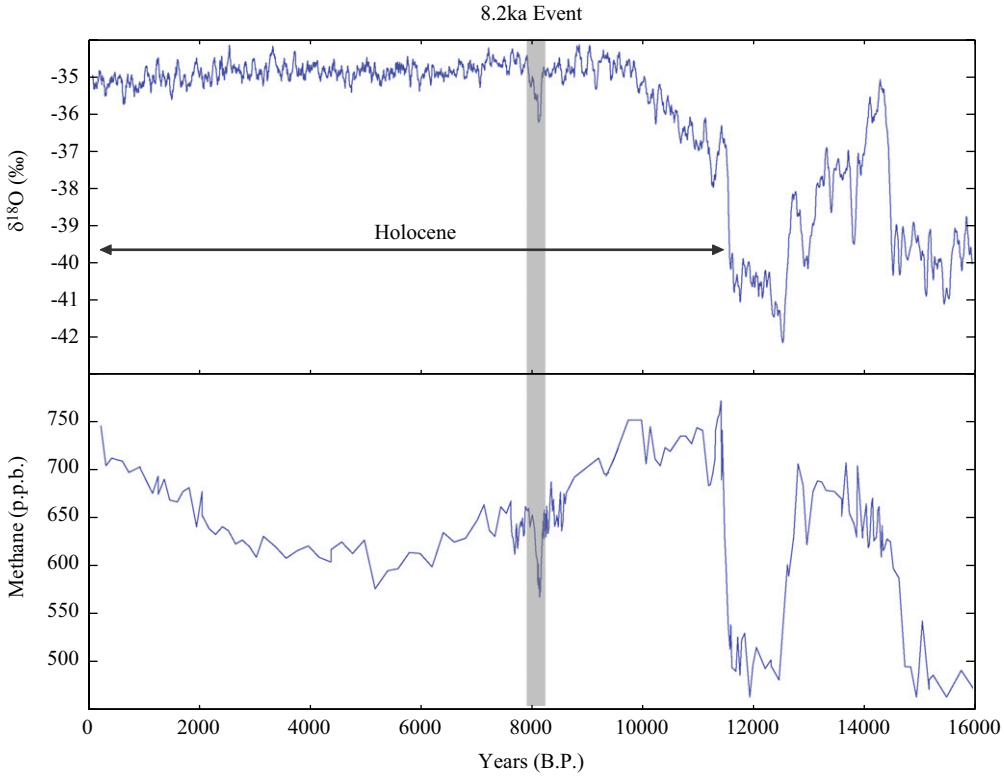

Fig. 1. Greenland ice core records covering the past 16,000 years (oxygen isotopes of ice and methane). Note that the 8.2 ka event punctuates the relatively warm and stable past ~11,000 years since the end of the last glaciation. (Top) GISP2 oxygen isotope record (Stuiver et al., 1995) with a 50-year moving average. The oxygen isotope record can be considered as a qualitative temperature record on this time scale. (Bottom) Methane concentration in ice cores. Owing to the different resolution and precision of methane records in GISP2 and GRIP, we used records from 175 to 7530 BP and 8762 to 9891 BP from GRIP (Blunier et al., 1995), and records from 7612 to 8617 BP (this study) and 10,153 to 16,507 BP from GISP2 (Brook et al., 2000).

This hypothesis holds that a massive outflow of fresh water ran out the Hudson Strait, to the North Atlantic, causing a slowdown of the meridional overturning circulation, which enabled wintertime sea ice cover to expand with consequent hemispheric cooling and drying especially surrounding the North Atlantic area (Alley and Agustsdottir, 2005; Ellison et al., 2006). An alternative hypothesis is that a millennial-scale cooling trend started a few centuries earlier than the 8.2 ka event (Rohling and Palike, 2005). A minor solar minimum coinciding with the 8.2 ka event (Muscheler et al., 2004) as a third hypothesis, may have forced the system to cross a threshold and may have triggered the 8.2 ka event (Bond et al., 1997, 2001).

In this study, we address the detailed timing and evolution of the 8.2k event by measuring nitrogen isotope ratios and methane concentration in trapped air in the GISP2 ice core. Atmospheric methane concentration can be viewed as a qualitative indicator of integrated terrestrial hydrological conditions in methane-producing regions, owing to the dominant methane source from wetland areas (~75% of natural emissions, Houweling et al., 2000). Nitrogen isotopes in trapped air in an ice core provide a signal of local temperature changes in Greenland (Severinghaus et al., 1998; Goujon et al., 2003; Landais et al., 2004). This method (Severinghaus et al., 1998; Severinghaus and Brook, 1999) provides an opportunity to precisely and directly assess the timing of abrupt climate change in Greenland with respect to changes in atmospheric methane, by comparing two gases in the same core. We also provide an improved estimate of the magnitude of the temperature change in central Greenland.

2. Materials and methods

We used the GISP2 ice core for our analyses. The resolution of nitrogen isotope data is 1 m (~10 year) from 1359.95 to 1458.95 m depth, corresponding to a gas age range of 7600–8600 BP (see below for the basis for this chronology). Replicate analyses (2–3 for each depth) were conducted for the entire record. Additional replicates were done in the intervals 1412.95–1416.95 m (6 replicates per depth) and 1417.97–1423.95 m (4 replicates per depth) to increase the confidence level of the temperature estimate for the 8.2 ka event. The total number of sampling depths is 96, and the total number of samples is 238. Ice samples were analyzed for nitrogen isotope ratios following the method described in Severinghaus et al. (2003) with some modifications described here. We (Kobashi et al., submitted for publication) developed a new method (“copper method”) for the simultaneous analysis of nitrogen and argon isotopes in air proportions, with oxygen removed from the air sample by exposure to hot copper. We used a

new mass spectrometer with 8 Faraday cups for simultaneous collection of the mass 28, 29, 36, and 40 beams (Finnigan Delta XP). For comparison purposes we also measured some argon data with a Finnigan MAT 252 (10 kV) by conventional methods as in Severinghaus et al. (2003).

