Quaternary Science Reviews 29 (2010) 2828–2838

Contents lists available at ScienceDirect

Quaternary Science Reviews

journal homepage: www.elsevier.com/locate/quascirev

Millennial-scale variability during the last glacial: The ice core record

E.W. Wolff [a,*], J. Chappellaz [b], T. Blunier [c], S.O. Rasmussen [c], A. Svensson [c]

[a] British Antarctic Survey, High Cross, Madingley Road, Cambridge, CB3 0ET, UK
[b] Laboratoire de Glaciologie et Géophysique de l'Environnement, CNRS-University of Grenoble, BP96, 38402 St. Martin d'Hères Cedex, France
[c] Niels Bohr Institute, Ice and Climate Research, University of Copenhagen, Juliane Maries vey 30, DK-2100 Copenhagen, Denmark

ARTICLE INFO

Article history:
Received 13 October 2009
Accepted 19 October 2009

ABSTRACT

The oxygen-isotope records from Greenland ice cores show a very strong, reproducible pattern of alternation between warm Greenland Interstadials (GI) and cold Greenland Stadials (GS) at millennial-scale during the last glacial period. Here we summarise what is known about this variability from ice core records. The typical cycle has a sawtooth pattern, with a very rapid warming event (occurring in a few decades), a slow cooling trend, and then a final fast cooling. 25 such events have been numbered in the last glacial. The recent GICC05 age scale provides the best available age scale that can be directly applied to this stratigraphy, and we summarise the timing of the warming events, and the length and strength of each event. The Greenland stratigraphy can be transferred to other records if we make assumptions about the contemporaneous nature of rapid events in different archives. Other parameters, such as the snow accumulation rate, and the concentration of terrestrial dust and sea salt recorded in the Greenland cores, also show a strong contrasting pattern between GI and GS. Methane concentrations are generally high during GI and lower during GS, with the increase from GS to GI occurring within a century. Antarctic ice cores show a different pattern: each GI has an Antarctic counterpart, but Antarctica appears to warm while Greenland is in a GS, and cool during GI. These changes are consistent with a mechanism involving ocean heat transport, but the rapid nature of warmings poses a challenge for modellers, while the rapid methane changes pose questions about the pattern of land biosphere emissions during the glacial that are also relevant for understanding glacial-interglacial methane variability.

1. Introduction

The most significant climate variability in the Quaternary record is the alternation between glacial and interglacial, occurring at approximately 100 ka periodicity in the most recent 800 ka. This signal is of global scale, and observed in all climate records, including the long Antarctic ice cores (Jouzel et al., 2007a) and marine sediments (Lisiecki and Raymo, 2005). There is a strong consensus that the underlying cause of these changes is orbital (i.e. due to external forcing from changes in the seasonal and latitudinal pattern of insolation), but amplified by a whole range of internal factors (such as changes in greenhouse gas concentration and in ice extent).

At sub-orbital periods, the most prominent scale of variability is the millennial variability that is particularly noticeable in the last glacial (Marine Isotope Stages (MIS) 4, 3 and 2, 73.5–14.7 calendar kyr BP, ka (Sanchez-Goni and Harrison, 2010)). The clearest manifestation of this variability are the Dansgaard-Oeschger (D-O) cycles observed in Greenland ice cores. Over the last 30 years, the structure and scale of these events has become clear, and a favoured (though not universally-shared) view of their origin has emerged. The D-O cycles show a very strong climate signal (typically about 50% of glacial-interglacial amplitude in Greenland temperature), with very abrupt transitions. Their signature has been seen in a range of climate records around the northern hemisphere, while counterparts with a different trend and shape are seen in Antarctica: thus they are of global scale, even if they are manifested differently in different regions. They are seen not only in climate but in measures related to biogeochemical cycles and atmospheric circulation. The Greenland D-O sequence has become the *de facto* event stratigraphy for the last glacial period, and a formally-recommended one for the period from 30 to 8 ka ago, including the termination (Björck et al., 1998; Lowe et al., 2008). For this reason, before embarking on a synthesis of how D-O cycles are seen in vegetation records, it is crucial to have a clear understanding of what is seen in ice cores.

We start with a description of the climate signal of D-O cycles during the last glacial in Greenland ice cores, and then consider some of the other signals seen in these cores, before moving on to discuss the counterpart millennial-scale signals seen in Antarctic

* Corresponding author.
E-mail address: ewwo@bas.ac.uk (E.W. Wolff).

0277-3791/$ – see front matter © 2009 Elsevier Ltd. All rights reserved.
doi:10.1016/j.quascirev.2009.10.013

ice cores. These summaries of climate will include a description of the changes in radiatively-important trace gases that are associated with millennial-scale changes, with a particular emphasis on methane. We will briefly touch on the question of whether similar variability occurred in earlier glacial periods, how it relates to interglacial millennial variability in ice cores, and how the ice core variability is related to that seen in other palaeorecords. A detailed description of the causes of such variability is outside the scope of this paper but we will briefly summarise the current hypotheses.

2. Millennial-scale variability in Greenland climate

The existence of a strong millennial-scale signal in Greenland ice core oxygen-isotope records was first clearly noted in the early 1980s, with the appearance of the DYE-3 core (Dansgaard et al., 1982) from south Greenland. The similarity of the pattern of climate during the last glacial period between the DYE-3 record and the Camp Century (northwest Greenland) core that had previously been obtained (Johnsen et al., 1972) strengthened the view that the oscillations seen were real climate signals of regional significance. However, it was the parallel cores drilled by the European GRIP (Johnsen et al., 1992; Dansgaard et al., 1993) and US GISP2 (Grootes et al., 1993) teams at or near Summit on top of the Greenland ice sheet that cemented the notion of a quasi-regular sequence of highly abrupt climate swings, which have become known as Dansgaard-Oeschger cycles. The different cores displaying a clear D-O sequence from Greenland are summarised in Table 1.

2.1. The pattern, structure and strength of D-O variability in Greenland temperature

The more recent NorthGRIP (NGRIP) ice core is now being seen as the standard reference core for two reasons. Firstly it is complete and continuous through the entire glacial and into the last interglacial (continuous sequence of 123 ka). Secondly it has the most complete multi-parameter layer-counted dating over the last 60 ka (Svensson et al., 2008).

The oxygen-isotope record from NGRIP (Fig. 1) covers the period from the present, through the last glacial (MIS 2-4), and beyond into MIS 5, which includes the glacial inception and the end of the last interglacial (identified with the Eemian in NW Europe, and with MIS 5E). It shows a sequence of 25 identified and numbered (Dansgaard et al., 1993; North Greenland Ice Core Project Members, 2004) D-O cycles, of which 18 (numbers 2–19) are in MIS 2-4. In the literature (e.g. Johnsen et al., 1992) the numbered warm periods are described as interstadials, and more specifically Greenland interstadials (GI or GIS) (see also Sanchez-Goni and Harrison, 2010). We use the notation of GIn for warm periods. The term interstadial is somewhat inappropriate because it contains an implication that these intervals are coeval with glacial retreats, but we maintain the usage in order to avoid introducing new terminology. There is some confusion in the literature in numbering the cold stages (GS) in between the GI, with cold events being given the same number as either the GI that preceded them or the one that followed them (Rousseau et al., 2006). In the absence of a clear concensus, we will avoid either usage by describing a GS as "the GS between GIn and GI($n+1$)". There is also an element of subjectivity in the numbering system for GI: for example, an apparently separate peak just after GI19 was not numbered, and GI15 and 17 each consist of doublet peaks. Finally we have followed the authors of NGRIP papers by using ages relative to 2000 AD, denoted as years (or ka) b2k. Where dates are compared to literature dates referenced to 1950, 50 years should be subtracted from the quoted b2k ages.

The glacial is populated by a series of alternations between cold and warm. More specifically, the typical D-O cycle (Fig. 1) begins with a very sharp D-O warming event (within decades, as discussed shortly). Temperature, as deduced from water isotopes, then cools slowly for a time, before plunging quite rapidly back to a cold baseline, which is maintained until the next rapid warming event. This pattern is generally seen in each cycle, irrespective of its length, although the shapes of earlier events are somewhat more complex, including at least one GI which lacks the sharp final cooling (GI23). The duration of events is variable, for example GI3 is only 300 years long (between the cold plateaus) while GI12 is more than 2500 years long; the earlier events, GI21 and 23, are particularly long. The length of the cold period plateaus between GI is also variable, from hundreds to thousands of years, and we note two periods (around 60–70 ka, and 25–15 ka, much of MIS 4 and 2) in which D-O activity appears particularly weak. The final numbered GI (GI1) is generally considered to be synonymous with the Bølling-Allerød warm period that precedes the Younger Dryas stadial. It therefore forms part of the termination. However, it has similar characteristics to other GI, and we consider it to be part of the general sequence.

Some authors have proposed that the D-O cycles are arranged into bundles, with a strong, wide GI being followed by successively weaker and shorter GI (Bond et al., 1993). This appears to be a good description of the sequence from GI12 to 9, and from 8 to 5, but it is not so obvious elsewhere in the sequence.

Until this point, we have considered the δ^{18}O record as if it was a quantitative recorder of Greenland air temperature. However, borehole thermometry showed very clearly that the change in temperature at Termination I was considerably larger than would be calculated from the modern spatial relationship between isotopes and temperature (Cuffey et al., 1995; Johnsen et al., 1995). There is now considerable evidence that the factor that translates an isotopic change to a temperature change is variable, probably due to changes in the seasonal distribution of precipitation (Jouzel et al., 2003; Krinner and Werner, 2003); we also cannot rule out that changes in the strength of the atmospheric inversion may account for a part of the difference between the temperature changes in the troposphere inferred from conventional interpretation of water isotopes, and temperature changes in surface snow (Cuffey and Clow, 1997) that are deduced from measurements of

Table 1
Locations and characteristics of Greenland ice cores in which the D-O sequence has been identified.

