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ABRUPT AND SUDDEN CLIMATIC TRANSITIONS AND FLUCTUATIONS: A REVIEW

J.G. LOCKWOOD*
4 Woodthorne Croft, Leeds LS17 8XQ, UK

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ABSTRACT

This review paper summarizes recent research work on abrupt climatic changes and oscillations. The climatic system is viewed as a dissipative, highly non-linear system, under non-equilibrium conditions, and, as such, should be expected to have some unusual properties. These unusual properties include bifurcation points with marked instability just before the point, magnification of semi-periodic oscillations around bifurcation points, and variations in the strength of teleconnections with distance from equilibrium. These properties are discussed and illustrated for the climatic system using both the historical, Holocene and glacial climatic records. It is found that there are abrupt climatic changes and oscillations on all time-scales. The amplitudes and frequencies of climate variability and teleconnections are found to vary between different time periods. A number of persistent oscillations exist, particularly one about 1500 years, but their amplitudes vary considerably between time periods. The Holocene appears to be no more climatically benign than the similar period in the Eemian. The importance of the North Atlantic thermohaline circulation for generating abrupt climatic changes in Europe, particularly in association with sudden pulses of fresh water, is illustrated. The concept of antiphase temperature changes between the North and South Atlantic is discussed. Externally generated abrupt climatic deteriorations owing to explosive volcanic eruptions and variations in solar irradiance are also discussed. Copyright © 2001 Royal Meteorological Society.

KEY WORDS: abrupt climatic changes; Dansgaard–Oeschger events; Europe; global warming; Heinrich events; Holocene; Little Ice Age; thermohaline circulation

1. INTRODUCTION

There has been a growing realization recently that the climate system has repeatedly switched, often in a matter of years to decades, between significantly different climatic modes (Berger and Labeyrie, 1987). Indeed, the National Research Council (1998) in its review of long-term climate changes comments: 'the long-held, implicit assumption that we live in a relatively stable climate system is thus no longer tenable'. The inherent non-linear nature of the atmosphere is manifest in interactions and fluctuations on a wide range of space and time-scales (Keller and Houghton, 2000). Thus, even on the annual scale, abrupt circulation changes are observed as seasonal changes progress. Many climatic switches take place on time-scales that are relatively short, and this makes them of significant societal relevance.

The period of regular instrumental records of global climate is relatively short at a little over 100 years. Even so, this record shows many climatic fluctuations, some abrupt or sudden, as well as slow drifts in climate. When high quality and resolution, proxy records are examined, climatic changes become apparent on many time-scales. Some of these may be termed abrupt or sudden in that they represent relatively rapid changes in otherwise comparatively stable conditions, but they can also be found superimposed on other much slower climatic changes. The picture that emerges is one of an unstable climatic system showing abrupt changes on every climatological time-scale.

* Correspondence to: 4 Woodthorne Croft, Leeds LS17 8XQ, UK; e-mail: jglockwood@claramail.com

Even though the basic forcing of the atmosphere is the seasonally smoothly varying input from solar radiation at the top of the atmosphere, the large-scale flow variations of the atmosphere include abrupt changes between seasons. Thus, in the 1950s, Yeh and Tao (1958) discovered a discontinuity in the seasonal variation of the atmospheric circulation over Southern Asia, and developed a concept of a 'seasonal jump' of the Asian monsoon system. This abrupt change in the Asian monsoon system is characterized by a sudden shift northward of the subtropical westerly jetstream in early June. Recently, Tao and Chen (1987) have examined the onset times for abrupt changes in the Asian monsoon, and Matsumoto (1992) has documented nearly simultaneous changes in circulation parameters over large regions in the monsoon area. On a slightly longer time-scale, Diaz and Quayle (1980) and Diaz (1986) described changes in climate regimes associated with surface temperature and precipitation during the 1920s and 1960s in North America. Rogers (1985) described the marked changes in sea-level circulation patterns of the northern North Atlantic during the 1920s. At the century to millennial time-scales, Thompson (1991) and Thompson *et al.* (1995) have provided evidence from ice-cores in tropical mountain ice caps suggesting that both the onset and termination of the 'Little Ice Age' occurred in an abrupt manner in South America. Throughout the Alps, the last 3000 years were characterized by repeated glacial events at intervals of 200–400 years which, in some cases, were comparable in extent to the Little Ice Age (Wanner *et al.*, 2000). The advances were quite sudden and rapid, with each main advance involving a number of minor fluctuations. The similarity between the scale of the various Holocene advances resulted in the later ones, notably those of the Little Ice Age, obscuring the earlier events which were, therefore, ignored until recently (Groves, 1988). On an even longer time-scale, sudden, short-lived oscillations (Dansgaard–Oeschger events (D/OE)) occurred many times during the last ice age, while short-lived cooling events (Heinrich events (HE)) appeared to occur every 7–13000 years (Adams *et al.*, 1999).

The climate system is a composite system consisting of five major interactive adjoint components: the atmosphere, the hydrosphere, including the oceans, the cryosphere, the lithosphere, and the biosphere. All the subsystems are open and non-isolated, as the atmosphere, hydrosphere, cryosphere and biosphere act as cascading systems linked by complex feedback processes. The climate system is subject to two main external forcings that condition its behaviour, solar radiation and the action of gravity. Solar radiation must be regarded as the primary forcing mechanism, as it provides almost all the energy that drives the climate system. The whole climate system can be regarded as continuously evolving, as solar radiation changes on diurnal, seasonal and long-term time-scales, with parts of the system leading and others lagging in time. Therefore, the subsystems of the climate system are not always in equilibrium with each other. Indeed, the climate system is a dissipative, highly non-linear system, with many sources of instabilities. Starting with a given initial state, the solutions of the equations that govern the dynamics of a non-linear system, such as the atmosphere, result in a set of long-term statistics. If all initial states ultimately lead to the same set of statistical properties, the system is ergodic or transitive. If, instead, there are two or more different sets of statistical properties, where some initial states lead to one set, while the other initial states lead to another, the system is called intransitive. If there are different sets of statistics that a system may assume in its evolution from different initial states through a long, but finite, period of time, the system is called almost intransitive (Lorenz, 1963, 1976, 1990). In the transitive case, the equilibrium climate statistics are both stable and unique. Long-term climate statistics will give a good description of the climate. In the almost intransitive case, the system in the course of its evolution will show finite periods during which distinctly different climatic regimes prevail. For this case, it may not be possible to produce a *single* set of long-term average values which are of significant use in describing the climate. Further, it may not be possible to even produce *multiple* sets of climatological statistics that, with any certainty, can be said to describe all aspects of the climate. The almost intransitive case arises because of internal feedbacks, or instabilities involving the different components of the climatic system. Because of this, Overpeck and Webb (2000) comment, while discussing recent climatic research, a new paradigm of climate variability is emerging. The climatic record shows rapid step-like shifts in climate variability that occur over decades or less, as well as climatic extremes (e.g. drought) that persist for decades.

The variability of climate can be expressed in terms of two basic modes: the forced variations which are the response of the climate system to changes in the external forcing, and the free variations owing to internal instabilities and feedbacks leading to non-linear interactions among the various components of the climate system (Peixoto and Oort, 1992). The external causes operate mostly by causing variations in the amount of solar radiation received by or absorbed by the Earth, and comprise variations in both astronomical (e.g. orbital parameters) and terrestrial forcings (e.g. atmospheric composition, aerosol loading). For example, the diurnal and seasonal variations in climate are related to external astronomical forcings operating via solar radiation, while ice-ages are related to changes in Earth orbital parameters. Volcanic eruptions are one example of a terrestrial forcing which may introduce abrupt climate modifications over a period of 2 or 3 years, while enhanced atmospheric CO_2 content is another. The internal free variations within the climate system are associated with both positive and negative feedback interactions between the atmosphere, oceans, cryosphere and biosphere. These feedbacks lead to instabilities or oscillations of the system on all time-scales, and can either operate independently or reinforce external forcings.

Stocker (1999) has discussed the classification of rapid climatic changes. Abrupt climatic transitions and fluctuations can be classified into the following types.

(a) Short term abrupt fluctuations and transitions about a mean or slowly varying state (both over the last 1000 years and over the Holocene). This can include El Niño–Southern Oscillation (ENSO), the North Atlantic Oscillation (NAO), variations in the Atlantic Thermocline circulation, etc.
(b) Sequences of large-scale abrupt changes, such as those observed at the end of the last ice age. This includes D/OE and HE, but the only real difference between these and the abrupt changes under (a) is that their amplitude is larger.
(c) Short-term rapid fluctuations due to explosive volcanic eruptions.
(d) Possible human-induced changes in the twentieth century. This includes greenhouse gas enhancement, which appears to be driving abrupt climatic changes at the end of the twentieth century.

The structure of the review is that the properties of highly non-linear systems are considered next, together with their relevance to climatology and also ocean circulation. Examples of abrupt climate changes are then discussed for both the Holocene and last ice age. As most climatological data are available for the twentieth century, this is considered first in some detail, followed by discussions of changes over longer time periods.

2. HIGHLY NON-LINEAR SYSTEMS FAR FROM EQUILIBRIUM

Glansdorff and Prigogine (1971), Nicolis and Prigogine (1977) and Prigogine and Stengers (1985) list a number of types of behaviour found in dissipative, highly non-linear systems, under non-equilibrium conditions. A basic theorem in probability theory, the law of large numbers, provides an estimate of the 'error' owing to fluctuations. As soon as the system becomes large enough, the law of large numbers enables a clear distinction to be made between mean values and fluctuations, and the latter may be neglected. However, in non-equilibrium processes, fluctuations can determine the global outcome, because instead of being corrections in the average values, fluctuations now modify those averages. In particular, as a system evolves, it may approach a bifurcation point where the fluctuations become abnormally high, as the system may 'choose' among various regimes. Fluctuations can even reach the same order of magnitude as the mean values, so that the distinction between fluctuations and mean values breaks down.