Our sampling depths are located in the transition zone between air bubbles and air clathrates (Bender et al., 1995). Unusual gas loss fractionation can be seen in the Ar/N_2 data as $\delta Ar/N_2$ values are abnormally enriched due to the partial formation of clathrates combined with gas loss from the bubbles (Bender et al., 1995) (Fig. 2). The higher variability of $\delta^{40}Ar$ may be partially introduced by the poorer linearity of the Finnigan Delta XP. The $\delta^{40}Ar$ data measured with the conventional method and with a modified "copper method" employing the Finnigan MAT 252 show better precision (Kobashi et al., submitted for publication). As the precision of $\delta^{40}Ar$ is so low, we do not attempt to interpret the argon data here (Fig. 2). However, the magnitude of the argon isotope shift for the 8.2 event is consistent with our temperature estimate within its uncertainty (Fig. 2). An additional difference from the Severinghaus et al. (2003) method is that the N_2:Ar ratio is 83, the air ratio, rather than 10. This higher ratio may have also contributed to lower $\delta^{40}Ar$ precision. In order to obtain a strong argon beam, the samples were run at an elevated inlet pressure of 208 mbar. These modifications degraded the linearity somewhat, as expected for this (3 kV) mass spectrometer source.

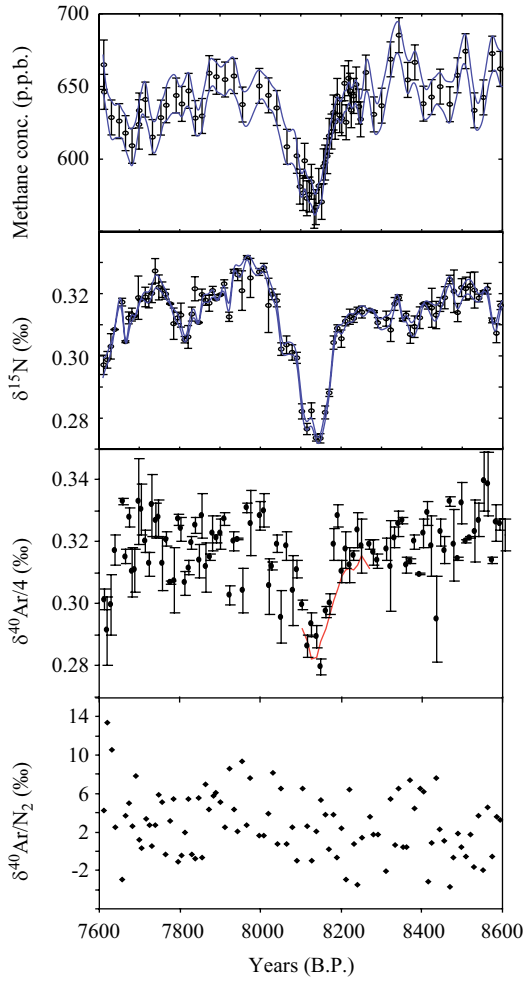

Fig. 2. Methane concentration, $\delta^{15}N$, $\delta^{40}Ar/4$, and $\delta^{40}Ar/N_2$ data. Circles represent average values. Error bars are 1σ. For methane and $\delta^{15}N$ data, blue lines are 1σ error range after spline fit (Enting, 1987) with 10-year cutoff period for $\delta^{15}N$ and 30-year cutoff period for methane. For $\delta^{40}Ar/4$ data, red line represents modeled $\delta^{40}Ar/4$ value (see text).

Two to six ~50 g ice samples from each depth are cut for isotopic analyses. The ice sample is placed in a glass flask with a Conflat flange and evacuated for 40 min while keeping the flask at −20 °C. Samples are then melted and warmed to room temperature (~22 °C) releasing the trapped air. The released air samples are transferred into a dip tube, cooled to 4 K using liquid helium. During the transfer, water vapor is removed by a glass water trap at −100 °C, and potential organic contaminants and CO_2 are removed by another glasstrap at −196 °C. Then, air passes through a copper mesh heated to 500 °C to remove the oxygen, which has an isotopic species ($^{18}O^{18}O$) with the same mass as ^{36}Ar. Finally, the last glass trap at −196 °C removes remaining CO_2 produced by the copper. Then, the air sample is introduced to the mass spectrometer for the measurement of argon and nitrogen isotopes, and Ar/N_2. The pooled standard deviation of all the replicated ice samples (samples taken from the same depth in the core) is 0.004‰ for $\delta^{15}N$. The estimated standard error of the mean of nitrogen isotope data ranges from 0.0028‰ to 0.0016‰ depending on the number of samples analyzed. Two data points were rejected owing to experimental reasons (large residual pressure after the gas extraction and a change in the location of the heated copper during the gas transfer).