Location	Latitude	Longitude	Core length/m	Thickness of ice covering glacial period (MIS 2–4) (m)	Reference
Camp Century	77.2°N	61.1°W	1387	100	(Johnsen et al., 1972; Dansgaard et al., 1982)
DYE-3	65.2°N	43.8°W	2037	140	(Dansgaard et al., 1982)
GISP2	72.6°N	38.5°W	3053	800	(Grootes et al., 1993)
GRIP	72.6°N	37.6°W	3029	800	(Johnsen et al., 1992; Dansgaard et al., 1993)
Renland	71.3°N	26.7°W	324	12	(Hansson, 1994)
NorthGRIP	75.2°N	42.5°W	3085	900	(North Greenland Ice Core Project Members, 2004)

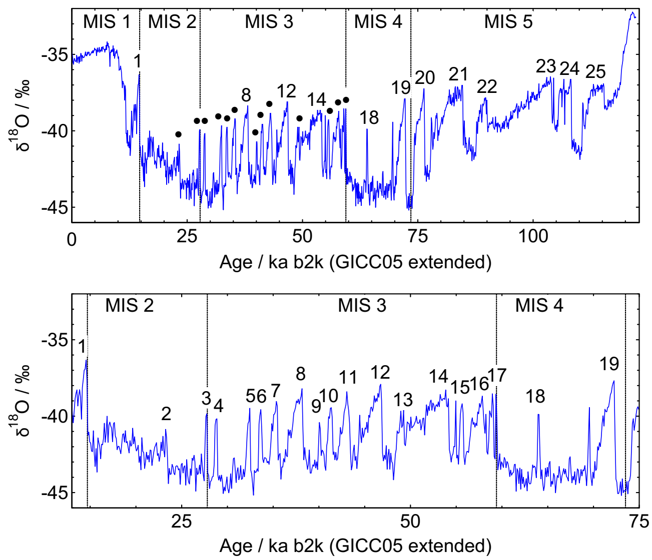

Fig. 1. Dansgaard-Oeschger events in the NGRIP ice core, Greenland. (a) From the present back to 123 ka ago; (b) MIS 2-4. The oxygen-isotope data (North Greenland Ice Core Project Members, 2004) smoothed to 100 year averages are presented here on the GICC05 age scale back to 60 ka (Svensson et al., 2008), and before that on the ss09sea modelled age scale (Johnsen et al., 2001) shifted (by −705 years) to match GICC05 at 60 ka. The notation b2k implies age before 2000 AD – remove 50 years for comparison with dates expressed as bp relative to 1950 AD. The numbers denote the standard notation for Greenland Interstadials (GI), with dots marking each intermediate numbered event in part (a).

borehole temperature (and indeed from isotopic fractionation, discussed next). Borehole temperature profiles do not conserve sufficient information to derive temperature changes for D-O warming events; however fortunately another elegant method exists.

Gases in the diffusive part of the firn column in an ice sheet fractionate due to concentration gradients, gravity and thermal gradients. During a rapid warming (or cooling), a temperature gradient is established between the top and bottom of the firn column, and this causes fractionation because gases diffuse faster than heat in polar firn. In particular, ^{15}N in N_2 will be enriched at the cold end of the column, by an amount that is physically related to the magnitude of the temperature change; $\delta^{15}N$ can be measured in air bubbles (Severinghaus and Brook, 1999). By measuring also $\delta^{40}Ar$, changes in firn thickness affecting both

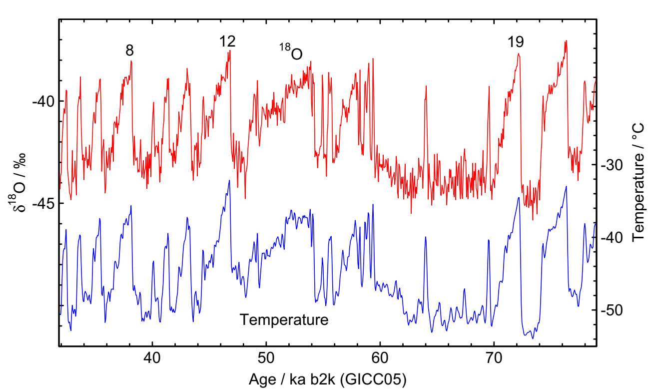

Fig. 2. Oxygen isotopes and temperature during several D-O cycles. The temperature has been deduced by fitting a firn densification and heat diffusion model to $\delta^{15}N$ data (Landais et al., 2004a; Huber et al., 2006). The temperature data, originally published on a different age scale, have been converted to GICC05.

isotopes through the gravitational effect can be estimated, allowing calculation of the actual temperature change corresponding to each rapid temperature change (because it only works if the temperature change is rapid, the method is most appropriate for the warmings). This method has now been used on a range of D-O warming events, and different authors used it to estimate for example that the warming at the start of GI1(in the GISP2 ice core) is 11 ± 3 °C (Grachev and Severinghaus, 2003), while that at the start of GI19 (at NGRIP) is as large as 16 ± 2.5 °C (Landais et al., 2004a). A slightly different method, also using $\delta^{15}N$, but where both the depth and magnitude of $\delta^{15}N$ changes are used as a constraint in a temperature/firn densification/gas diffusion model, has been applied to obtain continuous temperature estimates across D-O cycles (Lang et al., 1999; Huber et al., 2006). These methods of estimating the real temperature changes (Fig. 2) do not grossly alter the shape of the temperature evolution compared to that derived directly from $\delta^{18}O$; however, there are subtle changes such as the flatter top of GI14, or the relatively high temperature in GI12 compared to other GI. The most important finding from all these estimates is that the D-O warmings are immensely strong (Table 2), with the very fast part of the warming ranging from 8 °C up to 16 °C.

2.2. Dating, timing and pacing of Greenland D-O signals

It is important to establish a reliable age model for deriving the timing of D-O warming events. We restrict ourselves here to considering the three cores in which there is good resolution through the glacial period (Table 1). The age scales used for the GRIP ice core in the last glacial were based on rather simple ice flow models, with accumulation rates calculated, using an empirical relationship, from the oxygen-isotope data. The ss09sea model (Johnsen et al., 2001) is just such a model, where the isotope values were corrected for the isotopic value of seawater before they were used in the accumulation rate model. The GISP2 age scale was based on counting of annual layers in between 1 and 3 different parameters (Alley et al., 1997; Meese et al., 1997). The GRIP and GISP2 age scales showed significant discrepancies from each other (Southon, 2004), especially in the relative lengths of GI and GS, and from independently derived age scales (e.g. Shackleton et al., 2004).

A comprehensive effort has now been made to produce an age scale based on annual layer counts using mainly the NGRIP ice core. It is generally accepted (see e.g. Lowe et al., 2008) that this age scale (known as GICC05) is the "standard", despite some criticisms (Skinner, 2008), and we will therefore use it when possible in this paper. It is assumed that future age scales will provide a cross-reference to GICC05 so that the ages assigned to events in this paper can be readily translated.

The GICC05 is a multi-core, multi-parameter attempt at a fully layer-counted age scale. For the Holocene section, the DYE-3, GRIP and NGRIP cores were synchronised using the pattern of volcanic sulphate spikes, and the best records were used to perform annual layer counting (Vinther et al., 2006). Dating of the part of the record to 8 ka relies heavily on the seasonal cycle of oxygen isotopes, which is well-preserved in the high-accumulation rate DYE-3 core. Between 7.9 and 10.3 ka, GRIP high resolution chemistry data, backed by sections of water isotope data, were used. The remainder of the age scale relies only on the NGRIP core, using 7 parameters with a seasonal cycle to perform layer counting (Andersen et al., 2006; Rasmussen et al., 2006; Svensson et al., 2006; Svensson et al., 2008). The age scale comes with an uncertainty based on identification of layers considered to be uncertain years (counted as 0.5 ± 0.5 years) by the team doing the counting; the accumulated uncertainty is not strictly speaking a Gaussian error estimate, but has been treated as a 2σ error on the age scale.

Table 2
The age of D-O warming events in Greenland ice cores. Ages are given on the GICC05 age scale to 60 ka, and on an adjusted ss09sea age scale (with 705 years subtracted) beyond that, as discussed in the text. The ages are estimated for the rapid warming events at the start of GI. Durations of GI are given to the nearest century, from the rapid warming event to the bottom of the sharp final drop, omitting any estimate where the end of the GI is hard to pinpoint due to the complex structure of the event (such events labelled by *). The temperature jumps are those estimated from ^{15}N data wherever such estimates have been published, as described in the text, and for NGRIP except where specified. The values for the temperature jump at the start of GI 4-7 (in italics) do not use ^{15}N data, but are derived from an inversion of water isotope data: the values given here are those estimated in a table by Sanchez-Goni et al. (2008) based on a figure shown by Masson-Delmotte et al. (2005).

Start of which GI	Age/ka b2k	Quoted 1σ uncertainty	Age/ka b 1950	Reference	GI duration/ centuries	Magnitude of rapid T jump/°C	Reference citing magnitude
YD/PB	11.703	0.050	11.653	(Rasmussen et al., 2006)		10 ± 4 (GISP2)	(Grachev and Severinghaus, 2005)
1	14.692	0.093	14.642	(Rasmussen et al., 2006)	19	11 ± 3 (GISP2)	(Grachev and Severinghaus, 2003)
2	23.340	0.298	23.290	(Andersen et al., 2006)	1		
3	27.780	0.416	27.730	(Andersen et al., 2006)	3		
4	28.900	0.449	28.850	(Andersen et al., 2006)	3	*12 ± 5 (GRIP)*	(Sanchez-Goni et al., 2008)
5	32.500	0.566	32.450	(Andersen et al., 2006)	5	*7 ± 5 (GRIP)*	(Sanchez-Goni et al., 2008)
6	33.740	0.606	33.690	(Andersen et al., 2006)	4	*7 ± 5 (GRIP)*	(Sanchez-Goni et al., 2008)
7	35.480	0.661	35.430	(Andersen et al., 2006)	7	*9 ± 3 (GRIP)*	(Sanchez-Goni et al., 2008)
8	38.220	0.725	38.170	(Andersen et al., 2006)	16	11 (+3;−6)	(Huber et al., 2006)
9	40.160	0.790	40.110	(Andersen et al., 2006)	3	9 (2 steps) (+3;−6)	(Huber et al., 2006)
10	41.460	0.817	41.410	(Andersen et al., 2006)	7	11.5 (+3;−6)	(Huber et al., 2006)
11	43.340	0.868	43.290	(Svensson et al., 2008)	10	15 (+3;−6)	(Huber et al., 2006)
12	46.860	0.956	46.810	(Svensson et al., 2008)	26	12 ± 2.5	(Landais et al., 2004b)
13	49.280	1.015	49.230	(Svensson et al., 2008)	*	8 (+3;−6)	(Huber et al., 2006)
14	54.220	1.150	54.170	(Svensson et al., 2008)	*	12.5 (+3;−6)	(Huber et al., 2006)
15	55.800	1.196	55.750	(Svensson et al., 2008)	*	10 (+3;−6)	(Huber et al., 2006)
16	58.280	1.256	58.230	(Svensson et al., 2008)	*	9 (+3;−6)	(Huber et al., 2006)
17	59.440	1.287	59.390	(Svensson et al., 2008)	*	12 (+3;−6)	(Huber et al., 2006)
18	64.095	?	64.045	Estimated here	3	11 ± 2.5	(Landais et al., 2004a)
19	72.330	?	72.280	Estimated here	20	16 ± 2.5	(Landais et al., 2004a)
20	76.450	?	76.400	Estimated here	24	11 ± 2.5	(Landais et al., 2004a)

The GICC05 age scale is a clear methodological improvement on previous age scales, but it does have some weaknesses. Firstly, in GS, the layer thickness in the older part of the record is low, and the chemical parameters no longer show a seasonal cycle. As a result, the counting relies only on visual stratigraphy, solid electrical conductivity (ECM) and liquid conductivity during the GS (Svensson et al., 2008). Because these three parameters are closely related in origin, the level of independence in identifying missing or doubled annual layers is low. Overall, a strong assumption of the method is that the annual pattern in observed signals persists, even under glacial conditions. Finally, the estimates of uncertain years are based on criteria agreed by the counting team; the data presented in the different papers show several examples where it is obvious that an alternative but reasonable set of criteria could have been developed that would have given a different count and a different uncertainty.