Under non-equilibrium conditions, local events have repercussions throughout the whole system, with long-range correlations appearing at the precise point of transition from equilibrium to non-equilibrium conditions (e.g. Swinney and Gollub, 1978). The amplitudes of these long-range correlations are at first small, but increase with distance from equilibrium, and may become infinite at the bifurcation points. Long-range correlations (teleconnections) are indeed observed in the atmosphere, and the strength of these teleconnections is also observed to vary with time. Krishna Kumar *et al.* (1995) and Pant and Rupa

Kumar (1997) comment that there are significant variations over time in the thirty-one year sliding correlation coefficients between all-India summer monsoon rainfall (AISMR) and seven of its predictors (Figure 1), suggesting that the predictability of the monsoon itself may have secular variations. Most of the predictors show insignificant correlation coefficients with AISMR until about 1940, but all of them show significant correlation coefficients during the period 1942–1990. Northern Hemisphere January and February surface air temperatures even changed sign during the early part of the twentieth century. The sparse long records of Sahel rainfall suggest that conditions of the last couple of decades may be unprecedented in the context of the last several centuries, and also, that shifts from one variability state (e.g. 'wet') to another (e.g. 'dry') may occur in a couple of years (Overpeck and Webb, 2000). Hurrell (1995) and Hurrell and van Loon (1997) showed that the NAO exerts a strong influence on European winter climate. The NAO shows considerable variability at quasi-biennial and quasi-decadal time-scales, and the latter became especially pronounced during the second half of the twentieth century. Luterbacher *et al.* (1999) have found that the co-variability between the NAO and Eurasian circulation indices exhibit large decadal to century timescale variations, indicating that climate variability over the continent can be temporarily decoupled from the NAO. This was particularly so during the middle of the nineteenth century, when the circulation over the eastern North Atlantic at sea-level was temporally decoupled from the continental air flow. That the climate system variability is capable of undergoing significant change over time is also shown by records from Pacific corals spanning the past 1–4 centuries. They show synchronous shifts in the frequency domain among annual, interannual and multidecadal modes of Pacific variability, implying coupling across timescales (Cole *et al.*, 1993; Dunbar *et al.*, 1994). Allan (2000) reports that the ENSO phenomenon displays variations in its periodicity, with a dominant period of 3–4 years from 1870 to 1910, a 5–7-year periodicity from 1910 to 1920, and again from 1930 to 1960, and a period of 4–5 years from 1970 to 1990. Perhaps the most remarkable decadal variation identified in ENSO–streamflow relations is the decades-long changes in teleconnections when recent decades are compared with the period from about the 1920s into the 1950s. During the earlier epoch, large changes in the teleconnections of ENSO to tropical and extratropical streamflows are found, often amounting to near disappearances of those connections (Dettinger *et al.*, 2000). Similarly, data suggests that the modern relationship between El Niño and droughts in the southeastern US was absent before the 1920s (Overpeck and Webb, 2000). Lawrence (1965) reports that surface air temperatures from Edinburgh, Wakefield and Greenwich (UK) appear to be out of phase with Wolf Sunspot Numbers from about 1880 to 1930, but in phase from 1840 to 1880 and 1930 to 1960. Similarly, Hoyt and Schatten (1997) report that correlations

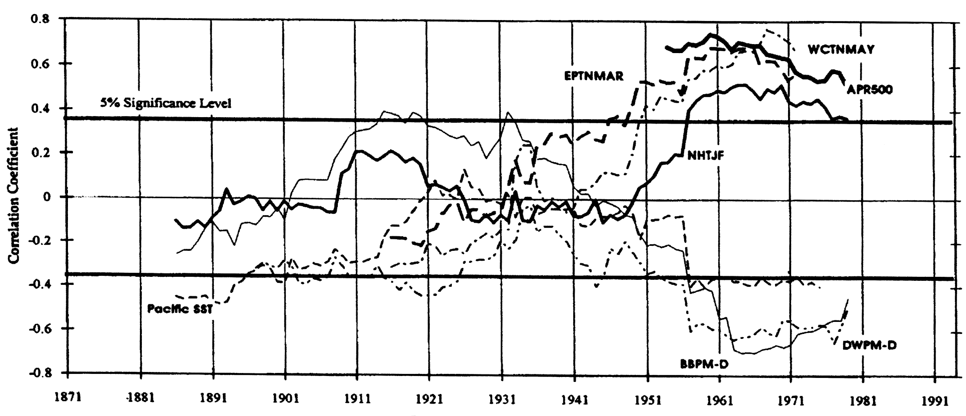

Figure 1. Thirty-one year sliding correlation coefficients between AISMR and seven of its predictors, showing secular variations in the relationships (from Krishna Kumar *et al.*, 1995, reproduced by permission of the Royal Meteorological Society). BBPM-D: Sea-level pressure at Bombay (March, April, May (MAM)–December, January, February (DJF)). DWPM-D: Sea-level pressure at Darwin (MAM-DJF). NHTJF: Northern Hemisphere surface air temperature (January + February). WCTNMAY: West-central India minimum surface air temperature, May. EPTNMAR: East peninsular India minimum surface air temperature, March. APR500: 500 hPa ridge location, April

between sunspot numbers, and many climatological time series show a reversal or failure around 1920, a time of an abrupt climate change. Palaeoclimatic data from around the tropical Pacific basin suggests that the climate of the mid-Holocene tropical Pacific was distinct from that of the last several centuries. Data from both the western and eastern Pacific suggest that interannual ENSO variability, as we now know it, was substantially reduced, or perhaps even absent (Overpeck and Webb, 2000).

The 'Feigenbaum sequence' is a remarkably simple road to 'chaos' that has attracted a lot of attention (e.g. Feigenbaum, 1980, 1983). It concerns any system whose behaviour is characterized by a very general feature, which is that for a determined range of parameter values, the system's behaviour is periodic, with a period T; beyond this range, the period becomes 2T, and beyond yet another critical threshold, the system needs 4T in order to repeat itself. The system is thus characterized by a succession of bifurcations, with successive period doubling. This route is characterized by universal numerical features independent of the mechanism involved, and the details of the governing equations, as long as the system possesses the qualitative property of period doubling. Other, more complex, systems exist that show more complex forms of behaviour, probably more typical of the atmosphere, as they move further away from equilibrium. For example, the Belousov–Zhabotinsky reaction in inorganic chemistry that consists of the oxidation of an organic acid by a potassium bromate in the presence of a suitable catalyst (Prigogine and Stengers, 1985). Various experimental conditions may be set up giving different forms of autoorganization within the same system (e.g. Winfree, 1974). In this case, the fluctuations in the system show both deterministic and stochastic elements, as the system moves further from equilibrium. Thus, the temporal oscillations of the ion Br^- as distance from equilibrium increases show first a homogeneous steady state, then sinusoidal oscillations, then complex periodic states through to mixed-mode oscillations with both chaotic and periodic characteristics (Figure 2). Because the movement from one stable state to another, as the distance from equilibrium increases, depends on universal numerical features rather than the actual mechanisms involved, it is not surprising that some of the curves look similar to climatological time series. For example, time series in Figures 4 and 7 are similar to the curves in Figure 2 showing chaos and mixed mode oscillations.

Gauthier (1999) notes that the Quaternary climate record tends to a unified geometric structure of superimposed aperiodic oscillations starting at the 11-year sunspot cycle, and spaced by powers of 2 in period through the major 90000 year glacial cycle. Climate cycles that do not fall in this structure typically

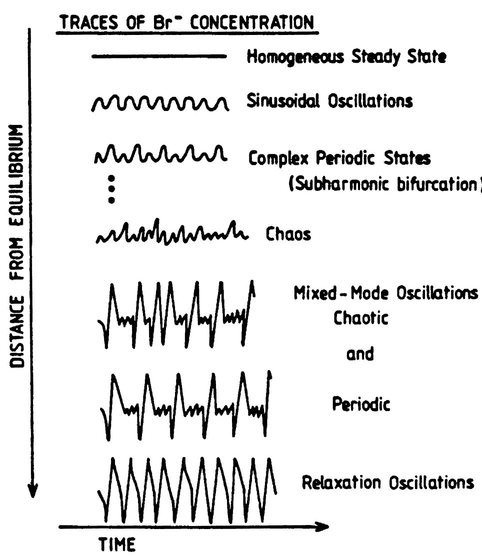

Figure 2. Temporal oscillations of the ion Br^- in the Belousov–Zhabotinski reaction. The figure represents a succession of regions corresponding to qualitative differences. This is a schematic representation. The experimental data indicate the existence of much more complicated sequences. (Figure 16 from *Order Out of Chaos*, by Prigogine and Stengers, 1985, reproduced by permission of the author)

correspond to harmonics of the structure oscillations. The average oscillation periods in years of the structure based on an estimate of 88.4 years for the Gleissberg solar cycle (quasi-periodic variation in sunspot numbers and other solar indices) are: 11.05, 22.1, 44.2, 88.4, 177, 354, 707, 1414, 2830, 5660, 11300, 22600, 45300 and 90500. The reported climate-cycle periods do not exactly match the average structure periods because the oscillations are aperiodic, and only have a tendency to cycle at the average periods. In the following discussion, climate cycles at around 22, 88, 177 and 1414 years appear frequently. For example, Schlesinger and Ramankutty (1994) report a temperature oscillation over the North Atlantic Ocean and its adjacent land areas with a period of 65–70 years. Allan (2000) reports that spectral analyses of global historical sea-surface temperature and mean sea-level pressure anomalies reveal significant climatic signals at about 2–2.5, 2.5–7, 11–13, 15–20, 20–30 and 60–80 years.

It is not the purpose of this paper to discuss the nature of non-linear chaotic states in the atmosphere, but the point needs to be stressed that the climate system is a non-linear system, often far from equilibrium and, therefore, should be expected to show many complex patterns with sudden jumps from one distribution pattern to the next.

2.1. Abrupt changes and the North Atlantic thermohaline circulation

Thermohaline circulations appear to be of major importance in explaining many abrupt climate oscillations. They also show many features of non-linear systems, so they need to be explored in some detail. When cooled, water with the salinity normal in the world's oceans becomes denser, but does not reach its maximum density until near its freezing point, at about -2°C. By contrast, fresh water is densest at about $+4$°C, so that when the surface is cooled below that temperature, the coldest water stays on top, and at 0°C, floating ice is formed. The salt water of the deep oceans, when cooled at the surface, goes into patterns of convection, the coldest and densest portions gradually sinking to the depths. Low density surface layers in the oceans can arise either because of surface heating, or the addition of relatively fresh continental runoff or precipitation onto the ocean surface. Surface heating in the tropics can be very effective in creating low density surface layers which continuously seek to spread over the entire surface of the ocean. This creates a 'thermohaline' circulation, whereby the warm, low density, surface water of tropical and sub-tropical areas spreads out towards higher latitudes, and after cooling by evaporation and sensible heat loss, leaves the surface and moves equator-ward at depth. This thermohaline circulation is important because it provides the major oceanic share of the transfer of warmth towards the poles.

In the cold oceans, sea-ice will only form when a layer of the ocean close to the surface has a relatively low salinity. The existence of this layer allows the temperature of the surface water to fall to freezing point, and ice to form, despite the lower levels of the ocean having a higher temperature. Fresh water runoff from rivers draining into the Arctic Ocean forms a thin layer of low-salinity water that covers the Arctic Ocean. The water balance of the Arctic Ocean is very approximately: net water vapour transport (precipitation–evaporation) at around 2×10^{15} kg year^{-1}, river runoff into the ocean at 4×10^{15} kg year^{-1}, outward sea-ice transport (mostly through the Denmark Strait) at -3×10^{15} kg year^{-1}, giving a net outflow of ocean water across 70°N of -3×10^{15} kg year^{-1} (Peixoto and Oort, 1992). It is in this relatively freshwater layer that the perennial Arctic sea-ice is formed. By contrast, in the Nordic Seas, the water is almost homogeneous right up to the surface, and this prevents the formation of sea-ice. Indeed, the North Atlantic Basin, which feeds surface water into the Nordic Seas, loses slightly more freshwater via evaporation (-1.21 m year^{-1}) than it gains from precipitation (0.87 m year^{-1}) and river runoff (0.21 m year^{-1}) (Bradley, 1999). In contrast, in the North Pacific, the net freshwater input is so large that it is mostly likely beyond the limit up to where deep water formation could be sustained (Rahmstorf, 1999).