For methane analyses, the resolution is 2 m (~20-year) from 1360.03 to 1460.03 m, except depths from 1410.03 to 1426.03 m, where resolution was increased to ~50 cm (~5-year). The method for methane analysis is described in Brook et al. (2000) with a slightly different extraction technique. We used a 6-port line with glass instead of stainless steel vessels, and a Shimadzu Gas chromatograph (GC). For each depth, we conducted duplicate analyses. Whenever the difference between the duplicates was more than 40 ppb we ran another set of duplicates, rejecting the first set. This occurred three times out of 77 duplicate runs. For two more samples one of the replicates was lost due to operator error. The mean difference between the replicates and the pooled standard deviation are 14 and 17 ppb, respectively. Twenty blanks were analyzed during the experiments, and the mean is 11.4 ± 8.4 ppb. The data agree with a lower resolution methane record (Spahni et al., 2003) for the 8.2k event from the GRIP ice core within its uncertainty.

The recent finding of a possible new methane source from terrestrial plants (Keppler et al., 2006) may change our understanding of the relative contribution of various atmospheric methane sources. As methane emissions from terrestrial plants are expected to positively correlate with

temperature, the significance of concentration changes in this study would not substantially change if a large plant methane source is verified.

In this study, we use the GISP2 visual-stratigraphic age scale (Alley et al., 1997b) for the ice age scale with minor modifications. By correlating the GRIP ^{10}Be ice-core data and tree ring ^{14}C, Muscheler et al. (2004) found that the minimum of the oxygen isotope record of ice for the 8.2 ka event occurred at an absolute age of ~8150 ± 20 years BP. To adjust our time scale to match this observation, the entire time scale is shifted by 110 years toward younger age. After this correction, we estimate that the uncertainty of the absolute ages of ice is ±30 years from 7600–8600 BP. However, relative age uncertainty, as applies to inferred duration in this study, should be much smaller (1%) (Alley et al., 1997b). The gas time scale is calculated from a firn densification model (Goujon et al., 2003). The gas–ice age difference (Δage) uncertainty is estimated to be ~10% (~20 years).

3. Evolution of Greenland temperature

3.1. Magnitude of Greenland cooling

The magnitude of temperature changes is often used as the most important climatic indicator and as a target for climate model simulations (LeGrande et al., 2006). There have been many attempts to quantify the magnitude of the temperature change for the 8.2 ka event in central Greenland. A conventional method uses oxygen isotopes of ice ($\delta^{18}O_{ice}$) calibrated by the empirical modern spatial relationship of temperature and $\delta^{18}O$ of precipitation. This slope of the temperature–isotope relationship is called the "spatial slope" or spatial sensitivity. Several studies, however, have shown that the temporal isotope–temperature sensitivity is quite variable and different from the spatial sensitivity (Cuffey et al., 1995; Jouzel et al., 1997). The 1.8‰ change of $\delta^{18}O_{ice}$ (decadal average) during the event in the GISP2 and GRIP ice cores corresponds to 2.7 °C cooling, using the spatial calibration (Johnsen et al., 1989). Alley et al. (1997a) calibrated the oxygen isotope record (GISP2) using the borehole-temperature estimate for the Holocene (Cuffey et al., 1995) assuming that the temperature sensitivity at a millennial time scale is valid for the decadal time scale, and estimated the magnitude of 8.2 ka cooling to be 6 ± 2 °C. Leuenberger et al. (1999) estimated the magnitude of the 8.2 ka cooling as 7.4 °C with a range of 5.4–11.7 °C using $\delta^{15}N$, a similar method as described here (see later discussion).

We used the nitrogen isotope data in trapped air in the GISP2 ice core (Fig. 2) to quantify the magnitude of temperature change in combination with a heat diffusion and firn densification model (Goujon et al., 2003). The nitrogen isotopic ratio ($^{15}N/^{14}N$) in the atmosphere is known to be constant for millions of years (Mariotti, 1983; Severinghaus et al., 1998). Thus, any isotopic deviations from atmosphere in trapped air in an ice core can be attributed to isotopic fractionation in the firn layer (porous top layer of ice sheet ~80 m thick). Two important processes cause fractionation in the firn, gravitational fractionation and thermal diffusion. Gravitational fractionation in the firn layer (Craig et al., 1988; Schwander, 1989) is mainly controlled by the thickness of the firn layer and the mass difference of the gas pair (typically $\delta^{15}N = \sim 0.31$‰ for our samples). Fractionation due to thermal diffusion occurs during abrupt warming or cooling, when a temperature gradient is generated between the top and base of the firn (Severinghaus et al., 1998). The magnitude of the fractionation is gas-dependent, and is calibrated with laboratory experiments (Grachev and Severinghaus, 2003a, b).

The variations in firn condition (e.g., firn thickness, density, and temperature profile) and associated isotopic fractionation can be explicitly calculated using a firn densification model (Schwander et al., 1988, 1993; Goujon et al., 2003) if surface temperature and accumulation rate are known. The model used here (Goujon et al., 2003) includes heat transfer through the firn and the ice down to the bottom of the ice sheet. The model uses as inputs the oxygen isotopes of ice (Stuiver et al., 1995) with an adjustable scaling α as a surface temperature proxy, and independently estimated accumulation rate from annual layer-counting with ice-flow corrections (Alley et al., 1997b; Cuffey and Clow, 1997). It calculates both firn thickness and thermal gradient in the firn and thus allows us to predict the nitrogen fractionation (Fig. 3). The best fit of our data with the modeled nitrogen isotope ratios (Fig. 3) was found with a decadal average cooling of 3.3 ± 0.5 °C, corresponding to an oxygen isotope temperature sensitivity

