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Greenland ice core records and rapid climate change

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Long ice cores from Greenland yield records of annually resolved climate change for the past ten to twenty thousand years, and decadal resolution for one hundred thousand years or more. These cores are ideally suited to determine the rapidity with which major climate changes occur. The termination of the Younger Dryas, which marks the end of the last glacial period, appears to have occurred in less than a human lifetime in terms of oxygen isotopic evidence (a proxy for temperature), in less than a generation (20 years) for dust content and deuterium excess (proxies for winds and sea-surface conditions), and in only a few years for the accumulation rate of snow. Similarly rapid changes have been observed for stadial-interstadial climate shifts (Dansgaard–Oeschger cycles) which punctuate the climate of the last glacial period. These changes appear to be too rapid to be attributed to external orbital forcings, and may result from internal instabilities in the Earth's atmosphere–ocean system or periodic massive iceberg discharges associated with ice sheet instability (Heinrich events). In contrast, the Holocene climate of the Arctic appears to have been relatively stable. However, the potential for unstable interglacials, with very rapid, shortlived climatic deteriorations, has been raised by results from the lower part of the GRIP ice core. These results have not been confirmed by other ice cores, notably the nearby GISP2 core. Evidence from other records of climate during the Eemian interglacial have yielded mixed results, and the potential for rapid climate change during interglacial periods remains one of the most intriguing gaps in our understanding of the nature of major Quaternary climate change.

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1. Introduction

Ice cores taken from the crests of large ice masses provide a continuous and high resolution record of climate change. Time series of up to several hundred thousand years, on a number of environmentally related parameters, have been obtained from ice cores over 3 km long drilled recently through the Greenland and Antarctic ice sheets (GRIP project members 1993; Grootes *et al.* 1993; Jouzel *et al.* 1993). Each successive annual increment of snowfall is buried and preserved in the cores, together with any contaminants, and variations in parameter values can be used to infer the nature of past environmental changes; for example, the oxygen isotopic composition of the ice yields information on palaeo-temperatures, air-bubbles provide evidence on the changing composition of the atmosphere, acidity levels record past volcanic

eruptions, and dust levels provide an index of storminess (Johnsen *et al.* 1995). It is the combination of long record length and high time resolution that makes ice cores one of the most effective sources of evidence on both the nature and the rate of environmental change on timescales of up to 10^5 years.

Two deep ice cores, the Greenland ice-core project (GRIP) and Greenland Ice Sheet project 2 (GISP2) cores, have recently been drilled by European and American teams at or close to the crest of the Greenland Ice Sheet. In this paper, the environmental record provided by these two Arctic ice cores is summarized, with particular emphasis on the very abrupt way in which climatic parameters have switched between relative warmth and cold on several timescales over the last 100–150 000 years. The rapidity and magnitude of these climatic shifts is in marked contrast to the relatively more subdued climatic changes of the past 10 000 years, and has served to emphasize the likely existence of significant thresholds in the behaviour of the linked atmosphere-ocean-cryosphere system. The relationship between these changes in climate at the summit of the Greenland Ice Sheet and the time-dependent behaviour of the North Atlantic ocean circulation is also discussed, together with possible correlations with climate change in North-West Europe.

2. The GRIP and GISP2 ice core records

(*a*) *Core sites and chronology*

The GRIP and GISP2 ice cores were drilled at or close to Summit, in central Greenland, in the early 1990s. The surface elevation at these sites is about 3200 m and the ice is over 3000 m thick. The GRIP core site is at the modern ice divide, whereas the GISP2 site is 28 km to the west. Optimal ice core sites are placed exactly at the ice crest, because here there is no horizontal component to ice flow, and thus ice deformation, especially close to the bed, does not take place as long as the ice divide itself does not migrate through time.

The timescales for the two ice cores are derived from the counting of annual layers of snow accumulation for the upper part of each core (e.g. back to about 14 500 years ago for GRIP). Deeper in the cores a modified steady-state ice flow model is used to calculate the age-depth relationship (i.e. the rate of core thinning resulting from depth-dependent vertical strain), assuming that the stratigraphy is undisturbed. Again taking the GRIP core as an example, model calculations are pinned to two characteristic features observed in other dated climate records (figure 1); first, the termination of the Younger Dryas event at 11 500 years ago and, secondly, marine isotope stage 5d at 110 000 years ago (Dansgaard *et al.* 1993).