It is possible to separate conceptually the oceanic wind-driven from the oceanic thermohaline circulation. While the wind-driven circulation is forced by wind stress variations, the thermohaline circulation is forced by density anomalies owing to air–sea fluxes of heat and freshwater inputs, and extends from the surface to the abyssal ocean. In reality, both circulations are coupled to each other, especially in the North Atlantic. The thermohaline circulation plays a crucial role in the climate system because of its large meridional heat transport, which roughly equals that of the atmosphere. For example, oceanic meridional transport of heat northward across 24°N (the aerial midpoint between the equator and pole) is estimated directly from oceanic measurements to be about 2 PW (Bryden *et al.*, 1991). This value represents about half of the net meridional heat transport required by the Earth's radiation budget at this latitude. The mean currents are responsible for most of the heat transport in the ocean, with the transient eddies contributing comparatively little. In the Northern Hemisphere 1.2 PW occurs in the North Atlantic and 0.8 PW in the North Pacific. The North Atlantic is of major importance because it is the dominant Northern Hemisphere site for the conversion of warm water to cold water in the thermohaline circulation. For example, at 24°N, about 17 Sv (1 Sv $= 10^6$ m^3 s^{-1}) of North Atlantic deep water (NADW) flows southward across 24°N at a temperature of $\sim$2.5°C, and is replaced by a northward flow of near-surface waters, which are some 16°C warmer; this implies a heat transport of 1.1 PW. In contrast, the wind-driven Florida current (23 Sv) and associated Ekman transport (6 Sv) recirculates towards the south at a barely lower temperature, giving a heat transport of only 0.1 PW (Roemmich and Wunsch, 1984). The wind-driven gyres are, furthermore, of limited latitudinal extent, in contrast to the deep ocean currents which provide an almost continuous flow from the Arctic to the Southern Ocean. Heat is transported northward throughout the entire Atlantic, including the Southern Hemisphere. About half of the heat delivered to the North Atlantic passes through and under the tropical Atlantic, rather than originating there (Pierrehumbert, 2000). Rahmstorf (1999) comments that the climatic effect of this thermohaline flow pattern is often illustrated by comparing the surface temperatures of the northern Atlantic with comparable latitudes of the less saline Pacific, as the former are 4–5°C warmer.

It was first suggested by Bjerknes (1964) that variations of the thermohaline circulation could lead to multi-year sea-surface temperature anomalies in the subpolar North Atlantic. Further, there are only a few localized areas of deep oceanic convection, generated by intense buoyancy loss owing to heat loss, but they have a controlling influence on the thermohaline circulation. These sites of intense deep convection are found in the North Atlantic located in the Greenland and Labrador Seas, and in the Southern Ocean near the Antarctic continent. It has been suggested that if all deep convection is interrupted, through, for example, a surface layer of freshwater, the thermohaline circulation breaks down in a matter of years, as suggested by early work by Stommel (1961) using a simple ocean model. Recent geochemical data (Lehman and Keigwin, 1992a,b; Veum *et al.*, 1992) have challenged the view that rapid climate fluctuations in the North Atlantic at the end of the last glacial were caused by the thermohaline circulation of the ocean being switched 'on' or 'off'. More recent data do not fit this interpretation, instead they show that even during cold periods, the thermohaline circulation did not stop; NADW flow continued at similar intensity to the present. However, the sites of deep convection appear to have shifted, and NADW flow was probably shallower than during warm periods. For example, Rahmstorf (1994), using a coupled ocean–atmosphere model, found that a brief freshwater pulse caused a drop in sea-surface temperature in the North Atlantic by up to 5°C within less than 10 years, but only a rearrangement of convection.

Stocker (1998, 1999, 2000), summarizing numerous modelling studies, describes how changes in the Atlantic Ocean, caused by sudden changes of the Atlantic's thermohaline circulation, results in an antiphase temperature change between the North and South Atlantic. A sudden increase of the northward meridional heat flux draws more heat from the south, and leads to a warming in the north that is synchronous with a cooling in the south. Similarly, a sudden reduction in the thermohaline circulation causes a cooling in the North Atlantic and a warming in the South. Stocker (1998) estimates that if the heat export stopped, with all else unchanged, the Southern Ocean would heat up by about 1.6°C per century. Broecker (2000) considers that this antiphasing is clearly seen in the deglaciation interval between 20 and 10000 years BP. During the first half of this period, Antarctica steadily warmed, but little change occurred in Greenland. Then, at the time when Greenland's climate underwent an abrupt warming, the warming in Antarctica stopped. Finally, at the onset of the Younger Dryas cold event in Greenland, the Antarctic warming resumed.

Stocker (2000) and Stocker and Marchal (2000) have used the concept of hysteresis to illustrate the possible responses of the ocean–atmosphere system to perturbations in the fresh water balance. Freshwater input in the regions of deep oceanic convection reduces the strength of the thermohaline circulation, but the system is stable, and reacts in a linear fashion, as long as threshold values are not crossed. An abrupt circulation change with an amplitude that no longer scales with the freshwater perturbation occurs if threshold values are crossed, but if the initial state of the ocean–atmosphere system is a unique equilibrium, the system jumps back to the original state once the freshwater perturbation has ceased. However, if other multiple equilibrium exist, the perturbation can cause an irreversible change.

The summary conclusions of many ocean–atmosphere numerical models (Stocker, 2000) are that:

(i) the Atlantic deep ocean circulation has multiple equilibrium;
(ii) transitions between them can occur on a decadal time-scale;
(iii) the stability of a particular thermohaline circulation is strongly influenced by the surface freshwater fluxes.

Numerical models of the North Atlantic thermohaline circulation show oscillations in the circulation that could be of great importance to European climate. Delworth *et al*. (1993) performed a long-time integration of a fully coupled atmosphere–ocean general circulation model (GCM) under present day greenhouse radiative forcing, and noted the existence of irregular oscillations in the strength of the thermohaline circulation with associated changes of sea-surface pattern in the North Atlantic. The period of the oscillation was about 50 years and appeared to be driven by density anomalies in the northern sinking region of the thermohaline circulation combined with much smaller density anomalies of opposite sign in the broad rising region to the south. The model generated sea-surface anomalies also induced model surface air temperature anomalies over the northern North Atlantic, the Arctic and northwest Europe. Using a three-dimensional ocean circulation model, Greatbatch and Zhang (1995) showed that under a constant zonally-uniform surface heat flux and a zero salt flux, the strength of the thermohaline circulation had about a 50-year oscillation which was very similar to that found by Delworth *et al*. (1993). The model sea-surface anomaly pattern was also like the observed pattern seen for the period 1950–1984. In this case, the oscillation is an internal one owing to a balance between convergence in the oscillatory part of the poleward heat transport, and changes in the local heat storage, and a similar balance must apply to the coupled model used by Delworth *et al*. Greatbatch and Zhang suggest that an oscillation of this kind may have played a role in the observed warming of the North Atlantic surface waters during the 1920s and 1930s and the subsequent cooling in the 1960s. Manabe and Stouffer (1995, 1997) used a coupled ocean–atmosphere model to examine the response of the thermohaline circulation and global climate to the effect of a sustained freshwater input across the North Atlantic similar to that which may have occurred at the end of the last ice age. These numerical experiments produced abrupt changes and large amplitude oscillations in both the thermohaline circulation and North Atlantic climate. In particular, large climate anomalies were predicted over the North Atlantic and across western Europe.

3. ABRUPT GLOBAL CLIMATIC CHANGES OF THE LAST 1000 YEARS

During the Holocene, there have been a number of rapid, widespread changes recorded in the paloeoclimatic records around the world, which are discussed in more detail in the next section. In the North Atlantic region, these changes have been spaced according to approximately the same 1500-year rhythm (1470 ± 500 years) as that found for the last glacial and earlier glacial periods (Bond *et al*., 1997; Campbell *et al*., 1998). For reasons to be discussed later, this rhythm seems amplified during the last ice age. The most recent cold event in these 1500-year cycles may have been the Little Ice Age (Bond *et al*., 1997). 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 (Adams *et al*., 1999). A recent Northern Hemisphere temperature reconstruction indicates an oscillating temperature drop from AD 1000 to 1850 of about 0.2°C with a subsequent and continuing warming of nearly 0.8°C (Mann *et al*., 1999).

The Little Ice Age is of particular climatological interest because, although much smaller in amplitude, it shared the geographic pattern of the Younger Dryas and the main glacial maximum (Broecker, 2000).

There were a series of post-medieval cool events, varying in intensity from one region to another, but there seems to have been a more widespread climatic deterioration after about 1200 (Lowell, 2000), which is generally referred to as the start of the Little Ice Age. There is general agreement that the Little Ice Age came to an abrupt end in the mid-nineteenth century (~ 1850) (Bradley, 1992, 1999). Pfister's (1985, 1992) studies in Switzerland indicate that, overall, the coldest conditions of the last 500 years were in the late seventeenth and nineteenth centuries, especially the early nineteenth century, which can be considered as the 'climatic pessimism' of the last 1000 years. Bradley (1999) comments that much of the 'global warming' registered since then represents a recovery from that low point in the early to mid-nineteenth century.

Mann *et al*. (1998, 1999) used a multiproxy data network to reconstruct Northern Hemisphere temperatures back to 1000 AD. They found a long-term cooling trend in the Northern Hemisphere prior to industrialization of −0.02°C per century. They consider that this cooling is possibly related to astronomical forcing, which is thought to have driven long-term temperatures downward since the mid-Holocene at a rate within the range from −0.01 to −0.04°C per century (Berger, 1988). The record shows significant century-scale variability which may be associated with solar irradiance variations (Lean *et al*., 1995; Mann *et al*., 1998). A robust spectral peak also appears centred at the 50–70-year period. The Mann *et al*. (1999) temperature reconstruction shows that the late eleventh, twelfth and fourteenth century rival mean twentieth century temperature levels, while cooling following the fourteenth century could be viewed as the initial onset of the Little Ice Age. Considerable spatial variability is evident, as in Lamb's (1965) original concept of a Mediaeval Warm Epoch with cooler as well as warmer conditions. For the Northern Hemisphere, temperature series both for the year 1998 and the decade 1989–1998 are documented as the warmest in the twentieth century instrumental record. The decade 1989–1998 is nearly two (decadal) standard errors warmer than 1166–1175, which is the next warmest decade prior to the twentieth century. Also the twentieth century (1900–1998) is nominally the warmest of the millennium, which is especially striking if viewed as deifying a long-term cooling trend associated with astronomical forcing.

3.1. Climate changes in the twentieth century

Global mean surface temperature has increased since the late nineteenth century, but not in a uniform manner. Karl *et al*. (2000) report that the global increase in temperature since 1880 occurred during two sustained periods, one beginning around 1910 and another during the 1970s (Table I). They consider that the rate of warming since 1976 is greater than the mean rate of warming averaged over the late nineteenth and twentieth centuries. It is less certain whether the rate of temperature change has been constant since 1976, or whether the recent string of record breaking temperatures over 1997 and 1998 represent yet another increase in the rate of temperature change.

