Progress in Physical Geography 23,1 (1999) pp. 1–36

Sudden climate transitions during the Quaternary

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Abstract: The time span of the past few million years has been punctuated by many rapid climate transitions, most of them on timescales of centuries to decades. The most detailed information is available for the Younger Dryas-to-Holocene stepwise change around 11 500 years ago, which seems to have occurred over a few decades. The speed of this change is probably representative of similar but less well studied climate transitions during the last few hundred thousand years. These include sudden cold events (Heinrich events/stadials), warm events (interstadials) and the beginning and ending of long warm phases, such as the Eemian interglacial. Detailed analysis of terrestrial and marine records of climate change will, however, be necessary before we can say confidently on what timescale these events occurred; they almost certainly did not take longer than a few centuries.

Various mechanisms, involving changes in ocean circulation and biotic productivity, changes in atmospheric concentrations of greenhouse gases and haze particles, and changes in snow and ice cover, have been invoked to explain sudden regional and global transitions. We do not know whether such changes could occur in the near future as a result of human effects on climate. Phenomena such as the Younger Dryas and Heinrich events might only occur in a 'glacial' world with much larger ice sheets and more extensive sea-ice cover. A major sudden cold event, however, did probably occur under global climate conditions similar to those of the present, during the Eemian interglacial around 122 000 years ago. Less intensive, but significant rapid climate changes also occurred during the present (Holocene) interglacial, with cold and dry phases occurring on a 1500-year cycle, and with climate transitions on a decade-to-century timescale. In the past few centuries, smaller transitions (such as the ending of the Little Ice Age

0309–1333(99)PP210RA

at about AD 1650) probably occurred over only a few decades at most. All evidence indicates that long-term climate change occurs in sudden jumps rather than incremental changes.

Key words: climate change, Heinrich events, interstadials, Quaternary, Younger Dryas.

I Introduction

Until a few decades ago it was generally thought that large-scale global and regional climate changes occurred gradually over a timescale of many centuries or millennia, scarcely perceptible during a human lifetime. The tendency of climate to change relatively suddenly has been one of the most surprising outcomes of the study of earth history, specifically that of the last 150 000 years (e.g., Taylor *et al*., 1993). Some and possibly most large climate changes (involving, for example, a regional change in mean annual temperature of several degrees celsius) occurred at most on a timescale of a few centuries, sometimes decades, and perhaps even just a few years.

The decadal timescale transitions would presumably have been noticeable to humans living at such times, and may have created difficulties or opportunities (e.g., the possibility of crossing exposed land bridges, before sea level could rise). Hodell *et al*. (1995) and Curtis *et al*. (1996), for instance, document the effects of climate change on the collapse of the Classic period of Mayan civilization and Thompson (1989) describes the influence of alternating wet and dry periods on the rise and fall of coastal and highland cultures of Ecuador and Peru. The beginning of crop agriculture in the middle east corresponds very closely in time with a sudden warming event marking the beginning of the Holocene (Wright, 1993). The burial in ice of the prehistoric mummified corpse of the famous ‘Iceman’ (e.g., Bahn and Everett, 1993) at the upper edge of an alpine glacier coincided with the initiation of a cold period (‘Neoglaciation’) after the Holocene climate optimum (Baroni and Orombelli, 1996). On longer timescales, evolution of modern humans has been linked to climatic changes in Africa (e.g., de Menocal, 1995). But the full implications of these sudden changes for biogeography and for the evolution of human cultures and biology have barely begun to be considered; there has simply not been time for the message to be absorbed by biogeographers, archaeologists and palaeoanthropologists, and this review is intended to assist in this process.

Sudden stepwise instability is also a disturbing scenario to be borne in mind when considering the effects that humans might have on the climate system through adding greenhouse gases. Judging by what we see from the past, conditions might gradually be building up to a ‘break point’ at which a dramatic change in the climate system will occur over just a decade or two, as a result of a seemingly innocuous trigger. The evidence for dramatic past changes on the timescale of centuries to decades will be the subject of this review.

II Broad climate variability: the background of oscillations on the timescale of tens of thousands of years

Climatic variability on the timescale of tens of thousands of years has turned out to be a predominant pattern in earth history. The last two and a half million years have been marked by many global climate oscillations between warmer and cooler conditions. This oscillating trend appears to be the continuation of a pattern of variability extending back well into the Tertiary period and possibly beyond (e.g., Kennett, 1996), but during the last few million years the length and the amplitude of these climate cycles have increased (e.g., Crowley and North, 1991; Hodell and Venz, 1992).

Large global interglacial–glacial–interglacial climate oscillations have been recurring at approximately a 100 000-year periodicity for the last 900 000 years (e.g., Berger *et al*., 1993; Mudelse and Schulz, 1997), though each individual cycle has had its own idiosyncrasies in terms of the timing and magnitude of changes (e.g., Lyle *et al*., 1992). As is usually the case with the study of the past, data become more scarce with increasing age (Winograd *et al*., 1997); even so, many detailed records are now becoming available (e.g., Linsley, 1996; Tzedakis *et al*., 1997). Extended records of atmospheric gas concentrations and polar temperatures may be expected from the continued deeper drilling of the Antarctic Vostok ice core (Jouzel *et al*., 1996; Petit *et al*., 1997).

The most recent large climate oscillation spanning the last 130 000 years (130 ka) has been the subject of the most intensive study, because it offers a relatively detailed climate record from the land, from the oceans and from the ice cores. In the last few years, a considerable amount of new data on the warm period known as the Eemian (the last major interglacial) has become available (e.g., Pewe *et al*., 1997). This interval has seen the global climate system switch (Figure 1) from warm interglacial (similar to present day) to cold glacial conditions, and back.

In stratigraphic terms, the last 130 ka oscillation began with the Eemian interglacial (isotope stage 5e, possibly including part of the series 5d–a; known as the Sangamon stage on land in the USA), followed by the colder last glacial period (stages 4–2), and the present Holocene interglacial beginning around 11 000 years ago (isotope stage 1; Table 1). There is considerable evidence that similar events and processes were at work in previous glacial–interglacial cycles over the past 900 000 years. The general background of continuous (but less dramatic) variability in the earth's climate system extends well back into the Tertiary and beyond (e.g., Crowley and North, 1991).

For the most recent Eemian-to-Holocene phase there remains significant ambiguity in terms of the errors in geological dating techniques, causing problems in correlation of events between different regions, so that it is difficult to know whether climate changes were truly synchronous globally. In many other respects, however, the record that has been assembled for this period is remarkably detailed. There are apparently few gaps in the record among the many long ocean cores, though land and lake records are much patchier. Indicator species on land (e.g., trees) may respond slowly, but oceanic unicellular plankton (such as diatoms, foraminifera and nannoplankton) has such short generation times that it can potentially give a precise picture of events, at least where sedimentation rates are sufficient. For instance, exciting new data can be expected from detailed analysis of the cores taken during Leg 172 of the international Ocean Drilling Program (Keigwin *et al*., 1998), which concentrated on high sedimentation rate sediment drifts in the North Atlantic.

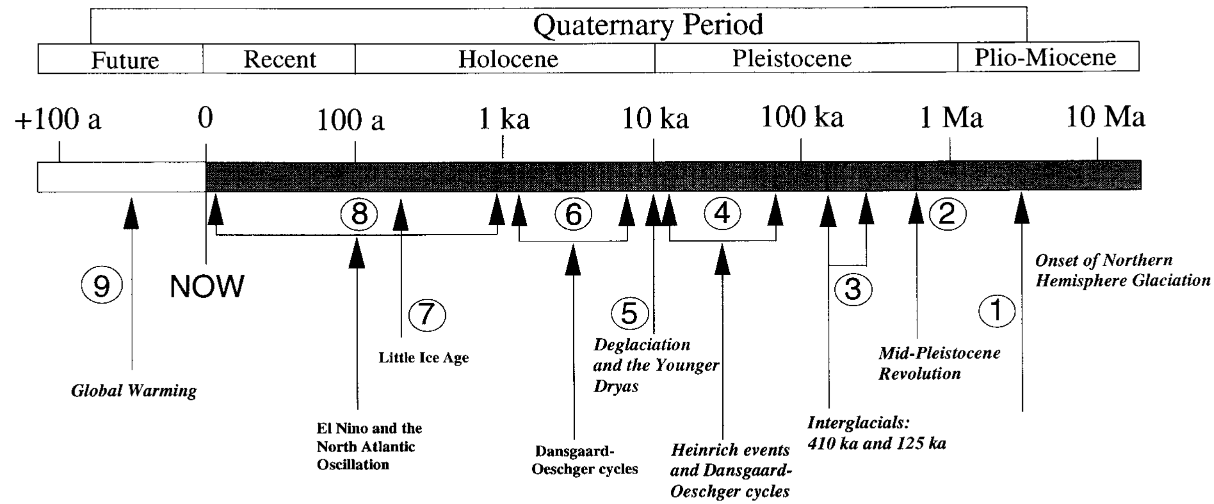

Figure 1 An outline of the major climate events during the Quaternary and late Tertiary. Most of the labelled events are referred to in the text. Note that the timescale is logarithmic:

1) Between 4 and 2.5 Ma ice sheets started to develop in the Northern Hemisphere, ushering in the strong glacial–interglacial cycles which are characteristic of the Quaternary period.
2) Prior to the mid-Pleistocene revolution the climate cycled between glacial–interglacial every 41 ka; afterwards it cycled every 100 ka. The external forcing of the climate did not change. This implies that the internal climate feedbacks must have altered, possibly due to reaching an atmospheric carbon dioxide threshold.
3) The two closest analogues to the present climate are the interglacial periods at 420–390 ka (oxygen isotope state 11) and 130–115 ka (oxygen isotope stag 5e, also known as the Eemian; see Figure 2).
4) Heinrich events and Dansgaard–Oeschger cycles (see Figures 3 and 4).
5) Deglaciation and the Younger Dryas event.
6) Dansgaard–Oeschger cycles and other climate events during the Holocene.
7) Little Ice Age (AD 1700), the most recent climate event which occurred throughout the Northern Hemisphere.
8) El Niño (~3–5 years) and North Atlantic climate oscillation (~10 years), which have occurred for at least the last 1000 years.
9) Anthropogenic global warming and changes to the global hydrological cycle.

Source: Adapted from Maslin (1998)

Table 1 The timescale of the last 130 000 years (130 ka) since the start of the Eemian warm period

Event	Age (yr) Martinson *et al.*	Imbrie *et al.*
Last deglaciation (event 2.0), termination i	12 050	12 000
Glacial maximum (event 2.2)	17 850	19 000
Boundary stage 2/3 (event 3.0)	24 110	24 000
Boundary stage 3/4 (event 4.0)	58 960	59 000
Boundary stage 4/5 (event 5.0)	73 910	71 000
Event 5.1	79 250	80 000
Event 5.2	90 950	87 000
Event 5.3	99 380	99 000
Event 5.4	110 790	107 000
Event 5.5	123 820	122 000
Stage 5/6 (event 6.0), termination ii	129 840	128 000

Note: Stage 5 is defined as an interglacial stage that contains three negative events (i.e., warm), labelled 5.1, 5.3 and 5.5; and two positive events – 'cold' – labelled 5.2 and 5.4. This numbering tends to supersede the 5a–5e names.
Source: Ages of isotope stages after Martinson *et al.* (1987) in the first column, after Imbrie *et al.* (1984) in the second column.