An alternative way to evaluate the age scale is to compare it with independent well-dated time markers. Such comparisons (Svensson et al., 2006; Svensson et al., 2008) confirm the GICC05 age scale, well within the margin of the quoted uncertainties. Specifically, the GICC05 counted age of two ash layers, at ~10 and 12 ka b2k, with well-known radiometric age, fits very well onto the IntCal04 calibration curve, and seems to pin the age scale at Termination I. The GICC05 age of the ^{10}Be maximum identified with the Laschamp magnetic event is 41.25 ± 0.8 ka b2k, compared to the most recent radiometric date for the geomagnetic excursion of 40.7 ± 0.95 ka (Singer et al., 2009). Assuming that the rapid events identified in $\delta^{18}O$ records from speleothems can be assumed synchronous with the D-O warming events, then agreement seems to be generally better than 500 years between the central GICC05 age and the preferred radiometric age, e.g from Hulu Cave (Wang et al., 2001), right back to 60 ka; slightly larger discrepancies (up to 800 years) between GICC05 and multiple speleothem records can be seen around GI10-12 (Fleitmann et al., 2009). In summary, despite some concerns over "unknown unknowns" in the age scale, comparisons with dated horizons in other records suggest that the GICC05 age scale is accurate well within its quoted 1σ uncertainty. This then validates the layer counting between horizons and suggests that the GICC05 estimates of the duration of events are likely to be rather precise.

The counted age scale extends only to 60 ka b2k. Beyond that, the model ss09sea age scale, as used in the original NGRIP publications (North Greenland Ice Core Project Members, 2004) exists, but it has an offset compared to GICC05 at 60 ka. We have therefore used the simplest method to splice the two age scales together by using ss09sea with 705 years removed for all ages older than 60 ka.

The ages of each D-O warming event (i.e of the start of the GI) on the GICC05 age scale, extended as described above, are given in Table 2. The ages of events beyond 60 ka must be treated with special caution because of the way the age scale was extended. We have also made estimates of the length of each event. This should be treated as a guide only, because it is sometimes difficult to pinpoint the sharp cooling that characterises the end of a GI (e.g. GI14).

A crucial question about the timing of each event is whether they occur at a fixed period, or stochastically. This is critical for understanding the likely causes of D-O cycles. Despite the excellent datasets and the clear nature of the events, it has proved difficult to get agreement on this issue. This reflects the fact that the detection of periodicity depends on the age scale used, that the shape of the events is not conducive to certain conventional approaches, and that most authors have confined themselves to a limited period so that the population of events to be tested is not very large. To overcome the second issue, several authors have used the waiting time between the sharp warmings as the statistic to be tested for periodicity. Using the GISP2 age scale, a period of 1470 years has been determined for the events of the last 50 ka by different authors using different methods (Grootes and Stuiver, 1997; Schulz, 2002; Rahmstorf, 2003), suggesting that D-O warming events occur at intervals that are multiples of this period. This may be the result of a forcing with the same period, such as solar forcing (Braun et al., 2005), operating either directly or in combination with a noisy system to produce stochastic resonance (Alley et al., 2001). However, the robustness of the periodicity is questioned, and it has been suggested that on the GICC05 age scale the recurrence times are indistinguishable from those expected with random occurrence (Ditlevsen et al., 2007). We consider this issue to be still unresolved.

3. Other signals in Greenland ice cores

Almost every parameter measured in ice cores varies strongly across D-O cycles. Here we summarise a few of them that indicate different aspects of the environment.

3.1. Snow accumulation rate, deuterium excess, and ice chemistry

Measured layer thicknesses, which can be strain-corrected to derive ice equivalent accumulation rates, varied in very close concert with $\delta^{18}O$ (Andersen et al., 2006; Svensson et al., 2008) across the D-O cycles of the last 60 ka. It was found that the derived median accumulation rates at NGRIP during the GI 3-10 were approximately double those of the intervening cold periods, and just over half those found in the present day (Andersen et al., 2006). Although a thermodynamic effect may be partly responsible (warmer air delivering more moisture), it has previously been suggested that the main cause of the changes is actually changes in storm tracks, suggesting a rapid shift to more northerly tracks at warming jumps, bringing more moisture to Greenland (Kapsner et al., 1995).

Deuterium excess (dxs) varies strongly with $\delta^{18}O$ across D-O cycles at Greenland sites (Masson-Delmotte et al., 2005; Jouzel et al., 2007b), with lower values during warm periods. The changes in dxs seem in some cases to be sharper than those in $\delta^{18}O$. Conventionally, dxs is used as an indicator of conditions in the water vapour source region, thus at first sight implying that very rapid changes occur at lower latitude, or else that rapid shifts occurred in the relative importance of different water sources (e.g. Atlantic versus Pacific). However, it seems to be difficult to reconcile the values seen with the conventional interpretation, and further work is needed to untangle the exact meaning of this parameter in Greenland ice.

Huge changes are observed in many chemicals measured across D-O cycles in Greenland cores, with higher values in cold periods than warm (Mayewski et al., 1994; Mayewski et al., 1997) (Fig. 3). The strongest signal is seen in all elements associated with terrestrial dust, such as Ca^{2+} (Fuhrer et al., 1999; Ruth et al., 2007). The Ca^{2+} or dust concentration typically decreases rapidly (decadal time scale) by a factor approximately 10 during a D-O warming event, stays low during the warm period (GI), and then increases more slowly (order 1 century) into the next cold period (GS). The decreases during warming actually seem to occur in a series of extremely rapid (1–2 year) steps (Fuhrer et al., 1999). The increase in alkaline dust is strongly associated with a decrease in acidity of the ice, as reflected in solid electrical logs of ice cores (Wolff et al., 1997). Because wet deposition dominates over dry deposition under current conditions, the concentration in ice is more relevant than the flux in deducing changes in atmospheric loading. Dry deposition probably had a greater role under glacial conditions, but it is unlikely that the change in snow accumulation rate between GI and GS played a significant role in the observed concentration changes. Dust, along with sea salt, has been extensively discussed in a recent review (Fischer et al., 2007b), and here we mainly summarise the conclusions of that paper.

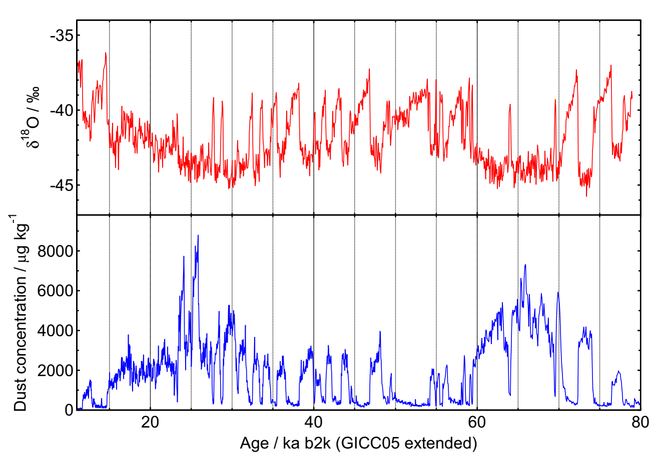

Fig. 3. Changes in terrestrial dust concentration in the NorthGRIP core across D-O cycles. The figure shows the $\delta^{18}O$ (smoothed to 60 year averages), and the dust concentration (Ruth et al., 2007) from the NorthGRIP ice core, Greenland.

Terrestrial dust arriving at Greenland in the last glacial period has been fingerprinted by geochemical measurements (Biscaye et al., 1997; Bory et al., 2003) as originating from arid and semi-arid regions in Asia (mainly China). Thus the large D-O changes in dust concentration in Greenland have to originate either from changes in conditions in the source region (aridity, presence or absence of stabilising vegetation, surface winds that mobilise dust), or in changes in the survival of dust during transport. Faster transport times or longer lifetimes against wet deposition due to a drier atmosphere in cold periods would tend to increase the amount of material reaching Greenland. Recent estimates (Fuhrer et al., 1999; Fischer et al., 2007b) suggest that changes during transport could have caused a factor 3 increase during GS compared to GI, requiring changes in source area or strength to provide a significant factor. However, the uncertainty on such estimates is very large. What is clear is that very significant changes in northern hemisphere atmospheric circulation across all longitude bands must have occurred at each D-O switch in order to cause the changes in dust.

Sea salt concentrations (such as Na^{+}) decrease by typically a factor of 5 during D-O warming events. Although it has been suggested that the Pacific could have been a major source during the last glacial period (de Angelis et al., 1997), most authors assume that the Atlantic remained the main source (it is very hard to see how such a large increase in concentration could be generated if there was a change to a more distal source). The factor 5 increase between GI and GS must therefore arise from an increased source strength and/or increased transport as for dust. Again, it seems likely that both factors played a role (Fischer et al., 2007b), and that the D-O sea salt variability results from an interplay between changes in sea ice extent and changes in atmospheric circulation patterns. Although changes in marine conditions are not a focus of this paper, the marine signals do provide an additional constraint, in parallel with terrestrial changes, on the overall climate pattern that is consistent with the evidence.

While there were dramatic changes in the terrestrial dust across D-O cycles, the changes for some other chemicals were much more modest. Ammonium, considered to be an indicator of soil and vegetation emissions from North America, also shows lower concentrations in the glacial compared to the Holocene, presumably in part because the Laurentide ice sheet covered much of the source area (Fuhrer et al., 1996). The ammonium concentration increases moderately at the transitions from GI to GS, probably due to changes in transport and deposition mechanisms. There is some evidence (Fuhrer et al., 1996) that ammonium concentrations tend to drift higher during long GI; this is certainly true during GI1, when the concentration approximately doubled. These increases during warm periods presumably reflect biomass in North America, perhaps related to decreasing ice cover and increasing temperatures. There were also strong increases in ammonium concentration during GI 21 and 23 in MIS 5, presumably reflecting warmer conditions and increased biomass.