(*b*) *General description of ice core stratigraphy*

Oxygen isotopic values and electrical conductivity measurements (ECM) for the whole length of the GRIP and GISP2 ice cores have been presented and compared in some detail (Dansgaard *et al.* 1993; Grootes *et al.* 1993; Taylor *et al.* 1993). The oxygen isotope profiles are expressed as $\partial^{18}O$ values, representing the relative difference in parts per thousand between the $^{18}O/^{16}O$ abundance ratios of ice core samples and standard mean ocean water (SMOW). The $\partial^{18}O$ value of snow is taken to reflect its temperature of formation, with higher $\partial^{18}O$ values indicating higher temperatures (Johnsen *et al.* 1989). ECM data vary with ice acidity, which can reflect volcanic
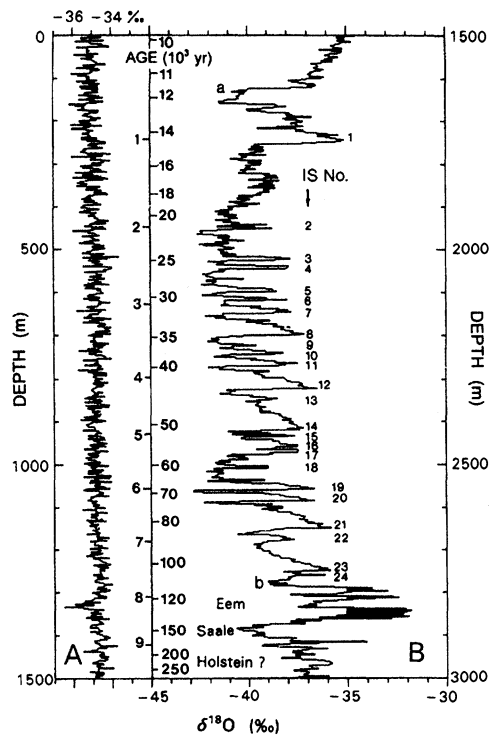
Figure 1. The GRIP $\partial^{18}O$ record plotted on a linear depth scale (from Dansgaard *et al.* 1993 and Johnsen *et al.* 1995). A represents the Holocene record, established by counting annual ice layers. B is the record from 10 000 years to more than 250 000 years, with a chronology derived from ice flow modelling. IS are interstadials.

events and changes in the concentration of alkaline dust (Taylor *et al.* 1993). Colder periods tend to be more dusty, leading to ice of lower conductivity.

The $\partial^{18}O$ and ECM records from the two Greenland drill sites are consistent for the upper 90% of the cores, but vary markedly in the 10% closest to the ice sheet bed (Grootes *et al.* 1993; Taylor *et al.* 1993). The core records can be divided into three parts. First, the upper 1500 m or so of the cores represents the Holocene or last 10 000 years. It is an interval of relatively stable climate, with mean $\partial^{18}O$ values of −34.7 ppt and −34.9 ppt for the GISP2 and GRIP cores, respectively (Grootes *et al.* 1993). The Holocene warm period is separated from the underlying isotopically more negative ice, indicating colder conditions, by abrupt shifts in both $\partial^{18}O$ and ECM values (figure 1). Detailed correlation between the two cores within the Holocene is difficult due to the small magnitude of Holocene isotopic fluctuations (less than 2 ppt). When compared with the remainder of the cores, the Holocene is seen as a period of unusual climatic stability (figure 1).

The second part of the cores is that between about 1500 m and 2750 m in depth, or about 103 000 years ago according to the GRIP timescale (Johnsen *et al.* 1995). Oxygen isotopic and ECM evidence suggests very much more variable climatic conditions for this part of both cores, where inter-core correlation is good down to 2750 m. This interval represents the last glaciation, with a series of colder stadials and warmer interstadials, similar to marine oxygen isotope stages 2 to 5d. Each of the 24 interstadials (IS) identified in the GRIP core is labelled in Figure 1. Only the most recent
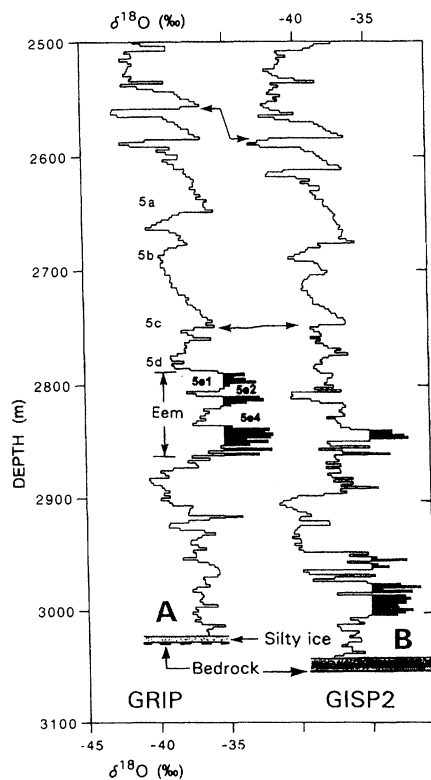
Figure 2. Comparison of the GRIP (A) and GISP2 (B) $\partial^{18}O$ records from the deepest parts of the ice cores (from Johnsen *et al.* 1995). $\partial^{18}O$ values higher than today are shaded black. Substages within marine isotope stage 5 are indicated. Note that ice of high $\partial^{18}O$ values is much closer to bedrock in the GRIP 2 than in the GRIP core (Johnsen *et al.* 1995).

interstadial, beginning at about 14 500 years ago, approaches the isotopic values of the Holocene, and this is rapidly replaced by the isotopically more negative, cold conditions of the Younger Dryas (figure 1).

