Int Journ Earth Sciences (1999) 88:365–374

REVIEW ARTICLE

T. F. Stocker

Abrupt climate changes: from the past to the future – a review

Received: 24 June 1998 / Accepted: 27 November 1998

Abstract A review of climatic variability is given with a focus on abrupt changes during the last glacial. Evidence from palaeoclimatic archives suggests that these were most likley due to reorganisations of the atmosphere–ocean system. The mechanisms responsible for these changes are presented. Their implication for future climate changes is discussed in light of recent climate model simulations.

Key words Abrupt climate change · Thermohaline circulation · Deglaciation · Paleoclimate · Younger Dryas

Introduction

Knowledge of past climate changes is important for two reasons. Firstly, the climate system exhibits natural variability on a range of time scales. Processes that govern this variability can only be understood if their temporal evolution is documented in the unperturbed climate of the past. Secondly, the range of possible climate states as a response to large perturbations has to be explored by reading the past records.

Systematic measurements of the simplest parameters characterizing climate (e.g. surface air temperature and sea-level pressure) only go back approximately 150 years. This was a time during which the atmospheric CO_2 concentrations were already significantly above the preindustrial level and the first anthropogenic influences on the climate system could be detected (Mann et al. 1998). To extend the time scale much further into the past, the information which is stored in palaeoclimatic archives needs to be accessed (Crowley and North 1991). These are environmental systems that are able to preserve information of environmental parameters in the past. They typically consist of a system in which material is deposited in a continuous way. Examples are the annual growth rings of corals or tree rings, sedimentary deposits in lakes, or the ocean or deposits of snow on the "permanent" ice covers of the polar and high-altitude sites.

Because environmental information is often present only in an encrypted form in such archives, special techniques have to be developed to translate measured variables (e.g. tree ring width or an isotopic composition of a coral growth ring), so-called proxy data, into climatically relevant data (e.g. temperature). The only archive that records a few key variables of the Earth system in a direct way are ice cores from the polar ice sheets. The chemical composition of precipitation is recorded in the annual layers of snow which is later compacted to ice. For instance, calcium concentrations remain unchanged for hundreds of thousands of years, whereas others, such as H_2O_2, can undergo chemical changes after a few centuries. Deposited snow is a porous medium in which air can circulate. During compression, in which firn and eventually ice is formed, air bubbles get trapped in the ice matrix (Schwander et al. 1993). These are minute samples of the air at the time of enclosure. Analysis of the enclosed air permits the reconstruction of past concentrations of the greenhouse gases such as CO_2, CH_4 and N_2O over the past 400,000 years (Raynaud et al. 1993). Information from these individual archives can be substantially augmented by synchronising records from different locations (Bender et al. 1994; Sowers and Bender 1995; Blunier et al. 1998).

Mechanisms of climate changes

Three types of climate changes

Climate changes are typically characterised by the natural time and length scales on which they occur

T. F. Stocker
Climate and Environmental Physics, Physics Institute,
University of Bern, Sidlerstrasse 5, CH-3012 Bern, Switzerland
Tel: +41-31-6314462, Fax: +41-31-6314405
e-mail: stocker@climate.unibe.ch

(Mitchell 1976). Analysing climate archives that extend back tens of thousands of years, three types of changes can be distinguished:

1. Natural fluctuations about a mean state such as the Holocene (Stocker 1996b). Examples are the El Niño Southern Oscillation (ENSO) and the North Atlantic Oscillation (NAO; Wallace and Gutzler 1981) with typical time and length scales of 3–15 years and 10^6–10^7 m, respectively.
2. Sequences of abrupt changes during the last glacial: Dansgaard/Oeschger (D/O) events (Oeschger et al. 1984) and Heinrich events (Heinrich 1988), and terminations of ice ages (Dansgaard et al. 1989). Typical time and length scales are 3–100 years and 10^6 m to global. The major difference to the first type is that amplitudes of changes are much larger.
3. Slow variations that are probably caused by changes in the parameters of the Earth's orbit. Milankovich's theory is one of the cornerstones of our understanding of Quaternary ice age (Imbrie et al. 1992; Berger et al. 1993; Loutre and Berger, in press), although important questions are still unanswered. Typical time scales are 20,000, 40,000, 100,000 and 400,000 years, and changes are global.

It is noteworthy that division into distinct types of variability helps identify and classify mechanisms, but the climate system is clearly more complex with fluctuations occurring on all spatial and temporal scales. It will be an important task of future research to investigate the interactions between scales that have been classically thought to be separated (e.g. sub-Milankovich and abrupt changes).

The aforementioned three types of climate changes are illustrated in the detailed climate history of the past 40,000 years (Fig. 1). The last ice age was characterised by a sequence of abrupt changes. This is documented in various palaeoclimatic archives ranging from tree rings as well as marine and lake sediments, peat bogs and tropical and polar ice cores (Broecker and Denton 1989). They are particularly well recorded in ice cores and marine sediments. Detailed analysis of ocean sediments suggests that the ocean plays a central role in these fluctuations (Bond and Lotti 1995). The mechanisms that are most likely responsible for these changes are discussed herein.

Natural variability during an epoch like the Holocene is more difficult to extract from ice cores or marine sediments because (a) amplitudes associated with these changes are much smaller, and (b) the temporal resolution of marine sediments is often limited. Only in the rarest cases and under special local conditions can annual layers be detected in marine sediments (Behl and Kennet 1996; Hughen et al. 1996). However, isotopic analysis of corals has the potential to reconstruct past El Niño events (Quinn et al. 1998), and the NAO has been recently detected in ice cores from Greenland (White et al. 1997; Appenzeller et al. 1998a; Appenzeller et al. 1998b). The interaction between the comparatively well-known dynamics in the Pacific equatorial region (ENSO) and the mid and high latitudes needs more attention in the future (Cane 1998; Stocker 1998).