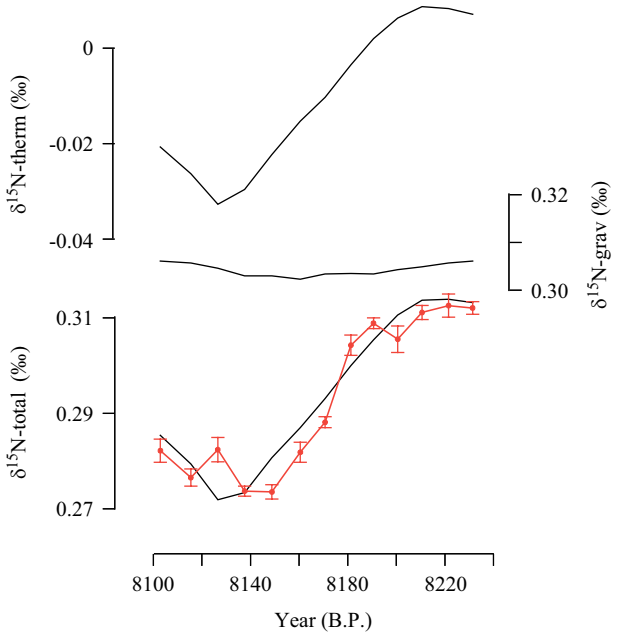

Fig. 3. Modeling results of the rapid cooling (see text). Black lines ($\delta^{15}N$-therm, $\delta^{15}N$-grav, and $\delta^{15}N$ -total) are model results. Red dots are data and average values, and error bars are standard errors of mean (1σ).

$\alpha = 0.55 \pm 0.05$‰/°C and $\beta = 17.34$ (range: 15.69–19)‰ [T(°C) = ($\delta^{18}O + \beta$)/α]. An important uncertainty regarding this temperature estimate comes from a possible change in the thickness of a near-surface convective zone, where gases mix with the atmosphere by wind pumping (Colbeck, 1989; Kawamura et al., 2006). Although we believe that the convective zone stayed relatively constant for the past 10,000 years in central Greenland, we cannot rule out an increase in convective zone thickness due to intensified wind during the event, which would bias the temperature estimate. A ±2 m change in the thickness of the convective zone during the cooling period corresponds to a $\delta^{15}N$ signal of ±0.009‰, and we therefore add ±1 °C uncertainty to the cooling estimate (thermal response of $\delta^{15}N$ is ~0.015‰ for 1 °C temperature gradient at ~−31°C: Grachev and Severinghaus, 2003b). Overall, we estimate that the magnitude of the cooling at central Greenland was 3.3±1.1 °C in the decadal average (difference between decadal averages before and after the cooling event). In particular, the magnitude of the cooling was twice as large as the Little Ice Age cooling (~1.6 °C after the Medieval Warm Period) (Dahl-Jensen et al., 1998), which took a few centuries to reach minimum temperature, but had substantial impacts on European society (Fagan, 2000). The magnitude of the 8.2 ka cooling on shorter time scales (1–5 years) may have been larger, but cannot be addressed by our technique due to smoothing of the gas record (Severinghaus et al., 1998).

This temperature estimate is smaller than previously published estimates. Notably, our estimate of temperature change is significantly smaller than that of Leuenberger et al. (1999) who used a similar technique with $\delta^{15}N$ data from the GRIP ice core to estimate that temperature changed by 7.4 °C with a range of 5.4–11.7 °C (Fig. 4). The experimental uncertainty of their data (pooled standard deviation for duplicate data = 0.030‰) is about 7-fold larger than our uncertainty of 0.004‰ (Fig. 4), and contains high values that are physically unrealistic. Clearly, the differences in experimental procedures introduced larger errors into Leuenberger et al.'s dataset, such as by air leaks during the gas extraction process or water introduced into the mass spectrometer (N_2H^+ forms in the presence of water and isobarically interferes with $^{15}N^{14}N^+$). As the signal for the 8.2k event (0.035‰) is rather small, the precision of the Leuenberger et al. dataset should make identification of the 8.2k event difficult. It is possible that the magnitudes of the 8.2 ka cooling at the GISP2 and GRIP sites are different. However, to draw such a conclusion, it would be necessary to reanalyze the GRIP ice core with higher precision.

Climate models employed to simulate the 8.2 ka event generally produce similar magnitude of cooling. Using the ECBILT-CLIO model, Renssen et al. (2001) estimated the cooling to be 2–5 °C in Greenland. Using CLIMBER-2, Bauer et al. (2004) estimated 3.6 °C in the North Atlantic sector (60–80°N). LeGrande et al. (2006) found 1.4 °C with a range of 1.9–1.1 °C cooling with a coupled model (Goddard Institute for Space Studies MODELE).

3.2. Time-evolution of Greenland temperature change

As mentioned above, the oxygen isotopes of ice may have a significant influence from changes in the seasonal distribution of precipitation and storm tracks, as well as other phenomena (Jouzel et al., 1997). The nitrogen isotopes provide an independent means to qualitatively assess the time-evolution of Greenland temperature under two plausible assumptions. The first assumption is that nitrogen signals mostly consist of the thermal signal from the temperature gradient ΔT between top and bottom of the firn layer. This is validated by the firn densification model, which shows that the thermal signal is about 10 times larger than the gravitational signal due to modeled firn thickness changes (Fig. 3). The second assumption is that the change in the temperature gradient ΔT in the firn layer is similar to the change in surface temperature. Because the heat conductivity of snow is much smaller than that of ice and the heat capacity of the ice sheet is quite large (Yen, 1981), ΔT should mimic surface temperature evolution on a decadal to centennial time scale. The firn densification–heat diffusion model (Goujon et al., 2003) with oxygen isotopes of ice as a surface temperature proxy supports the assertion that surface temperature changes