The remainder of both cores, below about 2750 m or so, shows a period over which oxygen isotope values return to a level equal to and sometimes higher than those of the Holocene. This is interpreted as the last, or Eemian interglacial, occurring at between about 115–130 000 years ago. Below this, and closest to the base of the cores, is ice more negative in $\partial^{18}O$. This is assumed to represent older ice from the Saalian glaciation and beyond.

Three key points can be made about the Greenland ice core records of the Eemian interglacial. First, and arguably the most novel inference from the $\partial^{18}O$ record in the GRIP ice core, was that of continuing climatic instability within the Eemian (figure 1). Unlike the Holocene, the last interglacial contained rapid isotopic shifts of up to 5 per mille, and the colder parts of the Eemian were isotopically similar to interstadials within the last glacial. This represents temperatures about 5 °C cooler than the Holocene (GRIP project members 1993). Secondly, the three warmest intervals of the Eemian were up to 4–5 °C warmer than those of today (Johnsen *et al.* 1995). These two contrasts between the Holocene and the Eemian appeared to have occurred despite the fact that the isotopic (e.g. deuterium excess) and chemical (e.g. concentrations of sodium, magnesium, sulphate and nitrate) signatures of the warm

stages of the Eemian are similar to those of the Holocene, suggesting a generally similar atmospheric circulation.

A third point, however, has cast some doubt on the interpretation of the Eemian record in the GRIP ice core. When the ∂^{18}O and ECM records from the GRIP and GISP2 cores were compared, there was a lack of detailed correlation below about 2750 m (figure 2), although some general trends are common to both cores (Grootes *et al.* 1993; Taylor *et al.* 1993). This implies that ice flow deformation is likely to have affected the chronology and stratigraphy of the lowest 10% of one or both of the cores in order to account for these observed differences. Inclined ice layering, observed below 2678 m and 2847 m in the GISP2 and GRIP cores, respectively, may be an indicator of deformed ice (Grootes *et al.* 1993; Taylor *et al.* 1993). The modern location of the GRIP core at the exact crest of the Greenland Ice Sheet means it is less susceptible to folding at depth than the GISP2 site, but modelling suggests that during the last glacial period the ice divide may have migrated eastwards by 10–50 km, also exposing the GRIP site to deformation close to the bed (Anandakrishnan *et al.* 1994). It appears likely that the GISP2 core is deformed from above the level at which Eemian ice first occurs, but Johnsen *et al.* (1995) argue that no large-scale overturn-folding appears to have disturbed the Eemian isotopic record in the GRIP core (because no two layers of exactly similar isotopic and chemical composition have been found). The GRIP core may therefore be a valid representation of climate change at least part-way into the Eemian above 2848 m (Johnsen *et al.* 1995). However, it is clear that additional work is required on the structural glaciology of the lowermost parts of both cores before the full climatic significance of the observed parameter variations can be assessed (Boulton 1993).

3. Abrupt environmental change and the ice core records

Evidence derived recently from both ice cores and marine sediments in Greenland and the North Atlantic region, respectively, suggests that marked climatic and other environmental changes have taken place over relatively short periods. The identification of layers of iceberg-rafted debris, 'Heinrich layers', in North Atlantic sediment cores at several times during the last 50 000 years is an example of the rapid collapse of the North American Laurentide Ice Sheet and the release of very large numbers of debris-laden icebergs into the North Atlantic (Bond *et al.* 1992). The duration of such events is on the order of 250–1000 years (MacAyeal 1993; Dowdeswell *et al.* 1995). However, it is the finer resolution provided by the ice-core record that allows the most detailed examination of the rapid nature of Late Quaternary climate change, demonstrating that a number of major climatic shifts in the North Atlantic region have occurred on timescales of clear relevance to human activities, sometimes as short as decades.