The general picture summarized here roughly reflects the present consensus gained from studies of ice cores, deep ocean cores, and terrestrial and lake sediments around the world. This consensus is itself subject to sudden jumps when new data are presented, or as more thorough reanalyses of previous data come forth; for this reason, our review is liable to be significantly out of date within only a few weeks or months of being written!

III The record of decade-to-century timescale changes during the last 130 000 years

1 The Eemian, or last interglacial

The last interglacial (also called the Eemian) has often been seen as a close counterpart of the present interglacial: sea surface temperatures were similar, and sea level was possibly somewhat higher (e.g., Linsley, 1996). Going by the principle that there is a general similarity between these two warm periods, the Eemian has been used to predict the duration of the present interglacial, and also to study the possibility of sudden climate variability occurring within the next few centuries or millennia. The dating of the warming event that begins the Eemian is still the subject of controversy, and this leads to greater uncertainty about what to make of it as an analogue for future climate change.

a Controversy over the timing of the last interglacial: The Eemian or marine oxygen isotope stage (MIS) 5e interglacial (Table 1 and Figure 1) began sometime between

130 and 140 ka (Martinson *et al*., 1987; Sarnthein and Tiedemann, 1990; Imbrie *et al*., 1993a; 1993b; Szabo *et al*., 1994; Stirling *et al*., 1995) with a warming phase (of uncertain duration) taking the earth out of an extreme glacial phase, into conditions generally warmer than those of today (Frenzel *et al*., 1992). Warming into the Eemian may have occurred in two major steps, similar to the last deglaciation (Seidenkrantz *et al*., 1995; 1996).

Though it was named after a warm-climate phase seen in the terrestrial pollen record of The Netherlands (e.g., Zagwijn, 1963; 1975), the first generally accepted numerical dates for the start of the Eemian came from the oceans (Shackleton, 1969), based on interpretation of benthic foraminiferal oxygen isotope data. Such oxygen isotope curves were later expanded and many curves were stacked; the timing of maximum ice volume as estimated from oxygen isotope values was then 'tuned' to astronomical variations in solar input at specific latitudes (see below). This timescale is usually called the SPECMAP curve (Imbrie *et al*., 1984; Martinson *et al*., 1987).

The warm MIS 5e, the warmest interval in the last 150 ka, was correlated with the warmest interval in pollen records, the Eemian. The land records were then dated by ^{14}C for the younger parts of the records and by correlation with the marine records for the older parts (e.g., Woillard, 1978; Woillard and Mook, 1982).

The age of the warming event (or events) at the beginning of stage 5e, however, is still under discussion. Work on deep-sea sediments (Sarnthein and Tiedemann, 1990; Imbrie *et al*., 1993a; 1993b; Maslin *et al*., 1998) and corals (Szabo *et al*., 1994; Stirling *et al*., 1995; Slowey *et al*., 1996) suggests that rapid warming could have started as early as 132 ka, while work on the Antarctic Vostok ice core suggests a possible initiation at 134 ka (Jouzel *et al*., 1993). Studies of an Alaskan site (the Eve Interglaciation Forest Bed) suggest that the warming definitely postdated 140 ka, because the tephra underlying this bed was dated at 140 ka. Uranium–thorium-dated records from a continental karst sediment in the southwestern USA (Devil's Hole; Winograd *et al*., 1988; 1992; 1997), however, suggest a much earlier start of warming at about 140 ka.

There has been much discussion about the reliability of dating of the marine isotope stages as compared to the Devil's Hole record (e.g., Imbrie *et al*., 1993a; 1993b), but new numerical dating methods have confirmed that both records are reliable (protactinium-231 dating of carbonates; Edwards *et al*., 1997). We must thus consider the possibility that warm conditions did not last for the same amount of time throughout the world. For instance, comparison of various land records suggests that warming may have occurred at different times in the Alps and in northern France (de Beaulieu and Reille, 1989).

Winograd *et al*. (1997) date the ending of the warm period at about the same time as the SPECMAP timescale, so that they require the duration of the Eemian warm period to have been about twice as long as in the SPECMAP scheme of events (i.e., lasting 22 ka rather than about 10 ka). We need to consider the possibility that warm intervals as seen in pollen records have a longer duration than periods of high sea level and low ice volume in the marine record: Kukla *et al*. (1997), for instance, suggest that the Eemian (as seen in the pollen records) started at about 130 ka, but ended much later than the end of MIS 5e, and that the duration of land interglacials thus is indeed longer than the period of low ice volume. Such a difference in duration is also apparent in the comparison of terrestrial and land records by Tzedakis *et al*. (1997: Figure 2).

If one accepts the contradictory picture obtained from comparing the Devil's Hole record and some of the terrestrial pollen records with other parts of the world, it seems that for thousands of years warm 'interglacial' type conditions in the mid-latitudes on land could have been occurring at the same time as much colder ocean conditions and expanded Arctic ice sheets. The Eemian, it appears, could have been a strange beast quite unlike our present interglacial phase. This confusion over the nature and duration of the Eemian adds to the difficulty in making simple, general comparisons with our present interglacial, and in interpreting the significance of some of the events seen in the marine and ice core records.

b Evidence of climate instability during the Eemian: Whatever the true timescale of the Eemian, what has generated most widespread interest (extending well beyond the usual geological community and into the popular media) is that there are indications of large-scale climate instability in the middle of the Eemian (e.g., Maslin and Tzedakis, 1996). This finding is alarming, as we are presently living in an interglacial period (called the Holocene) which closely resembles the Eemian in many respects. If sudden, dramatic climate changes could occur within the Eemian, then they could perhaps occur in the future during our present interglacial, especially if we perturb the system by adding greenhouse gases.

Initial evidence from the GRIP ice core (Dansgaard *et al.*, 1993; Taylor *et al.*, 1993) suggested that the Eemian was punctuated by many short-lived cold events. This is shown by variations in both the electrical conductivity (a proxy for windblown dust, with most dust indicating colder and more arid conditions) and the stable oxygen isotopes (a proxy for air temperature) of the ice. The cold events seemed to last a few thousand years, and the magnitude of cooling was similar to the difference between glacial and interglacial conditions; a very dramatic contrast in climate. Furthermore, the shifts between these warm and cold periods seemed to be extremely rapid, possibly occurring over a few decades or less.

A second ice core (GISP2) from the Greenland ice cap (Boulton, 1993) provided an almost identical climate record for the last 110 ka. GISP2 also shows apparent sudden climate jumps throughout the Eemian, but the two records diverge (Grootes *et al.*, 1993). Significantly, in GISP2 steeply inclined ice layers occur in this lower portion of the core, indicating that the ice has been disturbed, and that we cannot distinguish simple tilting from folding or slippage that would juxtapose ice of very different ages (Boulton, 1993). For this reason, the deeper GISP2 record has been interpreted as containing interglacial and glacial ice of indeterminate age, mixed in together by ice tectonics (Grootes *et al.*, 1993; Alley *et al.*, 1997a). It has been suggested that the deeper parts of the GRIP ice-core record (referred to above), including the crucial Eemian sequence, may also have been affected by ice tectonics (e.g., Boulton, 1993; Grootes *et al.*, 1993; Taylor *et al.*, 1993; Chapellaz *et al.*, 1997; Hammer *et al.*, 1997). Johnsen *et al.* (1995; 1997) reported layers tilted up to 20 ° within the marine isotope stage 5c (110 ka) section of the GRIP ice cores, precisely where the correlation between GRIP and GISP2 breaks down. The GISP2-GRIP Joint Workshop in Wolfeboro, New Hampshire, USA (September 1995), concluded from data on ice-trapped atmospheric CH_4 and isotopic data on other trapped gases such as He, ice oxygen isotopic composition ($\delta^{18}O$) from both Antarctica and Greenland, and from more detailed work on the structural properties of the cores, that both GRIP and GISP2 ice cores had suffered stratigraphic disturbances in ice older than 110 ka (Chapellaz *et al.*, 1997); Hammer *et al.*, 1997). More detailed analysis of the stratigraphically disturbed records of the GRIPS and GISP ice cores may, after careful reconstruction, still reveal information on the Eemian climate (Steffensen *et al*., 1997). There are plans to obtain high-resolution last interglacial records from both Greenland and the high accumulation rate West Antarctic Ice Sheet, which will add an interhemispheric dimension to the current Eemian debate.

Support for the occurrence of cold Eemian events was obtained from lake records from continental Europe (de Beaulieu and Reille, 1989; Guiot *et al*., 1993), the Massif Central in France (Thouveny *et al*., 1994) and Bispingen, Germany (Field *et al*., 1994). Oceanic records from the northeast Atlantic (McManus *et al*., 1994) and the Bahamas Outer Ridge (Keigwin *et al*., 1994) indicate very little or no climatic variability during the Eemian, but a cooling appears to be present in oceanic records from the Sulu Sea (Indonesia; Linsley, 1996). In contrast to this, records from both the Nordic seas and west of Ireland show a cooling and freshening of the North Atlantic in the middle of the Eemian somewhere between 122 and 125 ka (Cortijo *et al*., 1994; Fronval and Jansen, 1996). These Nordic records show highly variable surface-water conditions throughout the Eemian period. Records from slightly further south on the northwest European shelf suggest a similar picture of cold intervals during the Eemian (Seidenkrantz *et al*., 1995).

Evidence for a single sudden cool event during the Eemian is also present in pollen records from a lake in central Europe (Field *et al*., 1994), from loess sedimentology in central China (Zhisheng and Porter, 1997), and from ocean sediment records from ODP Site 658 in the eastern subtropical Atlantic (Maslin and Tzedakis, 1996; Maslin *et al*., 1996; 1998). Further support for the existence of the intra-Eemian cooling event comes indirectly from coral reef records. High-precision U-series coral dates from Western Australia indicate that the main global episode of coral reef building during the last interglacial period (dated between 130 and 117 ka) was confined to just a few thousand years between 127 to 122 ka (Stirling *et al*., 1995), thus ending at the beginning of the intra-Eemian cold event at about 122 ka (Figure 2).

Overall, the combined sources of evidence suggest that there was a cold and dry event near the middle of the Eemian, at about 122 ka, which was characterized by a change in circulation of the North Atlantic, a several-degree decline in the Nordic seas and Atlantic sea-surface temperatures, and an opening up of the west European forests to a mixture of steppe and trees. This intra-Eemian cold phase was less dramatic than had been suggested by the ice-core records, but still a major climatic change. Evidence from a high-resolution marine core record at Site ODP 658 (Maslin and Tzedakis, 1996) suggests that this event, which might possibly have come on in a few decades or less, lasted no more than 400 years. Afterwards climate recovered, but conditions did not return to the full warmth of the early Eemian 'optimum'. The event, however, may have been widespread but not global, as it apparently does not show up in cores from some parts of the world (e.g., Oppo *et al*., 1997).

The intra-Eemian event may have had a corresponding, though weaker, equivalent in the present interglacial (the Holocene). Around 4800–4500 yr, there was a 300-year long colder phase that resembled the intra-Eemian cooling event in that they both occurred after the interglacial peak and signalled the beginning of a trend of climatic deterioration (e.g., Baroni and Orombelli, 1996; Bond *et al*., 1997).