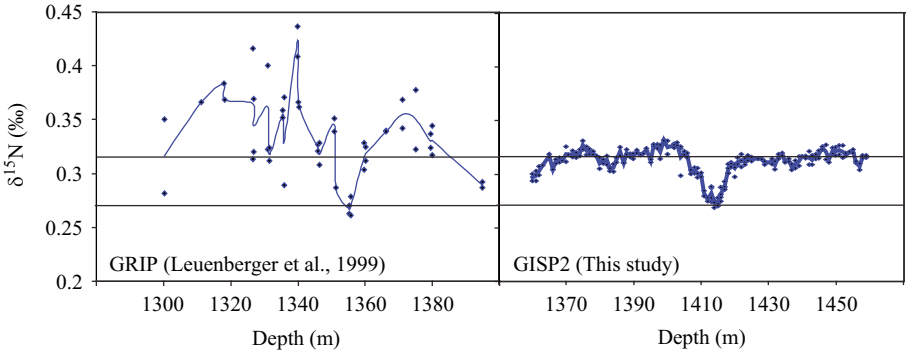

Fig. 4. Nitrogen isotope data from Leuenberger et al. (1999) and this study. Diamonds are individual data points, and blue solid lines represent average values.

should be qualitatively represented in nitrogen signals for the 8.2k event.

The surface temperature signal in the nitrogen isotope record is smoothed by gas diffusion and bubble close-off in the firn layer (Severinghaus et al., 1998). To reconstruct the surface temperature signals from smoothed signals in the nitrogen isotope record, we need to deconvolve the nitrogen record using some assumed smoothing function. The effect of the smoothing process can be represented by an age distribution (hereafter we call it a smoothing function). Spahni et al. (2003) developed smoothing functions for the central Greenland methane record (GRIP) for the 8.2 ka event, with two solutions, 'narrow' and 'wide' (referring to the width of the smoothing function) intended to represent the range of uncertainty. We adopt these functions for analysis of our data, under the plausible assumption that conditions at GRIP and GISP2 were similar.

Smoothing functions for nitrogen gas are obtained by modifying the smoothing functions ('wide' and 'narrow') for methane (Spahni et al., 2003) (methane diffuses 7% faster than nitrogen; Severinghaus et al., 1998). We can express the nitrogen isotope record M as $M(t) = \int_0^\infty C(\tau)A(t-\tau)\,d\tau$, where A is a qualitative surface temperature history, C is a smoothing function, t is time, and τ is a time index (Bendat and Piersol, 2000). Alternatively, we can write $\Im(M(\nu)) = \Im(C(\nu))\Im(A(\nu))$, where $\Im$ represents Fourier transform, and ν is frequency (Bendat and Piersol, 2000). Thus, qualitative surface temperature history $A(t)$ can be calculated as $A(t) = \Im^{-1}(\Im(M(\nu))/\Im(C(\nu)))$, where $\Im^{-1}$ represents the inverse Fourier transform. This Fourier transform method assumes perfect data; to account for experimental error we use the following technique.

To estimate the uncertainty of the deconvolution process, 100 realizations of synthetic nitrogen isotope records are generated by Monte Carlo simulation by adding white noise to the δ^{15}N data with a standard deviation of 0.004‰. These synthetic records are then linearly interpolated to obtain 1-yr-resolution time series. Then, these realizations are deconvolved with the two smoothing functions ('wide' and 'narrow'). The solutions are smoothed with a spline fit (Enting, 1987) with a 30-year cutoff period to eliminate higher frequency fluctuations (these ice core gas records do not possess high frequency climate information because of the smoothing processes), and the mean and standard deviation (1σ) of the 100 realizations at each time step are calculated for each smoothing function. This deconvolution process shifts the signals toward older ages, as expected, because the surface signals are recorded in ice with a time lag owing to the gas diffusion in the firn layer. To obtain a proper timing for the surface temperature history from the nitrogen isotope record, the deconvolved record needs to be shifted toward younger age. The magnitude of the shift can be calculated by integrating the smoothing functions. Accordingly, the age of data is shifted to younger age by 34 years for 'wide' and 22 years for 'narrow'. Upper and lower bounds (1σ) are presented (Fig. 5).

The deconvolved δ^{15}N (qualitative temperature history) suggests that the 8.2 ka event at Greenland started around 8175 BP on our time scale, and the cooling took ~20 years to reach the temperature minimum (Fig. 5). We emphasize that the cooling could have been faster than this due to