(a) The Younger Dryas

The time resolution of the GRIP and GISP2 ice cores has a relative accuracy of 1% or better over centuries at the depth of the Younger Dryas (about 1600 to 1700 m). The Younger Dryas, the last stadial before the Holocene, is bounded above and below by sharp changes in annual layer thickness, ∂^{18}O values and dust content and is about 1300 years in duration (figure 3). The upper boundary is at 11 640 ± 250 or 11 550 ± 70 calendar years BP (i.e. before 1950), according the GISP2 and GRIP timescales, respectively (Alley *et al.* 1993; Johnsen *et al.* 1992). The Younger Dryas
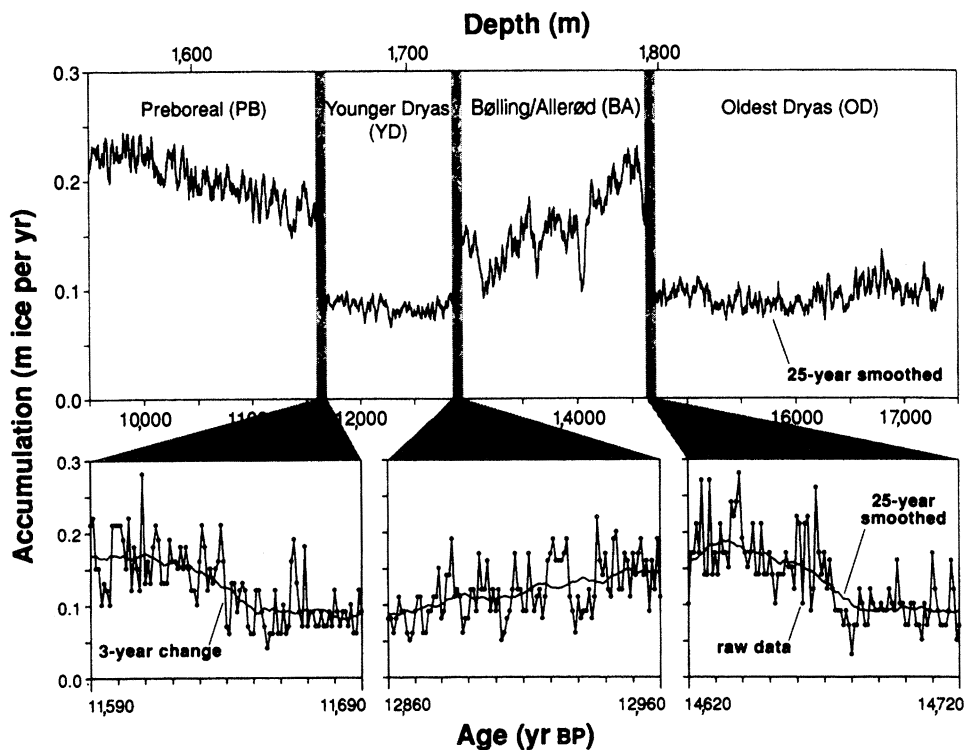
Figure 3. Accumulation rates in the GISP2 ice core between 10 000 and 17 000 years BP (from Alley *et al.* 1993). 25-year running means for accumulation are shown throughout the figure, with annual data superimposed in the lower three panels.

has more negative isotopic values, more dust and thinner ice layers than either the Preboreal above it or the underlying Bølling/Allerød interstadial (figure 3).

Alley *et al.* (1993) have emphasized the very abrupt termination of the Younger Dryas, pointing out that $\partial^{18}O$ values suggest termination over a period of about 50 years and that changing dust concentrations indicate a transition on the order of 20 years (Johnsen *et al.* 1992; Taylor *et al.* 1993). These rates confirm those seen in the Dye 3 core further south in Greenland by Dansgaard *et al.* (1989). However, detailed examination of changes in the thickness of annual ice layers, corrected for ice flow at depth, indicates a doubling of snow accumulation from the Younger Dryas to the Preboreal in as little as only three years (Alley *et al.* 1993). By a more conservative analysis, using 25 year running means for accumulation rates, accumulation doubled in 41 years. It is also pointed out that this very abrupt change is not a result of either enhanced shear-strain or thrust-faulting in this part of the ice core. The increase in accumulation rate between glacial conditions and the previous, Bølling/Allerød interstadial was also very abrupt, whereas the termination of that interstadial, marking the onset of the Younger Dryas, was less marked (figure 3).

Alley *et al.* (1993) suggest that the very rapidity of these shifts in environmental indicators places significant constraints on the nature of the climatic changes that can be responsible for them. Changes in accumulation may be linked to changing atmospheric dynamics, for example the duration or number of weather systems passing

over Greenland, and to changes in the thermohaline circulation in the North Atlantic (Alley *et al.* 1993).