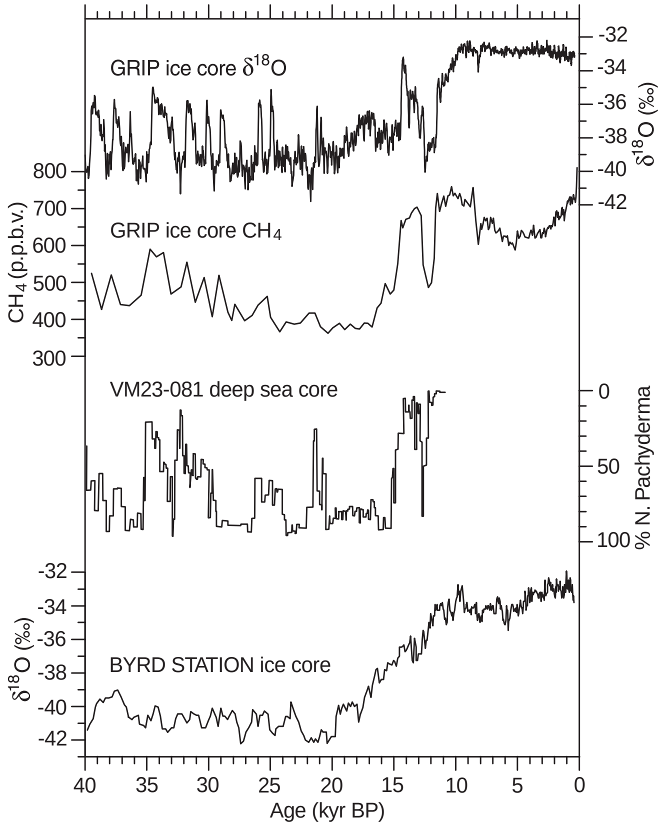

Fig. 1 Climatic change over the past 40,000 years as obtained from the measurement of d^{18}O on ice (from Johnsen et al. 1992, Dansgaard et al. 1993 and Hammer et al. 1994) and CH_4 on air of bubbles trapped in the ice core (from Chappellaz et al. 1993 and Blunier et al. 1995). Four different features of climate variability are evident in the different time series: (a) slow, astronomically forced transition from the glacial to the interglacial ($\delta^{18}O$ records and CH_4); (b) natural variability during the Holocene ($\delta^{18}O$ records); and (c) abrupt reorganisations before 25 kyr BP and during the Bolling/Allerod/Younger Dryas Period (all but Byrd Station core). (Courtesy of T. Blunier)

The ocean as a pacemaker of abrupt events

The ocean covers over 70% of the Earth's surface and the global transports of water and energy are strongly influenced by the ocean and its currents. Four major circulation types characterise the flow in an ocean basin on large spatial scales (Fig. 2). The general circulation is forced by the input of momentum through surface wind stress τ and by the flux of buoyancy, indicated by the vertical arrow S and uniform upwelling Q. The surface wind stress forces the wind-driven geostrophic circulation (WGC) which is intensified at the western boundary and forms the subtropical and subpolar

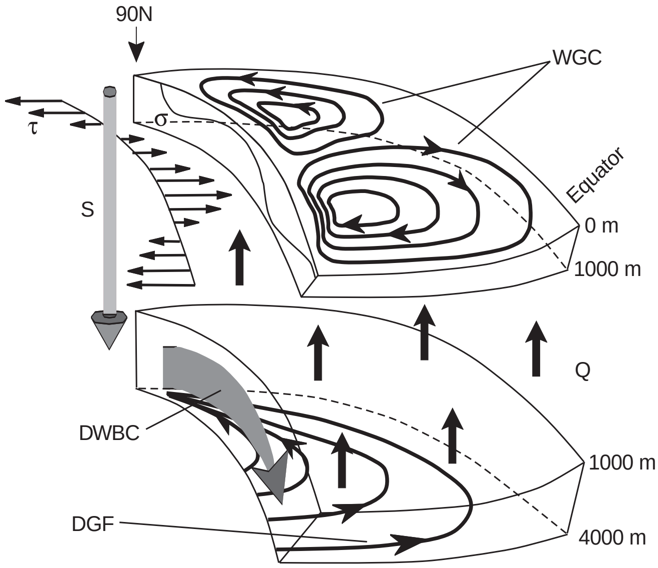

Fig. 2 View of the different types of steady-state circulations in a sectorial ocean basin extending from the equator to the pole with a longitudinal extent of approximately 60°. Wind stress τ drives a wind-driven gyre circulation (WGC) which shows western intensification due to the curvature of the rotating Earth. τ also causes Ekman suction in the northerly and Ekman pumping in the southerly upper layer giving a near-surface isopycnal surface s its typical shape: The isopycnal is shallow below the subpolar gyre and deep below the subtropical gyre. A source of newly formed deep water, *S*, feeds the deep ocean in which a deep western boundary current (DWBC) develops from which the deep geostrophic flow (DGF) of the interior is derived. The DGF flows northward to conserve potential vorticity while slowly upwelling. This results in a vertical mass flux *Q* that closes the flow. In reality, $Q<S_0$ in this sector and the DWBC is crossing the equator setting up a global circulation. (From Stocker 1996a)

gyres. Ekman pumping and Ekman suction change the local depth of the near-surface isopycnals σ which set up horizontal pressure gradients with associated geostrophic flows. They are responsible for the fact that the wind driven gyres do not extend all the way to the bottom but are compensated by sloping isopycnals in the top few hundred metres. In other words, the geostrophic velocities exhibit a vertical structure (Pedlosky 1996).

The source *S* feeds the deep western boundary current (DWBC) which flows southward and leaks into the deep interior where the deep geostrophic flow (DGF) is directed polewards at all latitudes. The DGF recirculates into the source area of the DWBC. There is a cross-interface mass flux *Q* into the upper 1000 m which supplies the mass lost due to *S* in the upper layer.

There are only a few locations in the ocean where new deep water is being formed. These are the Greenland–Iceland–Norwegian Seas in the north and the Weddell Sea in the south, and a few other, minor sites (Killworth 1983). The dynamics of a fluid moving on a rotating sphere dictates that the flow is confined to western boundary currents (Stommel 1958; Stommel and Arons 1960). The northern source is strong enough so that the current crosses the equator and penetrates eventually into the southern ocean. There, it mixes with the deep waters from the Weddell Sea and flows into the Indian and Pacific oceans where broad upwelling occurs. The global structure of the deep water paths was suggested by Stommel (1958); the return flow in the thermocline, preferentially via the "warm water route" around Africa, was first described by Gordon (1986). This global flow subsequently became known by the name "conveyor belt" (Broecker 1987; Broecker 1991), but the structure is far more complicated than a simple ribbon spanning the globe (Schmitz 1995). The metaphor of a continuous flow with a unique water mass identity is misleading. Whereas the abyssal flow can be traced from one ocean basin to the next, the thermocline flow is strongly influenced by other types of circulation such as the wind-driven gyre and Ekman circulations (Fig. 2).

Evaluations of the radiation balance at the top of the atmosphere show that the ocean–atmosphere system must transport heat towards the high latitudes where there is a net loss of energy over 1 year (Trenberth and Solomon 1994). Approximately half of that heat is carried by ocean currents (Macdonald and Wunsch 1996). In contrast to the other ocean basins, the meridional heat transport in the Atlantic Ocean is northward at all latitudes. Evaluation of oceanographic observations (Hall and Bryden 1982) as well as model simulations (Böning et al. 1996) indicate that the meridional heat transport in the Atlantic is due primarily to the meridional overturning circulation which carries warm near-surface waters northward and cold deep water southward. This is the deep circulation of the ocean that is driven by surface buoyancy fluxes and is referred to as the "thermohaline circulation" (Warren 1981). The wind-driven, near-surface circulations in the Atlantic do not transport significant amounts of heat polewards.