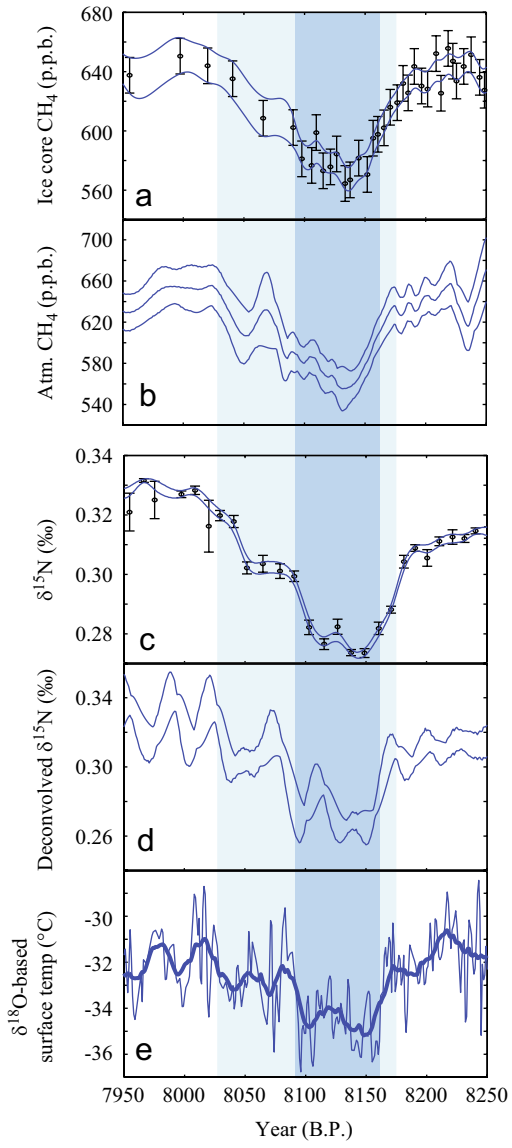

Fig. 5. Evolution of Greenland temperature and near-global wetland condition as inferred from δ^{15}N and methane. The blue area represents the duration of the event from 8175 to 8025 BP, and the dark blue area is the coldest period. (a) Methane record as for Fig. 2. The lines for the error range are calculated by Monte Carlo simulation and fitted with a 30 year cutoff period spline fit (Enting, 1987). (b) Reconstructed atmospheric methane concentration. Lines are mean and range of error (1σ) (see text for detail). (c) δ^{15}N record as for Fig. 2. The lines for error range are calculated by Monte Carlo simulation and spline-fitted with 10 year cutoff period (Enting, 1987). (d) Deconvolved δ^{15}N record. This qualitatively represents surface temperature change in central Greenland (see text). Lines are range of error (1σ). (e) Surface temperature change reconstructed from oxygen isotope record of ice (Stuiver et al., 1995). Thin and thick lines represent raw and smoothed (20-year running mean) records.

uncertainty in the smoothing functions and the inverse method. The decadal oxygen isotope record (thick blue line in Fig. 5e) shows a pattern with timings similar to the deconvolved nitrogen record ($r^2 = 0.63$; Fig. 5), which suggests that the oxygen isotopes of ice faithfully record the decadal surface temperature change during the 8.2 event. The coldest period lasted for ~60 years with one warm period of a few decades. Then, the temperature gradually increased for ~70 years with a brief period of rapid warming (Fig. 5). Thomas et al. (2006) found a similar duration of the event by compiling oxygen isotope and chemistry data from Greenland ice cores.

4. Evolution of the near-global 8.2 ka event

4.1. Methane concentration and emission history

How the 8.2 ka event evolved with time and space is important for understanding the mechanism of the event. The atmospheric methane concentration record provides a hint about the temporal evolution of the 8.2 ka event, because methane emissions are strongly affected by climate (Brook et al., 2000). The methane record thus can help to interpret the regional climatic signals found in other paleoclimate records. However, the ice-core methane record does not directly reflect atmospheric methane concentration; rather it is a smoothed record of atmospheric methane history due to diffusion and gradual bubble close-off. To reconstruct the atmospheric methane history, the ice-core methane record is deconvolved with the smoothing functions by the same procedure as for nitrogen isotopes (Fig. 5).

To reconstruct the methane emission history, we constructed a simple one-box model (the atmospheric methane concentration is calculated according to the equation $dM/dt = Q - S = Q - M/\tau$ with a constant atmospheric methane lifetime τ of 8 years). Here, we assume that the total methane emission before the event was the same as the estimated preindustrial methane emission of 221 ± 30 Tg (CH4)/yr, where 163 ± 32 is from wetlands, and 58.5 is from other natural sources (termites, wildfires, and wild animals) (Houweling et al., 2000). The preindustrial methane concentration (~700 ppb; Houweling et al., 2000) was higher than that prior to the 8.2 ka event. Therefore, the estimated emissions should be considered upper bounds. We note that recent discovery of aerobic methane production from plants (Keppler et al., 2006) may account for a fraction of this "wetland" source, but with little impact on our conclusions. The methane lifetime of $\sim 8 \pm 1$ years is estimated from the total atmospheric methane loading before the event of 1757 Tg (= 2.767[Tg/ppb] × 635[ppb]) (Etheridge et al., 1998; Fung et al., 1991) divided by the total methane emission with the ± 1 yr uncertainty obtained from the emissions uncertainty. We assume that the lifetime stayed constant through the event. This atmospheric mixing process can be represented as a smoothing function derived by forcing the model with a delta function. Two smoothing functions representing atmospheric reservoir B and firn diffusion C are derived and convolved to produce one smoothing function D: $\Im(D(v)) = \Im(B(v))\Im(C(v))$. The deconvolution of the ice-core methane record with this combined function D is conducted with the same procedure as for the nitrogen isotope record. The deconvolved results are converted to methane emissions under the assumption of constant lifetime of methane.

Results show that during the event, the atmospheric methane concentration decreased by 80 ± 25 ppb, from 635 to 555 ± 18 ppb. The firn layer smoothing reduced this magnitude by ~15% in the ice core record. The model suggests that methane emissions decreased by 32 ± 14 Tg (CH_4)/yr, from 220 to 188 ± 10 Tg (CH_4)/yr or by $15 \pm 5\%$. The reduction started around 8175 BP coincident with the temperature cooling at Greenland (Fig. 5). It took about 40 years to reach minimum methane emission values, although large uncertainty makes this figure tentative, and soon after atmospheric methane concentration started to increase (Fig. 5). During the period of gradual increase, reconstructed atmospheric methane concentration shows a spike-like increase around 8070 BP (Fig. 5), suggesting a brief period of increased methane emission, and by extension, warming and wetting. The spike is also observed in the Greenland temperature record suggesting that the short term warmer and wetter event was a large scale climatic excursion during the 8.2 ka event (Fig. 5). The total length of the 8.2 ka event was roughly 150 years (Fig. 5). Note that the magnitude or speed of methane concentration change on shorter time scales (such as interannual) may have been larger or faster, but cannot be addressed by our technique due to the inherent information loss by smoothing of the gas record.