(b) *The last glaciation*

Isotopic values for the last glaciation switch between 24 intervals of relatively high and low $\partial^{18}O$ (figure 1), each lasting between several hundred to a few thousand years (Grootes *et al.* 1993). Variations are on the order of 4-6 per mille, which implies a temperature change of 7–8 °C. Full glacial conditions at the core sites were 10–13 °C colder than during the Holocene (Grootes *et al.* 1993). The intervals of less negative $\partial^{18}O$ values, representing warmer interstadials, lasted between about 500 and 2000 years (Johnsen *et al.* 1992). The isotopic records from the new central Greenland ice cores correlate well with similar evidence from other parts of the Greenland Ice Sheet, where shorter $\partial^{18}O$ time series for the last forty or so thousand years from Camp Century, Dye 3 and Renland also show an irregular series of stadials and interstadials (Johnsen *et al.* 1992).

An important element to this pattern of environmental change during glacial conditions back to about 100 000 years ago (figure 1) is that of the abruptness of the shift from cold to warmer temperatures at the onset of interstadials (Johnsen *et al.* 1992; Bond *et al.* 1993; Grootes *et al.* 1993). This shift may take place within as little as a few decades (Johnsen *et al.* 1992). By contrast, the initial part at least of the return to cooler stadial conditions is more gradual, sometimes followed by a more rapid fall to full stadial conditions (Grootes *et al.* 1993). The combination of repetitive interstadial and stadial patterns of isotopic variation has been termed a 'Dansgaard–Oeschger cycle', with each cycle having a characteristic saw-tooth form of very rapid warming to interstadials followed by slower cooling into the next stadial (Bond *et al.* 1993). These packages of warmer and colder oscillations within the last glacial, and in particular the sudden shift to warmer conditions, may be linked to the periodic rapid discharge of large numbers of icebergs into the North Atlantic (Bond *et al.* 1993), an issue that is discussed below.

(c) *The Eemian Interglacial*

If the $\partial^{18}O$ variations in the GRIP ice core during the last or Eemian interglacial represent a climatic proxy rather than the effects of ice deformation at depth, then several very rapid shifts in interglacial climate are implied within marine isotope Stage 5e (GRIP project members 1993). Substages 5e1, 5e3 and 5e5 are the warmest parts of the Eemian interglacial, and relatively unbroken intervals of warm climate within them lasted about 3000 years, less than 1000 years and 2000 years, respectively (figure 2). Each warm substage began and terminated with relative rapidity (on the order of decades), giving the Eemian record a variability quite unlike that observed in the very consistent $\partial^{18}O$ values from the Holocene (figure 1).

There are apparently several events of extreme rapidity within the interglacial record inferred from the GRIP core (figure 2). A catastrophic cooling lasting only about 70 years is located with Substage 5e1 (GRIP project members 1993). There are indications of an extremely rapid initiation of warming, followed by a less marked recovery, associated with this event. This is particularly apparent in the measurements of dust concentrations, which have been analysed with millimetre resolution and are affected less than other parameters by smoothing through diffusion at depth. There are also several major oscillations lasting up to about 750 years during the first 8000
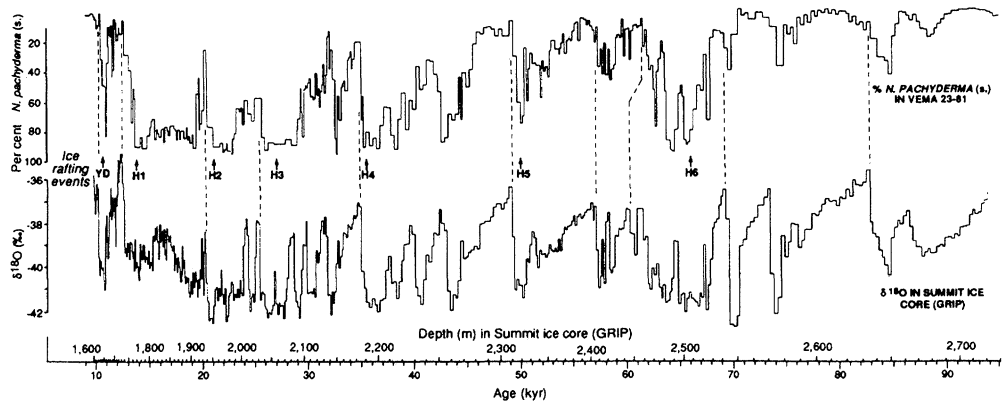
Figure 4. Comparison of the GRIP ice-core $\partial^{18}O$ record for the last 90 000 years with a proxy record for North Atlantic sea-surface temperatures and the occurrence of major ice-rafting events (adapted from Bond *et al.* 1993). Marine sediment core VEMA 23-81 is located in the North Atlantic west of Ireland and south of Iceland. Variations in the percentage of the foraminiferal species *N. pachyderma (s.)* correlate closely with sea-surface temperatures, and ice-rafting events are made up of Heinrich layers H-1 to H-6 and a further event during the Younger Dryas (YD).

years of the Eemian, where the cooling linked with each event of low $\partial^{18}O$ values is on the order of 10–14 °C (GRIP project members 1993).