Other large and relatively sudden cool and arid phases (occurring against a background of similar-to-present conditions) seem to have affected some of the

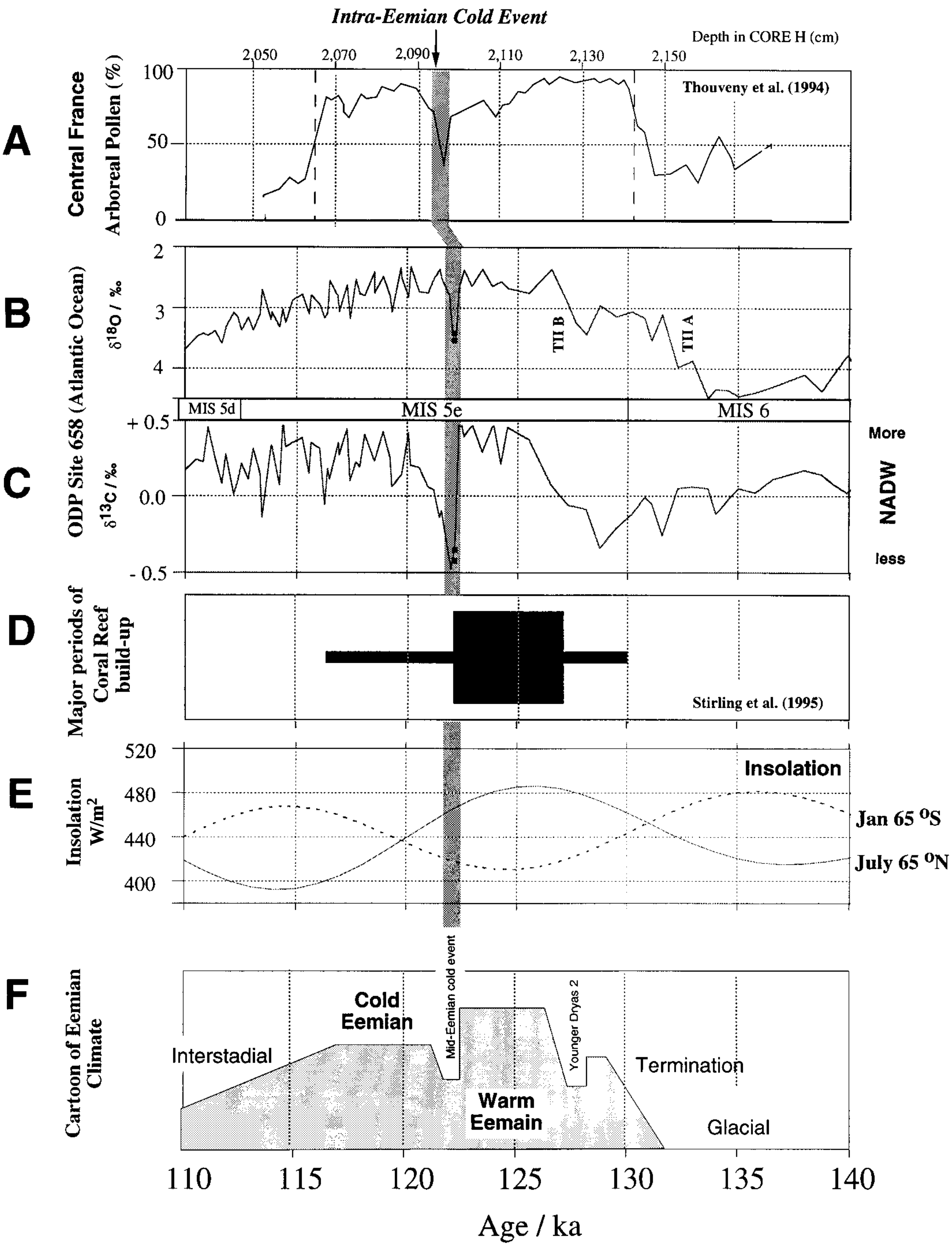

Figure 2 The intra-Eemian cold event, seen from various records that span the Eemian interglacial
Source: Adapted from Maslin and Tzedakis (1996)

previous interglacials before about 200 000 years ago (Winograd *et al*., 1997). Again, the speed with which these climate transitions occurred is unclear through lack of detailed time resolution in these older records, but these changes may have occurred over only a few decades.

2 Sudden transitions after 115 000 years ago

According to the marine records, the Eemian interglacial ended with a rapid cooling event about 110 000 years ago (e.g., Imbrie *et al*., 1984; Frenzel and Bludau, 1987; Martinson *et al*., 1987), which also shows up in ice cores and pollen records from across Eurasia. Adkins *et al*. (1997) suggested that the final cooling event took less than 400 years, and it might have been much more rapid.

Following the end of the Eemian, a large number of other sudden changes and short-term warm and cold alternations have been recognized; apparently many or all of these occurred on a global or at least a regional scale (e.g., Dansgaard *et al*., 1993). The most extreme of these fluctuations are the warm interstadials and the cold Heinrich events. These are most prominent in the ice-core record of Greenland, deep-sea cores from the North Atlantic, and in the pollen records of Europe and North America, suggesting that they were most intense in the North Atlantic region (e.g., Bond *et al*., 1992; 1993).

a Interstadials: Sudden and short-lived warm events occurred many times during the generally colder conditions that prevailed between 110 000 and 10 000 years ago (isotope stages 2–5.4; Dansgaard *et al*., 1993). First picked up as brief influxes of warm climate plants and insects into the glacial tundra zone of northern Europe, they are known as ‘interstadials’ to distinguish them from the cold phases or ‘stadials’ (e.g., Lowe and Walker, 1984). The interstadials show up strongly in the Greenland ice-core records (e.g., Mayewski *et al*., 1997). Between 115 000 and 14 000 years ago, there are 24 of these warm events recognized in these cores (where they are called ‘Dansgaard–Oeschger events’; e.g., Bond *et al*., 1993; Dansgaard *et al*., 1993; Bond and Lotti, 1995; Mayewski *et al*., 1997; Taylor *et al*., 1993; 1997). Each warm interstadial is linked to a cold interstadial and these so-called ‘Dansgaard–Oeschger’ cycles last approximately 1500 years.

Short-lived and/or moist warm phases, coeval with interstadials, appear in the eastern Pacific (Behl and Kennett, 1996), western Siberia, the Arabian Sea (Sirocko *et al*., 1993; Schulz *et al*., 1998), and possibly also in central China (Behl and Kennett, 1996). The duration of each interstadial for the younger parts of the records can be counted in ice cores from the annual snow layers or (rather less precisely) from the thickness of sediment accumulated in an ocean bed core. In marginal ocean basins with low oxygenation, annual layers may be preserved in the sediment (e.g., Behl and Kennett, 1996; Ocean Drilling Program Leg 169S, Saanich Inlet, British Columbia, Canada).

Ice-core and ocean data suggest that interstadials both began and ended suddenly, though in general the ‘jump’ in climate at the start of an interstadial was followed by a more gradual decline involving a stepwise series of smaller cooling events and often a fairly large terminal cooling event which returned conditions to the colder ‘glacial’ state (e.g., Rasmussen *et al*., 1997). From the ice-core evidence from Greenland, warming into each interstadial occurred over a few decades or less, and the overall duration of some of these warm phases may have been just a few decades, while others vary in length from a few centuries to nearly two thousand years (e.g., Mayewski *et al.*, 1997).

b Heinrich events: Opposite in sign to the interstadials are extreme and short-lived cold events, known as 'Heinrich events' (e.g., Heinrich, 1988; Bond *et al.*, 1992; Grousset *et al.*, 1993; Andrews *et al.*, 1994; see review in Andrews, 1998), which were first recognized as periods with very intense ice-rafting in the North Atlantic (Ruddiman, 1977). These events occurred against the general background of the glacial climate and represent the climatic effects of massive surges of icebergs into the North Atlantic. The several massive ice-rafting events (e.g., Heinrich, 1988; Broecker *et al.*, 1992; Bond *et al.*, 1992; 1993; Maslin *et al.*, 1995; Rasmussen *et al.*, 1997; Andrews, 1998) (Table 2) show up in the Greenland ice cores as a further 3–6 °C drop in temperature from the already cold glacial climate (Bond *et al.*, 1993; Maslin *et al.*, 1995; Mayewski *et al.*, 1997). Many of these events have also been picked up as particularly cold and arid intervals in European and North American pollen records (Grimm *et al.*, 1993). It is still debated whether these events are caused by internal ice-sheet dynamics (MacAyeal, 1993a; 1993b) or by periodic external climate changes (Broecker, 1995) possibly caused by harmonics of the orbital parameters. The freshwater input via the melting icebergs during the Heinrich events was so large that it caused a collapse of deep-water formation (Seidov and Maslin, in press; Stocker, in press).

There is still considerable discussion as to the exact region of provenance of the icebergs participating in the Heinrich events (e.g., Gwiazda *et al.*, 1996a; 1996b; Lehman, 1996; Revel *et al.*, 1996; Bond *et al.*, 1997). At North Atlantic mid-latitudes, the circulation was changed (see Figure 3) and oceanic surface productivity dropped precipitously (Thomas *et al.*, 1995). At least some of these events may also have affected the climate further afield from Greenland – giving cold, arid conditions as far away as central China and Antarctica (Thompson *et al.*, 1989). Figure 4 summarizes the locations where events synchronous with the Heinrich events have been described. Sources for these include marine data, pollen records from Pacific Northwest and European lakes and cores from the Arabian Sea (Schulz *et al.*, 1998). The effect of Heinrich events must, thus,

Table 2 Timing of major Heinrich events during the last 130 000 years (ages in calendar years)

Event	Age
YD or H0	12.2
H1	16.8
H2	24.1
H3	30.1
H4	35.9
H5	50.0
H6	66.0

Notes:
YD: Younger Dryas; H: Heinrich event.
Source: After Bond *et al.*, 1997, for H0–H3; after Bond *et al.*, 1993, for H4–H6.