The idea that the ocean has more than just a regulating or damping effect on climate changes is old. The American geologist, T.C. Chamberlin, who made significant contributions to the understanding of the ice ages, wrote in 1906 (Chamberlin 1906):
"In an endeavor to find some measure of the rate of the abysmal circulation, it became clear that the agencies which influence the deep-sea movements in opposite phases were very nearly balanced. From this sprang the suggestion that, if their relative values were changed to the extent implied by geological evidence, there might be a reversal of the direction of the deep-sea circulation, and that this might throw light on some of the strange climatic phenomena of the past and give us a new means of forecast of climatic states in the future".

This statement contains the important ingredients that are responsible for what we presently call "multiple equilibria" of the ocean circulation. The topic was then dormant for approximately 50 years until Stommel (1961) observed two stable equilibrium states

in a simple model of the thermohaline circulation. In his conclusions he wrote:
"One wonders whether other, very different states of flow are permissible in the ocean or some estuaries and if such a system might jump into one of these with a sufficient perturbation; if so, the system is inherently fraught with possibilities for speculation about climatic change".

These early insights were forgotten for another 20 years until marine sediment cores were analysed in sufficiently high temporal resolution. One example is the study by Ruddiman and McIntyre (1981) who showed that the Younger Dryas (YD) cold event (12,700–11,550 years BP) was associated with a rapid southward movement of the North Atlantic polar front by more than 20° latitude. This, combined with evidence of CO_2 changes in ice cores, led Oeschger et al. (1984) to suggest that rapid changes in ocean circulation are responsible for the variations observed in the palaeoclimatic archives. The "target" for modellers was formulated by Broecker et al. (1985), and the first three-dimensional ocean model showing multiple equilibria followed soon (Bryan 1986). It was this study that finally convinced researchers that the ocean plays a central and *active* role in the climate system.

Bryan (1986) reasoned that the water sinking in the high latitudes is fed from the lower latitudes where the salinity is higher. This operates as a positive feedback mechanism for deep-water formation. Assume that this surface mass flux is temporarily reduced: the salinity then decreases in high latitudes and deep-water formation is reduced. This in turn reduces the advection of low-latitude high-salinity waters further and results in a rapid switch-off of the thermohaline circulation on time scales of decades. A "polar halocline catastrophe" ensues which is characterised by very low surface salinities in the former sinking regions (Bryan 1986). As a result, the thermohaline circulation is interrupted or even reversed implying large changes in the meridional heat fluxes, in sea-surface temperature and in sea-surface salinity. Such different states could also be realised in two-dimensional thermohaline models (Marotzke et al. 1988; Wright and Stocker 1991; Stocker and Wright 1991), multi-basin three-dimensional ocean models (Marotzke and Willebrand 1991; Mikolajewicz and Maier-Reimer 1994; Hughes and Weaver 1994; Rahmstorf and Willebrand 1995) and coupled atmosphere–ocean models (Manabe and Stouffer 1988; Schiller et al. 1997; Fanning and Weaver 1997).

All these models essentially exhibit a universal hysteresis behaviour (Fig. 3). This is due to the non-linear processes that govern the atmosphere–ocean fluxes of buoyancy which are the drivers for the thermohaline circulation (Stocker and Wright 1991). Anomalies of sea-surface temperatures are removed efficiently by anomalies in the atmosphere–ocean heat fluxes because of the strong correlation between these two quantities. On the other hand, sea-surface salinity anomalies are not correlated with anomalies in the net surface freshwater fluxes (evaporation–precipitation–runoff). Therefore, there is a large difference between the response time of these different anomalies (Rooth 1982). The important result of hysteretic behaviour is that certain perturbations in the climate system can induce abrupt, non-linear and sometimes irreversible reorganisations in the atmosphere–ocean system.

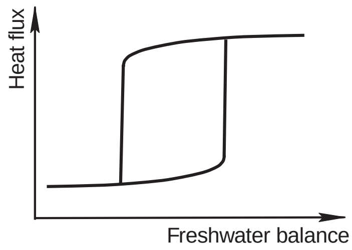

Fig. 3 The ocean–atmosphere system is a non-linear physical system that can exhibit hysteresis behaviour of the deep circulation in the ocean (Stocker and Wright 1991). Depending on the surface freshwater balance of the Atlantic Ocean, the meridional heat flux in the Atlantic is not unique and multiple equilibria exist. Changes are linear as long as they remain on the same branch of the hysteresis loop. If certain threshold values in the atmosphere–ocean system are passed, the climate state can change abruptly by switching from one branch to the other. This is a robust feature of the climate system as demonstrated by the entire hierarchy of climate models (Stocker and Wright 1991)

Dansgaard/Oeschger, Heinrich events and the Younger Dryas

Abrupt changes have become a focus in palaeoclimatic research during the past decade (Broecker and Denton 1989), because of new analytical methods and the availability of high-resolution records. Abrupt changes characterised the glacial climate: approximately 23 of these, now referred to as Dansgaard/Oeschger events, were found in the Greenland ice cores (Dansgaard et al. 1993), and both marine sediments (Bond et al. 1992; Bond et al. 1993; Bond and Lotti 1995; Hughen et al. 1996) as well as terrestrial records (Grimm et al. 1993) from the northern hemisphere show similar changes.

The typical evolution of these abrupt events as recorded in Greenland ice cores is as follows: a rapid warming, completed usually within a few decades, is followed by gradual cooling over a few hundred years, and a distinct cold phase afterwards for a few centuries to a millennium. Some of these events are more pronounced occurring on a time scale of approximately 6000–8000 years. In the marine records the cold phase of these events is associated with a layer of ice-rafted debris. These events are referred to as Heinrich events (Heinrich 1988; Broecker et al. 1992; Bond et al. 1992).

Particularly the isotopic records from Greenland ice cores suggest that three to four Dansgaard/Oeschger cycles group into long-term cooling trends punctuated by abrupt warming (Bond et al. 1993); they are referred to as "Bond cycles" (Broecker 1994). It appears that they may represent an important mode of variability during the glacial, because CO_2 exhibits small variations that correlate with these cycles (Stauffer et al. 1998; Marchal et al. 1998a). The last of these abrupt warmings was the termination of YD at approximately 11,550 years BP (Dansgaard et al. 1989; Taylor et al. 1993) within less than a decade. Another fast warming transition occurred at approximately 14,500 years BP, the first northern hemispheric warming after the last glacial maximum. However, these last two events occur during the deglaciation and other important effects (disappearing ice sheets, rising sea level) must be considered as well.