Fu *et al*. (1999) have described the abrupt warming during the 1920s and 1930s. In this case, the strongest warming occurred in North Atlantic Ocean sea-surface temperatures during winter, being distinct, but more gradual, in the other oceans and seasons. The Northern Hemisphere continental surface temperature record shows that both middle and high latitudes also experienced rapid warming in the 1920s. The main feature of the annual mean sea-surface change in the Atlantic Ocean are:

1. an abrupt warming in most latitudes in the period from the late 1920s to early 1930s;
2. comparing different latitudes, the warming signal achieves higher significance levels in the northern part of the Atlantic, and tends to be weaker in the South Atlantic;
3. the warming is detected earlier near the equator than at higher latitudes, while the latest warming appears at high latitudes of the South Atlantic.

Przybylak (2000) found that in the Arctic that the highest temperatures since the beginning of the instrumental observations occurred clearly in the 1930s, and can be attributed to changes in the atmospheric circulation. For Arctic temperatures, the most important factor is a change in atmospheric circulation over the North Atlantic.

Table I. Some abrupt late twentieth century climate changes

Global temperature changes (Karl *et al.*, 2000)	
1880–1912	−0.38°C/century
1912–1941	1.20°C/century
1941–1976	−0.27°C/century
1976–1998	1.96°C/century

Global sea-surface changes
Over the period 1984–1996, global sea-surface temperature signals from satellites show an increase at a rate of +0.005°C/year. Upward trends are notable for their statistical significance for latitudes between 5°S and 20°N. The peak of this increase during the past 13 years is found at 5°N at a rate of +0.5°C/decade. Evidence of warming is found in the mid-latitude Northern Hemisphere, while estimates from the Southern Hemisphere, though strongly indicative of compensatory cooling in the region, are found to be not as reliable (antiphase temperature change?). 1997 sea-surface temperature data for the South Atlantic continued to show a most dramatic decrease, especially from the equator to 35°S until the strongest El Niño of the modern record began in March 1997 (Strong *et al.*, 2000).

Arctic
Surface air temperature (1901–1977) shows a warming in most arctic regions, except for the Labrador Sea, where cooling is observed. The annual time series of the leading surface air temperature eigenvector mode indicates decadal variability embedded with a small warming trend since the beginning of the twentieth century. Since the 1970s, the arctic region has experienced three strong warming events: 1970s, 1980s and 1990s. The 1990s event was the strongest and longest (Wang and Ikeda, 2000).
Several oceanographic cruises indicate that the Arctic Ocean has undergone related significant changes since about 1990, including an increase in the temperature and a decrease in depth of the Atlantic water temperature maximum and a shift in the frontal circulation between eastern and western water types (McPhee *et al.*, 1998).
Comparison of sea-ice draft data acquired on submarine cruises between 1993 and 1997 with similar data acquired between 1958 and 1976 indicates that the mean ice draft at the end of the melt season has decreased by about 1.3 m in most of the deep water portion of the Arctic, from 3.1 m in 1958–1976 to 1.8 m in the 1990s (Rothrock and Maykut, 1999).
There is evidence for a recent change in the link between the NAO and Arctic sea-ice export through the Fram Strait during wintertime (December, January, February, March). No correlation between the two series is found from 1958, but the correlation increased significantly between 1978 and 1997 (Hilmer and Jung, 2000).

Tropics
El Niño occurrences revealed in the pre-1976 instrumental record have a frequency spectrum almost identical to that of palaeo-El Niños recorded in coral sequences from the previous major interglacial period (Eemian, about 124 000 years BP), whereas the post 1976 spectrum is distinctly different (Hughen *et al.*, 1999).
New coral data show an abrupt increase in ^{14}C in the upwelling season starting in 1976, precisely the year when the instrumental record begins to show a change in the character of El Niño. This increase has been interpreted as indicative of a shift in the equatorial thermocline structure (Pierrehumbert, 2000).

These changes in North Atlantic sea-surface temperatures are associated with changes in the winter (December–February) atmospheric circulation patterns. During the period preceding the time of rapid warming, the circulation was already characterized by an enhancement of the North Atlantic high and Icelandic low. During the abrupt warming, a further strengthening of the atmospheric circulation is evident, which would be expected to lead to stronger westerly and southwesterly winds in the North Atlantic region. In the warm period following the abrupt changes, the strength of the atmospheric circulation decreases. This weakening may be related to a reduction in the mean meridional thermal gradient associated with the greater amplitude of the warming at higher latitudes. Concomitant to the weakening of the westerlies and trade wind systems, the Asian monsoon troughs deepened substantially, a situation generally favourable to the development of active monsoons. Fu *et al.* (1999) suggest that the combination of these two features, enhanced continental monsoons and implied lowered vertical wind shear over the oceans, would tend to enhance the rainfall and the release of latent heat in the tropics.

This, in turn, would strengthen the Hadley and Walker vertical circulations, which may have been at least partly responsible for greater aridity in subtropical land areas of both hemispheres during this period.

The behaviour of the Asian monsoon shows marked changes during the periods of rapid warming in the 1920s (Fu *et al*., 1999). The number of 'Break Monsoon Days' (BMD) during July and August over India may be used as a summer monsoon index. An abrupt change of the Indian summer monsoon occurred in the early 1920s, when the index of annual number of BMD shows a shift from an inactive (high BMD) to an active monsoon period (low BMD) (Figure 3). This pattern persisted during the time from 1935 to 1943 when air temperature was rising. Since the mid-1960s, the monsoon again entered into an inactive period until 1970, during which time a cooling tendency in the Northern Hemisphere is evident. During the late 1970s, the monsoon became active again, associated with further global warming.

Figure 4 presents global average annual precipitation, expressed as departures from the mean, based upon 5899 stations, for the period from 1900 to 1993. While annual precipitation amongst stations can vary between 0 and 11 900 mm, the range averaged for the globe rarely exceeds $\pm$ 50 mm (910–1135 mm) (Bryant, 1997). Bryant comments that the global record displays many of the characteristics of a chaotic system, with rainfall variance changing continually. Global rainfall fluctuations from year to year were greatest before the abrupt warming in the 1920s, and smallest during the warming period of the 1930s and 1940s. They also become larger again just before the abrupt warming in the 1970s. Two abrupt global rainfall fluctuations of interest occurred in 1952 and 1980–1983. After 1952, rainfall rose globally by 60 mm in 3 years, associated with the large La Niña event of 1950–1951. Between 1980 and 1983, global rainfall fell by 50 mm, associated with the strong ENSO event of 1982–1983.

There is considerable evidence from climatological records for an abrupt global warming in the 1970s (Table I). Allan (2000) has summarized the contemporary research suggesting that the ENSO phenomenon has undergone a change in mode and nature since the mid-1970s, with a predominance of El Niño over La Niña phases. There is mounting evidence from high elevation ice caps and glaciers for

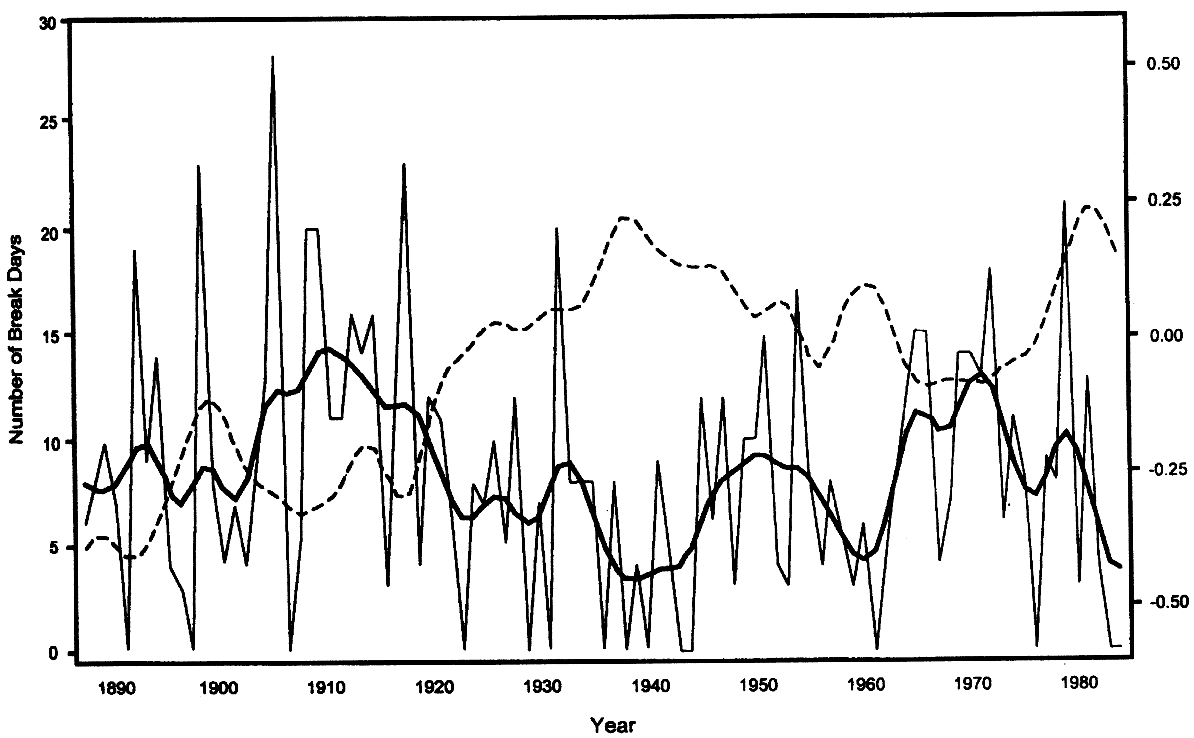

Figure 3. Annual number of monsoon break days during the months of July and August over India, 1888–1984 (after Parthasarathy and Pant, 1985) (thin solid line with heavy line 9-point smoothed). Dashed curve illustrates the 9-point smoothed Northern Hemisphere annual temperature (data from Jones *et al*., 1986). Figure from Changes in atmospheric circulation over Northern Hemisphere Oceans associated with the rapid warming of the 1920s, *International Journal of Climatology*, **19**: 581–606, C Fu, HF Diaz, D Dong and JO Fletcher, 1999, reproduced by permission of John Wiley & Sons, Ltd)