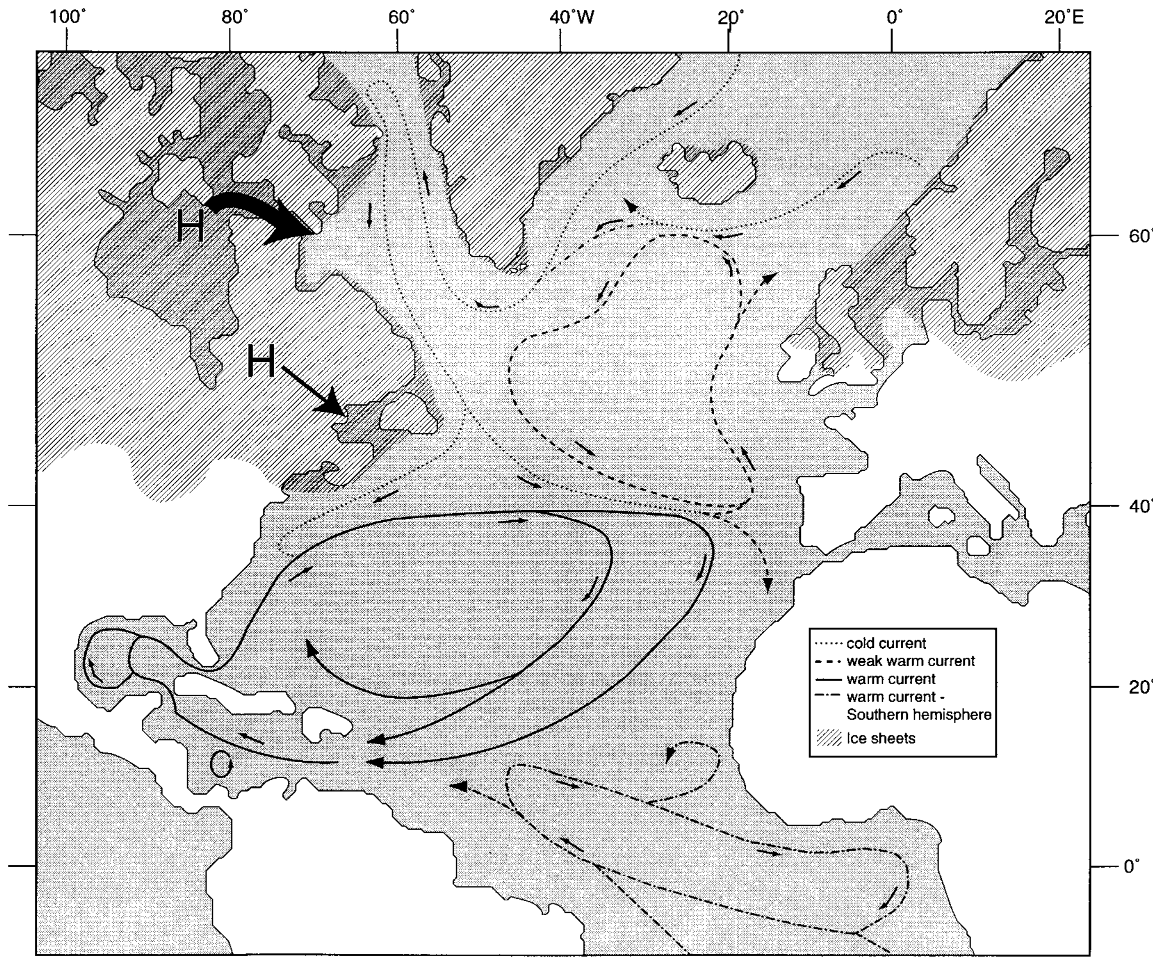

Figure 3 Summary of the palaeo-reconstructed surface ocean circulation of the North Atlantic during the Last Glacial Maximum. 'H' represents the possible source areas of the icebergs which flooded the North Atlantic during the Heinrich events
Source: Adapted from Maslin *et al*. (1997)

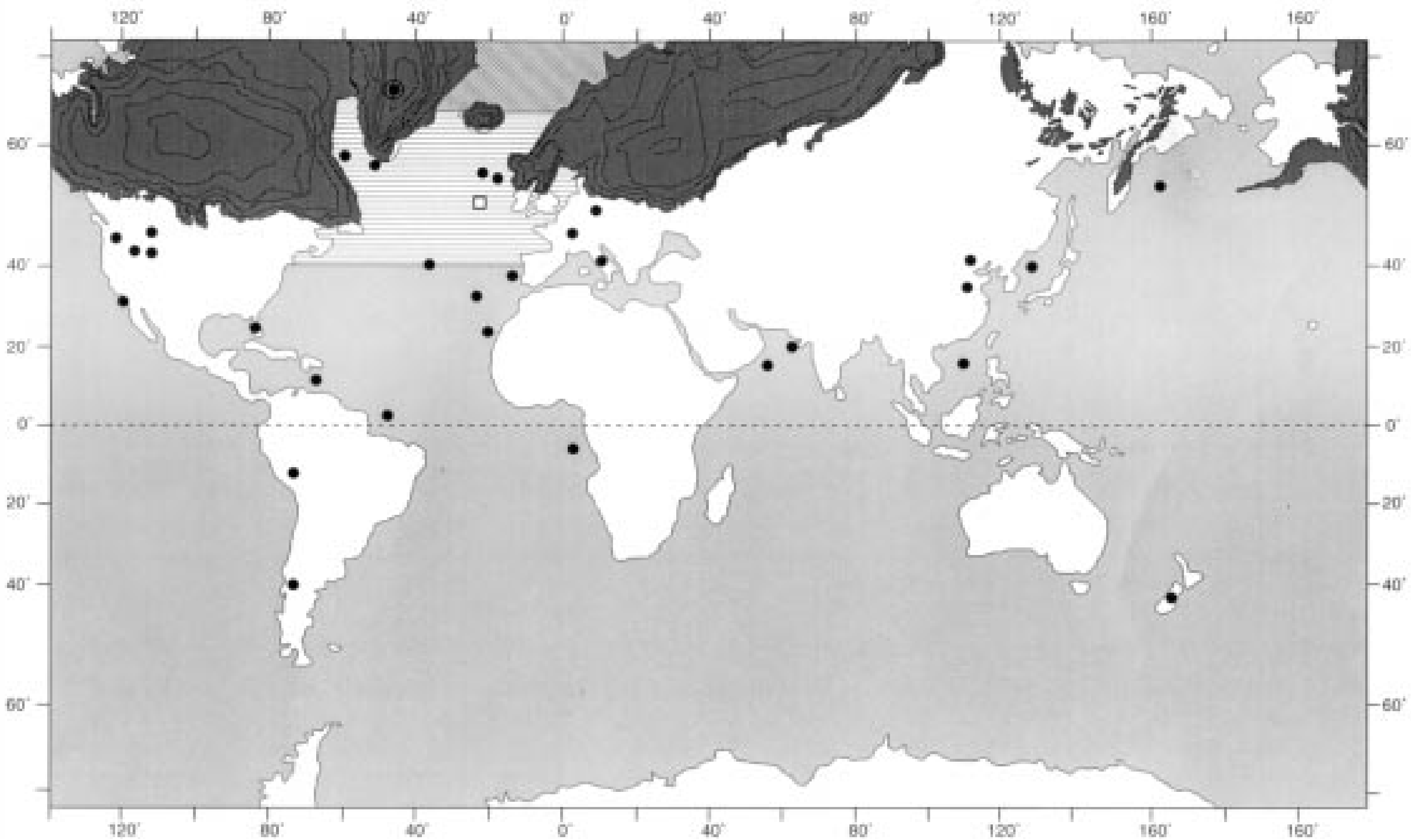

Figure 4 A compilation of current locations where the climatic influence of the Heinrich events has been described. The reconstructed ice-sheet boundaries are based on CLIMAP (1976; 1981)

have been global, although the climate shift may have been smaller outside the North Atlantic region.

The last Heinrich event (known as H1) *sensu stricto* occurred just after the Last Glacial Maximum and seems to mark the extreme cold and aridity that occurred in many parts of the world around 17 000–15 000 years ago. The Younger Dryas cold phase (see below) may also be regarded as a 'Heinrich' climate event (it is sometimes now referred to as H0), and as it has been studied in considerable detail it may give clues to a general pattern true to all Heinrich climate events. The detailed timescale on which most of these ice-rafting and climate change events began and ended is uncertain, though it cannot have been very much longer than several decades because the events themselves lasted no more than a few centuries (e.g., François and Bacon, 1993; Dowdeswell *et al.*, 1995; Mayewski *et al.*, 1997). By analogy with the well studied Younger Dryas event (see below), Heinrich events may be regarded as possibly beginning and ending with sudden climate 'jumps' taking just a few decades. However, this idea remains tentative.

Detailed studies of the timing of the Heinrich events suggest that they occurred about once every 7–13 ka during oxygen isotope stages 4, 3 and 2. Throughout the same period there were also smaller Dansgaard–Oeschger cycles, which seem to occur every 1–3 ka, with a mean cycle length of about 1500 years (Bond *et al.*, 1997). Similar but lower amplitude 1500-year oscillations occur during the Holocene interglacial (Campbell *et al.*, 1998; see below) as well as during earlier glacial and interglacial stages (Oppo *et al.*, 1998). Heinrich events only occur when there is an ice sheet on the North American continent that is big enough to collapse dramatically. Raymo *et al.* (1998) suggests that these dramatic climate events were not confined to the last few large glacial–interglacial cycles: in a relatively high-resolution North Atlantic ocean core, large sudden climate events (resembling Heinrich events and interstadials) occurred during the cooler parts of the climate cycles through the last 1.5 Myr. This would confirm that dramatic instability is a 'normal' part of the earth's climate system and that instability is not confined merely to the extreme glacial–interglacial oscillations that have operated only for the last 900 000 years.

c The Younger Dryas: The Younger Dryas cold event at about 12 900–11 500 years ago seems to have had the general features of a Heinrich event (e.g., Bond *et al.*, 1997; Severinghaus *et al.*, 1998). The sudden onset and ending of the Younger Dryas have been studied in detail in the ice-core and sediment records on land and in the sea (e.g., Björck *et al.*, 1996; Hughen *et al.*, 1996). A detailed study of Greenland ice cores (Taylor *et al.*, 1997) suggests that the main Younger Dryas-to-Holocene warming took several decades in the Arctic, but was marked by a series of warming steps, each taking less than 5 years. About half of the warming was concentrated into a single period of less than 15 years. The occurrence of a rapid rise in atmospheric methane concentration at the same time suggests that the warming and moistening of climate (causing more methane output from swamps and other biotic sources) were globally synchronized (Fuhrer and Legrand, 1997; Meeker *et al.*, 1997). According to data from the Greenland ice cores, conditions remained slightly cooler than present for a while after the main warming period; 'normal' Holocene warmth was not reached for a further 1500 years (around 10 000 calendar years ago). It is not yet clear if the general pattern of the transition between the Younger Dryas and Holocene is representative of other rapid warming and cooling events in the past 110 000 years, but similar events seem to have occurred at the beginning of the Eemian (Seidenkrantz *et al.*, 1995).

d Sudden climate transitions since the start of the Holocene: Following the sudden start of the Holocene, there have been a number of rapid, widespread climate changes recorded from the palaeoclimatic record around the world. The Greenland ice-core data again show a clear record of these events (O'Brien *et al.*, 1996; Mayewski, 1997). At least in the North Atlantic region, these changes seem to have been paced according to approximately the same 1500-year rhythm as that found for the last glacial and earlier glacial periods (Bond *et al.*, 1997; Campbell *et al.*, 1998). At the coldest point of each 1500-year cycle surface temperatures of the North Atlantic were generally about 2 °C cooler than at the warmest part, representing a fairly substantial change in climate. Regional or global fluctuations of this order would be major events if they were suddenly to affect the present-day world with its high population and finely balanced food production. It is uncertain whether these climate cycles indeed extended around the world or were generally confined to the region around the North Atlantic, but an event at 8.2 ka (see below) seems to have been widespread. There is still uncertainty in exact time correlations for other events.