Abrupt changes are also found in variations of the atmospheric concentration of methane (Chappellaz et al. 1993a, 1993b), and in isotopic compositions of foraminifera from marine sediments (Bond et al. 1992; Bond et al. 1993; Cortijo et al. 1995; Vidal et al., in press). Icebergs from different locations around the North Atlantic disintegrated and brought large amounts of debris which were deposited on the ocean floor. The meltwater reduced the sea-surface density with the likely effect that deep-water formation in the North Atlantic was weakened. Rapid climatic changes are strongest in the North Atlantic, but concomitant signals are also found clearly at lower latitudes, in a lake in Florida (Grimm et al. 1993) and in the Cariaco Basin (Hughen et al. 1996), as well as in the Pacific region (Behl and Kennet 1996). In contrast, the climatic record of Antarctica is relatively quiet during that period with the exception of two important climatic events (Blunier et al. 1998).

The period of the last termination (21–10 kyr BP), the deglaciation (Alley and Clark 1999), contains a complicated sequence of abrupt changes both in the north (Greenland) and in the south (Antarctica; Fig. 4). Warming starts in both hemispheres around 21 kyr BP but is soon interrupted in the north, probably due to Heinrich event 1. While warming continues in the south, an abrupt warming occurs in the north at 14.5 kyr BP. This is most likely linked to a sudden onset of deep-water formation. Because a strong thermohaline circulation in the Atlantic draws heat from the southern ocean (Crowley 1992), a cooling should follow in the south. This is what is found in the isotopic record in Antarctica: the warming trend is interrupted and the Antarctic cold reversal begins. Due to the abrupt warming in the south, ice sheet melting and the hydrological cycle then increase and lead to an enhanced supply of freshwater to the Atlantic ocean. In the simple model of the interhemispheric seesaw (Broecker 1998; Stocker, in press) this would result in a gradual reduction in the Atlantic thermohaline circulation, implying gradual cooling during the Bølling/Allerød phase. Whereas this is consistent with the isotopic record from Greenland ice cores, the data of Marchitto et al. (1998) suggest a strengthening of North Atlantic deep water. If deep water is being formed at more southerly latitudes, less heat is delivered to the high northern latitudes resulting in a cooling (Rahmstorf 1994). Therefore, both amount and location of deep-water formation determine high-latitude temperatures in the North Atlantic region.

An extreme cold phase, the YD, then starts in the north, at which time the warming resumes in the south. This particular climatic evolution, superimposed on the long-term glacial–interglacial warming and opposite in both hemispheres, was suggested before (Charles et al. 1996; Jouzel et al. 1995) but became apparent when ice cores from Greenland and central Antarctica were synchronised using methane (Blunier et al. 1997). There is strong evidence that this last cold episode was due to a complete shut-down of the thermohaline circulation in the North Atlantic, at least during its initial few hundred years (Hughen et al. 1998). After another 1200 years, YD terminates within a few decades and the thermohaline circulation resumes again, drawing heat from the southern ocean. Around that same time, the long-term warming in the south comes to a close. There are therefore three instances (14.5, 12.7 and 11.6 kyr BP) when the climate system exhibits strong coupling of the two hemispheres.

There are numerous pieces of circumstantial evidence that support the hypothesis of a complete shut-down, at least during the first few centuries, of the thermohaline circulation during abrupt coolings. For example, at the end of YD, a strong ^{14}C-plateau is registered in tree rings and varved marine sediments (Hughen et al. 1998). Model simulations show that such plateaus can be generated (Stocker and Wright 1996), but the detailed evolution of ^{14}C during YD still remains to be explained by models. For example, the early decrease in atmospheric radiocarbon suggested by the data (Hughen et al. 1998) cannot yet be reproduced by models. The fact that CO_2 does not decrease during YD (Fig. 4) is another indirect indication that cooling was not global but that the sea surface must have warmed in large areas outside the North Atlantic region (Schiller et al. 1997). Model simulations are broadly consistent with the palaeoclimatic reconstructions (Marchal et al. 1999). However, as more palaeoclimatic archives are being analysed, further questions arise. The recent isotopic record of Taylor Dome (Antarctica) indicates an abrupt shift around the time of the Bølling/Allerød transition in the north (Steig et al. 1998). This may be interpreted as an abrupt warming which is synchronous to that in Greenland. However, synchronisation is disturbed again, as a clear YD signal is not observed in the Taylor Dome isotopic record. This is a major puzzle and calls for a better understanding of the regional changes on the Antarctic continent.

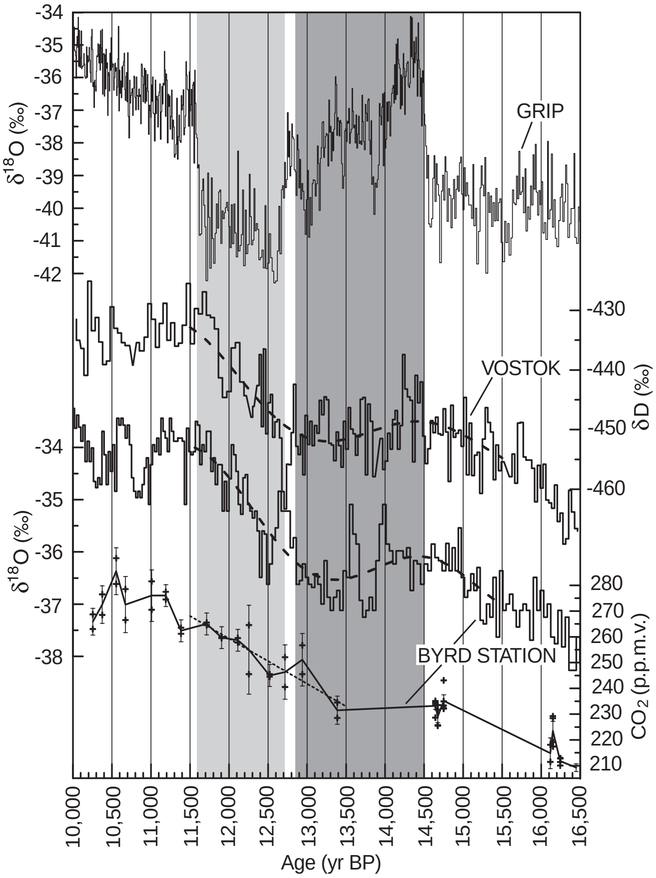

Fig. 4 High-resolution climate records based on polar ice cores from Greenland and Antarctica during a sequence of abrupt climate changes in the northern hemisphere. Changes in the isotopic composition of the water molecule, $\delta^{18}O$ and δD, indicate temperature variations. Atmospheric CO_2 is measured in bubbles enclosed in Antarctic ice from Byrd station (Blunier et al. 1997) and increases almost linearly during Younger Dryas. (From Blunier et al. 1997)

Future changes in a warmer climate

Palaeoclimatic research is the key to the quantitative estimation of future changes that may be caused by the human influence on the Earth's system. The increase in the major anthropogenic greenhouse gases (CO_2, CH_4 and N_2O) since the beginning of industrialisation could be convincingly demonstrated on the basis of reconstructions from ice cores (Raynaud et al. 1993). Since approximately 1750 the atmospheric concentration of CO_2 has increased by over 30% (Fig. 5). Through the burning of fossil fuels, land use change and cement production, approximately 7×10^{12} kg of carbon are emitted into the atmosphere every year of which approximately one third remains in the atmosphere, one third is taken up by the ocean and the rest is probably currently taken up by the terrestrial biosphere.