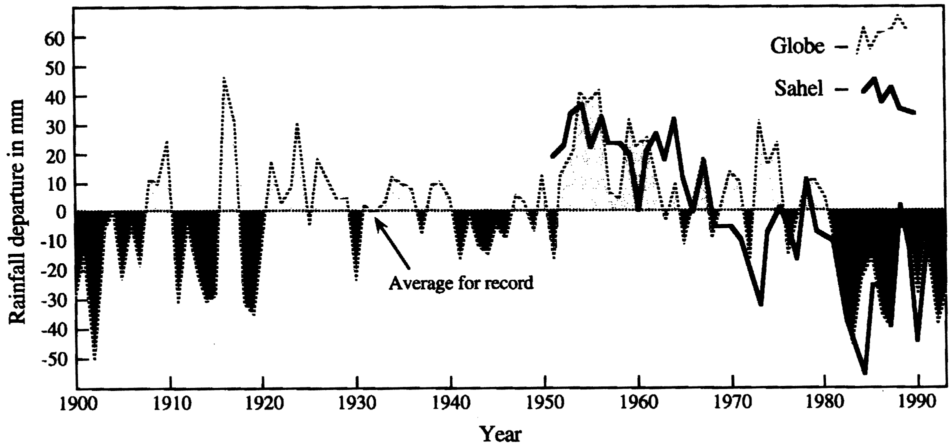

Figure 4. Annual deviations in global precipitation, 1900–1993 (data from NOAA, 1995). Values are referenced to the average for this timespan. Superimposed are scaled deviations for the Sahel region of Africa, 1950–1991 (data from Hulme, 1994). Figure from E Bryant, *Climate Process and Change*, 1997, reproduced by permission of Cambridge University Press)

recent, strong warming in the tropics (National Research Council, 1998). These ice masses are particularly sensitive to small changes in ambient temperatures, as they already exist very close to the melting point. For example, in 1993, two cores were drilled through an ice mass in the north-central Andes (9°S, 77°W, 6048 m) (Thompson *et al*., 1995). The data from these cores indicate that the nineteenth and twentieth centuries were the warmest in the 5000 years, and also reveal an accelerated rate of warming since 1970. These results are typical of high altitude ice masses in both tropical South America and Africa. Also, Satellite Advanced Very High Resolution Radiometer-based data, which began in the early 1970s, indicates more extensive snow cover over both Eurasia and North America in the 1970s to mid-1980s, relative to the later part of the time series (Robinson *et al*., 1993; Walland and Simmonds, 1997). The decline in snow cover occurs over a span of approximately five years, from 1985–1986 to 1989–1990. Walland and Simmonds (1997) found significant co-variability between Eurasia and North America, with the Eurasian signal lagging behind the North American signal by over a year.

The possibilty that El Niño episode durations or 'spell lengths' have increased with global warming has been highlighted by the study by Trenberth and Hoar (1996), which emphasized the unusually long nature of the 1990–1995 warm eposide. Mann *et al.* (2000) report that this spell does not appear especially unusual when viewed in the context of certain previous periods, such as the early nineteenth century, during which there were between three and five 5-year-long warm events. The early nineteenth century period associated with the greatest probability of warm spells is associated with the coldest global temperatures back to 1650, and an apparent breakdown of interannual ENSO variability. Mann *et al.* (2000) comment that the unusually protacted 1990–1995 event may, in turn, be signalling a breakdown of interannual ENSO variability at the 'warm global temperature' boundary of its dynamical regime.

The recent Arctic climate record is anomalous, in that, while there is some evidence of ocean warming and ice retreat, this is not , as yet, reflected in the Arctic air temperatures. Przybylak (2000) reports that the mean rate of warming for the period 1991–1995 was 2–3 times lower in the Arctic than the global average, and that the temperature levels observed in Greenland in the last 10–20 years are similar to those observed in the nineteenth century. A cooling of 0.9°C has occurred between 1970 and 1995 in the Labrador Sea water. This water mass is currently colder, fresher, and larger than ever before recorded, in observations extending back to the 1930s (Read and Gould, 1992; Lazier, 1995).

Direct measurements of decadal or longer term variability of the deep ocean circulation are not readily available, but there are some observations for the North Atlantic. Bacon (1998) has compiled hydrographic conditions off Greenland, and has concluded that there is good evidence for decadal variations in overflow from the Nordic Seas (Figure 5), which he links to the NAO index. Observations from Ocean Weather Station Bravo (56°30′N; 51°00′W) show that deep convection occurred almost every

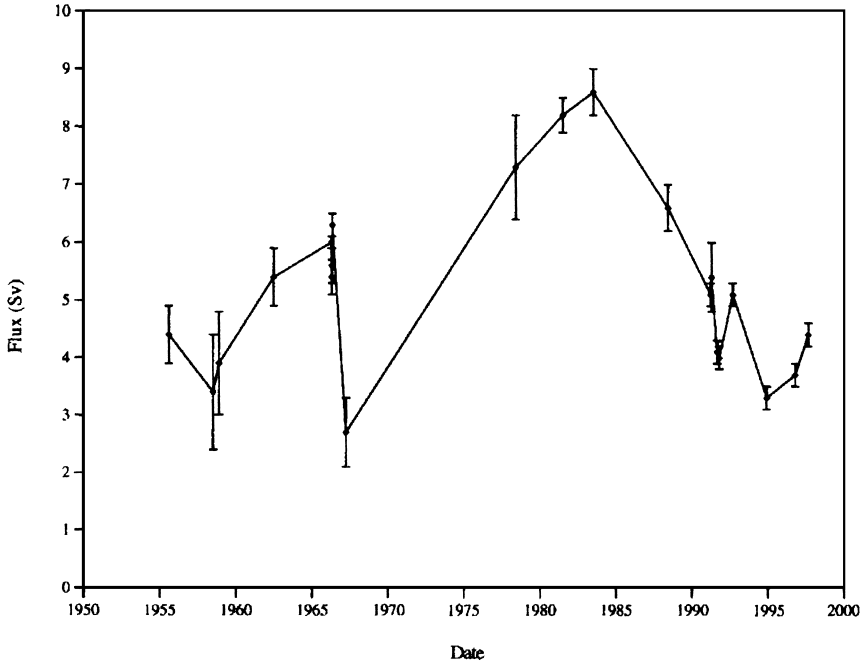

Figure 5. Transport of the deep western boundary current flowing southward off south-east Greenland, derived from various hydrographic sections since 1955. (Figure reproduced with permission from *Nature*, S Bacon, Decadal variability in the outflow from the Nordic Seas to the deep Atlantic Ocean, *Nature*, **394**: 871–874, © 1998 Macmillan Magazines Limited)

winter until 1967, but then stopped for several years until convective activity resumed in 1972. The interruption of convection was a consequence of the so-called Great Salinity Anomaly (GSA) (Dickson *et al.*, 1988). The GSA appeared as a significant surface freshwater anomaly in about 1969 in the Labrador Sea. It can be traced moving eastward across the subpolar gyre, into the Norwegian Sea and ending up near Fram Strait more than 10 years later (National Research Council, 1998). The GSA was probably caused by an increase in sea-ice export from the Arctic, and there is evidence that a similar event occurred in the early 1980s (Lazier, 1995). For the Greenland Sea, Schlosser *et al.* (1991) have concluded from tracer measurements that deep convection was reduced by 80% during the 1980s. Deep water changes at 24°N in the Atlantic may reflect reduced flow of Labrador Sea water, possibly as a consequence of the interruption of convection, and a freshening of deep waters formed in the Greenland–Iceland–Norwegian Sea (Bryden *et al.*, 1996). During the last glacial maximum, the rate of NADW export from the Atlantic was probably similar to today's (Yu *et al.*, 1996), but convection sites were shifted to the south, and NADW flow was shallower. Several times, during the glacial period, the NADW circulation appears to have collapsed rapidly (within about a decade), probably owing to an inflow of meltwater or a surge of icebergs (HE) into the North Atlantic (MacAyeal, 1993; Sarnthein *et al.*, 1994, 1995; Bjorck *et al.* 1996).

Understanding the linkages between heat and freshwater transport in the ocean is a fundamental requirement for solving many of the puzzles of climate variability (National Research Council, 1998). Mysak *et al.* (1990) have argued that the GSA, in part, may have been remotely generated by prior anomalously large freshwater runoffs into the western Arctic from North America during the mid-1960s. They also proposed that the GSA may be part of a sequence of atmosphere, hydrological and oceanic events in the Arctic and northern North Atlantic that form an interdecadal (approximately, 20 years) climate cycle that can be described in terms of a negative feedback loop. Stocker and Mysak (1992) have found in the spectrum of the 318-year central England temperatures record a significant peak at 24 years. They also found in the spectrum of a 370-year Koch ice index time series (number of weeks per year when ice affected the coast of Iceland) a significant peak at 27 years. The proposed mechanism is that fluctuations in the Mackenzie River discharge tend to produce sea-ice cover anomalies of similar sign in the Beaufort and Chukchi Seas region about 1 year later. Sea-ice anomalies formed to the north of Yukon and Alaska move into the Greenland Sea about 3 or 4 years later by the Transpolar drift stream. Lastly, there is a close relationship between positive sea-ice extents and freshwater anomalies in the Greenland Sea that can partly close down the Atlantic thermohaline circulation.

3.2. The Little Ice Age

Climatic changes during the Little Ice Age (roughly AD 1200–1850) caused a worldwide expansion of glaciers (Lowell, 2000). Three thousand 200 year long records of the movement of the Great Aletsch Glacier show that it varies very slowly with a response time of about 75 years. These results, combined with the 460 year long record of the movement of the rather small, rapidly reacting Lower Grindelwald Glacier show surprisingly regular, almost periodical fluctuations between glacier hostile (warm/dry) and friendly (cool/wet) periods. Phases with positive mass balances occurred around every 200–400 years. Wanner *et al*. (2000) recommend that these quasi-regular glacier advances be called Little Ice Age Type Events (LIATEs). Three striking LIATEs were observed during the last 750 years, at around 1300–1380, 1570–1640 and 1810–1850. Barber *et al*. (1999) have examined macrofossil data from peat cores through Fallahogy Bog, Northern Ireland and Moine Mhor, Cairngorm Mountains, Scotland. The two wettest periods at Fallahogy Bog are around AD 725–800 and the particularly wet period between AD 1685 and 1845, a short wet period was also identified around AD 160. The record from Moine Mhor shows similar changes (within the limits of the dating methods) to those at Fallahogy, with an especially wet period between AD 1680 and 1850. At both bogs, the start and end of the AD 1680–1850 wet period is particularly abrupt. Barber *et al*. also found an approximately 200-year cycle in sediment records from Lochan Uaine, Cairngorms.

Records are reasonably complete for the final LIATE phase, unlike the two earlier phases. Broecker (2000) comments that snowlines in both the north and the south temperate zones during the last LIATE (1810–1850) were about 100 m lower than they were in 1975. This difference is comparable with a snowline lowering of 900–950 m during full glacial time and to a roughly 350 m lowering during the Younger Dryas. Broecker (2000) considers that the last LIATE, although much smaller in amplitude, shared the geographic pattern of the Younger Dryas and the main glacial maximum. The question, therefore, arises as to whether the Little Ice Age was an event similar in nature to the Younger Dryas.

In the last sections, it was suggested that incursions of fresh water into the North Atlantic could cause marked changes in the thermohaline circulation and thus a fall in North Atlantic sea-surface temperatures. This fall in sea-surface temperatures could, in turn, force changes in western European climate. Records show that, from 1650 to 1890, Iceland was surrounded by sea-ice for an average of 2 months per year, but by 1920, the waters around Iceland were free of sea-ice for the entire year. Some data on North Atlantic sea-surface temperatures can be gained by studying records of cod fisheries. This is because the abundance of cod seems to be limited by the 2°C water temperature isotherm owing to kidney failure in colder water. Optimum conditions for cod are in waters at around 4–7°C. Using fishery evidence such as this, Lamb (1988) concluded that the ocean surface in the region of the Faeroe Islands, and between there and Iceland, was about 5°C colder in the 30-year mean for 1675–1704 than the twentieth century average. In contrast, around AD 1000, the North Atlantic climate was relatively mild, and sea-ice restricted in extent (Ogilvie *et al*., 2000). Broecker (2000) considers that CFC-11 inventories, and related tracer evidence, is consistent with a reduction in deep water formation during the Little Ice Age.