Computer simulations using a global ocean model have mimicked these Eemian interglacial climatic states and the abrupt transitions between them through sudden shifts from one circulation pattern to another (Weaver & Hughes 1994). These sudden switches in the model ocean circulation were thought to be driven by a hydrological cycle which was assumed to be stronger during the Eemian interglacial than the Holocene, because the latter reached less high temperatures. However, two reservations are tied to this modelling (Broecker 1994); the first is that the Eemian environmental fluctuations in the GRIP core may simply be an artefact of the deformation of ice at depth and, secondly, the results produced by Weaver and Hughes' ocean model are likely to be heavily dependent on the surface boundary conditions that are used.

4. Ice cores and North Atlantic palaeoceanography

The causes of the observed variations in environmental parameters at the GRIP and GISP2 ice core sites in Greenland are often sought in the adjacent North Atlantic ocean. The good correlation between shifts in the thermohaline circulation of the North Atlantic during the last glacial period and the Greenland $\partial^{18}O$ record give evidence in support of this (Keigwin *et al.* 1994; McManus *et al.* 1994). The apparent link between pulses of rapid iceberg calving into the North Atlantic and the abrupt inception of warm interstadial conditions also suggests a strong ocean–atmosphere–cryosphere coupling (Bond *et al.* 1993).

(a) North Atlantic thermohaline circulation

The GRIP ice core record from Greenland has been compared with both the surface and deep-water circulation of the North Atlantic. Time series data on sea-surface temperatures and deep water circulation have been obtained through the analysis of planktonic and benthic foraminifera from marine sediment cores (Bond *et al.* 1993;
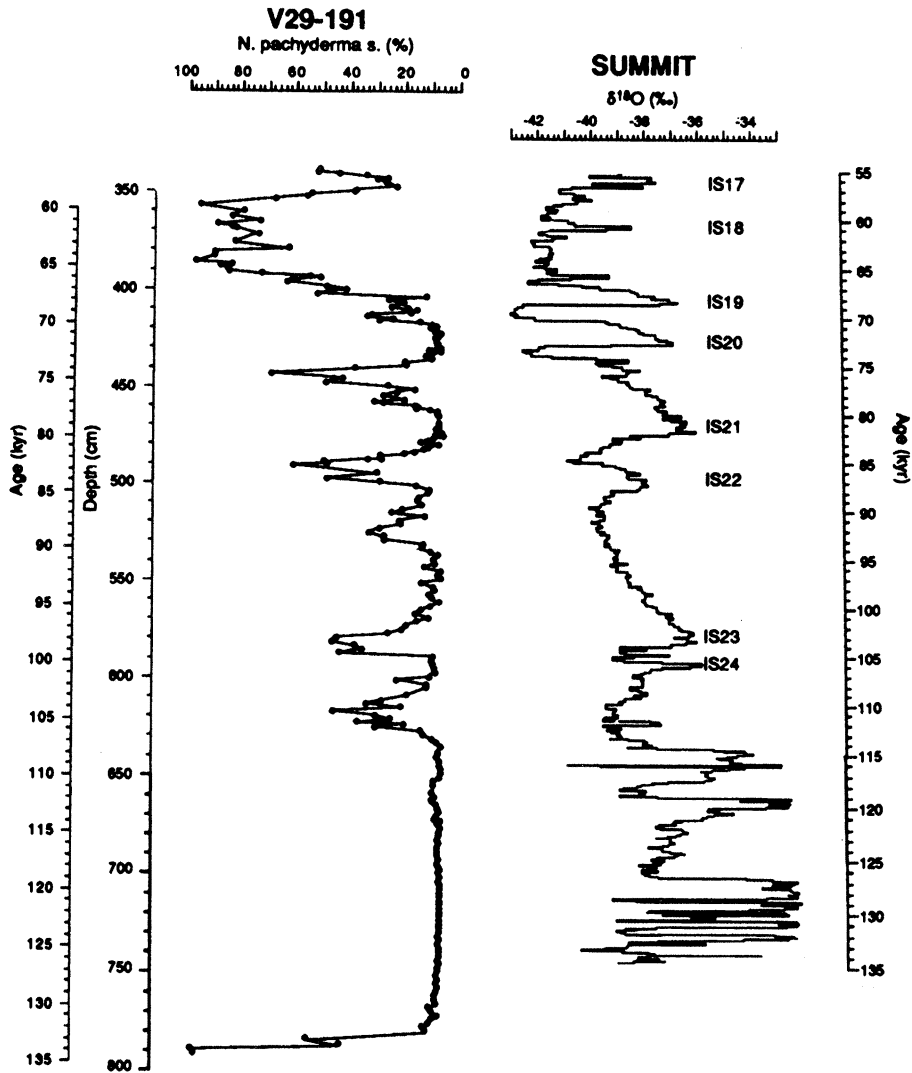
Figure 5. Comparison of the GRIP ice core record from Summit in Greenland with a proxy record for North Atlantic sea-surface temperatures for the period 60 000 to 135 000 years BP (adapted from McManus *et al.* 1994). Marine sediment core V23-191 is located in the North Atlantic west of Ireland and south of Iceland. Variations in the percentage of the foraminiferal species *N. pachyderma (s.)* correlate closely with sea-surface temperatures. Interstadials 17 to 24 are labelled. Note the lack of correlation for the last interglacial between the records, and the very low variability in *N. pachyderma (s.)* percentage.