Carbon dioxide influences the radiation balance of the atmosphere. Climate sensitivity has been estimated at approximately 1.5–4.5 °C of global mean temperature increase for a doubling of CO_2 (IPCC 1996). This estimate is based on three-dimensional atmospheric general circulation models. However, significant regional differences are expected, e.g. the polar amplification of surface air temperatures due to increased melting of ice and snow cover in the summer.

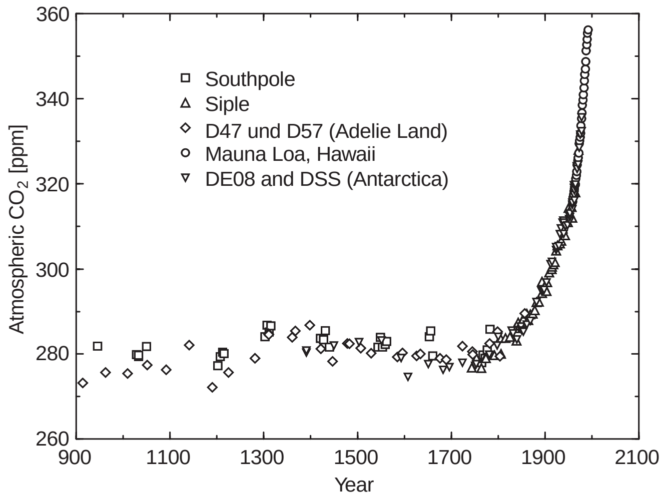

Fig. 5 Reconstruction of the industrial increase of atmospheric CO_2 from measurements in various Antarctic ice cores (Neftel 1985; Siegenthaler et al. 1988; Barnola et al. 1995; Etheridge et al. 1996). Before 1750, the concentration was approximately 280 ppm; in 1997 it was over 360 ppm. (From Joos 1996)

More recent simulations also indicate that ocean circulation is influenced by a warming of the atmosphere. Manabe and Stouffer (1993) showed that the thermohaline circulation in the North Atlantic decreases in strength. Between two and four times of the preindustrial CO_2 concentration, a threshold value is passed and the thermohaline circulation ceases completely. The atmosphere–ocean system changes into a second, structurally different climate state, not unlike those observed during the brief cold events of the last glacial. The reason for this non-linear behaviour, again, is associated with the hysteresis discussed herein (Fig. 3). Model simulations suggest that this transition may be irreversible. The question of irreversibility, however, is still difficult to address, as it depends on the exact position of the present state of the atmosphere–ocean system on the hysteresis curve. This, in turn, is strongly dependent on the model used and the exact parametrisations of sub-grid scale processes such as mixing and deep-water formation. Generally, coupled integrations exceeding several thousand years are necessary to determine whether or not the thermohaline circulation recovers (Kerr 1998).

A simplified, coupled climate model with the same global climate sensitivity has confirmed these findings (Stocker and Schmittner 1997). It is further noted that not only the final concentration is a critical quantity with respect to irreversible changes of the Atlantic thermohaline circulation. Also the rate of increase is important (Fig. 6). If CO_2 increases by 1% per year up

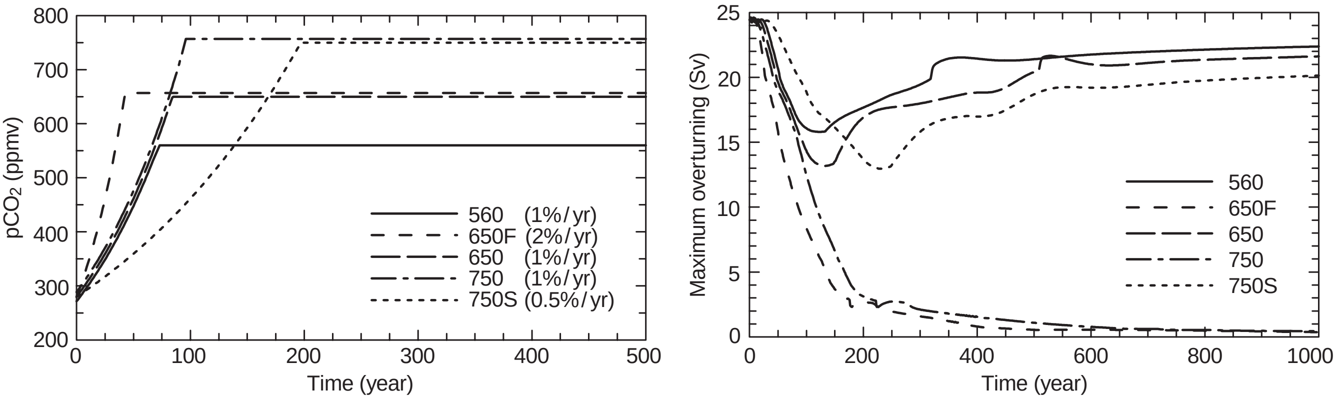

Fig. 6 a Prescribed increase of atmospheric CO_2 until a maximum concentration is reached. The rates of increase are 1% per year (exps. 560, 650, 750) and are slightly higher than the increase of CO_2 in the 1980s. Additional experiments are performed at 0.5% per year (exp. 750 S) and 2% per year (exp. 650F). **b** Evolution of the overturning volume transport of the thermohaline circulation in the North Atlantic. For an increase of 1% per year the threshold value is between 650 und 750 ppm. When passed, the thermohaline circulation decreases and a new stable state is reached. The circulation recovers, if the CO_2 increase is slower (exp. 750S), or it collapses if it is faster (exp. 650F). (From Stocker and Schmittner 1997)