Hoyt and Schatten (1997) have reviewed the evidence for slight variations in solar output over the last 1000 years. They find evidence of two and possibly three, grand minima: (1) the Maunder Minimum from 1645 to 1715, (2) the Sporer Minimum around 1500, and (3) an unnamed minimum around 1350. A grand maximum in solar activity near AD 1200 is also evident. Wanner *et al*. (1995) have used multi-proxy data to investigate the climate of the North Atlantic and Western Europe during the late Maunder Minimum period. They found that the Late Maunder Minimum was a relatively cool and dry period, with low ocean temperatures and a large sea-ice extent, although Alpine glaciers did not grow during this time because of the relatively low snowfall. A comparison of the winter weather types of the period 1675–1704 with the period 1961–1990 shows that the late Maunder Minimum was characterized by strong sea-level pressure reversals, with high pressure centres over northern or northwestern Europe, and large outbreaks of northeasterly cold continental air. Andronova and Schlesinger (2000) used a simple climate/ocean model, with radiative-forcing of changes in solar radiation, to simulate the observed temperature changes over the last 142 years. They found a minimum in solar forcing around 1810, with a small irregular increase up to the present day. Their results indicate that the observed warming over 1856–1990 was predominantly owing to anthropogenic greenhouse gas radiative forcing, plus an unexplained residue warming, with the increased solar output contributing a warming less than half that contributed by the anthropogenic factor (if the solar irradiance varied as construed), and with volcanoes contributing a small cooling.

4. HOLOCENE

Western Europe is particularly sensitive to changes in the North Atlantic sea-surface temperature arising from changes in the thermohaline circulation (Manabe and Stouffer, 1995, 1997). In particular, the glaciers of the Alpine region are very sensitive to climate, and are, therefore, good indicators of climatic changes (Lowell, 2000). Patzelt (1974) has produced the list below of Holocene cold phases in the Eastern Alps; these are of interest because the Alps are downwind of the North Atlantic, which is a source of European climatic variations (Hurrell, 1995). They may also be compared with ice-rafted debris concentrations in the North Atlantic, as these observations provide another indicator of glacial activity at tidewater level (Lowell, 2000). The coarse size of lithic fragments requires that they be transported to the open ocean via icebergs, and thus, their abundance reflects glacial activity. Lowell (2000) comments that Alpine glaciers peak as ice rafted debris increases, whereas the ice sheets peak during or at the end of the major ice rafting events. Studies of sediment grain size by Bianchi and McCave (1999) indicate systematic thermohaline circulation changes in the Holocene. These authors argued that, for every millennial-scale cold-episode since about 8000 years ago, there was a decrease in the flux of Iceland–Scotland overflow water, one of the components of NADW.

Holocene cold phases in Europe are as follows.

1. The Schlaten, comparable in scale with the Little Ice Age, occurred probably about 9400 years BP. North Atlantic ice rafted debris reported at about 9500 years BP.
2. The Venediger, occurring between about 8700 and 8000 years BP, deposited lateral and terminal moraines outside the Little Ice Age moraines. It was characterized by two main advances, and may have been even more complex, for pollen profiles suggest repeated strong depressions of the treeline about this time. Ice rafted debris reported about 8200 years BP.
3. The Frosnitz or Larstig advances, between 6600 and 6000 years BP, are marked by non-arboreal pollen increasing by up to 50%. The moraines are outside those of the Little Ice Age, but are not large, suggesting that the advance was not long sustained. Ice rafted debris reported about 5900 years BP.
4. The Rotmoos in the Otzal, dated by Patzelt (1974) as between 5300 and 4300 years BP, is just a minor climatic deterioration. Ice rafted debris reported about 4500 years BP.
5. The Lobben, occurring between about 3500 and 3100 years BP, was associated with glaciers in the Vendigergruppe and Stubaier Alpen, advancing 100–150 m outside the Little Ice Age moraines.
6. Goschener I, comparable in extent with glacial advances of recent centuries, covers the period 2900–2300 years BP. Ice rafted debris at about 3000 years BP.
7. Goschener II was a period of expansion from roughly 1700 to 1300 years BP, that is from the second to the sixth centuries AD. Ice rafted debris at about 1500 years BP.

Patzelt (1974) emphasized that this sequence and the time limits of individual cold phases are likely to be subject to modification as a result of further work.

Adams *et al*. (1999) consider that the event at 8200 years BP 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. They also comment that smaller, but also sudden and widespread, changes to drier or moister conditions have been noted for many parts of the world since about 5000 years ago. Overpeck and Webb (2000) comment that the climate of the mid-Holocene (approximately 7000–5000 years BP) was quite distinct from that of the present day. Data from mid-Holocene corals reveal that western Pacific surface waters were substantially warmer and saltier than present. When compared with a variety of palaeoclimatic data from around the tropical Pacific ocean, it becomes apparent that the climate of the mid-Holocene tropical Pacific was distinct from that of the last several centuries (Markgraf and Diaz, 2000). Data from both the western and eastern Pacific suggest that interannual ENSO variability was substantially reduced, or perhaps even absent. Rodbell *et al*. (1999), using laminae deposited in an alpine lake in Ecuador argue that the modern El Niño period was established about 5000 years ago, with a major change from variability on decadal and longer timescales, to variability with a characteristic timescale of 2–8.5 years. Overpeck and Webb (2000) comment that this suggests that significant shifts in variability could take place in the future, even with modest greenhouse warming. Similarly, Holmes *et al*. (1999), in their study of the Manga Grasslands of northeast Nigeria report that wet conditions prevailed during much of the early to mid-Holocene, but a marked deterioration in climate commenced around 4100 years BP, leading to the formation of the present-day semi-arid landscape.

4.1. Was the Holocene climate uniquely benign?

It is sometimes proposed that the Holocene climate was uniquely stable and benign in contrast to the supposedly highly variable Eemian (last interglacial) climates. Kukla (1997) considers that this conclusion is questionable. He presents evidence that, with very high probability, the Eemian lasted considerably longer than the Holocene, and that the climates of the elapsed part of the current interglacial in many parts of the world were far from uniform, and certainly no more benign, than in the corresponding segment of the last interglacial. He quotes Watkins (1971), who in his paper on geomagnetic polarity events comments that 'the evidence for short events will be forthcoming, whether or not they have actually occurred'. Watkins contents that it is more difficult to prove that a given behaviour has not taken place than to show that it did. Therefore, the publication of data confirming an original mistaken finding is considerably more likely (the reinforcement syndrome) than publication of the data which would disprove it. Thus, Kukla (1997) asks the question 'was the Eemian climate so different from the Holocene and so much more unstable? Or are we witnessing an attack of the reinforcement syndrome on the Eemian period? He claims that the record of the so-called Eemian interval in both the GRIP and GISP ice-cores from Greenland are probably mechanically disturbed by ice tectonics and therefore do not properly represent the Eemian climate. A duration of 20000 years has been proposed for the last interglacial in La Grande Pile (Guiot *et al*., 1989) based on comparison of the pollen spectra with the pollen record of the deep sea core SU 8132 off Portugal (Turon, 1984). Kukla (1997) comments that there are additional observations which indicate that the Eemian in France lasted about twice as long as the elapsed part of the Holocene. The fluctuations of the tree pollen ratio in La Grande Pile (Woillard, 1978) appears to mimic closely the variation of the cold water foraminifer *Neogloboquadrina pachyderma sinistra* in deep sea core V29-191 situated in eastern North Atlantic at 54°16′N and 16°47′W (Figure 6). The equivalent position of the elapsed part of the Holocene is also shown on Figure 6; it is seen that it corresponds to a relatively benign period at the start of the Eemian.

5. ABRUPT TRANSITIONS OVER THE LAST 115000 YEARS

Evidence from Greenland ice-cores and Atlantic Ocean sediments suggest that the patterns and rates of deep oceanic circulation have undergone large changes over the past several hundred thousand years.

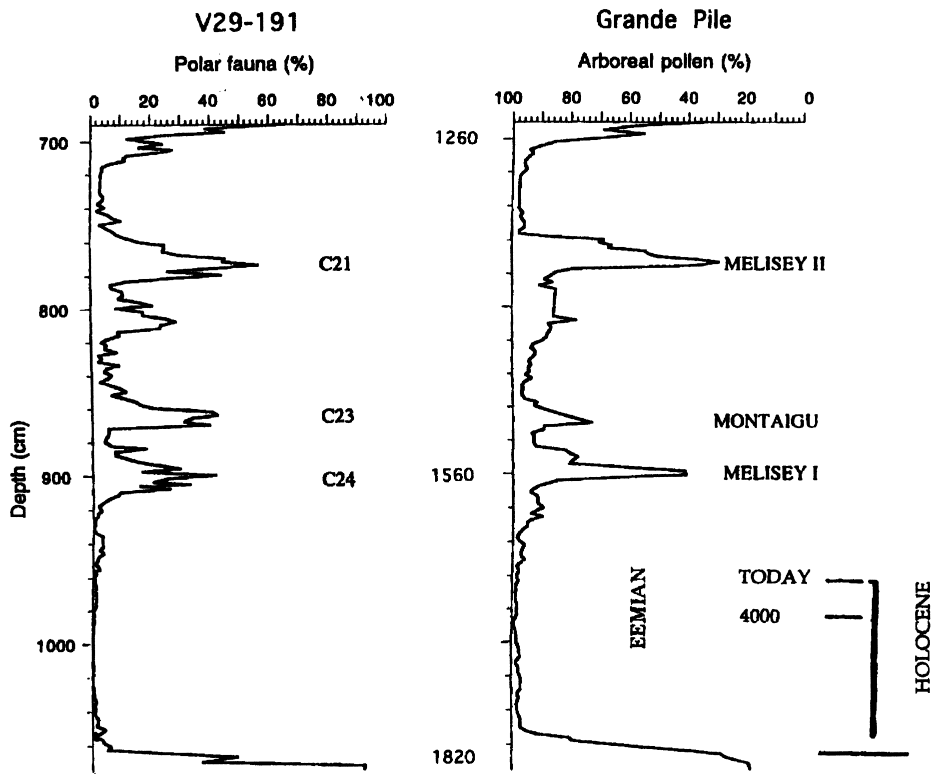

Figure 6. Percentage of left coiling *Neogloboquadrina pachyderma sinistral* in the North Atlantic core V29-191 (C21-24, cold water intrusion events from McManus *et al*., 1994), compared with the percentage of arboreal pollen in La Grande pollen record (Woillard, 1978). Plotted on the original depth scale and aligned at the MIS 4/5 and 5/6 boundaries assuming linear depth to time ratio. The equivalent position of the elapsed part of the Holocene and the 4000 BP desiccation event in the glacial/interglacial cycle is shown on the right (from G Kukla, 1997, Was the Holocene climate uniquely benign or is the Eemian catching cold from the reinforcement syndrome virus. In *Third Millennium BC Climate Change and Old World Collapse*, HN Dalfes, G Kukla, H Weiss (eds), pp. 699–710 (figure 1), reproduced by permission of Springer-Verlag GmbH & Co. KG)

These changes can be interpreted to be the result of the recurrent reorganization or cessation and initiation of the Atlantic thermohaline circulation that may have taken place in a matter of decades or less (e.g. Broecker and Denton, 1989; Grootes, 1995; Severinghaus *et al*., 1998).