3.2. Speed and phasing of different changes at D-O warming events

There should be clues to the processes occurring at rapid climate changes in the phasing of changes in local temperature in Greenland ($\delta^{18}O$), precipitation (layer thickness), input of Asian dust (Ca, dust), and input of sea salt (Na), as well as changes in methane concentration (discussed in Section 3.3). Detailed study of this phasing has been attempted for the final warming out of the Younger Dryas (Taylor et al., 1997; Steffensen et al., 2008), for both ends of GI1 (Steffensen et al., 2008), and for the warming into GI8 (Thomas et al., 2009). Interestingly, deuterium excess seems to undergo the fastest changes, switching within 1–3 years both at warmings, and at the cooling at the end of GI1 into the Younger Dryas (Steffensen et al., 2008). This presumably reflects some fast switch in conditions at the moisture source or of the storm track, although the difficulties in interpreting dxs in Greenland preclude a firm conclusion.

For the cooling at the end of GI1, $\delta^{18}O$, ice layer thickness, and Ca and dust concentration all changed slowly over 1–2 centuries between about 12.9–12.7 ka b2k, with dxs changing rapidly at 12.9 ka b2k (Steffensen et al., 2008). Apart from the early change in dxs, there is no clear phase difference between the other components. For the warmings, changes happen much faster: at the start of GI1 and GI8, the transition in all components is completed within 40 years, with dxs changing particularly fast. Applying ramps to either the linear or logarithmic values at the transitions in each component, it was found that, for the start of GI1, the start of the ramp defining the decrease in dust began earlier than the changes in $\delta^{18}O$ and other components reflecting high latitude processes (Steffensen et al., 2008). A similar conclusion was reached for GI8 (Thomas et al., 2009). It was suggested that such phasing was indicative of changes starting at low latitudes, leading to an initial change in atmospheric circulation, and finally a change in Greenland climate. However, we caution against over-interpreting this result. All the signals being observed are quite noisy, which makes it hard to determine when a transition starts precisely; diffusion of water isotopes in the firn may also matter at very fine resolution. It is not clear if an alternative model to a ramp would still lead to a statistically significant lead for dust. The warmings at the start of GI3 and GI20 have been investigated (Fuhrer et al., 1999) in detail for just $\delta^{18}O$ and Ca: they suggest that a model in which Ca changes in steps might be more appropriate. They confirm that the first step in each change certainly occurs early in the sequence of events, but it would be hard to state clearly that it occurs before other components start to change. More transitions should be investigated, but the evidence suggests that all components change within a very short period at warmings, and that this, rather than a small lead by dust, would be a safer target for models testing causes of the change.

3.3. Methane

Methane is of particular relevance for this journal issue, because its concentration in the atmosphere is clearly strongly connected to aspects of vegetation and soils. Wetlands have long been considered the major natural source of methane, so that changes in wetland extent (and their CH_4 fluxes) or climate must play a role in changing CH_4 concentrations. Biomass burning is another one of the larger natural sources of CH_4. The main sink for CH_4 is oxidation by OH; particularly during glacial periods, when methane concentrations are lower, the OH concentration is sensitive to the concentrations of volatile organic compounds (VOCs) emitted from mainly tropical vegetation. Thus any attempt at bottom-up modelling to understand changes in atmospheric CH_4 across D-O cycles relies on correctly assessing the changing patterns of vegetation across them, with a particular emphasis on wetlands.

Methane concentrations are high during GI and low in GS (Fig. 4) (Brook et al., 1996; Brook et al., 2000; Flückiger et al., 2004), typically with increases of 100–200 ppbv at the transitions between the two. This increase is usually >50% of the glacial-interglacial amplitude; thus understanding the causes of D-O methane variability is essential for understanding glacial-interglacial changes also. Although there is an interhemispheric gradient (higher concentrations in the northern hemisphere), fast changes are seen in both hemispheres, and must occur almost simultaneously; indeed this feature is used to synchronise Antarctic and Greenland ice core chronologies. (e.g. Blunier and Brook, 2001; EPICA Community Members, 2006).

There are a number of important features of the relationship between D-O cycles as seen in CH_4, and as seen in Greenland $\delta^{18}O$. Firstly, the shape of warm events is not the same. Most events in $\delta^{18}O$ are sawtoothed in shape, with highest values at the start, ramping down during the warm event. Although this feature is somewhat modulated in the records of derived temperature shown in Fig. 2, it is still the dominant pattern. CH_4 shows a variety of patterns, including events in which concentrations increase during the warm period: GI1 is the clearest example of this contrasting trend, but it can also be seen in other events. This is perhaps indicative of a threshold behaviour, whereby the D-O switch initiates an increase in methane, and the source can be further enhanced as long as the temperature remains high, even if it is decreasing.

There is no direct link between the magnitude of temperature change and the amplitude of methane change (for times and cores in which there is sufficient resolution in the methane record to capture the full amplitude of change). The methane jump at the start of GI8 is around 150 ppbv, for example, and values in GI8 settle at about 130 ppbv higher than those of the preceding GS (Figs. 4 and 5); in contrast for GI19, with its much larger temperature change (Figs. 2 and 5), the initial methane jump is only around 120 ppbv (Blunier and Brook, 2001; Flückiger et al., 2004; Grachev

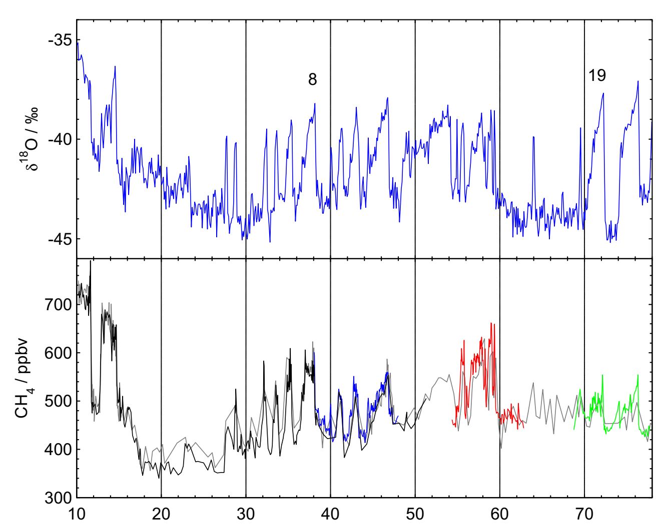

Fig. 4. Methane across D-O cycles. The figure shows methane and Greenland $\delta^{18}O$ across the D-O cycles of the last glacial. Methane data are from GISP2 (grey (Blunier and Brook, 2001)), GRIP (black (Blunier and Brook, 2001, compiling data from earlier papers) and green (Flückiger et al., 2004)), and NorthGRIP (blue (Flückiger et al., 2004) and red (Huber et al., 2006)). Ages have been transferred approximately to the NorthGRIP GICC05 age scale using matches between GRIP and GISP2 (Blunier and Brook, 2001), and between GRIP and NorthGRIP (Huber et al., 2006; Rasmussen et al., 2008). D-O events 8 and 19 are marked. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

et al., 2009), and values in GI19 settle at only about 70 ppbv higher than those of the preceding GS. It has been pointed out that the pattern of the mean amplitude of the CH_4 D-O signal resembles that of precession (Brook et al., 1996; Flückiger et al., 2004).

At a particular depth in an ice core, the gas record is younger than the record of components preserved in ice, because the bubbles only close off at depth (typically 60–100 m, depending on site). However, in Greenland the gas and ice records can be synchronised rather well, because rapid temperature changes are represented both in $\delta^{18}O$ in ice, and in $\delta^{15}N$ in N_2 and $\delta^{40}Ar$ in the gas phase (through the thermal fractionation effect, as discussed in Section 2.1). Using this method, it has been shown that the start of the change in CH_4 at some D-O warming events occurs within 30 years of the start of the change in Greenland temperature (i.e. almost simultaneously) (Severinghaus et al., 1998; Flückiger et al., 2004); for other events, a range of 25–70 years has been quoted (Huber et al., 2006). The CH_4 change at warmings tends to be completed in about a century, and any explanation for the change must reproduce this feature: slow changes as ice sheets retreat cannot be responsible for fast changes of this sort.

For readers who wish to use the date of D-O events given in Table 2, we note that the quoted dates refer approximately to the mid point of the fast rise in $\delta^{18}O$. Since this rise occurs in a few decades, while the rise in CH_4 lasts for around a century, the mid point of the methane rise, will be of order 50 years later than the quoted date (remembering that these dates refer only to the GICC05 age scale, and have a much greater uncertainty than this when converted to other age scales).

There is some limited evidence that constrains the causes of CH_4 changes across D-O cycles. It has been suggested that the sharp increase of CH_4 at the start of GIs is caused by the release of methane from marine hydrates (Kennett et al., 2000). Measurements of δD in CH_4 across the start of GI8, GI1, and the end of the Younger Dryas show a tendency towards more negative values, in contrast to what would be predicted if a release of marine methane hydrates was a significant contributor (Sowers, 2006). Recent measurements of ^{14}C in CH_4 over the Younger Dryas-PreBoreal transition confirm that the additional CH_4 is not mainly of fossil origin (Petrenko et al., 2009). $\delta^{13}CH_4$ decreases significantly during the slow methane increase leading up to the warming event at the start of GI1, but the change at the fast jump in CH_4 is small (1 permil), suggesting a change in the strength of sources with the same isotopic content, and/or changes in sinks, rather than the introduction of a new class of source (Fischer et al., 2008).