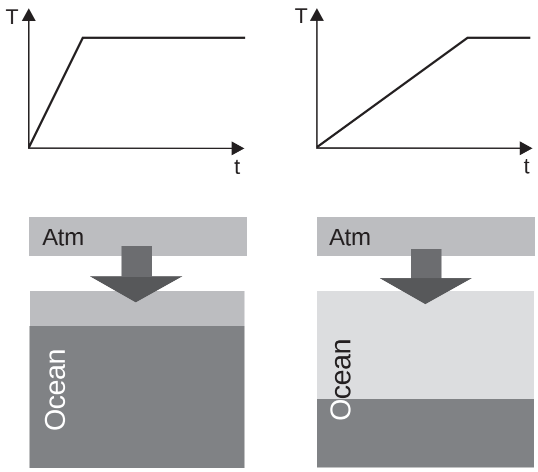

Fig. 7 Illustration of the rate-sensitive response of the thermohaline circulation. The downward transport of excess heat is limited by the vertical mixing in the ocean. For a fast increase of the atmospheric warming (*left*), the surface layers of the ocean must heat up more. This implies a greater reduction of density and an overall stabilisation of the water column. If the warming proceeds more slowly (*right*), the temperature increase distributes over a deeper layer. The near-surface density reduction is therefore smaller and the stability of the water column is not significantly altered

to 750 ppm, the deep circulation turns off permanently. For a slower rate of only 0.5% per year, the system sustains the increase of CO_2 to the same final value without a transition to the second equilibrium state. Similarly, the circulation is destabilised if the CO_2 increase is faster (exps. 650 und 650F in Fig. 6).

The reason for this rate-sensitive response of the ocean–atmosphere system is the speed of the uptake of heat by the ocean (Fig. 7). Vertical mixing processes near the ocean surface (advection, diffusion and convection) determine the downward heat flux into the ocean. If the warming is fast, then the ocean is not capable of mixing the heat down fast enough and a large reduction in sea-surface density follows. This in turn slows down the sinking of waters in the high latitudes and leads to a reduction in the thermohaline circulation. The hydrological cycle, i.e. the meridional flux of freshwater, is also increased in a warmer climate. This is crucial in finally bringing the thermohaline circulation to a collapse. The reduction in the associated meridional heat flux in the ocean causes a cooling in the North Atlantic region that is superimposed on the warming due to the increase in CO_2. This illustrates clearly that global warming is probably associated with large regional differences.

Conclusion

The Earth is a dynamical system the fate of which is determined by external forces (insolation and the orbital parameters) as well as internal processes. Climate changes have influenced the Earth's system at all times. However, complex and resource-intensive human societies are a relatively recent addition to this planet. Humans are not well experienced with climatic change, because during the past 5000 years, the late Holocene, climate has been stable. The last abrupt reorganisation happened approximately 8200 years ago (Alley et al. 1997), but its amplitude was already much smaller than the amplitude of the series of abrupt events during the last glacial. Moreover, human society is vulnerable to much smaller fluctuations such as the little ice age (Pfister et al. 1996). Burning fossil fuels is beginning to influence the global climate in a detectable manner (Mann et al. 1998; Hegerl et al. 1997) and larger changes, not unlike those of the past, can no longer be excluded.

Climate model simulations indicate that processes of change are of the same nature as those that were responsible for past changes. The investigation of past climatic changes thus exhibits the full dynamical properties of Earth's system. Threshold values are a fundamental property of this nonlinear system. Therefore, irreversible changes can happen if perturbations achieve sufficiently large amplitudes. It must be a primary goal of future climate research to determine exactly how large these threshold values are.

Acknowledgements I thank A. Berger and J. Adkins for insightful reviews. Financial support of the Swiss National Science Foundation and the Swiss Priority Program of Environment is acknowledged.

References

Alley RB, Clark PU (1999) The deglaciation of the northern hemisphere: a global perspective. Ann Rev Earth Planet Sci 27:149–182

Alley RB, Mayewski PA, Sowers T, Stuiver M, Taylor KC, Clark PU (1997) Holocene climatic instability: a prominent, widespread event 8200 yr ago. Geology 25:483–486

Appenzeller C, Schwander J, Sommer S, Stocker TF (1998a) The North Atlantic Oscillation and its imprint on precipitation and ice accumulation in Greenland. Geophys Res Lett 25:1939–1942

Appenzeller C, Stocker TF, Anklin M (1998b) North Atlantic Oscillation recorded in Greenland ice cores. Science 282:446-449

Barnola J-M, Anklin M, Procheron J, Raynaud D, Schwander J, Stauffer B (1995) CO_2. evolution during the last millennium as recorded by Antarctic and Greenland ice. Tellus 47B: 264–272

Behl RJ, Kennet JP (1996) Brief interstadial events in the Santa Barbara basin, NE Pacific, during the past 60 kyr. Nature 379:243–246

Bender M, Sowers T, Dickson M-L, Orchardo J, Grootes P, Mayewski PA, Meese DA (1994) Climate correlations between Greenland and Antarctica during the past 100,000 years. Nature 372:663–666

Berger A, Loutre M-F, Tricot C (1993) Insolation and Earth's orbital perlods. J Geophys Res 98:10341–10362

Blunier T, Chappellaz J, Schwander J, Stauffer B, Raynaud D (1995) Variations in the atmospheric methane concentration during the Holocene. Nature 374:46–49

Blunier T, Schwander J, Stauffer B, Stocker T, Dällenbach A, Indermühle A, Tschumi J, Chappellaz J, Raynaud D, Barnola J-M (1997) Timing of temperature variations during the last deglaciation in Antarctica and the atmospheric CO_2 increase with respect to the Younger Dryas event. Geophys Res Lett 24:2683–2686

Blunier T, Chappellaz J, Schwander J, Dällenbach A, Stauffer B, Stocker TF, Raynaud D, Jouzel J, Clausen HB, Hammer CU, Johnsen SJ (1998) Asynchrony of Antarctic and Greenland climate change during the last glacial period. Nature 394:739–743

Bond G, Heinrich H, Broecker W, Labeyrie L, McManus J, Andrews J, Huon S, Jantschik R, Clasen S, Simet C, Tedesco K, Klas M, Bonani G, Ivy S (1992) Evidence for massive discharges of icebergs into the North Atlantic Ocean during the last glacial period. Nature 360:245–249

Bond G et al. (1993) Correlations between climate records from North Atlantic sediments and Greenland ice. Nature 365:143–147

Bond GC, Lotti R (1995) Iceberg discharges into the North Atlantic on millennial time scales during the last glaciation. Science 267:1005–1010

Böning C, Bryan FO, Holland WR, Döscher R (1996) Deep-water formation and meridional overturning in a high-resolution model of the North Atlantic. J Phys Oceanogr 26:1142–1164

Broecker WS (1987) The biggest chill. Natural Hist 96:74–82

Broecker WS (1991) The great ocean conveyor. Oceanography 4:79–89

Broecker WS (1994) Massive iceberg discharges as triggers for global climate change. Nature 372:421–424