Sudden and short-lived climate oscillations giving rise to warm events occurred many times during the generally colder conditions that prevailed during the last glacial period between 110000 and 10000 years ago (Figure 7). They are often known as 'interstadials' to distinguish them from the cold phases or 'stadials' (e.g. Lowe and Walker, 1984). Between 115000 and 14000 years ago, there are 24 of these oscillations recognized in the Greenland ice-core records, where they are called 'Dansgaard–Oeschger oscillations' (Bond *et al*., 1993; Dansgaard *et al*., 1993; Taylor *et al*., 1993, 1997; Bond and Lotti, 1995; Mayewski *et al*., 1997). Some workers view Dansgaard–Oeschger oscillations as being just sudden warmings in an otherwise cold climate (e.g. Adams *et al*., 1999). In the presence of these oscillations, it can be difficult to define the mean climatic state, which is unlikely to be at the cold extreme of the oscillations. It is better, therefore, to view the Dansgaard–Oeschger oscillations as being oscillations of the climatic system about an extremely ill-defined mean state. Each oscillation contains a warm interstadial which is linked to a cold stadial, and they tend to come in bundles of three or four with decreasing recurrence time and duration, with the first in the bundle being significantly longer (Stocker and Marchal, 2000). Ice-core and ocean data suggest that the oscillations began and ended suddenly, though in general, the 'jump' in climate at the start of an oscillation was followed by a more gradual decline involving a stepwise series of smaller cooling events and often a fairly large terminal cooling event

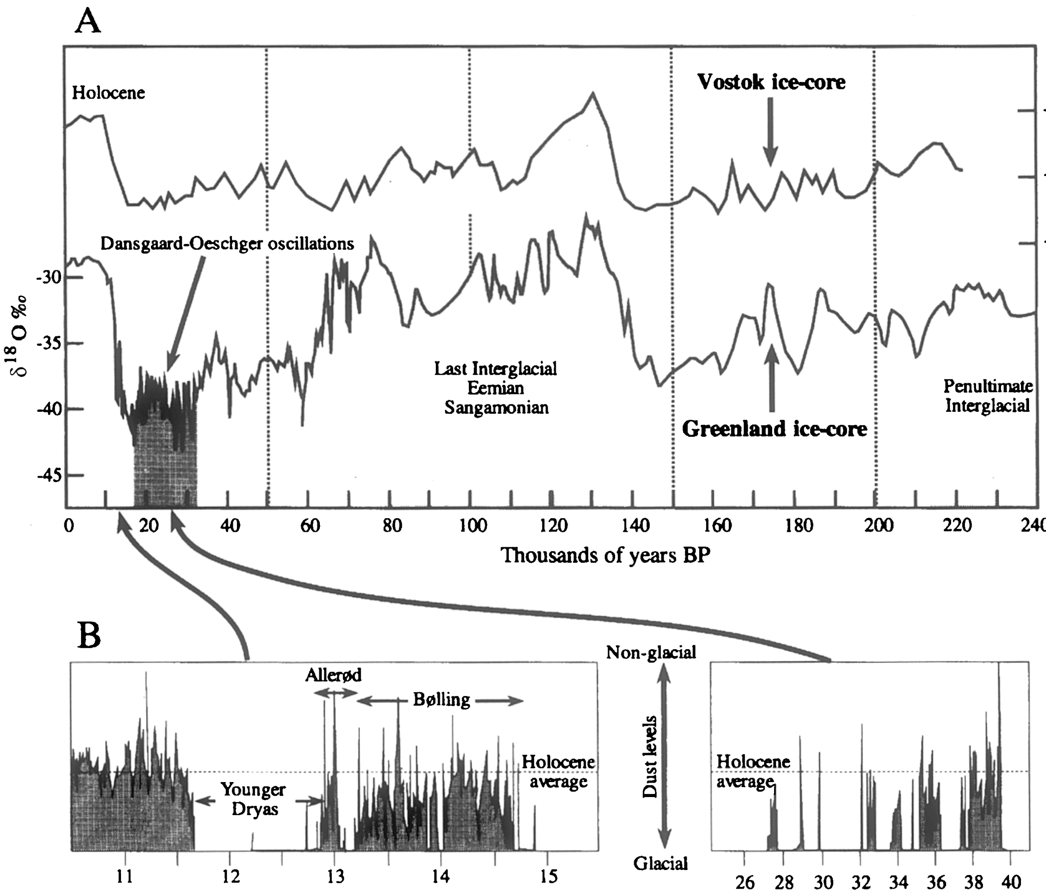

Figure 7. Late Pleistocene climatic records: (A) record from the Greenland and Antarctic icecaps (based on Dansgaard *et al*., 1993). The Dansgaard–Oeschger oscillations are marked. (B) detailed records of electrical conductivity measurements from the Greenland icecap (based on Taylor *et al*., 1993). These latter measurements are a proxy for the continental dust content falling on this icecap (from E Bryant, *Climate Process and Change*, 1997, reproduced by permission of Cambridge University Press)

which returned conditions to the colder ‘glacial’ state (Adams *et al*., 1999). From the ice-core evidence from Greenland, warming into each oscillation 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 2000 years (Mayewski *et al*., 1997). Short-lived warm phases, within warm interstadials, appear in the eastern Pacific (Behl and Kennett, 1996), western Siberia and the Arabian Sea (Sirocko *et al*., 1993; Schulz *et al*., 1998). Climatic temperature signals are weak or absent in the Southern Hemisphere and Antarctica (Stocker and Marchal, 2000).

Of apparently different nature to Dansgaard–Oeschger oscillations are extreme and short-lived cold events, known as ‘Heinrich events’ (HE) (Heinrich, 1988; Bond *et al*., 1992; Grousset *et al*., 1993; Andrews *et al*., 1994; Andrews, 1998). These events occurred against the general background of the glacial climate, and represent the climatic effects of massive surges of fresh water and icebergs from melting ice sheets into the North Atlantic, causing substantial changes in the thermohaline circulation. Several massive ice-rafting events show up in the Greenland ice-cores as a further 3 to 6°C drop in temperature from already cold glacial conditions (Figure 8) (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). They are associated with antiphase warming in the Southern Hemisphere (Stocker and Marchal, 2000) and a complete stop in NADW formation, together with a cessation of heat export from the Southern Ocean. The most recent HE is known as the Younger Dryas, and appears as a time of glacial re-advance in Europe after the end of the main ice-age.

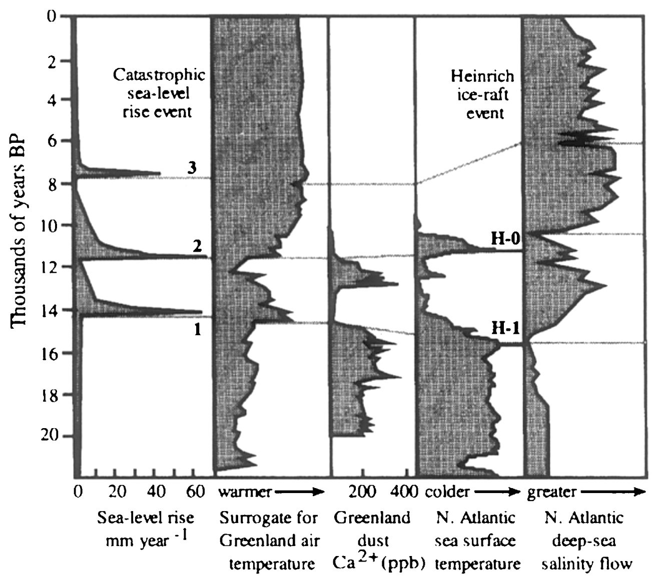

Figure 8. Atmosphere–ocean–icecap relationships for the Northern Hemisphere during the last deglaciation. All panels do not align in time because of errors in dating (Blanchon and Shaw, 1995). Figure from E Bryant, *Climate Process and Change*, 1997, reproduced by permission of Cambridge University Press

HE appear to have occurred during the last glacial about once every 7–13000 years, and are climatic coolings forced by the influx of fresh water and icebergs into the North Atlantic. Throughout the same period, there were also smaller Dansgaard–Oeschger cycles, which seem to occur every 1–3000 years, 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), as well as during earlier glacial and interglacial stages (Oppo *et al.*, 1998). Raymo *et al.* (1998) comment that in a relatively high-resolution North Atlantic ocean core, large sudden climate events resembling HE and interstadials occurred during the cooler parts of the climate cycles through the last 1.5 million years. Comment was made earlier that close to bifurcation points, system fluctuations become abnormally high since the system may 'choose' among various regimes. This was noted in the discussion of global average annual precipitation just before the abrupt warming in the 1920s. A similar process could operate for the Dansgaard–Oaschger oscillations, which are always present as low amplitude climatic oscillations, but reach high amplitudes just before climate system bifurcations, such as at the end of a glacial period.

Stocker and Marchal (2000) have considered the relationships between D/OE and HE. Comparison with the marine record suggests that the longer series of D/OE are preceded by HE. The synchronization of Greenland and Antarctic ice-cores suggests that a HE and the subsequent long series of D/OE form a unified sequence with associated global climatic signals. Stocker and Marchal (2000) call this sequence a 'Heinrich–Dansgaard/Oeschger tandem (H–D/O tandem)' and consider that there are at least three, and possibly five, examples during the last glacial period.

The following idealized scenarios are proposed by Stocker and Marchal (2000) to explain both D/OE and H–D/O tandems. Before a D/OE, Greenland temperatures are cold. This implies that the North Atlantic thermohaline circulation was reduced and that, probably, deep water formation in the Greenland–Iceland–Norwegian Sea was largely reduced or stopped, but it may have continued to the south of Iceland and in the Labrador Sea. D/OE would then be attributable to an abrupt switch-on of the Atlantic thermohaline circulation by resuming deep water formation in the Greenland–Iceland–Norwegian Sea. The warming would enhance the melting of the surrounding ice sheets and, therefore, freshwater production, leading to a subsequent reduction of the thermohaline circulation and associated cooling. The relatively short recurrence time of D/OE and small ice rafted debris layer thickness suggest that the amount of meltwater was not large. The partial shutdown, however, would be sufficient to induce a substantial cooling in the North Atlantic region. Because deep water formation is still active at other locations, the changes in the thermohaline circulation are too small to activate teleconnections with the Southern Hemisphere.