Keigwin *et al.* 1994; McManus *et al.* 1994). Reconstructed sea-surface temperatures for the last 90 000 years show a series of warm-cold fluctuations on timescales of thousands of years, with particularly sharp terminations to the cold phases (figure 4). Bond *et al.* (1993) show that these fluctuations correlate closely with Dansgaard–Oeschger cycles in the ice core record, and suggest that the abrupt shift in ocean surface temperatures at the end of stadials is likely to represent a rise of several degrees over a few decades.

Using the same foraminiferal indicator of sea-surface temperatures, McManus *et al.* (1994) have analysed the North Atlantic record between 65–135 000 years ago, into the Eemian interglacial. Correlation with the GRIP ice core record is strong until about 105 000 years ago (figure 5), which is also about the time that the GRIP and GISP2 $\partial^{18}O$ records cease to correlate. Below this level, the marine sediment cores show a rather uniform sea-surface temperature field during the Eemian (figure 5), without the rapid shifts indicated in the GRIP ice core (McManus *et al.* 1994). This finding is of particular significance, since shifts in the North Atlantic ocean circulation pattern are often invoked as a cause of the observed $\partial^{18}O$ variations in Greenland. It implies one of the following: (i) that the GRIP record represents relatively localized climatic changes; (ii) that it is affected significantly by deformation, or (iii) that the ocean sediment records lack the temporal detail required to capture the Eemian climatic shifts.

Comparison of the GRIP $\partial^{18}O$ evidence with that from isotopic studies of benthic foraminifera from a western north Atlantic deep-sea sediment core shows a close correlation with ice core data for the period 70 000 to about 105 000 years ago (Keigwin *et al.* 1994). However, there is again no evidence for rapid shifts in deep marine isotopic parameters, which are a proxy for North Atlantic Deep Water production, during the Eemian interglacial (Keigwin *et al.* 1994). The benthic foraminiferal data reflect the shifting balance between North Atlantic Deep Water and Antarctic Bottom Water in the North Atlantic. This deep water flux is particularly important in maintaining the northward flux of heat in the surface waters of the North Atlantic. Keigwin *et al.* (1994) show that carbon isotope and carbonate concentration measurements reveal a more detailed signal than benthic foraminiferal $\partial^{18}O$ values. Using these parameters, they found no indication of variability in the deep ocean record of North Atlantic Deep Water in the Eemian, with the implication that variability in the GRIP ice core must have an alternative source.

(*b*) *Heinrich events in the North Atlantic and rapid climate change*

The record of sea-surface temperatures in North Atlantic cores not only correlates closely with the GRIP ice core time series during the last glaciation, but also appears to have clear links with ice rafted pulses or Heinrich events (figure 4). It appears that these rapid iceberg discharge events (MacAyeal 1993; Dowdeswell *et al.* 1995), and the lowered surface salinities associated with the melting of this ice (Bond *et al.* 1992), occurred during particularly cold stadial conditions (figure 4). The Heinrich events tend to occur, with the exception of H-1, close to the end of the cold period of Dansgaard–Oeschger cycles. The rapid warming, following the release of very large volumes of ice into the North Atlantic, may be associated with increasing surface salinities after the cessation of iceberg production. This increase, Bond *et al.* (1993) suggest, could be large enough to enhance the thermohaline circulation in the North Atlantic and the northward heat transfer that is associated with it.