Broecker WS (1998) Paleocean circulation during the last deglaciation: a bipolar seasaw? Paleoceanography 13:119–121

Broecker WS, Peteet DM, Rind D (1985) Does the ocean-atmosphere system have more than one stable mode of operation? Nature 315:21–25

Broecker WS, Denton GH (1989) The role of ocean-atmosphere reorganizations in glacial cycles. Geochim Cosmochim Acta 53:2465–2501

Broecker WS, Bond G, Klas M, Clark E, McManus J (1992) Origin of the northern Atlantic's Heinrich events. Clim Dyn 6:265–273

Bryan F (1986) High-latitude salinity effects and interhemispheric thermohaline circulations. Nature 323:301–304

Cane MA (1998) A role for the tropics. Science 282:59–61

Chamberlin TC (1906) On a possible reversal of deep-sea circulation and its influence on geologic climates. J Geol 14:363–373

Chappellaz J, Blunier T, Raynaud D, Barnola JM, Schwander J, Stauffer B (1993) Synchronous changes in atmospheric CH_4 and Greenland climate between 40 and 8 kyr BP. Nature 366:443–445

Charles CD, Lynch-Stieglitz J, Ninnemann US, Fairbanks R (1996) Climate connections between the hemisphere revealed by deep sea sediment core/ice core correlations. Earth Planet Sci Lett 142:19–27

Cortijo E, Yiou P, Labeyrie L, Cremer M (1995) Sedimentary record of rapid climatic variability in the North Atlantic during the last glacial cycle. Paleoceanography 10:911–926

Crowley TJ (1992) North Atlantic deep water cools the southern hemisphere. Paleoceanography 7:489–497

Crowley TJ, North GR (1991) Paleoclimatology. Number 18 in Oxford monographs on geology and geophysics. Oxford University Press, Oxford

Dansgaard W, White JWC, Johnsen SJ (1989) The abrupt termination of the Younger Dryas climate event. Nature 339:532–534

Dansgaard W, Johnsen SJ, Clausen HB, Dahl-Jensen D, Gundestrup NS, Hammer CU, Hvidberg CS, Steffensen JP, Sveinbjornsdottir AE, Jouzel J, Bond G (1993) Evidence for general instability of past climate from a 250-kyr ice-core record. Nature 364:218–220

Etheridge DM, Steele LP, Langenfelds RL, Francey RJ, Barnola J-M, Morgan VI (1996) Natural and anthropogenic changes in atmospheric CO_2 over the last 1000 years from air in antarctic ice and firn. J Geophys Res 101:4115–4128

Fanning AF, Weaver AJ (1997) Temporal-geographical meltwater influences on the North Atlantic conveyor: implications for the Younger Dryas. Paleoceanography 12:307–320

Gordon AL (1986) Interocean exchange of thermocline water. J Geophys Res 91:5037–5046

Grimm EC, Jacobson GL Jr, Watts WA, Hanson BCS, Maasch KA (1993) A 50,000-year record of climate oscillations from Florida and its temporal correlation with the Heinrich events. Science 261:198–200

Hall MM, Bryden HL (1982) Direct estimates and mechanisms of ocean heat transport. Deep-Sea Res 29:339–359

Hammer CU, Clausen HB, Langway CC (1994) Electrical conductivity method (ECM) stratigraphic dating of the Byrd Station ice core, Antarctica. Ann Glaciol 20:115–120

Hegerl GC, Hasselmann K, Cubasch U, Mitchell JFB, Roeckner E, Voss R, Waszkewitz J (1997) Multi-fingerprint detection and attribution analysis of greenhouse gas, greenhouse gas-plus-aerosol and solar forced climate change. Clim Dyn 13:613–634

Heinrich H (1988) Origin and consequences of cyclic ice rafting in the Northeast Atlantic Ocean during the past 130,000 years. Quaternary Res 29:142–152

Hughen KA, Overpeck JT, Peterson LC, Trumbore S (1996) Rapid climate changes in the tropical Atlantic region during the last deglaciation. Nature 380:51–54

Hughen KA, Overpeck JT, Lehman SC, Kashgarian M, Southon J, Peterson LC, Alley R, Sigman DM (1998) Deglacial changes in ocean circulation from an extended radiocarbon calibration. Nature 391:65–68

Hughes TMC, Weaver AJ (1994) Multiple equilibria of an asymmetric two-basin ocean model. J Phys Oceanogr 24:619–637

Imbrie J, Boyle EA, Clemens SC, Duffy A, Howard WR, Kukla G, Kutzbach J, Martinson DG, McIntyre A, Mix AC, Molfino B, Morley JJ, Peterson LC, Pisias NG, Prell WL, Raymo ME, Shackleton NJ, Toggweiler JR (1992) On the structure and origin of major glaciation cycles. 1. Linear responses to Milankovitch forcing. Paleoceanography 7:701–738

IPCC (1996) Climate Change 1995: the science of climate change. Intergovernmental Panel on Climate Change, Cambridge University Press, Cambridge

Johnsen SJ, Clausen HB, Dansgaard W, Fuhrer K, Gundestrup N, Hammer CU, Iversen P, Jouzel J, Stauffer B, Steffensen JP (1992) Irregular glacial interstadials recorded in a new Greenland ice core. Nature 359:311–313

Joos F (1996) The atmospheric carbon dioxide perturbation. Europhys News 27:213–218

Jouzel J, Vaikmae R, Petit JR, Martin M, Duclos Y, Stievenard M, Lorius C, Toots M, Mélières MA, Burckle LH, Barkov NI, Kotlyakov VM (1995) The two-step shape and timing of the last deglaciation in Antarctica. Clim Dyn 11:151–161

Kerr RA (1998) Warming's unpleasant surprise: shivering in the greenhouse? Science 281:156–158

Killworth PD (1983) Deep convection in the world ocean. Rev Geophys Space Phys 21:1–26

Loutre MF, Berger A Future climate changes: Are we entering an exceptionally long interglacial? Clim Change (in press)

Macdonald AM, Wunsch C (1996) An estimate of global ocean circulation and heat fluxes. Nature 382:436–439

Manabe S, Stouffer RJ (1988) Two stable equilibria of a coupled ocean-atmosphere model. J Clim 1:841–866

Manabe S, Stouffer RJ (1993) Century-scale effects of increased atmospheric CO_2 on the ocean-atmosphere system. Nature 364:215–218

Mann ME, Bradley RS, Hughes MK (1998) Global-scale temperature patterns and climate forcing over the past six centuries. Nature 392:779–787

Marchal O, Stocker TF, Joos F (1998) Impact of oceanic reorganisations on the marine carbon cycle and atmospheric pCO_2. Paleoceanography 13:225–244