The gradient in methane concentration between Greenland and Antarctica can also be used to assess the relative strength of northern high latitude sources (large gradient) and tropical sources (small gradient). Such work suggested that the change between GS

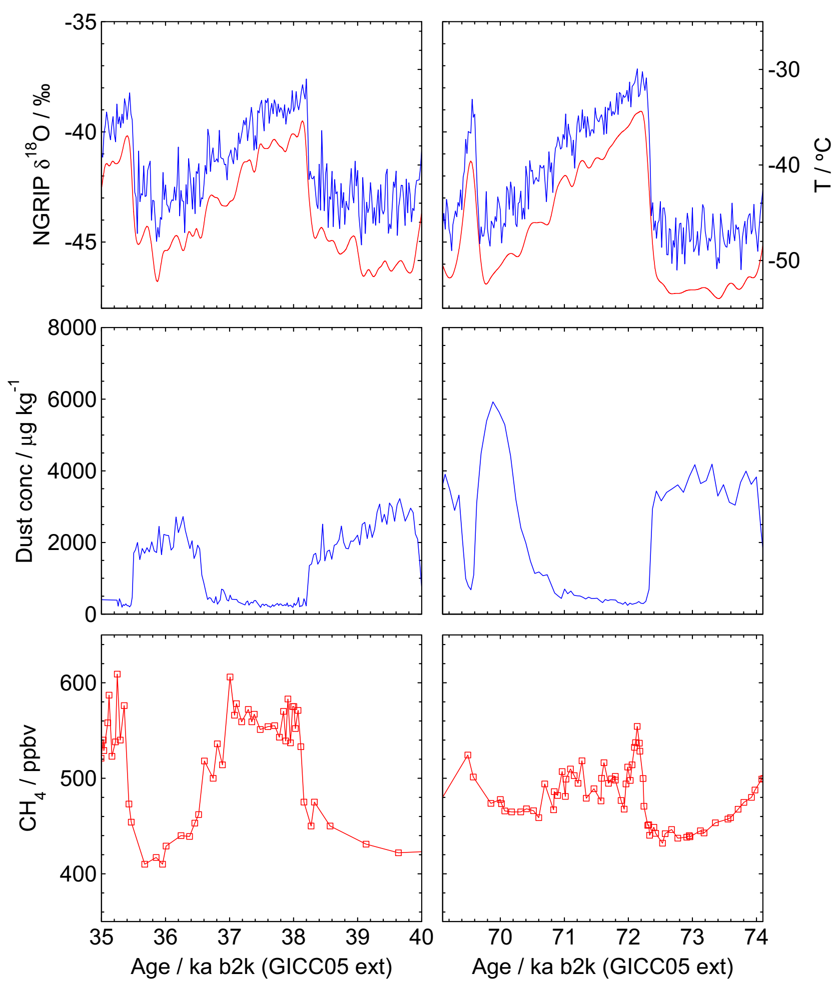

Fig. 5. The contrast between GI8 and GI19. The left hand panels show GI8 and its surrounds, while the right hand panels show GI19 and its surrounds. The top panels show NorthGRIP $\delta^{18}O$ in blue (left hand axis), and the smoother estimated temperature record (Huber et al., 2006) in red (right hand axes). The mid panels are dust (Ruth et al., 2007) from NorthGRIP. The lower panels show CH_4 from GRIP (Blunier and Brook, 2001; Flückiger et al., 2004). All records have been placed on the GICC05 extended age scale. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

and GI methane concentrations, averaged over several events, was mainly due to an increase in high latitude sources (Brook et al., 2000; Dällenbach et al., 2000). On the other hand, similar reasoning suggests that the decrease at the end of GI1 involves mainly tropical effects. The occurrence of rapid changes in $\delta^{18}O$ of O_2 has been used to suggest that significant changes in low latitude monsoon rainfall occurred (Severinghaus et al., 2009) at many D-O cycles, implying that there is a significant tropical contribution to D-O CH_4 increases. Still, a good working hypothesis seems to be that the rapid changes in atmospheric circulation and temperature at high latitudes led to a greatly increased source for wetland emissions, including for example thermokarst lakes (Walter et al., 2007), and that this explains a large part of the D-O changes.

4. Millennial-scale variability in Antarctic ice cores

A much more muted and gradual millennial-scale variability is observed in the $\delta^{18}O$ or δD records of temperature in the last glacial period in Antarctic ice cores (Blunier and Brook, 2001; EPICA Community Members, 2006). It has been possible to synchronise Greenland and Antarctic records using the simultaneous D-O jumps in CH_4 concentration. In fact, this synchronises the gas records; the ice and gas record in Greenland must then be put on a common scale using the finding that the Greenland temperature rise and the CH_4 rise are close to synchronous at the multi-decadal level. There are some issues in synchronising the Antarctic gas and ice age scales, but this error will be rather small for sites such as Byrd Station (Blunier and Brook, 2001) where the snow accumulation rate is high, and the gas age-ice age difference is small. However, the error could be between 400 and 800 years for a site with lower accumulation rate such as EPICA Dronning Maud Land (EDML) (EPICA Community Members, 2006). Around GI10, the Greenland and low-accumulation Antarctic records have recently been synchronised using the ^{10}Be peak associated with the Laschamp event (Raisbeck et al., 2007). This is a promising approach, but has not yet been applied to other records, or to any other time periods in the glacial.

These synchronisation efforts have shown that: (1) there is an Antarctic temperature counterpart to every one of the Greenland D-O warmings (EPICA Community Members, 2006) (Fig. 6); and (2) for the most prominent Antarctic warmings, corresponding to the longest-duration GI, there is a clear phase relationship: while Greenland is cold, Antarctica warms; as soon as Greenland jumps into its GI state, Antarctica cools again (Blunier and Brook, 2001). This phasing leads to a more triangular pattern in Antarctica in contrast to the sawtooth pattern of Greenland, with the top of the Antarctic triangle coinciding with the peak of the Greenland sawtooth. The uncertainty in synchronisation for the EDML record was too great for any conclusion to be drawn about phasing for the shorter GI, but the working assumption is that the same phasing will be seen during these intervals. The larger events in Antarctica were originally called A events (A1–A7); all the Antarctic counterparts have now been re-named as Antarctic Isotopic Maxima (AIM) (EPICA Community Members, 2006), with the numbering set such that AIMn starts during the GS preceding GIn. Thus the rising limb of AIM 1 precedes GI1, which is more or less synchronous with the Antarctic Cold Reversal (ACR).

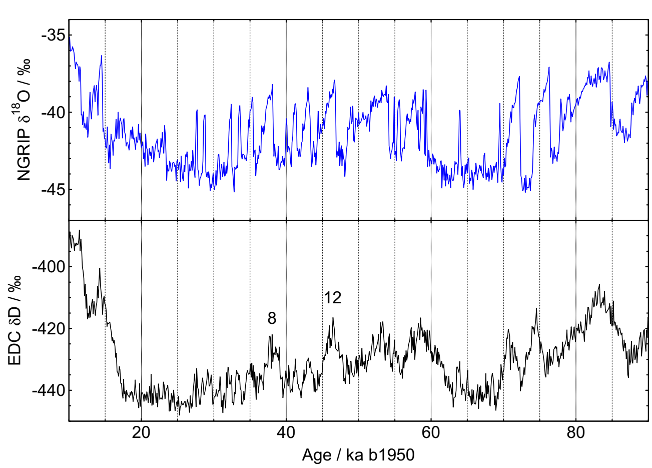

Fig. 6. D-O cycles in Greenland and AIM in Antarctica. NGRIP is shown on the GICC05 age scale, Dome C on the EDC3 age scale (Jouzel et al., 2007a). In this case, no attempt has been made to synchronise the age scales, as was done for shallower parts of this section in earlier work (EPICA Community Members, 2006). Two of the larger AIM are numbered on the figure.

A long cold period in Greenland is associated with a large-amplitude AIM in Antarctica (EPICA Community Members, 2006). This makes sense if Antarctica warms at a more or less constant rate during Greenland cold periods, so that greater amplitudes are reached when warming lasts longer. The pattern of events we have described here is consistent with ideas about a bipolar seesaw, in which reductions in ocean heat transport during GS allow the Southern Ocean to warm gradually, and the resumption of full heat transport during GI leads to a loss of heat from the Southern Ocean heat reservoir (Stocker and Johnsen, 2003).

A number of other parameters measured in Antarctic ice cores vary in time with the temperature proxy – for example the non-sea-salt Ca flux is higher in cold periods (Fischer et al., 2007a), reducing to almost Holocene values in the warm parts of AIM (Röthlisberger et al., 2002). It is important to note that, although dust in Greenland and Antarctica appear in phase at orbital scales, they have the same out-of-phase relationship at millennial-scales as do the respective temperature signals. Of particular interest is the CO_2 concentration, which shows a pattern very similar to Antarctic temperature, at least for the larger AIM (A events) (Ahn and Brook, 2007; Ahn and Brook, 2008). A significant and time-varying lag of CO_2 by 720 ± 370 years was deduced, although the CO_2 and Antarctic temperature maxima appear in phase within the resolution of existing measurements. It is readily seen that CO_2 starts to increase well before the large CH_4 jumps, confirming that CO_2 changes are linked to Southern Ocean changes, and that they are not involved in any direct way as triggers for the D-O warmings. Having said this, we should point out that the mainly slow increase in CO_2 across the last termination does include two rapid periods of change (by a few ppmv) in the CO_2 record, corresponding to the fast temperature rises in Greenland (Monnin et al., 2001).

N_2O is also of great interest, as it is produced both in marine (notably upwelling areas) and continental (mainly tropical) environments, through nitrification and denitrification processes. It shows large changes associated with both small and large GI. The amplitude of N_2O changes appear linearly related with the amplitude of Greenland warmings, but does not bear a precessional modulation such as observed for CH_4 (Flückiger et al., 2004). For long-lasting GI, N_2O starts to increase well before Greenland temperature and CH_4, but later than Antarctic temperature. Such a particular phasing and amplitude of a trace gas emitted both by the ocean and vegetation/soils should make a good test of coupled climate/biogeochemical models (Schmittner and Galbraith, 2008).

5. Context and causes of millennial-scale variability during the past glacial

Variability reminiscent of the Greenland pattern is seen in numerous other records in the northern hemisphere, and the connections between the records allow speculation about mechanisms. It is beyond the scope of this paper to discuss these, but we list some of the most well-known records because correlations with these records allow other stratigraphic markers to be placed in the Greenland ice core stratigraphy.

North Atlantic $\delta^{18}O$ (and SSTs) show very similar variability to Greenland ice cores; as an example we quote the record from MD95-2042 from the Portuguese margin, in which the planktonic $\delta^{18}O$ resembles Greenland ice core $\delta^{18}O$, while the benthic record resembles Antarctic ice core $\delta^{18}O$ or δD (Shackleton et al., 2000). As discussed elsewhere (Sanchez-Goni and Harrison, 2010), the assumption that the North Atlantic planktonic $\delta^{18}O$ jumps are contemporaneous with D-O events allows the Heinrich layers observed in North Atlantic sediments to be placed in the Greenland stratigraphy. Similarly, the $\delta^{18}O$ of speleothems in Hulu Cave in China (Wang et al., 2001) has a very similar pattern to Greenland ice core $\delta^{18}O$, and the assumption of synchroneity in this case allows radiometric dates to be used to check the Greenland age scale. Finally, the pattern of colour reflectance changes and other measures in highly-resolved marine sediments from the Cariaco Basin, Venezuela (Peterson et al., 2000), can also be matched to Greenland ice core $\delta^{18}O$, and this (again making assumptions of synchroneity) opens up an additional possibility to check the ice core dating against ^{14}C-dated material.