4.2. Geographical area of methane source changes

An important question concerns the actual locations where changes in natural methane emissions occurred. Recent studies (Walter et al., 2001a; Christensen et al., 2003) suggest that methane emission is strongly controlled by temperature change. Walter et al. (2001a) found that a global temperature change of ± 1°C leads to ~20% changes in methane emission. Christensen et al. (2003) estimate that a 2 °C change of summer temperature over the area where most of the northern wetlands are located results in a 45% change in methane emission. Both studies with different approaches agree that emission sensitivity to temperature is roughly ~20% for a 1 °C change. However, the recent finding of methane emissions from plants (Keppler et al., 2006) may substantially revise the figures described here.

At present, ~70% of natural wetland methane emissions come from low latitudes, and only 25% of methane emissions come from wetlands north of 30°N, despite larger wetland areas in the north (Hein et al., 1997; Walter et al., 2001a, b). The past inter-hemispheric gradient of

atmospheric methane concentration from two polar ice cores (Greenland and Antarctica) can provide constraints on the locations of past methane emissions (Chappellaz et al., 1997; Etheridge et al., 1998; Brook et al., 2000). Chappellaz et al. (1997) found that during the early Holocene the northern (90°–30°N) contribution may have been slightly higher (~30%) and the tropical (30°N–30°S) contribution smaller (~60%; the numbers are average values of two different periods, 5000–7000 BP and 9500–11,500 BP). This corresponds to ~50 Tg/yr from the north and ~100 Tg/yr from the tropics, and the rest (10 Tg/yr) from the south.

If we assume that during the 8.2 ka event average northern (90°–30°N) temperature cooled by 1–2 °C as inferred from many paleo-data from Europe (Wiersma and Renssen, 2006), this would correspond to a 10–20 Tg/yr methane emission reduction from the northern area, which is smaller than our estimate for the methane emission reduction during the 8.2 ka event. Although large uncertainty for our methane emission estimates precludes a firm conclusion, this may imply that the impacts of the event extended beyond the northern area. Furthermore, recent compilations of paleoclimate data from around the globe (Alley and Agustsdottir, 2005; Morrill and Jacobsen, 2005; Wiersma and Renssen, 2006) suggest a near-global cooling and drying event. The European region experienced strong cooling and hydrologic changes, and the Sahara and western Asian monsoon region experienced drying. Cooling probably occurred in North America, with drying in the US Great Plains (Alley and Agustsdottir, 2005; Morrill and Jacobsen, 2005). The southward shift of the intertropical convergence zone (ITCZ) as inferred from the Cariaco Basin record (Hughen et al., 1996; Haug et al., 2001) may indicate that a large tropical area experienced changes in precipitation (Alley and Agustsdottir, 2005; Morrill and Jacobsen, 2005). A recent study showed a weakened Asian Monsoon during the 8.2 ka event from oxygen isotopes of stalagmites from Dongge Cave in southern China (Wang et al., 2005). LeGrande et al. (2006) calculated methane changes for the 8.2 ka event using local regression models coupled with a general circulation model (GCM). They found significant methane emission reduction over North America and Europe by 11% (8%) with North Atlantic Deep Water formation reduced by 60% (40%). Their estimates are slightly less than our estimate of $15 \pm 5\%$ but agree within the uncertainty. In conclusion, our methane data are broadly consistent with this widespread geographical footprint and model result of the event's impact on temperature and precipitation.

5. Timing of the 8.2 ka event

Although many high-resolution proxy data for the early Holocene reveal an abrupt event at around 8.2 ka, the apparent age and duration are different in different locations, probably owing to the difficulties of all paleoclimate records with age uncertainty, continuity, and regional climatic influences (Alley and Agustsdottir, 2005; Morrill and Jacobsen, 2005; Rohling and Palike, 2005). Therefore, it is difficult to assess the relative timing of climatic events between different regions inferred from paleoclimatic proxies. The simultaneous analysis of nitrogen isotopes and methane in a single ice core circumvents these problems, and allows us to compare the timing of northern high-latitude and large-scale climate change in methane source regions (Severinghaus et al., 1998; Severinghaus and Brook, 1999).

The methane and nitrogen isotope records show very similar timing through the 8.2 ka event ($r = 0.91$ for the period of 8000–8250 BP, Fig. 6). We found that the maximum correlation coefficient ($r = 0.95$) occurs when the methane age is shifted by 7.6 ± 4 (1σ) years toward older ages, suggesting that the methane signal lags the nitrogen signal. Uncertainty was estimated using a Monte Carlo procedure. We carried out 5000 realizations in which we added white noise to the data with a standard deviation of 17 ppb for methane and 0.004 per mil for $\delta^{15}N$, and found the maximum of the lagged correlation.

