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Past and future reorganizations in the climate system

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Abstract

High-resolution records of past climatic changes during the last glacial have revealed a number of abrupt changes on time scales of decades or less. Climate models suggest that the deep ocean circulation has the potential to act as a pacemaker of such changes. Based on results from ice cores from both polar regions, and the reference to a common time scale based on the methane record, it is suggested that the ocean is involved in the 24 Dansgaard-Oeschger events. For the longer events, northern and southern hemispheres are strongly coupled and exhibit climate changes of opposite sign. For the shorter events, the hemispheres are not coupled. The specific global response depends upon the forcing, and probably, the state of the ocean prior to the onset of these events. While such abrupt climate changes appear to be caused by a unique mechanism (changes in the sea surface freshwater balance), models suggest that the response of the ocean circulation depends on the amplitude and temporal evolution of the perturbation.

1. The ocean's role in the global heat budget

The ocean covers over 70% of the Earth's surface, the heat capacity of the seasonally active ocean layers exceeds that of the atmosphere by a factor of over 30, and the ocean is the biggest of the fast exchanging carbon reservoirs. The transport of water and energy is strongly influenced by the ocean and its currents. Continental run-off compensates the loss of water by net evaporation over the entire ocean and is estimated at about 1.15 Sv (1 Sv = 1 Sverdrup = $10^6 m^3 s^{-1}$) (Webster, 1994). This implies a renewal time of water in the atmosphere of only 11 days compared to that in the entire ocean of over 3000 years. In spite of the long renewal time, the ocean is also influential for fast climate change.

The radiation balance at the top of the atmosphere, averaged over 1 yr, requires a poleward heat transport of about 5–6 PW (1 PW = 10^{15}W) in order to balance the net radiative loss in the high latitudes (Trenberth and Solomon, 1994). The atmosphere transports about half of this heat; the rest is carried by ocean currents. Recent inverse calculations (Macdonald and Wunsch, 1996; Macdonald, 1998) indicate, that the maximum heat transport of the global ocean amounts to 1–2.5 PW. The largest magnitudes occur in the Pacific of the southern hemisphere, while the maximum transports in Atlantic and Pacific are very similar in the northern hemisphere in spite of the much larger Pacific basin. While for the Pacific the transport is poleward in both hemispheres, the transports in the Atlantic and Indian oceans exhibit no symmetry about the equator. The annual-mean transport in the Indian is southward at all latitudes with a strong seasonal variation due to the monsoon.

The meridional heat transport in the Atlantic Ocean is northward at all latitudes, e.g. about 0.25 PW at 30°S (Rintoul, 1991). This implies that the southern ocean is supplying heat to the Atlantic, which constitutes an important thermal coupling of these ocean basins.

The wind-driven, near-surface circulations in the Atlantic do not transport sufficient heat meridionally to account for the above estimates and their latitudinal structure. For example, at 24°N the meridional volume transport is about 30 Sv (Florida Current) and the east–west temperature contrast between the western boundary and the basin interior is small (Roemmich, 1980). 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 primarily due to the meridional overturning circulation which carries warm near-surface waters northward and cold deep water southward.

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2. The thermohaline circulation

Temperature and salinity determine the density of sea water. Density changes (thermal and haline) caused by atmosphere-ocean fluxes and the circulation itself are the principal drivers of the *thermo-haline* circulation (henceforth THC) (Warren, 1981). Intensive cooling at high latitudes, as well as production of sea ice promote deep water formation. Deep waters derive from localised regions in the northern and southern high latitudes, mainly in the Greenland-Norwegian-Iceland, the Labrador and the Weddell Seas (Killworth, 1983). Deep water must eventually upwell and it does so, most likely, on basin scale but not uniformly (Olbers et al., 1985).

Hydrographic surveys suggested indirectly that the deep circulation is confined to western boundaries (Defant, 1941). The dynamical mechanism of the steady-state circulation in the deep ocean interior is the conservation of total angular momentum of a water column (Stommel, 1958; Stommel and Arons, 1960). Away from boundary layers and topographic irregularities, pressure gradients nearly balance the Coriolis force. Upwelling requires convergent flow at depth which results in poleward motion in the interior of ocean basins below about 1 km. The deep interior flow is therefore generally *towards* the sources of deep water formation (Fig. 1), and it feeds from the deep western boundary currents along the continental margins which are the principal pathways of newly formed deep water. These intense currents are confined to within less than 100 km width (Lee et al., 1996). The classical concept of (Stommel and Arons, 1960) has been refined: non-uniform upwelling (Pedlosky, 1992), meridional topography (Kawase, 1993; Pedlosky and Chapman, 1993) or ridges and gaps (Pedlosky, 1994) all modify the inerior flow both in direction and strength.

The deep circulation is not confined to an individual hemisphere or ocean basin (Stommel, 1958), and the possibility was indicated that North Atlantic Deep Water (NADW) could be traced as far as the North Pacific (Fig. 1). Inter-hemispheric mass exchange occurs in narrow western boundary currents mediated by lateral and equatorial wave processes (Kawase, 1987). A water mass analysis (Gordon, 1986) and extensive geochemical measurements (GEOSECS, 1987) have confirmed this view. The circulation has later become widely known as the 'global conveyor belt' (Broecker, 1987; Broecker, 1991; Schmitz, 1995) with a volume transport of about 13–20 Sv in the Atlantic (Dickson and Brown, 1994; Macdonald, 1998). The appealing metaphor of a continuous flow from the deep water formation areas through the abyss, upwelling into the thermocline and back into the high-latitude formation areas is too simplistic. It remains a popularization of the complex and unsteady dynamics that is responsible for the deep circulation.