The Greenland ice core record does not extend into earlier glacials, so the existence of GI events in earlier periods can only be inferred (Siddall et al., 2006). However, given the strong link between rapid jumps in CH_4 and D-O warming events during the last glacial, and the fact that similar rapid jumps in CH_4 are ubiquitous in the ice core records from Antarctica during earlier glacials (Loulergue et al., 2008), there are strong indications that D-O cycle variability must have occurred in earlier glacials. Additionally, AIM variability similar to that seen in the last glacial is observed throughout the last 800 ka (Jouzel et al., 2007a).

There is also the question of whether millennial-scale variability similar to that of the glacial D-O cycles occurs during the Holocene (Bond et al., 2001; Risebrobakken et al., 2003). There is no sign of any significant variability at millennial-scale, or of changes with the rapidity of D-O warming events, in most of the Holocene in the $\delta^{18}O$ records from Greenland. Millennial-scale variability has been demonstrated using the parameter of xsK (O'Brien et al., 1995), but the meaning of this parameter is not clear. There are however some rapid climate changes in the early Holocene: most notably the 8.2 ka event shows a very clear reduction in $\delta^{18}O$ and other parameters (including one of global character: CH_4) in Greenland ice cores (Thomas et al., 2007), and there is a similar event at 9.3 ka (Rasmussen et al., 2007).

The pattern of change seen in Greenland ice cores in the last glacial period, and the contrasting counterparts seen in Antarctica, are consistent with the idea that the changes are due to changes in the pattern of ocean heat transport (e.g. Stocker and Johnsen, 2003), probably caused by inputs of freshwater or icebergs into the North Atlantic. However, such inputs are best-suited to explaining fast cooling events (such as the 8.2 ka event discussed above), and a spontaneous or forced resumption of a stronger meridional overturning circulation that would have to be responsible for the extremely rapid warming events seen in Greenland.

6. Conclusions and summary

We have summarised what is known about millennial-scale variability during the last glacial period from the ice core record, and documented the pattern and timing of events (Fig. 1 and Table 2). The Greenland stratigraphy provides a compelling template for millennial-scale climate changes during the glacial because it offers numerous sharp and reproducible stratigraphic boundaries. This template can be used only if there is an authoritative and accepted age scale, and the stratigraphy can be transferred to other records. The emergence of the layer-counted GICC05 age scale to 60 ka provides the best available age scale, and is likely to be the standard for some years to come. There remains a problem to decide how to treat the ice older than 60 ka – an age scale that incorporates radiometrically-dated speleothem observations would seem an obvious candidate in the long term.

Multiple parameters, representing different parts of the environmental system and measured in the Greenland core, change almost simultaneously with $\delta^{18}O$. This supports the idea that rapid changes in marine, terrestrial or speleothem archives with a similar pattern to that seen in Greenland can be assumed to be synchronous. Once this assumption is made, it allows the Greenland stratigraphy and chronology to be checked against and transferred to the marine and terrestrial realm. However, transferring the stratigraphy to southern hemisphere records is likely to be much more difficult, because Antarctic ice cores show much more subdued counterparts in a different phase to the Greenland signal.

The record of methane in ice cores shows rapid increases at the same time as the rapid warmings in Greenland. However, the amplitude of methane change at each GI has a different pattern to that of the amplitude of temperature change. N_2O, produced both in the marine and continental realms, also shows large changes when Greenland warms, linearly related with the duration of warming events. Understanding this rapid millennial-scale variability in methane (and nitrous oxide) is essential to understanding the glacial-interglacial changes, and clearly calls for an understanding of how land biosphere emissions varied with time.

Millennial-scale variability is dominant during the last glacial period, and may even hold clues to the termination of glacials (Wolff et al., 2009). The existence of extreme warmings that occur within decades poses questions to modellers, who have conventionally concentrated on the cooling phases, where a clear and testable hypothesis can be applied. Clear documentation of the spatial pattern of change across the events that are so obvious in Greenland is essential if their causes are to be understood.

Acknowledgments

This paper was conceived and developed at workshops sponsored by the UK QUEST (Quantifying Uncertainty in the Earth System) programme of the Natural Environment Research Council in support of the QUEST/INSU project DESIRE (Dynamics of the Earth System and the Ice-core Record, NERC grant NE/E007600/1) and is a contribution to the ongoing work of the QUEST Working Group on Abrupt Climate Change. EWW and JC were funded by QUEST/INSU-DESIRE and by the BAS core project CACHE-DRAM (EWW) and the ANR project IMPACT-BOREAL (JC). We thank Sandy Harrison, Jean Jouzel, Maria Fernanda Sanchez Goñi, Jeff Severinghaus and Renato Spahni for helpful comments.

References

Ahn, J., Brook, E.J., 2008. Atmospheric CO2 and climate on millennial time scales during the last glacial period. Science 322 (5898), 83–85. doi:10.1126/science.1160832.

Ahn, J.H., Brook, E.J., 2007. Atmospheric CO2 and climate from 65 to 30 ka BP. Geophysical Research Letters 34 (10), L10703. doi:10.1029/2007gl029551.

Alley, R.B., Anandakrishnan, S., Jung, P., 2001. Stochastic resonance in the North Atlantic. Paleoceanography 16 (2), 190–198.

Alley, R.B., Shuman, C.A., Meese, D.A., Gow, A.J., Taylor, K.C., Cuffey, K.M., Fitzpatrick, J.J., Grootes, P.M., Zielinski, G.A., Ram, M., Spinelli, G., Elder, B., 1997. Visual-stratigraphic dating of the GISP2 ice core: basis, reproducibility, and application. Journal of Geophysical Research 102 (C12), 26367–26381.

Andersen, K.K., Svensson, A., Johnsen, S., Rasmussen, S.O., Bigler, M., Rothlisberger, R., Ruth, U., Siggaard-Andersen, M.L., Steffensen, J.P., Dahl-Jensen, D., Vinther, B.M., Clausen, H.B., 2006. The Greenland ice core chronology

2005, 15–42 kyr. part i: constructing the time scale. Quaternary Science Reviews 25 (23–24), 3246–3257.

Biscaye, P.E., Grousset, F.E., Revel, M., VanderGaast, S., Zielinski, G.A., Vaars, A., Kukla, G., 1997. Asian provenance of glacial dust (stage 2) in the Greenland Ice Sheet Project 2 Ice Core, Summit, Greenland. Journal of Geophysical Research 102 (C12), 26765–26781.

Björck, S., Walker, M.J.C., Cwynar, C., Johnsen, S., Knudsen, K.-L., Lowe, J.J., Wohlfarth, B., et al., 1998. An event stratigraphy for the last termination in the North Atlantic region based on the Greenland ice-core record: a proposal by the INTIMATE group. Journal of Quaternary Science 13 (4), 283–292.

Blunier, T., Brook, E.J., 2001. Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. Science 291 (5501), 109–112.

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

Bond, G., Kromer, B., Beer, J., Muscheler, R., Evans, M.N., Showers, W., Hoffmann, S., Lotti-Bond, R., Hajdas, I., Bonani, G., 2001. Persistent solar influence on north Atlantic climate during the Holocene. Science 294 (5549), 2130–2136.

Bory, A.J.M., Biscaye, P.E., Piotrowski, A.M., Steffensen, J.P., 2003. Regional variability of ice core dust composition and provenance in Greenland. Geochemistry Geophysics Geosystems 4 (12), 1107. doi:10.1029/2003GC000627.

Braun, H., Christl, M., Rahmstorf, S., Ganopolski, A., Mangini, A., Kubatzki, C., Roth, K., Kromer, B., 2005. Possible solar origin of the 1,470-year glacial climate cycle demonstrated in a coupled model. Nature 438 (7065), 208–211.

Brook, E.J., Harder, S., Severinghaus, J., Steig, E.J., Sucher, C.M., 2000. On the origin and timing of rapid changes in atmospheric methane during the last glacial period. Global Biogeochemical Cycles 14 (2), 559–572.

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

Cuffey, K.M., Clow, G.D., 1997. Temperature, accumulation, and ice sheet elevation in central Greenland through the last deglacial transition. Journal of Geophysical Research 102 (C12), 26383–26396.

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

Dällenbach, A., Blunier, T., Flückiger, J., Stauffer, B., Chappellaz, J., Raynaud, D., 2000. Changes in the atmospheric CH4 gradient between Greenland and Antarctica during the Last Glacial and the transition to the Holocene. Geophysical Research Letters 27 (7), 1005–1008.

Dansgaard, W., Clausen, H.B., Gundestrup, N., Hammer, C.U., Johnsen, S.F., Kristinsdottir, P.M., Reeh, N., 1982. A new Greenland deep ice core. Science 218 (4579), 1273–1277.

Dansgaard, W., Johnsen, S.J., Clausen, H.B., Dahl-Jensen, D., Gundestrup, N.S., Hammer, C.U., Hvidberg, C.S., Steffensen, J.P., Sveinbjornsdottir, A.E., Jouzel, J., Bond, G., 1993. Evidence for general instability of past climate from a 250 kyr ice-core record. Nature 364, 218–220.

de Angelis, M., Steffensen, J.P., Legrand, M., Clausen, H., Hammer, C., 1997. Primary aerosol (sea salt and soil dust) deposited in Greenland ice during the last climatic cycle: comparison with east Antarctic records. Journal of Geophysical Research 102 (C12), 26681–26698.

Ditlevsen, P.D., Andersen, K.K., Svensson, A., 2007. The DO-climate events are probably noise induced: statistical investigation of the claimed 1470 years cycle. Climate of the Past 3 (1), 129–134.

EPICA Community Members, 2006. One-to-one hemispheric coupling of millennial polar climate variability during the last glacial. Nature 444, 195–198.

Fischer, H., Fundel, F., Ruth, U., Twarloh, B., Wegner, A., Udisti, R., Becagli, S., Castellano, E., Morganti, A., Severi, M., Wolff, E.W., Littot, G.C., Rothlisberger, R., Mulvaney, R., Hutterli, M.A., Kaufmann, P., Federer, U., Lambert, F., Bigler, M., Hansson, M., Jonsell, U., de Angelis, M., Gabrielli, P., Boutron, C., Siggaard-Andersen, M.L., Steffensen, J.P., Barbante, C., Gaspari, V., Wagenbach, D., 2007a. Reconstruction of millennial changes in transport, dust emission and regional differences in sea ice coverage using the deep EPICA ice cores from the Atlantic and Indian Ocean sector of Antarctica. Earth and Planetary Science Letters 260 (1–2), 340–354.