However, to compare accurately the near-global climate signal in methane and the Greenland climate signal in nitrogen isotopes, we have to account for two non-climatic processes that can create differences in phasing of the two gases in an ice core. The first is the atmospheric methane reservoir effect. Owing to the large load of methane in the atmosphere, a change in methane emissions takes time to be reflected in the atmospheric methane concentration (e-folding time of ~8 years). The second is due to differences in the diffusion rate of methane and nitrogen gases. As these gases diffuse through the firn layer, a change at the surface takes 20–30 years on average to be recorded in the ice (Spahni et al., 2003). The difference in the speed of diffusion between methane and nitrogen creates an apparent methane lead relative to the nitrogen signal by 1–2 years (Severinghaus et al., 1998; Spahni et al., 2003).

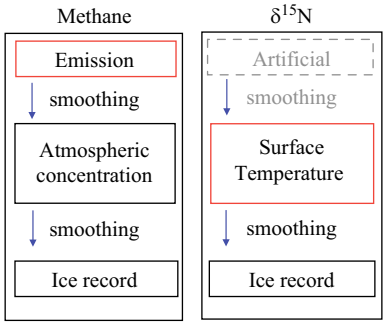

Fig. 6. Filtering processes for nitrogen isotopes and methane. The climatic signal (emission change) in methane goes through two natural filtering processes while the climatic signal (surface temperature change) in nitrogen isotopes goes through only one filtering process. Red rectangles indicate the places where climatic signals start. For accurate comparison of timing, nitrogen signals are processed by a fictitious atmospheric reservoir (see text).

To more precisely evaluate the timing of climatic changes in methane and nitrogen isotopes, we process the data so that both signals experience the same filtering (Fig. 6). Because climatic signals in methane experience two filtering processes before being trapped in ice, and the signal to noise ratio of data are lower than for nitrogen isotopes, it is more difficult to obtain a precise climatic signal (i.e., emission change) by deconvolution. Instead, we process the nitrogen isotope signal to be comparable to the methane signal by the following method. First, the age of the nitrogen signal is shifted by 1–2 year toward older age to account for the difference in speed of diffusion. Second, an atmospheric reservoir effect was added to the nitrogen isotope signal. To do so, the nitrogen signal is scaled to be comparable with the methane record for optimal comparison of the timing (Fig. 7). Then, the scaled nitrogen signal is converted to methane emission using the relationship (modeled methane emission (Tg/yr) = 2.767 × [the scaled nitrogen signal]/8 yr (methane lifetime)). Then, the emission scenario derived from nitrogen record is smoothed with an atmospheric mixing model of the form ($dM/dt = Q - S = Q - M/\tau$, where M represents total mass of CH_4 in atmosphere, Q is source strength, S is sink strength, and τ is a constant atmospheric methane lifetime of 8 years). The methane and adjusted nitrogen isotope signals show synchronous timing with a maximum correlation coefficient of $r = 0.96$ with 0–1 year shift (Fig. 7). Thus we conclude that cooling and drying in many parts of the northern hemisphere and tropics occurred within ±4 years of climate changes in Greenland recorded by the nitrogen isotope record.

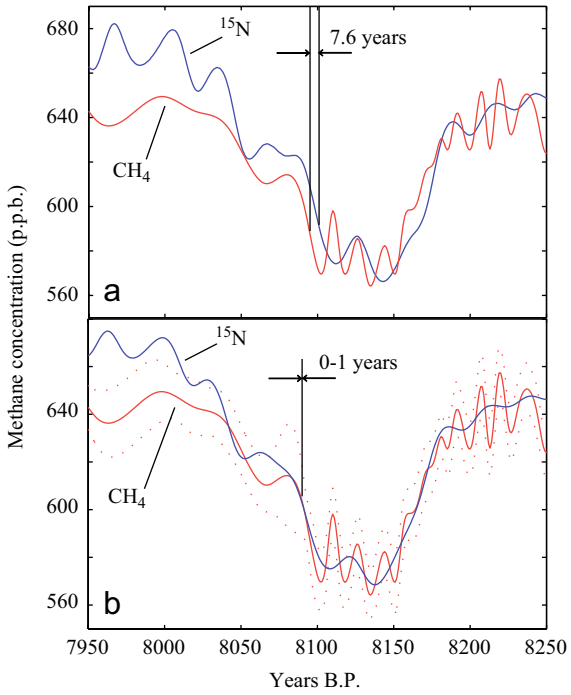

Fig. 7. Timing of the 8.2 ka event. (a) Red and blue lines represent methane concentration and scaled nitrogen isotope data, respectively, after a spline fit (Enting, 1987) with cut-off period of 10 yr, corresponding to the resolution of data. (b) Same as (a) except the nitrogen signal has been processed to mimic the methane atmospheric reservoir effect (see text). Red and blue lines represent the methane and the processed nitrogen signal, respectively. Dotted lines are error ranges (1σ) for methane data. Note that the methane signal lags the nitrogen signal by 7.6 ± 4 years in the ice core record (a), but after the nitrogen signal is processed to account for the 8-year lifetime of methane, the two signals show almost identical timing (see text).

6. Conclusion

A large number of paleoclimatic records over a hemispheric area show a large and abrupt climate change around 8200 years BP. However, the duration and general character of the event have been ambiguous. Here, we provide a precise characterization and timing of the event using methane and nitrogen isotopes in trapped air in a Greenland ice core. Climate change in Greenland and at a hemispheric scale was simultaneous (within ±4 years) as supported by climate model results (LeGrande et al., 2006). The event started around 8175 ± 30 years BP, and it took less than ~20 years to reach the coldest period, with a magnitude of cooling of 3.3 ± 1.1 °C in central Greenland. After ~60 years of maximum cold, climate gradually recovered for ~70 years to a similar state as before the event. The total duration of the event was roughly 150 years.

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