5. Greenland ice cores and other Eemian climate records

The interglacial record of abrupt climate change derived from the GRIP core is not mirrored in the long Vostok ice core record from East Antarctica (Jouzel *et al.* 1993), which is the only existing ice core dataset of comparable length. However, this lack of correlation is not unexpected if it is assumed that interglacial fluctuations in climate at the summit of the Greenland Ice Sheet are linked to reorganizations of

the oceanic and atmospheric circulation in the North Atlantic region, rather than to global factors. Indeed, the Antarctic ice core records respond only weakly to well-established climate changes in the North, such as the Dansgaard–Oeschger Events and the Younger Dryas cooling.

A problem in comparing the Greenland ice core records with marine and terrestrial evidence for climate change is that the latter are of lower time resolution and have a potentially long response time for changes in biotic systems (Larsen *et al.* 1995). Even so, there is little evidence in western European lowland pollen diagrams of the large-scale climatic fluctuations indicated in the last interglacial ∂^{18}O record in the GRIP ice core. In the long Grande Pile peat bog record, for example, the Eemian appears as an interval of continuous warmth in Europe (Woillard 1978). By contrast, a number of marked fluctuations in pollen data are indicated for the last glaciation, although providing by no means as detailed a record as those from ice cores and North Atlantic sediments. This interpretation of a stable interglacial pollen record for the Eemian is supported by the continuous presence of species which are unable to withstand long periods of extreme winter cold, for example holly and ivy (Johnsen *et al.* 1995).

However, Johnsen *et al.* (1995) point out that palaeobotanical evidence from higher altitudes in Europe may provide a more sensitive record of interglacial environmental change, since climatic thresholds are more easily passed in such locations. Eemian records of magnetic susceptibility, pollen and organic carbon from lake deposits in the French Massif Central suggest two rapid cooling events during the interglacial, similar to those in the GRIP core (Thouveny *et al.* 1994).

It is concluded that the GRIP ∂^{18}O record on the climate of the last interglacial has no immediate parallel in proxy environmental datasets from other areas (Johnsen *et al.* 1995). Three possible interpretations of this situation are given by Johnsen *et al.* First, the GRIP core may be deformed stratigraphically. Secondly, the evidence for environmental change at Summit may represent simply a climatic instability in a small portion of the north-western North Atlantic. Finally, interglacial climate changes may have taken place in a wider region, but are subdued and thus difficult to pinpoint using proxy indicators of limited temporal and environmental sensitivity.

6. Conclusions

The major changes which define climatic variability over the past million years are the glacial–interglacial cycles. Until recently, the existing paradigm was that such large changes, 5 °C or more shifts in average regional to global temperatures, occur over centuries to millennia, the timescale of orbital solar forcings. Recent evidence from ice cores has challenged this viewpoint. Glacial-interglacial scale changes in temperature, inferred from stable oxygen isotope ratios, have occurred in 50 years, or less than a human lifetime (figures 1 and 2). Changes in dust concentrations, which reflect dust source variability and wind strengths, can occur even faster, in 20 years. Shifts in deuterium excess, an isotopic parameter reflecting sea surface conditions, have also occurred in 20 years or less. Changes in the accumulation rate of snow, which reflect atmospheric temperature and storm frequencies, can occur faster still. A doubling of accumulation accompanying the end of the Younger Dryas occurred in only a few years (figure 3). These large, rapid changes are in contrast to the last 11 000 years which, in the Arctic, is a time of relatively stable climate (figure 1).

While the cause or causes of these changes is unknown, the speed implies a shift in atmospheric circulation, probably accompanied and amplified by changes in ocean circulation in the North Atlantic (figures 4 and 5). Suggested causes, such as periods of very rapid iceberg production, or Heinrich events, have usually relied on the existence of large amounts of ice on land, implying that rapid climate shifts are characteristic of glacial, and not of interglacial, periods. From an anthropogenic perspective, this is a fortuitous situation. For this reason, the results from the GRIP ice core, with its highly variable record of Eemian interglacial climate (figure 2), are particularly important. While this result has not been duplicated in another ice core, particularly not in the GISP2 core which is only 28 km away from the GRIP site, neither has there been conclusive evidence that the GRIP record is fatally flawed by folding of the ice. Other evidence from ocean sediment records (figure 5) and land-based pollen records during the Eemian have yielded mixed results, with the current weight of evidence perhaps falling on the side of a stable Eemian. It would be unwise to draw conclusions about the variability of interglacial climate, however. The temporal detail available in ice cores is difficult to obtain in other sedimentary records, and another ice core from Greenland may be required to resolve the issue. Until then, as human activity continues to modify the atmospheric system, possible future scenarios for climate change should include one which incorporates major shifts on timescales of years to decades.

The identification of abrupt climate change in the polar regions is dependent on a large number of data sources and measurements, many of them undertaken in difficult field conditions. We thank all of the investigators who have dedicated their time to collecting, analysing and interpreting the thousands of observations reviewed here.

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