Fischer, H., Siggaard-Andersen, M.L., Ruth, U., Rothlisberger, R., Wolff, E.W., 2007b. Glacial-interglacial changes in mineral dust and sea salt records in polar ice cores: sources, transport, deposition. Reviews of Geophysics 45, RG1002. doi:10.1029/2005RG000192.

Fischer, H., Behrens, M., Bock, M., Richter, U., Schmitt, J., Loulergue, L., Chappellaz, J., Spahni, R., Blunier, T., Leuenberger, M., Stocker, T.F., 2008. Changing boreal methane sources and constant biomass burning during the last termination. Nature 452 (7189), 864–867. doi:10.1038/nature06825.

Fleitmann, D., Cheng, H., Badertscher, S., Edwards, R.L., Mudelsee, M., Göktürk, O.M., Fankhauser, A., Pickering, R., Raible, C.C., Matter, A., Kramers, J., Tüysüz, O., 2009. Timing and climatic impact of Greenland interstadials recorded in stalagmites from northern Turkey. Geophysical Research Letters 36 (L19707). doi:10.1029/2009GL040050.

Flückiger, J., Blunier, T., Stauffer, B., Chappellaz, M., Spahni, R., Kawamura, K., Schwander, J., Stocker, T.F., Dahl-Jensen, D., 2004. N2O and CH4 variations during the last glacial epoch: insight into global processes. Global Biogeochemical Cycles 18 (1) GB1020.

Fuhrer, K., Neftel, A., Anklin, M., Staffelbach, T., Legrand, M., 1996. High-resolution ammonium ice core record covering a complete glacial-interglacial cycle. Journal of Geophysical Research 101 (D2), 4147–4164.

Fuhrer, K., Wolff, E.W., Johnsen, S.J., 1999. Timescales for dust variability in the Greenland Ice Core Project (GRIP) ice core in the last 100,000 years. Journal of Geophysical Research 104 (D24), 31043–31052.

Grachev, A.M., Brook, E.J., Severinghaus, J.P., Pisias, N.G., 2009. Relative timing and variability of atmospheric methane and GISP2 oxygen isotopes between 68 and 86 ka. Global Biogeochemical Cycles 23. doi:10.1029/2008gb003330.

Grachev, A.M., Severinghaus, J.P., 2003. Determining the thermal diffusion factor for Ar-40/Ar-36 in air to aid paleoreconstruction of abrupt climate change. Journal of Physical Chemistry A 107 (23), 4636–4642.

Grachev, A.M., Severinghaus, J.P., 2005. A revised $+10 \pm 4\,^{\circ}C$ magnitude of the abrupt change in Greenland temperature at the Younger Dryas termination using published GISP2 gas isotope data and air thermal diffusion constants. Quaternary Science Reviews 24 (5–6), 513–519. doi:10.1016/j.quascirev.2004.10.016.

Grootes, P.M., Stuiver, M., 1997. Oxygen 18/16 variability in Greenland snow and ice with 10(-3)- to 10(5)-year time resolution. Journal of Geophysical Research 102 (C12), 26455–26470.

Grootes, P.M., Stuiver, M., White, J.W.C., Johnsen, S., Jouzel, J., 1993. Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores. Nature 366, 552–554.

Hansson, M.E., 1994. The Renland ice core: a Northern Hemisphere record of aerosol composition over 120,000 years. Tellus 46B, 390–418.

Huber, C., Leuenberger, M., Spahni, R., Flückiger, J., Schwander, J., Stocker, T.F., Johnsen, S., Landals, A., Jouzel, J., 2006. Isotope calibrated Greenland temperature record over Marine Isotope Stage 3 and its relation to CH4. Earth and Planetary Science Letters 243 (3–4), 504–519.

Johnsen, S.J., Clausen, H.B., Dansgaard, W., Fuhrer, K., Gundestrup, N., Hammer, C.U., Iversen, P., Jouzel, J., Stauffer, B., Steffensen, J.P., 1992. Irregular glacial interstadials recorded in a new Greenland ice core. Nature 359, 311–313.

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

Johnsen, S.J., Dahl-Jensen, D., Gundestrup, N., Steffensen, J.P., Clausen, H.B., Miller, H., Masson-Delmotte, V., Sveinbjornsdottir, A.E., White, J., 2001. Oxygen isotope and palaeotemperature records from six Greenland ice-core stations: camp Century, Dye-3, GRIP, GISP2, Renland and NorthGRIP. Journal of Quaternary Science 16 (4), 299–307.

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

Jouzel, J., Masson-Delmotte, V., Cattani, O., Dreyfus, G., Falourd, S., Hoffmann, G., Nouet, J., Barnola, J.M., Chappellaz, J., Fischer, H., Gallet, J.C., Johnsen, S., Leuenberger, M., Loulergue, L., Luethi, D., Oerter, H., Parrenin, F., Raisbeck, G., Raynaud, D., Schwander, J., Spahni, R., Souchez, R., Selmo, E., Schilt, A., Steffensen, J.P., Stenni, B., Stauffer, B., Stocker, T., Tison, J.-L., Werner, M., Wolff, E.W., 2007a. Orbital and millennial Antarctic climate variability over the last 800,000 years. Science 317, 793–796.

Jouzel, J., Stievenard, M., Johnsen, S.J., Landais, A., Masson-Delmotte, V., Sveinbjornsdottir, A., Vimeux, F., von Grafenstein, U., White, J.W.C., 2007b. The GRIP deuterium-excess record. Quaternary Science Reviews 26 (1–2), 1–17.

Jouzel, J., Vimeux, F., Caillon, N., Delaygue, G., Hoffmann, G., Masson-Delmotte, V., Parrenin, F., 2003. Magnitude of isotope/temperature scaling for interpretation of central Antarctic ice cores. Journal of Geophysical Research 108 (D12), 4361. doi:10.1029/2002JD002677.

Kapsner, W.R., Alley, R.B., Shuman, C.A., Anandakrishnan, S., Grootes, P.M., 1995. Dominant influence of atmospheric circulation on snow accumulation in Greenland over the past 18,000 years. Nature 373, 52–54.

Kennett, J.P., Cannariato, K.G., Hendy, I.L., Behl, R.J., 2000. Carbon isotopic evidence for methane hydrate instability during quaternary interstadials. Science 288 (5463), 128–133.

Krinner, G., Werner, M., 2003. Impact of precipitation seasonality changes on isotopic signals in polar ice cores: a multi-model analysis. Earth and Planetary Science Letters 216, 525–538.

Landais, A., Barnola, J.M., Masson-Delmotte, V., Jouzel, J., Chappellaz, J., Caillon, N., Huber, C., Leuenberger, M., Johnsen, S.J., 2004a. A continuous record of temperature evolution over a sequence of Dansgaard-Oeschger events during Marine Isotopic Stage 4 (76 to 62 kyr BP). Geophysical Research Letters 31 (22). doi:10.1029/2004gl021193.

Landais, A., Caillon, N., Goujon, C., Grachev, A.M., Barnola, J.M., Chappellaz, J., Jouzel, J., Masson-Delmotte, V., Leuenberger, M., 2004b. Quantification of rapid temperature change during DO event 12 and phasing with methane inferred from air isotopic measurements. Earth and Planetary Science Letters 225 (1–2), 221–232. doi:10.1016/j.espl.2004.06.009.

Lang, C., Leuenberger, M., Schwander, J., Johnsen, S., 1999. $16\,^{\circ}C$ rapid temperature variation in central greenland 70,000 years ago. Science 286, 934–937.

Lisiecki, L.E., Raymo, M.E., 2005. A Pliocene-Pleistocene stack of 57 globally distributed benthic delta O-18 records. Paleoceanography 20 (1), PA1003. doi:1010.1029/2004PA001071.

Loulergue, L., Schilt, A., Spahni, R., Masson-Delmotte, V., Blunier, T., Lemieux, B., Barnola, J.M., Raynaud, D., Stocker, T.F., Chappellaz, J., 2008. Orbital and millennial-scale features of atmospheric CH4 over the last 800,000 years. Nature 453, 383–386.

Lowe, J.J., Rasmussen, S.O., Bjorck, S., Hoek, W.Z., Steffensen, J.P., Walker, M.J.C., Yu, Z.C., Grp, I., 2008. Synchronisation of palaeoenvironmental events in the North Atlantic region during the Last Termination: a revised protocol recommended by the INTIMATE group. Quaternary Science Reviews 27 (1–2), 6–17. doi:10.1016/j.quascirev.2007.09.016.

Masson-Delmotte, V., Jouzel, J., Landais, A., Stievenard, M., Johnsen, S.J., White, J.W.C., Werner, M., Sveinbjornsdottir, A., Fuhrer, K., 2005. GRIP deuterium excess reveals rapid and orbital-scale changes in Greenland moisture origin. Science 309 (5731), 118–121.

Mayewski, P.A., Meeker, L.D., Twickler, M.S., Whitlow, S., Yang, Q.Z., Lyons, W.B., Prentice, M., 1997. Major features and forcing of high-latitude northern hemisphere atmospheric circulation using a 110,000-year-long glaciochemical series. Journal of Geophysical Research 102 (C12), 26345–26366.

Mayewski, P.A., Meeker, L.D., Whitlow, S., Twickler, M.S., Morrison, M.C., Bloomfield, P., Bond, G.C., Alley, R.B., Gow, A.J., Grootes, P.M., Meese, D.A., Ram, M., Taylor, K.C., Wumkes, W., 1994. Changes in atmospheric circulation and ocean ice cover over the North Atlantic during the last 41,000 years. Science 263, 1747–1751.

Meese, D.A., Gow, A.J., Alley, R.B., Zielinski, G.A., Grootes, P.M., Ram, M., Taylor, K.C., Mayewski, P.A., Bolzan, J.F., 1997. The Greenland Ice Sheet Project 2 depth-age scale: methods and results. Journal of Geophysical Research 102 (C12), 26411–26423.

Monnin, E., Indermuhle, A., Dallenbach, A., Fluckiger, J., Stauffer, B., Stocker, T.F., Raynaud, D., Barnola, J.M., 2001. Atmospheric CO2 concentrations over the last glacial termination. Science 291 (5501), 112–114.

North Greenland Ice Core Project Members, 2004. High-resolution record of Northern Hemisphere climate extending into the last interglacial period. Nature 431 (7005), 147–151.

O'Brien, S.R., Mayewski, P.A., Meeker, L.D., Meese, D.A., Twickler, M.S., Whitlow, S.I., 1995. Complexity of Holocene climate as reconstructed from a Greenland ice core. Science 270, 1962–1964.