The event at 8.2 ka is the most striking sudden cooling event during the Holocene, giving widespread cool, dry conditions lasting perhaps 200 years before a rapid return to climates warmer and generally moister than the present. This event is clearly detectable in the Greenland ice cores, where the cooling seems to have been about half-way as severe as the younger Dryas-to-Holocene difference (Alley *et al.*, 1997b; Mayewski *et al.*, 1997). No detailed assessment of the speed of change involved has been made, but the short duration of these events at least suggests changes that took only a few decades or less to occur. Records from north Africa across southern Asia show markedly more arid conditions involving a failure of the summer monsoon rains, although age correlations are not always reliable (e.g., Sirocko *et al.*, 1993). Cold and/or aridity also seems to have hit northernmost South America, eastern North America and parts of northwest Europe (Alley *et al.*, 1997b). Thinking of our densely populated present-day world, we can only hope that no similar event occurs in the near future.

Smaller, but also sudden and widespread, changes to drier or moister conditions have also been noted for many parts of the world for the second half of the Holocene, since about 5000 years ago (e.g., Dorale *et al.*, 1992). One fairly strong arid event occurred about 4000 years ago across northern Africa and southern Asia. It remains to be seen whether these later events will eventually fit into a consistent global 1500-year pattern of cold/arid events, the last of which may have been the Little Ice Age which ended about 350 years ago (see below; Bradley and Jones, 1992; Bond *et al.*, 1997).

The effects of mid-Holocene climate fluctuations on regional ecology are still being worked out. In Holocene pollen records, elm (*Ulmus*) declined in Europe (about 5000 ^{14}C yr or 5700 calendar yr ago) and hemlock (*Tsuga*) in North America (about 4700 ^{14}C yr or 5300 calendar yr ago). Both vegetation changes have been attributed to specific pathogen attacks (Rackham, 1980; Peglar, 1993), but with the evidence from the Eemian it may now be worth considering these declines in terms of climate deterioration, or at least its effect on the spread of epidemics (A. Parker, pers. comm.).

e Sudden climate jumps in the more recent past: Different sources seem to suggest differing speeds and intensities for Holocene climate events. The Little Ice Age began in late medieval times and played a role in extinguishing Norse colonies on Greenland (e.g., Barlow *et al*., 1997). It ended at about AD 1650 (Bradley and Jones, 1992), and may have been the most rapid and largest change in polar circulation during the Holocene according to chemical indicators of windblown sea salt in the GISP2 ice core (O'Brien *et al*., 1996; Mayewski *et al*., 1997). The event even shows up in high-resolution marine records in the northern Atlantic (Keigwin, 1996), but may have been of less importance in other regions (Bradley and Jones, 1992). This would seem to imply an event which was clearly intense in some regions (such as the dry phase around 8200 years ago across so many low and mid-latitude regions), but was not so important in other regions (such as near to the poles). The Little Ice Age was thus a climate oscillation (fairly small by comparison with many of the events recorded in ice cores and sediment records) which gave cooler conditions over the lands around the North Atlantic between about 700 and 200 years ago.

Additional, smaller changes are observed in the detailed Greenland ice-cap record, but not all the rapid changes observed in the Greenland ice cap correspond to noticeable climate changes elsewhere. For example, a warming of 4 °C per decade was observed in an ice core from northern Greenland for the 1920s (Dansgaard *et al*., 1989), but this corresponded to a global shift of 0.5 °C or less. There is some evidence that this event may have been widespread (Thompson *et al*., 1993), but by no means has it been demonstrated to have been global. For this reason it is always desirable to have sources of evidence from other regions before invoking a broad, dramatic climate shift. What this relatively recent climate shift does suggest though, is that the climate system tends to undergo most of its changes in sudden jumps, even if those changes are relatively small against the background of those seen during the Quaternary. This is further evidence that if and when the next climate shift occurs, it will not be a gradual century-on-century change but rather a sudden step-function that will begin suddenly and occur over a decade or two.

IV The mechanisms behind sudden climate transitions

It is still unclear how the climate on a regional or even global scale can change as rapidly as present evidence suggests. It appears that the climate system is more delicately balanced than had previously been thought, linked by a cascade of powerful mechanisms that can amplify a small initial change into a much larger shift in temperature and aridity (e.g., Rind and Overpeck, 1993). At present, the thinking of climatologists tends to emphasize several key components.

1 North Atlantic circulation as trigger or amplifier of rapid climate changes

The circulation of the North Atlantic Ocean probably plays a major role in either triggering or amplifying rapid climate changes in the historical and recent geological record (e.g., Keigwin *et al*., 1994; Shaffer and Bendtsen, 1994; Broecker, 1995; Jones *et al*., 1996; Rahmstorf *et al*., 1996; Seidov and Maslin, in press; Stocker, in press). The North

Atlantic has a peculiar circulation pattern: the north–east trending Gulf Stream carries warm and relatively salty surface water from the Gulf of Mexico up to the Nordic seas. Upon reaching these regions, the surface waters cool sufficiently to be dense enough to sink into the deep ocean. The 'pull' exerted by this deep-water formation is thought to help maintain the strength of the warm Gulf Stream, ensuring a current of warm tropical water into the North Atlantic that sends mild air masses across to the European continent (e.g. Rahmstorf *et al*., 1996; Stocker, in press).

If the sinking process in the North Atlantic were to diminish or cease, the weakening of the warm Gulf Stream would mean that Europe had colder winters (e.g., Broecker, 1995). However, the presence of the Gulf Stream does not give markedly warmer summers in Europe – more the opposite, in fact – so a shutting off of the mild Gulf Stream air masses does not in itself explain why summers also became colder during sudden cooling events (and why ice masses started to build up on land due to winter snows failing to melt during summer).

In the North Atlantic itself, sea ice would form more readily in the cooler winter waters due to a shut-off of the Gulf Stream, and for a greater part of the year the ice would form a continuous lid over the North Atlantic. A lid of sea ice over the North Atlantic would last for a greater proportion of the year; this would reflect back solar heat, leading to cooler summers on the adjacent landmass, as well as to colder winters (e.g., Jones *et al*., 1996; Overpeck *et al*., 1997). With cooler summers, snow cover would last longer into the spring, further cooling the climate by reflecting back the sun's heat. The immediate result of all this would be a European and west Siberian climate that was substantially colder, and substantially drier because the air that reached Europe would carry less moisture, having come from a cold sea-ice surface rather than the warm Gulf Stream waters.

After an initial rapid cooling, the colder summers would also tend to allow the snow to build up year-on-year into a Scandinavian ice sheet, and as the ice built up it would reflect more of the sun's heat, further cooling the land surface, and giving a massive high pressure zone that would be even more effective at diverting Gulf Stream air and moisture away from the mid-latitudes of Europe. This would reinforce a much colder regional climate.

The trigger for a sudden 'switching off' or a strong decrease in rate of deep-water formation in the North Atlantic must be found in a decrease in density of surface waters in the areas of sinking in the northern Atlantic Ocean. Such a decrease in density would result from changes in salinity (addition of freshwater from rivers, precipitation or meltwater), and/or increased temperatures (Dickson *et al*., 1988; Rahmstorf *et al*., 1996). For example, an exceptionally wet year on the landmasses which drain into the Arctic Sea (Siberia, Canada, Alaska) would lead to such a decreased density. Ocean circulation modelling studies suggest that a relatively small increase in freshwater flux (called 'polar halocline catastrophe') to the Arctic Sea could cause deep-water production in the North Atlantic to cease (e.g., Mikolajewicz and Maier-Reimer, 1994; Rahmstorf, 1994; Rahmstorf *et al*., 1996).

During glacial phases, the trigger for a shut-off or a decrease in deep-water formation could be as follows: (1) an increase in the amount of melting icebergs entering the North Atlantic, e.g., Heinrich events (Figure 3); (2) the sudden emptying into the northern seas of a lake formed along the edge of a large ice sheet on land (for instance, the very large ice-dammed lake that existed in western Siberia); or (3) a diversion of a meltwater stream from the North American Laurentide Ice Sheet through the Gulf of St Lawrence, as seems to have occurred as part of the trigger for the Younger Dryas cold event (e.g., Kennett, 1990; Berger and Jansen, 1995). A pulse of fresh (melt) water would dilute the dense salty Gulf Stream and form a temporary lid that stops the sinking of water that helps drive the Gulf Stream. The Gulf Stream could then weaken, and its northern end (the North Atlantic Drift) could switch off altogether, breaking the 'conveyor belt' (Figure 5) and allowing an extensive sea ice cap to form across the North Atlantic, which would prevent the ocean current from starting up again at its previous strength. Theoretically, the whole process could occur very rapidly, in the space of just a few decades or even several years. The result could be a very sudden climate change to colder conditions, as has happened many times in the area around the North Atlantic during the last 130 ka. The models used by Seidov and Maslin (in press) show that Heinrich events did cause a total collapse of the deep-water thermohaline conveyor belt. However, this cessation of the conveyor is independent of the origin and magnitude of the meltwater produced during a Heinrich event as the essential prerequisite for cessation is that enough meltwater is transported to the central northern North Atlantic convection sites. Modelled meltwater events that were constrained within the Nordic seas did reduce North Atlantic deep-water (NADW) production but did not cause it to collapse; this may also provide a mechanism for the Dansgaard–Oeschger events.

The sudden switch off of the NADW can also occur in the opposite direction, for example, if warmer summers caused the sea ice to melt back to a critical point where the sea-ice lid vanished; the NADW could just as rapidly start up again. Indeed, following an initial cooling event the evaporation of water vapour in the tropical Atlantic could result in an 'oscillator' whereby the salinity of Atlantic Ocean surface water (unable to sink into the North Atlantic because of the lid of sea ice) built up to a point where strong sinking began to occur anyway at the edges of the sea-ice zone (e.g., Broecker, 1995). The onset of sinking could result in a renewed northward flux of warm water and air to the North Atlantic, giving a sudden switch to warmer climates, as is observed many times within the record of the last 130 ka.

The process of switching off or greatly diminishing the flow of the Gulf Stream would not only affect Europe. Antarctica would become even colder than it is at present, because much of the heat it receives now ultimately comes from Gulf Stream water that sinks in the North Atlantic, travels in a sort of river down the western side of the deep Atlantic Basin and then at least partially resurfaces just off the bays of the Antarctic coastline (e.g., Schmitz and McCarthey, 1993; Stocker, in press). Even though this water is only a few degrees above freezing when it reaches the surface, it is much warmer than the adjacent Antarctic continent, helping to melt back some of the sea ice that forms around Antarctica in the ice-free regions called polynyas (Comiso and Gordon, 1987). The effect of switching off the deep-water heat source would be cooler air and a greater sea-ice extent around Antarctica, reflecting more sunlight and further cooling of the region. However, the North Atlantic deep water takes several hundred years to travel from its place of origin to the Antarctic coast, so it could only produce an effect a few centuries after the change occurred in the north.