Marchal O, Stocker TF, Joos F, Indermühle A, Blunier T, Tschumi J (1999) Modelling the concentration of atmospheric CO_2 during the Younger Dryas climate event. Clim Dyn 15:341–354

Marchitto TM, Curry WB, Oppo DW (1998) Millennial-scale changes in North Atlantic circulation since the last glaciation. Nature 393:557–561

Marotzke J, Willebrand J (1991) Multiple equilibria of the global thermohaline circulation. J Phys Oceanogr 21:1372–1385

Marotzke J, Welander P, Willebrand J (1988) Instability and multiple equilibria in a meridional-plane model of the thermohaline circulation. Tellus 40A: 162–172

Mikolajewicz U, Maier-Reimer E (1994) Mixed boundary conditions in ocean general circulation models and their influence on the stability of the model's conveyor belt. J Geophys Res 99:22633–22644

Mitchell JM (1976) An overview of climatic variability and its causal mechanisms. Quaternary Res 6:481–493

Neftel A, Moor E, Oeschger H, Stauffer B (1985) Evidence from polar ice cores for the increase in atmospheric CO_2 in the past two centuries. Nature 315:45–47

Oeschger H, Beer J, Siegenthaler U, Stauffer B, Dansgaard W, Langway CC (1984) Late glacial climate history from ice cores. Geophys Monogr 29:299–306

Pedlosky J (1996) Ocean circulation theory. Springer, Berlin Heidelberg New York

Pfister C, Schwarz-Zanetti G, Wegmann M (1996) Winter-severity in Europe: the fourteenth century. Clim Change 34:91–108

Quinn TM, Crowley TJ, Taylor FW, Henin C, Joannot P, Join Y (1998) A multicentury stable isotope record from a New Caledonia coral: interannual and decadal sea surface temperature variability in the southwest Pacific since 1657 A.D. Paleoceanography 13:412–426

Rahmstorf S (1994) Rapid climate transitions in a coupled ocean-atmosphere model. Nature 372:82–85

Rahmstorf S, Willebrand J (1995) The role of temperature feedback in stabilizing the thermohaline circulation. J Phys Oceanogr 25:787–805

Raynaud D, Jouzel J, Barnola JM, Chappellaz J, Delmas RJ, Lorius C (1993) The ice record of greenhouse gases. Science 259:926–934

Rooth C (1982) Hydrology and ocean circulation. Prog Oceanogr 11:131–149

Ruddiman WF, McIntyre A (1981) The mode and mechanism of the last deglaciation: oceanic evidence. Quaternary Res 16:125–134

Schiller A, Mikolajewicz U, Voss R (1997) The stability of the North Atlantic thermohaline circulation in a coupled ocean-atmosphere general circulation model. Clim Dyn 13:325–347

Schmitz WJ (1995) On the interbasin-scale thermohaline circulation. Rev Geophys 33:151–173

Schwander J, Barnola J-M, Andrié C, Leuenberger M, Ludin A, Raynaud D, Stauffer B (1993) The age of the air in the firn and the ice at Summit, Greenland. J Geophys Res 98:2831–2838

Siegenthaler U, Friedli H, Loetscher H, Moor E, Neftel A, Oeschger H, Stauffer B (1988) Stable-isotope ratios and concentration of CO_2 in air from polar ice cores. Ann Glaciol 10:1–6

Sowers T, Bender M (1995) Climate records covering the last deglaciation. Science 269:210–213

Stauffer B, Blunier T, Dällenbach A, Indermühle A, Schwander J, Stocker T, Tschumi J, Chappellaz J, Raynaud D, Hammer CU, Clausen HB (1998) Atmospheric CO_2 and millennial-scale climate change during the last glacial period. Nature 392:59–62

Steig EJ, Brook EJ, White JWC, Sucher CM, Bender ML, Lehman SJ, Morse DL, Waddington ED, Clow GD (1998) Synchronous climate changes in Antarctica and the North Atlantic. Science 282:92–95

Stocker TF (1996a) The ocean in the climate system: observing and modeling its variability. In: Boutron CF (ed) European Research Course on Atmospheres: Physics and chemistry of the atmospheres, of the earth and other objects of the solar system, vol 2, pp 39–90. Editions de Physique, Les Ulis, France

Stocker TF (1996b) An overview of century time-scale variability in the climate system: observations and models. In: Anderson DLT, Willebrand J (eds) Decadal climate variability: dynamics and predictability. Volume I 44 of NATO ASI. Springer, Berlin Heidelberg New York, pp 379–406

Stocker TF (1998) The sea saw effect. Science 282:61–62

Stocker TF Past and future reorganisations in the climate system. Quaternary Sci Rev (in press)

Stocker TF, Schmittner A (1997) Influence of CO_2 emission rates on the stability of the thermohaline circulation. Nature 388:862–865

Stocker TF, Wright DG (1991) Rapid transitions of the ocean's deep circulation induced by changes in surface water fluxes. Nature 351:729–732

Stocker TF, Wright DG (1996) Rapid changes in ocean circulation and atmospheric radiocarbon. Paleoceanography 11:773–796

Stommel H (1958) The abyssal circulation. Deep-Sea Res 5:80–82

Stommel H (1961) Thermohaline convection with two stable regimes of flow. Tellus 13:224–241

Stommel H, Arons AB (1960) On the abyssal circulation of the world ocean. I. Stationary planetary flow patterns on a sphere. Deep-Sea Res 6:140–154

Taylor KC, Lamorey GW, Doyle GA, Alley RB, Grootes PM, Mayewski PA, White JWC, Barlow LK (1993) The "flickering switch" of late Pleistocene climate change. Nature 361:432–436

Trenberth KE, Solomon A (1994) The global heat balance: heat transports in the atmosphere and ocean. Clim Dyn 10:107–134

Vidal L, Schneider R, Marchal O, Bickert T, Stocker TF, Wefer G Link between the North and South Atlantic during the Heinrich events of the last glacial period. Clim Dyn (in press)

Wallace JM, Gutzler DS (1981) Teleconnections in the geopotential height field during the northern hemisphere winter. Monthly Weather Rev 109:784–812

Warren BA (1981) Deep circulation of the world ocean. In: Warren BA, Wunsch C (eds) Evolution of physical oceanography: scientific surveys in honor of Henry Stommel. MIT Press, Cambridge, Mass., pp 6–41

White JWC, Barlow LK, Fisher D, Grootes P, Jouzel J, Johnsen SJ, Stuiver M, Clausen H (1997) The climate signal in the stable isotopes of snow from Summit, Greenland: results of comparisons with modern climate observations. J Geophys Res 102:26425–26439

Wright DG, Stocker TF (1991) A zonally averaged ocean model for the thermohaline circulation. Part 1. Model development and flow dynamics. J Phys Oceanogr 21:1713–1724