Stocker and Marchal (2000) speculate that, after a number of D/OE, the accumulated meltwater caused a sea-level rise that was sufficient to destabilize a large number of marine ice shelves (McCabe and Clark, 1998). Once these shelves had broken loose, they would free the ice streams behind them, cause a large meltwater discharge, and trigger an HE. The large meltwater discharge would shut down the North Atlantic thermohaline circulation, and cause a massive cooling in the Atlantic region. Because the thermohaline circulation has largely stopped in the North Atlantic, heat is no longer exported northwards from the Southern Ocean, resulting in a gradual warming in the south and an antiphase climatic change with the north. This is the bipolar seesaw effect (Stocker, 1998) described earlier where a warming in the Southern Hemisphere is synchronous with an HE in the Northern Hemisphere. After a few centuries, the freshwater anomaly in the North Atlantic has been mixed away, sea-surface density has increased, and the thermohaline circulation resumes. Stocker and Marchal (2000) comment that numerical models suggest that this switch-on is rapid, and would be registered as an abrupt warming in the north representing the D/O part of the H–D/O tandem. Because of the large loss of ice mass at the HE, there would be several millennia of reduced melting activity. The D/OE of H–D/O tandem lasts, therefore, longer, and it would take some time until the next meltwater event occurs initiating the next bundle of D/OE.

6. VOLCANIC ERUPTIONS AND ABRUPT CLIMATIC DETERIORATIONS

There are several types of volcanic eruptions, but the only ones of climatological interest are those involving violent explosions because they can throw vast quantities of material to great heights. The particulate matter from such eruptions is largely volcanic ash and sulphates, especially impure sulphuric acid droplets. The gases include water vapour, carbon dioxide and sulphur dioxide, but much of the sulphur dioxide is oxidized and hydrated over time to form additional sulphuric acid droplets. Tropospheric material originating from explosive eruptions either falls out rapidly under gravity, or is washed out in a few days by rain or snowfall. Volcanic material in the stratosphere consists of fine particles, liquid droplets and gases. As it is not washed out by rainfall (there are no clouds in the stratosphere), stratospheric volcanic material can have a residence time of several years, and if the eruption is in equatorial latitudes, spread over the whole Earth.

Because of their relatively large size compared with the wavelength of the incident short-wave radiation, volcanic dust particles and droplets in the stratosphere predominantly scatter solar radiation in the direction of the incident beam (Mie scattering), with only a very small percentage being reflected back to space. The result is that the direct beam radiation from the Sun is reduced much more substantially by scattering than is the total solar radiation falling on a horizontal surface, as the bulk of the radiation scattered from the direct beam re-appears at the horizontal surface as diffuse radiation. Indeed, often the changes in the direct and diffuse components almost completely cancel out and only a very slight fall in total radiation is detectable at the ground surface. Therefore, the impacts on climate of explosive volcanic eruptions are not as large as at might at first be expected. The magnitude of any climatic effect depends on the volume of material ejected and the ejection height, the latitude of the volcano (equatorial eruptions are most significant), the prevailing stratospheric circulation, and the amount of sulphur dioxide emitted. It appears that, for many major eruptions, the effects on large-scale temperature averages are indistinguishable from the background noise after 2–3 years (Kelly and Sear, 1984; Sear *et al*., 1987; Bradley, 1988). Major climatic effects are, therefore, restricted to approximately 2 or 3 years after the explosive eruption, but can be very significant over this short period. For example, climate models calculate a maximum cooling of about 0.5°C following the 1991 eruption of Mount Pinatubo, while the same models would yield a cooling of 3–4°C if the aerosols remained present indefinitely (Hansen *et al*., 1997).

Following the eruption of Mount Pinatubo in June 1991, clear-sky solar radiation incident at the Earth's surface was reduced by about 3% for several months because of aerosol scattering (Dutton and Christy, 1992). That is, the aerosols reflected to space, sufficient sunlight to cause a negative forcing of about 3 Wm^{-2}; by comparison, none of the other natural and anthropogenic radiative forcings that are known alter radiative forcing by more than 0.1 Wm^{-2} on a 2-year timescale (Hansen *et al*., 1997). Parker *et al*. (1996) report that global mean temperatures were reduced, by up to 0.5°C at the surface, and 0.6°C in the troposphere, for some months in mid-1992. Seasonal series of land-surface air temperatures indicate that the cooling in 1992, which probably resulted from the Mount Pinatubo eruption, was particularly marked in summer and autumn in the Northern Hemisphere. Robock and Mao (1995) found that this maximum cooling pattern, in the Northern Hemisphere summer of the year after the eruption, agrees with that after the five other largest volcanic eruptions since 1883. There is some evidence that explosive eruptions can influence long-term mean temperatures. Andronova and Schlesinger (2000) consider that the 1904–1944 warming was predominately owing to a residual factor plus volcanic forcing, the latter as a result of the decrease in volcanic eruptions during 1904–1944, relative to the late nineteenth century. Kelly *et al*. (1996) report evidence of limited higher latitude warming, and a major change in the atmospheric circulation over the Northern Hemisphere during the first winter after equatorial eruptions.

The largest climatic impact of explosive eruptions is in the summer following the eruption (Robock and Mao, 1995). The resulting lower temperatures can have a serious adverse influence on the growing season and agricultural production. The occurrence of temperatures below freezing leaves a permanent anatomically distinct record in tree rings. Frost early in the growing season causes changes in the early-wood part of a tree rings; frost late in the growing season causes damage to the small thick-walled cells of the latewood. LaMarche and Hirschboeck (1984) used bristlecone pines at seven locations in the western USA to identify 'notable frost-ring events' which caused widespread latewood damage. These data were compared with Lamb's geographically weighed Dust Veil Index, according to which 19 major volcanic events occurred in the period 1500–1963. Ten of these were followed by notable frost events in the same year, or within 1 or 2 years afterwards. In ten other cases, however, a frost occurred without a preceding volcanic event, and seven volcanic events had no associated frost event. The observed number of joint occurrences was found to be six times that expected by chance. There are other causes for sudden short cold events. For example, the extraordinary hard European winters in the 1690s were connected with extreme outbreaks of cold continental air from Siberia to central Europe (Wanner *et al*., 1995).

The sequence of weather and crop reactions in the New England States of the USA during the summer of 1816, after the eruption of Tambora in 1815, is an example of an extreme year in a period of unusually severe conditions (Post, 1977). July 1816 averaged 3°C below normal, making this the coldest July in American meteorological history (Groves, 1988). Though the mean temperature in 1816 was below the normal for the period 1816–1838, it was not as cold as in 1836 or 1837, the years following the eruption of Coseguina. It was the low temperatures in the growing season of 1816 which made it such an outstanding year. Hoyt and Schatten (1997) speculate that a slightly dimmer sun in the early 1800s may have contributed to this prolonged cold spell, and it is this combined with the eruption of Tambora that caused the poor summer of 1816 in New England.

Explosive volcanic eruptions can, therefore, cause abrupt climatic deteriorations in the temperate summer for 1 or 2 years after the eruption. These abrupt deteriorations could cause severe damage to crops as they occur in the northern temperate growing season. Kelly *et al*. (1996) comment that the magnitude and duration of the global cooling is not sufficient, however, to obscure any signal owing to greenhouse gas enhancement in the twenty-first century.

7. FINAL COMMENTS

Investigations of the properties of systems which are far from equilibrium show that they have a number of unusual properties. In particular, as the distance from equilibrium increases, they can develop complex oscillations with both chaotic and periodic characteristics. They also may show bifurcation points where the system may 'choose' among various regimes. Under non-equilibrium conditions, local events have repercussions throughout the whole system. These long-range correlations are at first small, but increase with distance from equilibrium, and may become infinite at bifurcation points. Some aspects of non-equilibrium systems can be found in the climatic system. On the climatological scale, it exhibits abrupt jumps in the long-term rate of temperature change, which are often associated with changes in atmospheric flow patterns. For example, in the twentieth century, such abrupt changes have been noted around 1912 and 1976 (Table I). Numerous examples have been quoted of changes in both climatic teleconnections and variability over time, suggesting that these are not stable, but vary with the state of the climatic system.

An interesting question is where is equilibrium for the climate system, and is it the same for all parts of the system? In many climates, the equilibrium state is probably found when the atmospheric flow patterns are weakest, local conditions dominate, and teleconnections are non-existent. For middle latitudes, the strength of the zonal circulation can be taken as a measure of the distance from equilibrium. Strong zonal westerlies (or high NAO index) are associated with strong meridional temperature gradients and, in particular, with anomalous polar cold and topical warmth. Anomalously cold polar and temperate regions, and relatively warm tropics are most strongly developed in glacial periods, and these, therefore, could represent conditions when the temperate latitude atmosphere is farthest from equilibrium. It is under glacial conditions that oscillations and abrupt changes in temperate climates, such as the Dansgaard–Oeschger oscillations, are most marked.

In contrast, a warmer tropical atmosphere is capable of holding more water vapour, and is, therefore, associated with enhanced rainfall and a greater release of latent heat. This increased release of latent heat results in a more vigorous tropical circulation, which, therefore, moves further from equilibrium. Under these conditions, the tropical atmosphere could be more subject to abrupt changes and significant teleconnections. Mann *et al.* (2000) comment that there is a clear breakdown in interannual variability during the early and mid-nineteenth century in both the Niño-3 index and the global temperature field. It was also commented earlier that Luterbacher *et al.* (1999) found during the middle of the nineteenth century that the circulation over the eastern North Atlantic was temporally decoupled from the continental air flow. Mann *et al.* (2000) consider that this amplitude modulation is intriguing because this period is also the coldest in terms of Northern Hemisphere mean temperature, associated with the Maunder minimum period of relatively low apparent levels of solar irradiance, and preceding the marked twentieth century greenhouse gas increase. It suggests the possibility that interannual ENSO variability is a feature of a relatively warm climate, and thus, enhanced greenhouse forcing might be expected to increase the amplitude of interannual ENSO variability. Mann *et al.* (2000) comment that there may be an upper limit on the global temperature regime within which interannual ENSO variability can reside. It was noted earlier that it appears to have vanished in the relatively warm mid-Holocene. Indeed, numerical experiments by Knutson *et al.* (1997) show a decrease in the interannual ENSO-like variability in an enhanced greenhouse climate. Marika *et al.* (2000) report numerical simulations for double CO_2 conditions that show the variance of the North Atlantic thermohaline circulation is reduced to 7% of its simulated value under present day forcing. This decrease is cause by relatively low arctic sea-ice export.

Numerical simulations by Manabe and Stouffer (1993) showed, for the North Atlantic, that between two and four times the preindustrial CO_2 concentration, a threshold value is passed and the thermohaline circulation ceases completely. The atmosphere–ocean system changes into a climate state not unlike those observed during the brief cold events of the last glacial or the Little Ice Age. Stocker and Schmittner (1997) further note that not only is the final CO_2 concentration critical with respect to irreversible changes of the Atlantic thermohaline circulation, but so also is the rate of CO_2 increase. For an increase of 1% per year (slightly higher than in the 1980s), the threshold value is between 650 and 750 ppm. When this threshold is passed the thermohaline circulation decreases and a new stable state is reached, that is, it represents some form of bifurcation point. For a slower rate of only 0.5% per year, the circulation recovers, while it collapses if it is faster.

The study of abrupt climatic changes is very new and raises many interesting and important questions about the way in which the climatic system operates. Climatic changes in the past have not always taken place in a slow, smooth manner. It is most unlikely that future changes associated with the present observed global warming will be smooth. We could, therefore, be in for some climatological surprises!

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