It is not known what delay was present in the various climate changes that occurred between the north Atlantic region and Antarctica, but it is generally thought that other (relatively indirect) climate mechanisms, such as greenhouse gases in the atmosphere,

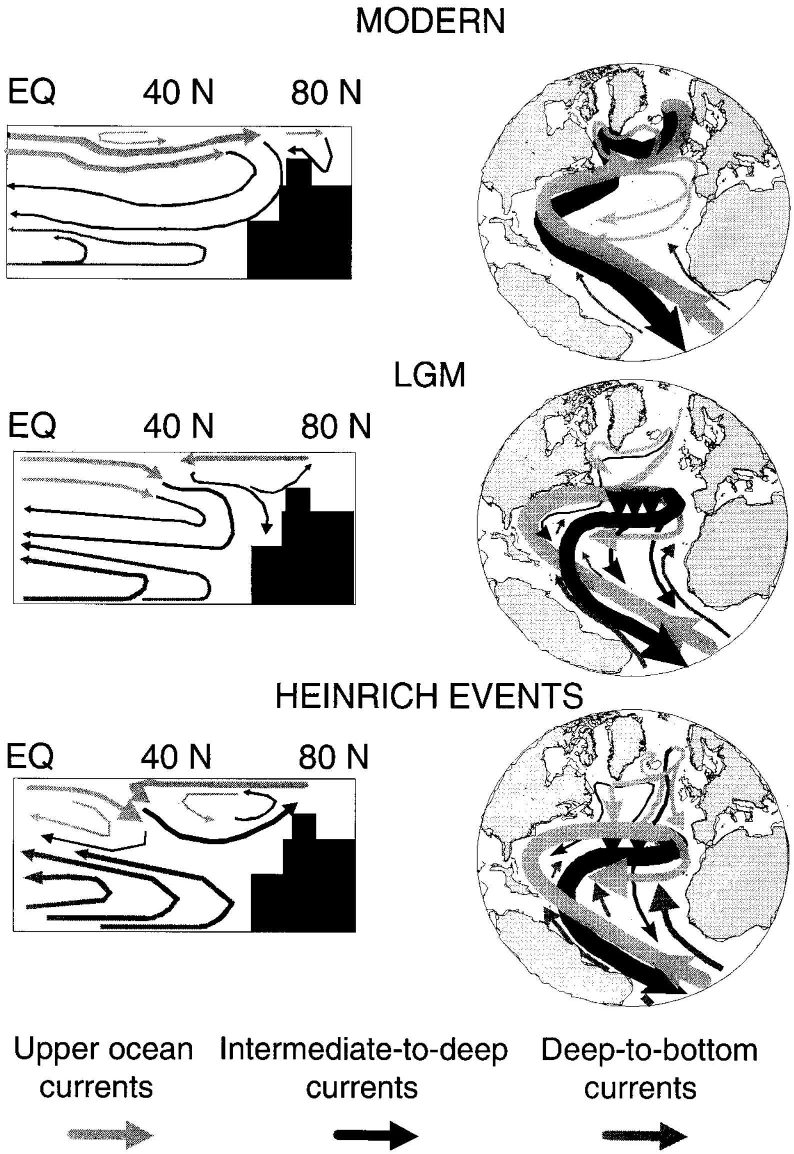

Figure 5 North Atlantic conveyor during interglacial (modern Holocene), glacial and Heinrich event scenarios. Typical overturning patterns are shown on the left panel (see also Table 1), and corresponding three-dimensional circulation patterns are sketched on the right (EQ is the equator). The key role of the North Atlantic current zonality and deep convection is emphasized. Note that there is no deep convection during Heinrich events
Source: Adapted from Seidov and Maslin (in press)

linked these two far-flung regions and sometimes produced closely synchronized changes (i.e., within a few centuries of one another).

Although the end of the Last Glacial and various other sudden climate events such as Heinrich events do show up in the Antarctic ice record, not all large changes show such a closely linked occurrence and timing around the world. For example, there is no clear trace of the Younger Dryas in the Vostok ice core from Antarctica (Chapellaz *et al.*, 1993; Broecker, 1998), and the warming at the start of the Eemian also does not seem to be closely linked to the timing of the warming at northern latitudes (Sowers *et al.*, 1993).

In general during glacial phases, deep-water formation in the Nordic seas would have ceased or diminished. In the most intense cold phases the deep-water formation area moved south of the British Isles, to approximately 45 °N (e.g., Duplessy *et al.*, 1984; Imbrie *et al.*, 1992; Maslin *et al.*, 1997). Glacial deep-water formation seems to have been weaker than at present and seems to have penetrated to mid-depths rather than to the deepest ocean basins. This was at least partly probably because the whole surface of the Atlantic Ocean (even the tropics) was cooler. As a result, there was less evaporation from the ocean's surface and the water that did reach northwards was less briny and thus less dense; consequently it was less able to sink when it reached the cold edge of the sea-ice zone. An initial slowdown of North Atlantic circulation may have been the initial trigger for a set of amplifying factors (see below) that rapidly led to a cooling of the tropical Atlantic, reinforcing the sluggish state of the glacial-age Gulf Stream.

The idea of Gulf Stream slowdowns as a mechanism in climate change is not merely theoretical. There is evidence from the study of ocean sediments that deep-water formation in the North Atlantic was diminished during the sudden cold Heinrich events and other colder phases of the last 130 ka, including the Younger Dryas phase (e.g., Fairbanks, 1989; Kennett, 1990; Maslin *et al.*, 1995). The process also 'switched on' rapidly at times when climates suddenly warmed around the North Atlantic Basin, such as at the beginning of interstadials or the beginning of the present interglacial (Rasmussen *et al.*, 1997). Deep-water formation decreased when the climate was cooling towards the end of an interstadial, and it diminished suddenly with the final cooling event that marked the end of the interstadial (Rasmussen *et al.*, 1997). This happened over a period of less than 300 years at the beginning of the Younger Dryas (e.g., Berger and Jansen, 1995).

The intra-Eemian cooling event – if it really did happen – is perhaps one of the most alarming of the changes observed from recent climate history, because it occurred in the midst of an interglacial not too dissimilar from our own. There are strong signs that a rapid cut-off in North Atlantic deep-water formation was associated with this climate change (Maslin and Tzedakis, 1996). This event near 122 ka seems to have coincided with intense cooling and freshening in the seas west of Ireland and in the northern Norwegian Sea (Cortijo *et al.*, 1994; Fronval and Jansen, 1996). Today any similar reduction in the salinity of the sea surface in these two areas would be enough to result in a reduction of deep-water formation in the Nordic seas (e.g., Bryan, 1986; Dickson *et al.*, 1988); Seidov and Maslin, 1996; in press). This reduction could explain why Europe (and perhaps other areas) cooled so dramatically (Maslin *et al.*, 1998).

What might have caused this freshening of the North Atlantic seas? The freshwater could have come from melting icebergs, from increased rain and snowfall in the region, or a change in surface currents that brought in fresher water from another part of the ocean. There is no evidence of an increased supply of melting icebergs (Keigwin *et al.*

1994; McManus *et al.*, 1994) during this phase. Instead, one or more of the following factors seems a more likely cause: (1) greater incursion of relatively 'fresh' North Pacific water through the Bering Strait and around the Arctic seas to the North Atlantic during times of raised sea level during the Eemian (Shaffer and Bendtsen, 1994); (2) enhanced precipitation over the North Atlantic and the Nordic, due to the greater amount of solar heat reaching the northern latitudes during the summer at that time (due to changes in the earth's orbital parameters; see below); and (3) the closure of the northern Baltic Sea link to the North Atlantic Ocean, which would ensure that more freshwater ended up in the area (van Andel and Tzedakis, 1996).

We do not yet know how widespread the event was, or what caused it, but its occurrence at least points to the NADW formation being a 'weak point' in the interglacial climate system (e.g., Rahmstorf *et al.*, 1996; Broecker, 1997a; 1997b), with the potential to affect climate rapidly in the North Atlantic region and perhaps elsewhere. This is rather worrying, for we are presently in the midst of an interglacial phase not greatly unlike the Eemian.

Broader changes in temperature and rainfall over much of the world are thought likely to have occurred as a result of a switching on or off of the deep North Atlantic circulation (Rind and Overpeck, 1993; Jones *et al.*, 1996). These changes would result in amplification by the feedback mechanisms suggested below. As evidence of such a broader link to global climate, over recent years changes in the monsoon-belt climates of Africa and Asia have been observed to occur in association with decadal-scale phases of weaker North Atlantic circulation (e.g., Hurrell, 1995; 1996). By extrapolation, it is generally thought that bigger changes in the North Atlantic circulation would result in correspondingly larger changes in climates in the monsoon belts and in other parts of the world (e.g., Sirocko *et al.*, 1993).

In addition to this relatively direct effect of deep water on North Atlantic and Antarctic climate, other subtle effects on global climate would be expected to result from a sudden change in North Atlantic circulation, or indeed they may themselves trigger a change in the North Atlantic circulation by their effects on atmospheric processes. These include the interaction with global carbon dioxide concentrations, dust content and surface reflectivity (albedo).

2 Carbon dioxide and methane concentration as a feedback in sudden changes

Analysis of bubbles in ice cores shows that at the peak of glacial phases, the atmospheric CO_2 concentration was about 30% lower than during interglacial conditions (e.g., Jouzel *et al.*, 1993). We do not at present know whether the lower glacial CO_2 levels were a cause or an effect of the ice ages. Lower CO_2 levels might result from changes in plankton productivity (e.g., Lyle *et al.*, 1988; see also discussion in Thomas *et al.*, 1995), drawing more carbon down out of the atmosphere once climate began to cool. Terrestrial biomass reservoirs seem to move in the opposite direction, releasing carbon to the atmosphere with the onset of glaciation (Adams and Faure, 1998).

The lower carbon dioxide concentrations resulting from greater ocean carbon storage would cool the atmosphere, and allow more snow and ice to accumulate on land. Relatively rapid changes in climate, occurring over a few thousand years, could have resulted from changes in the atmospheric CO_2 concentration (e.g., Broecker, 1997a;

1997b). The actual importance of carbon dioxide in terms of the climate system is unknown, though computer climate simulations tend to suggest that it directly cooled the world by less than 1°C on average. Due to amplification of this change by various feedback factors within the climate system (such as the water vapour content) the resulting change in global climate could have been more than 2 °C (e.g., Houghton, 1994). In addition, warming as a result of increased atmospheric CO_2 levels is greater at higher latitudes, possibly inducing snow and ice melt.

A problem with invoking atmospheric carbon dioxide levels as a causal factor in sudden climate changes is that they seem to have varied too slowly, following on the timescale of millennia what often occurred on the timescale of decades – but the resolution of our records may not be good enough to resolve this question. Methane, a less important greenhouse gas, was also 50% lower during glacial phases (e.g., Sowers *et al.*, 1993; Chapellaz *et al.*, 1997), probably due to reduced biological activity on the colder, drier land surfaces (Meeker *et al.*, 1997). Methane levels seem to have increased rapidly in concentration in association with changes in climate, reaching Holocene levels in around 150 years or less during the global climate warming at the end of the Younger Dryas, around 11 500 years ago (Taylor *et al.*, 1997).

Such sudden rises in atmospheric methane concentration were probably not important in affecting climate; the warming effect of a 50% change in methane would have been much less than an equivalent change in CO_2, because methane is at such a low overall concentration in the atmosphere. It has been suggested that another mechanism, involving sudden and short-lived releases of massive amounts of methane from the ocean floors, could sometimes have resulted in rapid warming phases. These do not leave any trace in terms of raised methane levels in the ice-core data, where the trapped gas bubbles generally only indicate methane concentrations at a time resolution of centuries rather than the few years or decades that such a 'methane pulse' might last for. However, ice cores from areas where the ice sheet built up particularly rapidly (Chapellaz *et al.*, 1993; Taylor *et al.*, 1997) show detailed time resolution of the record of methane concentration in the atmosphere, and fail to show evidence of sudden 'bursts' of methane. However, the idea does remain a possibility, to at least some biogeochemists (Brook *et al.*, 1996).