Peterson, L.C., Haug, G.H., Hughen, K.A., Rohl, U., 2000. Rapid changes in the hydrologic cycle of the tropical Atlantic during the last glacial. Science 290 (5498), 1947–1951.

Petrenko, V.V., Smith, A.M., Brook, E.J., Lowe, D., Riedel, K., Brailsford, G., Hua, Q., Schaefer, H., Reeh, N., Weiss, R.F., Etheridge, D., Severinghaus, J.P., 2009. $^{14}CH_4$ measurements in Greenland Ice: investigating Last Glacial Termination CH_4 Sources. Science 324 (5926), 506–508. doi:10.1126/science.1168909.

Rahmstorf, S., 2003. Timing of abrupt climate change: a precise clock. Geophysical Research Letters 30 (10), 1510. doi:10.1029/2003GL017115.

Raisbeck, G.M., Yiou, F., Jouzel, J., Stocker, T.F., 2007. Direct North-South synchronization of abrupt climate change record in ice cores using Beryllium 10. Climate of the Past 3, 541–547.

Rasmussen, S.O., Andersen, K.K., Svensson, A.M., Steffensen, J.P., Vinther, B.M., Clausen, H.B., Siggaard-Andersen, M.L., Johnsen, S.J., Larsen, L.B., Dahl-Jensen, D., Bigler, M., Rothlisberger, R., Fischer, H., Goto-Azuma, K., Hansson, M.E., Ruth, U., 2006. A new Greenland ice core chronology for the last glacial termination. Journal of Geophysical Research 111 (D6), D061202. doi:10.1029/2005JD006079.

Rasmussen, S.O., Selerstad, I.K., Andersen, K.K., Bigler, M., Dahl-Jensen, D., Johnsen, S.J., 2008. Synchronization of the NGRIP, GRIP, and GISP2 ice cores across MIS 2 and palaeoclimatic implications. Quaternary Science Reviews 27 (1–2), 18–28. doi:10.1016/j.quascirev.2007.01.016.

Rasmussen, S.O., Vinther, B.M., Clausen, H.B., Andersen, K.K., 2007. Early Holocene climate oscillations recorded in three Greenland ice cores. Quaternary Science Reviews 26, 1907–1914. doi:10.1016/j.quascirev.2007.06.015.

Risebrobakken, B., Jansen, E., Andersson, C., Mjelde, E., Hevroy, K., 2003. A high-resolution study of Holocene paleoclimatic and paleoceanographic changes in the Nordic Seas. Paleoceanography 18 (1), 1017.

Röthlisberger, R., Mulvaney, R., Wolff, E.W., Hutterli, M., Bigler, M., Sommer, S., Jouzel, J., 2002. Dust and sea salt variability in central East Antarctica (Dome C) over the last 45 kyrs and its implications for southern high-latitude climate. Geophysical Research Letters 29 (20), 1963. doi:10.1029/2002GL015186.

Rousseau, D.D., Kukla, G., McManus, J., 2006. What is what in the ice and the ocean? Quaternary Science Reviews 25 (17–18), 2025–2030.

Ruth, U., Bigler, M., Rothlisberger, R., Siggaard-Andersen, M.L., Kipfstuhl, S., Goto-Azuma, K., Hansson, M.E., Johnsen, S.J., Lu, H.Y., Steffensen, J.P., 2007. Ice core evidence for a very tight link between North Atlantic and east Asian glacial climate. Geophysical Research Letters 34 (3), L03706. doi:10.1029/2006GL027876.

Sanchez-Goni, M.F., Harrison, S.P., 2010. Millennial-scale climate variability and vegetation changes during the Last Glacial: Concepts and terminology. Quaternary Science Reviews 29 (21-22), 2823–2827.

Sanchez-Goni, M.F., Landais, A., Fletcher, W.J., Naughton, F., Desprat, S., Duprat, J., 2008. Contrasting impacts of Dansgaard-Oeschger events over a western European latitudinal transect modulated by orbital parameters. Quaternary Science Reviews 27 (11–12), 1136–1151.

Schmittner, A., Galbraith, E.D., 2008. Glacial greenhouse-gas fluctuations controlled by ocean circulation changes. Nature 456, 373–376.

Schulz, M., 2002. On the 1470-year pacing of Dansgaard-Oeschger warm events. Paleoceanography 17 (2). doi:10.1029/2000pa000571.

Severinghaus, J.P., Beaudette, R., Headly, M.A., Taylor, K., Brook, E.J., 2009. Oxygen-18 of O_2 records the impact of abrupt climate change on the terrestrial biosphere. Science 324 (5933), 1431–1434. doi:10.1126/science.1169473.

Severinghaus, J.P., Brook, E.J., 1999. Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice. Science 286, 930–934.

Severinghaus, J.P., Sowers, T., Brook, E.J., Alley, R.B., Bender, M.L., 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.

Shackleton, N.J., Fairbanks, R.G., Chiu, T.C., Parrenin, F., 2004. Absolute calibration of the Greenland time scale: implications for Antarctic time scales and for Delta C-14. Quaternary Science Reviews 23 (14–15), 1513–1522.

Shackleton, N.J., Hall, M.A., Vincent, E., 2000. Phase relationships between millennial-scale events 64,000–24,000 years ago. Paleoceanography 15 (6), 565–569.

Siddall, M., Stocker, T.F., Blunier, T., Spahni, R., McManus, J.F., Bard, E., 2006. Using a maximum simplicity paleoclimate model to simulate millennial variability during the last four glacial periods. Quaternary Science Reviews 25 (23–24), 3185–3197. doi:10.1016/j.quascirev.2005.12.014.

Singer, B.S., Guillou, H., Jicha, B.R., Laj, C., Kissel, C., Beard, B.L., Johnson, C.M., 2009. $^{40}Ar/^{39}Ar$, K–Ar and $^{230}Th–^{238}U$ dating of the Laschamp excursion: a radiometric tie-point for ice core and climate chronologies. Earth and Planetary Science Letters 286 (1–2), 80–88. doi:10.1016/j.epsl.2009.06.030.

Skinner, L.C., 2008. Revisiting the absolute calibration of the Greenland ice-core age-scales. Climate of the Past 4 (4), 295–302.

Southon, J., 2004. A radiocarbon perspective on Greenland ice-core chronologies: can we use ice cores for C-14 calibration? Radiocarbon 46 (3), 1239–1259.

Sowers, T., 2006. Late quaternary atmospheric CH4 isotope record suggests marine clathrates are stable. Science 311 (5762), 838–840.

Steffensen, J.P., Andersen, K.K., Bigler, M., Clausen, H.B., Dahl-Jensen, D., Fischer, H., Goto-Azuma, K., Hansson, M., Johnsen, S.J., Jouzel, J., Masson-Delmotte, V., Popp, T., Rasmussen, S.O., Rothlisberger, R., Ruth, U., Stauffer, B., Siggaard-Andersen, M.L., Sveinbjírnsdottir, A.E., Svensson, A., White, J.W.C., 2008. High-resolution Greenland ice core data show abrupt climate change happens in few years. Science 321, 680–684.

Stocker, T.F., Johnsen, S.J., 2003. A minimum thermodynamic model for the bipolar seesaw. Paleoceanography 18 (4), 1087.

Svensson, A., Andersen, K.K., Bigler, M., Clausen, H.B., Dahl-Jensen, D., Davies, S.M., Johnsen, S.J., Muscheler, R., Parrenin, F., Rasmussen, S.O., Rothlisberger, R., Seierstad, I., Steffensen, J.P., Vinther, B.M., 2008. A 60,000 year Greenland stratigraphic ice core chronology. Climate of the Past 4, 47–57.

Svensson, A., Andersen, K.K., Bigler, M., Clausen, H.B., Dahl-Jensen, D., Davies, S.M., Johnsen, S.J., Muscheler, R., Rasmussen, S.O., Rothlisberger, R., Steffensen, J.P., Vinther, B.M., 2006. The Greenland Ice Core Chronology 2005, 15–42 ka. Part 2: comparison to other records. Quaternary Science Reviews 25 (23–24), 3258–3267.

Taylor, K.C., Mayewski, P.A., Alley, R.B., Brook, E.J., Gow, A.J., Grootes, P.M., Meese, D.A., Saltzman, E.S., Severinghaus, J.P., Twickler, M.S., White, J.W.C., Whitlow, S., Zielinski, G.A., 1997. The Holocene-Younger Dryas transition recorded at Summit, Greenland. Science 278, 825–827.

Thomas, E.R., Wolff, E.W., Mulvaney, R., Steffensen, J.P., Johnsen, S.J., Arrowsmith, C., 2009. Anatomy of a Dansgaard-Oeschger warming transition – high-resolution analysis of the NGRIP ice core. Journal of Geophysical Research 114, D08102. doi:10.1029/2008JD011215.

Thomas, E.R., Wolff, E.W., Mulvaney, R., Steffensen, J.P., Johnsen, S.J., Arrowsmith, C., White, J.W.C., Vaughn, B., Popp, T., 2007. The 8.2 kyr event from Greenland ice cores. Quaternary Science Reviews 26 (1–2), 70–81. doi:10.1016/j.quascirev.2006.07.017.

Vinther, B.M., Clausen, H.B., Johnsen, S.J., Rasmussen, S.O., Andersen, K.K., Buchardt, S.L., Dahl-Jensen, D., Seierstad, I.K., Siggaard-Andersen, M.L., Steffensen, J.P., Svensson, A., Olsen, J., Heinemeier, J., 2006. A synchronized dating of three Greenland ice cores throughout the Holocene. Journal of Geophysical Research 111 (D13), D13102. doi:10.1029/2005JD006921.

Walter, K.M., Edwards, M.E., Grosse, G., Zimov, S.A., Chapin, F.S., 2007. Thermokarst lakes as a source of atmospheric CH4 during the last deglaciation. Science 318 (5850), 633–636.

Wang, Y.J., Cheng, H., Edwards, R.L., An, Z.S., Wu, J.Y., Shen, C.C., Dorale, J.A., 2001. A high-resolution absolute-dated Late Pleistocene monsoon record from Hulu Cave, China. Science 294 (5550), 2345–2348.

Wolff, E.W., Fischer, H., Rothlisberger, R., 2009. Glacial terminations as southern warmings without northern control. Nature Geoscience 2, 206–209.

Wolff, E.W., Moore, J.C., Clausen, H.B., Hammer, C.U., 1997. Climatic implications of background acidity and other chemistry derived from electrical studies of the Greenland Ice Core Project ice core. Journal of Geophysical Research 102 (C12), 26325–26332.