3 Surface reflectivity (albedo) of ice, snow and vegetation

The intensely white surface of sea ice and snow reflects back much of the sun's heat, hence keeping the surface cool. Presently, about a third of the heat received from the sun is reflected back into space, and changes in this proportion thus have the potential to influence global climate strongly (e.g., Crowley and North, 1991). In general the ice cover on the sea, and the snow cover on the land, have the potential to set off rapid climate changes because they can either appear or disappear rapidly given the right circumstances. Ice sheets are more permanent objects which, whilst they reflect a large proportion of the sunlight that falls upon them, take hundreds of years to melt or build up because of their sheer size. When present, sea ice or snow can have a major effect in cooling regional and global climates, but with a slight change in conditions (e.g., just a slightly warmer summer) they will each disappear rapidly, giving a much greater warming effect because sunlight is now absorbed by the much darker sea or land cover underneath. In an unusually cold year, the opposite could happen, with snow staying on the ground throughout the summer, itself resulting in a cooler summer climate. A runaway change in snow or sea ice (positive feedback) could thus be an important amplifier or trigger for a major change in global temperature. Slow changes over millennia or centuries could bring the climate a break point or threshold, involving a runaway change in snow and ice reflectivity over a few decades. These slow background changes might include variations in the earth's orbit (affecting summer sunlight intensity), or gradual changes in carbon dioxide concentration, or in the northern forest cover which affects the amount of snow that is exposed to sunlight.

It is possible that the relatively long-lived ice sheets might occasionally help bring about very rapid changes in climate, by rapidly 'surging' outwards into the sea and giving rise to large numbers of icebergs that would reflect back the sun's heat and rapidly cool the climate (e.g., MacAyeal, 1993a; 1993b). The intensely cold Heinrich events that punctuated the Last Ice Age were initially thought to be caused by sudden slippage of the Laurentide Ice Sheet that covered most of Canada. It appears, however, that all the separate ice sheets around the North Atlantic surged outwards simultaneously, and that their outwards movement probably thus represents a secondary response to an initial climate cooling (e.g., a change in the deep-water formation system in the North Atlantic) rather than the initial trigger (Bond *et al*., 1997). This does not mean that ice surges and icebergs were irrelevant in the extreme cold of Heinrich events; by their albedo effects they may have helped to intensify and temporarily stabilize a cooling event that would have occurred anyway. However, this amplification may have occurred decades or centuries after the initial 'step function' event associated with the rapid cooling.

Another, possibly neglected, factor in rapid regional or global climate changes may be the shifts in the albedo of the land surface that result from changes in vegetation or algal cover on desert and polar desert surfaces. An initial spreading of dark-coloured soil surface algae or lichens following a particularly warm or moist year might provide a 'kick' to the climate system by absorbing more sunlight and thus warming the climate, and also reducing the dust flux from the soil surface to the atmosphere (see below). Larger vascular plants and mosses might have the same effect on the timescale of years or decades. Detailed analysis of the ending of the Younger Dryas (Taylor *et al*., 1997) suggests that warming occurred around 20 years earlier in lower and mid-latitudes, perhaps due to some initial change in vegetation or snow cover affecting land surface albedo. Some of the earlier climate warming events during the last 130 ka show similar signs of changes in dust flux followed by changes in high-latitude temperature (Raymo, pers. comm.).

4 Water vapour as a feedback in sudden changes

Water vapour is a more important greenhouse gas than carbon dioxide and, as its atmospheric concentration can vary rapidly, it could have been a major trigger or amplifier in many sudden climate changes. For example, a change in sea-ice extent or in carbon dioxide would be expected to affect the flux of water vapour into the atmosphere from the oceans, possibly amplifying climate changes. Large, rapid changes in vegetation cover might also have added to these changes in water vapour flux to the atmosphere. W.S. Broecker suggested in a recent lecture (1997) that water vapour may act as a global 'messenger', co-ordinating rapid climate changes, many of which seem to have occurred all around the world fairly simultaneously, or in close succession. Broecker notes the evidence for large changes in the water vapour content of the atmosphere in terms of changes in the water vapour content of the atmosphere in terms of changes in the ^{18}O content of tropical high Andean ice cores (Thompson *et al*., 1995).

5 Dust and particulates as a feedback in sudden changes

Particles of mineral dust, plus the aerosols formed from fires and from chemicals evaporating out of vegetation and the oceans, may also be a major feedback in co-ordinating and amplifying sudden large climate fluctuations. Ice cores from Greenland (Mayewski *et al*., 1997; Ram and Koenig, 1997; Taylor *et al*., 1997), Antarctica (Jouzel *et al*., 1996) and tropical mountain glaciers (Thompson *et al*., 1995) show greater concentrations of mineral dust during colder phases. This suggests that there was more dust in the earth's atmosphere during cold periods than during warm phases. It seems that the atmospheric content of dust and sulphate particles changed very rapidly, over just a few decades, during sudden climate transitions in the Greenland ice-core record (Taylor *et al*., 1997). At least some of these sulphate particles may have been derived from large volcanic explosions, which can induce global cooling on short timescales. The rapid transition between interglacial stage 5a and glacial stage 4, for instance, occurred coeval with the huge eruption of Toba volcano in Indonesia (e.g., Linsley, 1996).

The drier and colder the world gets, the more desert there is and the higher the wind speeds, sending more desert dust into the atmosphere where it may reinforce the cold and dryness by forming stable 'inversion' layers that block sunlight and prevent rain-giving convective processes. A run of wet years in the monsoon belt could trigger rapid revegetation of desert surfaces by vascular plants or algae, and a sudden decrease in the amount of dust blown into the atmosphere. Less dust could help make conditions still warmer and wetter, pushing the climate system rapidly in a particular direction (though dust and other particles might actually tend to warm the surface if they blow over lighter-coloured areas covered by snow or ice; Overpeck *et al.*, 1997).

It has even been proposed that variations in the influx of dust from outer space (interplanetary dust particles, IDPs) could have played a role in triggering the large-scale glacial/interglacial alternations at 100 ka periodicities (Farley and Patterson, 1995; Muller and MacDonald, 1997). It was proposed that the accretion rate of IPDs might be linked to the varying inclination of the earth's orbit with respect to the invariable plane of the solar system. It now appears, however, that this process is not likely to have been of influence on ice ages on earth (Kortenkamp and Dermott, 1998).

Haze production is a poorly understood but potentially very significant factor in triggering or amplifying sudden climate changes. Given what is known of the present-day patterns of emission of haze-producing compounds from land vegetation, decade-to-century timescale changes in vegetation distribution and activity could have resulted in rapid changes in global haze production. This factor eludes the sedimentological record, and it may be impossible to test it with observations from the past.

6 Seasonal sunlight intensity as a background to sudden changes

A major background factor in pacing climate switches on timescales of tens of thousands of years seems to have been the set of 'Milankovitch' rhythms in seasonal sunlight distribution or insolation (Imbrie *et al*., 1992; 1993a; 1993b). Although the insolation values change gradually over thousands of years, they may take the earth's climate to a 'break point' at which other factors will begin to amplify change into a sudden transition.

There are three Milankovitch cycles. 'Eccentricity' is the variation in the earth's orbit (from elliptical to nearly circular) that occurs at periodicities of about 100 ka. Eccentricity alters the total amount of solar radiation received on the earth. Obliquity is the second cycle. This is the variation in the degree of tilt of the earth's axis and it has a periodicity of 42 ka. The third is 'precession', which is the timing of the seasons relative to the earth's elliptical track (i.e., nearer and further from the sun) and this cycle has periodicities of 19 and 23ka. The latter two rhythms alter the relative amount of solar radiation reaching the earth's Northern and Southern Hemispheres during summer and winter. The times when summer sunlight in the Northern Hemisphere is strong (but when winter sunlight is correspondingly weak) tend to be the times when the rapid global transition from glacial to interglacial conditions occurs.

Glacial–interglacial cycles roughly follow the 100 000-year timescale of the last 900 ka where the three different rhythms have varied the temperatures of the North Hemisphere summer. The lesser individual rhythms, however, can also be detected in the temperature record of the 19 and 42 ka timescales. The timing of interglacial onset, in fact, tends more closely to follow multiples of the 19 ka cycle than an exact correspondence to the 100 ka cycle (Imbrie *et al*., 1992; 1993a; 1993b). This is thought to be the result of the effects of summer temperatures on the various factors mentioned above – for example, it ensures the melting back of snow and sea-ice in summer, which helps the earth absorb more solar radiation and thus to heat up further. It is generally accepted that the effect of changes in heat budget as a result of the Milankovitch variations by themselves is not enough to bring about the large, rapid changes in climate that follow these rhythms in seasonal sunlight. Some set of positive feedback factors – directly or indirectly linked to the seasonal insolation changes – must be involved in bringing the earth out of glacial and into interglacial conditions.

It is important to note, however, that most of the very rapid climate transitions during the last 100 000 years do not show any clear association in timing with the background Milankovitch rhythms, especially the fluctuations at periodicities below 10 ka. In these cases their ultimate trigger must lie in other factors – probably a combination of many processes that sometimes line up to set the climate system on a runaway course in either the direction of cooling or warming.

V Could dramatic decadal timescale climate transitions occur in the near future?

We have shown in this review that, over the last 150 ka, there have been numerous large climate changes that have occurred on the timescale of individual human lifetimes – for example, the end of the Younger Dryas. Other substantial climate shifts documented in the Quaternary took, at most, a few centuries. However, we are still limited in our understanding of these rapid events. The first problem is that, currently, there are only a few sites where we can obtain records with a decadal or annual resolution. The best records we have at the moment are the ice cores, but future results from the IMAGES programme (Figure 6) and the Ocean Drilling Program (e.g., Saanich Inlet (western Canada; leg 169A) and northern Atlantic (leg 172)), will provide additional exciting high-resolution data. The international palaeoclimate community is aware of the need for concerted action to obtain high-resolution data and, under the IGBP-PAGES programme, terrestrial, ice and oceanic data are being collected (see Figure 6). Figure 6 shows the combination of the current international programmes operating. The second problem is our limited ability to develop age models, which makes comparison of different regions and different data sets difficult. Blunier *et al.* (1998) argue that they can compare Greenland and Antarctic ice cores using fluctuations in atmospheric methane as the age control. They argue that Greenland warming during a Dansgaard–Oeschger event lags 1–2ka behind Antarctic warming. However, it is currently impossible to undertake this type of study using ocean or terrestrial data. The third problem concerns finding suitable mechanisms. Currently, we have a basic understanding of variations greater than 10 ka, which are due to orbital forcing (see section IV 6), and of variations of less than a decade – the El Niño South Oscillation (ENSO), the North Atlantic Oscillation (NAO) and solar cycles. We, however, have no understanding of what may cause climate cycles that occur between 10 years and 10 ka (see Figure 7). There are three possible causes of millennial and submillennial events: (1) harmonics of astronomical forcing; (2) superdecadal or century-scale ENSOs or NAOs; or (3) a new mechanism entirely.

It is, thus, extremely difficult to quantify the risk of a sudden switch in global or regional climate because the mechanisms behind all past climate changes (sudden or otherwise) are incompletely understood. However, they appear to be real, and relatively small-scale changes in North Atlantic salinity have been observed and studied in the last few decades (Dickson *et al.*, 1988). Fluctuations in surface water characteristics and precipitation patterns in that region vary on decadal timescales, with variations in the strength of high-pressure areas over the Azores and Iceland (the NAO; Hurrell, 1995; 1996) providing an observed apparent link between salinity and climate fluctuations. The fear is that relatively small anthropogenic changes in high-latitude temperature as a result of increased concentrations of greenhouse gases might switch North Atlantic circulation and alter the course of the Gulf Stream during such natural fluctuations.

As we do not know how often decadal timescale changes occurred in the recent geological past, we are handicapped in trying to find mechanisms which might be used for forecasting future events. Even if we knew everything there was to know about past climate mechanisms, it is likely that we would still not be able to forecast such events confidently into the future. This is because the system will have been influenced by probabilistic processes (due to the chaotic nature of the ocean climate system, with runaway changes coming from minuscule differences in initial conditions – e.g., Crowley and North, 1991). Hence, it is not justifiable to talk in terms of what 'definitely' will or will not happen in the future – even though the public and policy-makers are looking for certainties. All we can reasonably do is set out what the current understanding is, acknowledging that this understanding is limited and may turn out to be

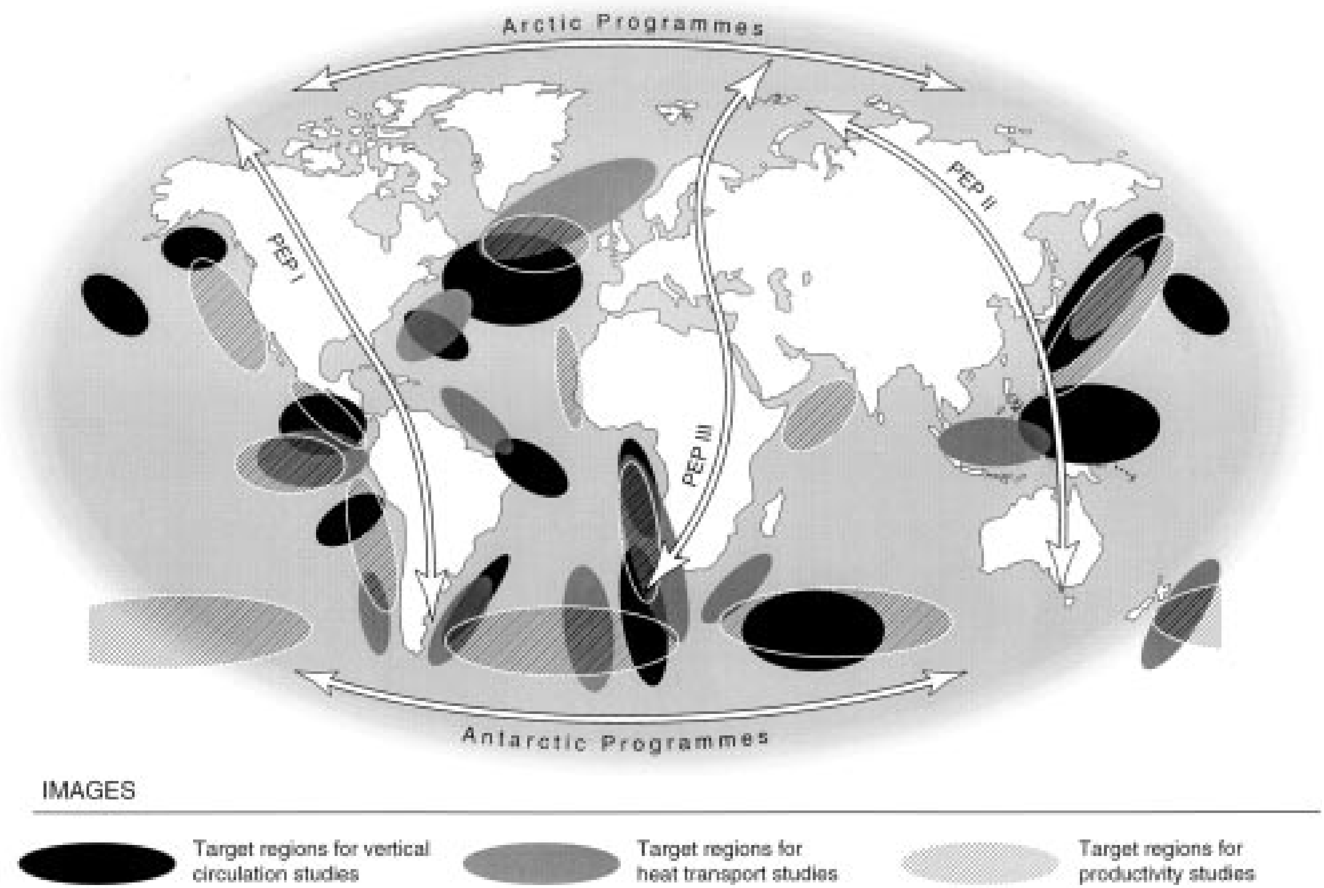

Figure 6 Global map of the planned and ongoing palaeoclimate studies. These programmes include the IGBP-PAGES Pole–Equator (PEP) transacts, the Arctic and Antarctic programmes and IMAGES (indicated by the ocean ellipses). Another major international programme which contributes greatly to palaeoclimatology is the Ocean Drilling Program
Source: Adapted from Maslin and Berger (1997)

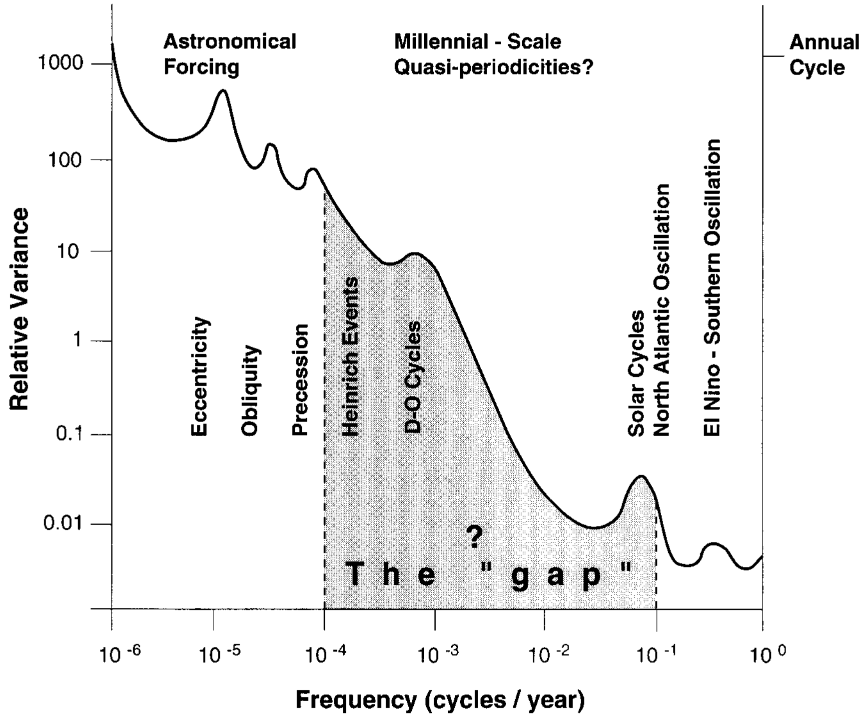

Figure 7 Spectrum of climatic variance, illustrating our knowledge gap, i.e., the lack of suitable 'climate change' mechanisms on the scale of 10 years to 10 ka
Source: Adapted from Overpeck (pers. comm.)

wrong in certain key respects. We can then talk in terms of probabilities of particular events occurring.

There is a possibility that most of the climate instability seen in the recent geological past is not relevant to our immediate future because it represents a different system – a 'glacial' state almost certainly characterized by a different pattern of deep-sea circulation (e.g., Rahmstorf *et al.*, 1996). Most of the rapid climate transitions during the last 150 000 years seem to have occurred against the background of a world that had a larger northern ice-sheet extent than at present, perhaps indicating that in this glacial mode the climate is predisposed to be more unstable than in our present interglacial state (e.g., Rind and Chandler, 1991).

However, there were at least some rapid climate transitions which occurred when ice-sheet extent was no greater than at present, such as the apparently widespread late Holocene cool/arid events at 8200 yr BP, at around 3800 yr BP and another cool event around 2600 yr BP (although the time taken for onset of these later Holocene changes in regional and global climates does not yet seem to have been determined).

Various large, full interglacial climate changes during the Holocene and certain earlier interglacials (e.g., the Eemian and the Holstein interglacials in Europe; Winograd *et al.*, 1997) that show up in the Greenland ice cap also seem to correlate with genuinely large climate shifts in Europe and elsewhere. Whether they occurred over decades, centuries or thousands of years, they offer a worrying analogue for what might happen if greenhouse gas emissions continue unchecked. Judging by its past behaviour under both glacial (e.g., the ending of the Younger Dryas) and interglacial conditions (e.g., the various Holocene climate oscillations leading up to the twentieth century; Alley *et al.*, 1997b), the climate has a tendency to remain stable for most of the time and then suddenly 'flip' – at least sometimes over just a few decades – as a result of the influence of the various triggering and feedback mechanisms. Such observations suggest that even without anthropogenic climate modification there is always an axe hanging over our head in the form of random, very large-scale changes in the natural climate system: a possibility policy-makers should perhaps bear in mind when considering contingency plans and international treaties designed to cope with sudden famines on a greater scale than any experienced in written history. By disturbing the system, humans may simply be increasing the likelihood of sudden events which might have occurred anyway.

Another source of evidence seems to underline the potential importance of sudden climate changes in the coming centuries and millennia. Computer modelling studies of the (still incompletely understood) NADW formation system suggest that this system is, indeed, sensitive to quite small changes in freshwater runoff from the adjacent continents, whether from river fluxes or meltwater from ice caps (Rahmstorf *et al.*, 1996). Some scenarios in which atmospheric carbon dioxide levels are allowed to rise to several times higher than at present result in increased runoff from rivers entering the Arctic Basin and a rapid weakening of the Gulf Stream resulting in colder conditions (especially in winter) across much of Europe. Simply doubling the amount of carbon dioxide in the atmosphere could be enough to set off such a change (Broecker, 1997b). Whilst these are only preliminary models, and thus subject to revision as more work is done, they do seem to point in the same direction as the ancient climate record in suggesting that sudden shutdowns or intensifications of the Gulf Stream circulation might occur under full interglacial conditions, and be brought on by the disturbance caused by rising greenhouse gas levels. To paraphrase W.S. Broecker: 'Climate is an ill-tempered beast, and we are poking it with sticks.'

Acknowledgements

We would like to thank the many of our colleagues who have offered helpful advice on the Quaternary record of climate history. Jonathan Adams is supported by a Eugene P. Wigner fellowship from the Lockheed–Martin Corporation, at the Environmental Sciences Division of Oak Ridge National Laboratory. Mark Maslin would like to thank Catherine Pyke and Nick Mann for creating the figures for this article.

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