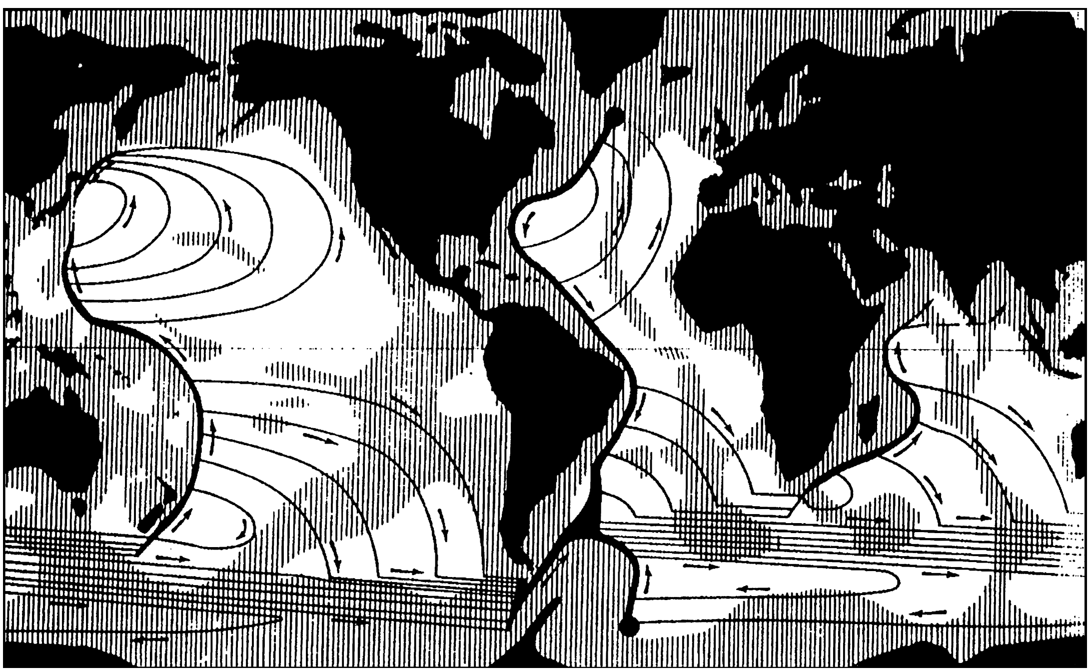

Fig. 1. Reproduction of Stommel's classic picture of the deep flow that encompasses the Earth (Stommel, 1958). Two sources of newly formed deep water, south-east of Greenland and in the Weddell Sea, which are compensated by uniform global upwelling, give rise to deep western boundary currents flowing southward in the Atlantic and northward in the Indian and Pacific Oceans. These currents feed the basin interior in which the waters recirculate towards the sources of deep water formation. Because of the absence of deep water sources in the Pacific and Indian Oceans, the deep western boundary currents are northward except in the North Pacific where the recirculation itself feeds a southward current. (Reproduced from Stommel, 1958).

A misleading feature of 'global conveyor belts' is their upper, near-surface branches in which the water returns in the thermocline to the deep water formation areas. In the upper 500 m the fluid participates in the wind-driven circulation, and the surface fluxes of heat and moisture quickly erase spatial coherency of such near-surface paths. The currents and processes that eventually lead to waters with the characteristics of North Atlantic Deep Water are complex and involve water mass conversion in both the Greenland (Dickson and Brown, 1994) and Labrador Seas (Sy et al., 1997).

3. Models and mechanisms of thermohaline changes

The American geologist T. C. Chamberlin, who has made significant contributions to the understanding of the ice ages, wrote already 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 today is called "multiple equilibria" of the ocean circulation. However, this idea was probably common knowledge among the natural scientists of that epoch (Bommeli, 1898). Stommel (1961) formulated a simple conceptual model of the THC and found that the number of stable equilibria depended on the characteristic time scale of air–sea heat fluxes. This was confirmed with a three-dimensional ocean general circulation model (Bryan, 1986), and many subsequent studies (see Weaver and Hughes, 1992, for a review) highlighted the relevance of the deep ocean circulation for abrupt, large-scale climate change. The major findings are: (i) the deep circulation has multiple equilibria; (ii) transitions between them can occur on a decadal time scale; and (iii) the stability of a particular circulation is strongly influenced by the surface freshwater fluxes.

Common for non-linear physical systems with multiple equilibrium states, the ocean exhibits a hysteresis behaviour (Fig. 2) first shown in a zonally averaged ocean circulation model (Stocker and Wright, 1991) and later confirmed with comprehensive 3-dimensional ocean general circulation models (Mikolajewicz and Maier-Reimer, 1994; Rahmstorf, 1995). In the simplest case, depending on the value of a control variable (e.g. the freshwater balance of the North Atlantic), the climate variable (e.g. sea surface temperature in the North Atlantic) can take on two different values. The existence of hysteresis implies three qualitatively different responses to a given perturbation in the control variable. If critical thresholds of the control variable are not passed, perturbations result in only small, linear changes of the climate variable (Fig. 2a). Such changes remain reversible: once the perturbation is switched off, the system evolves back to the previous state. If the initial state is located closer to a threshold value, the same perturbation results in an abrupt change following a phase of linear changes. This corresponds to a large-scale reorganisation of the deep ocean circulation. Such changes are either reversible (Fig. 2b) or irreversible (Fig. 2c) transitions to a different equilibrium state. A permanent change of the climate variable thus results, if the initial equilibrium state is not unique and the perturbation exceeds the threshold. Changes are irreversible, even though the

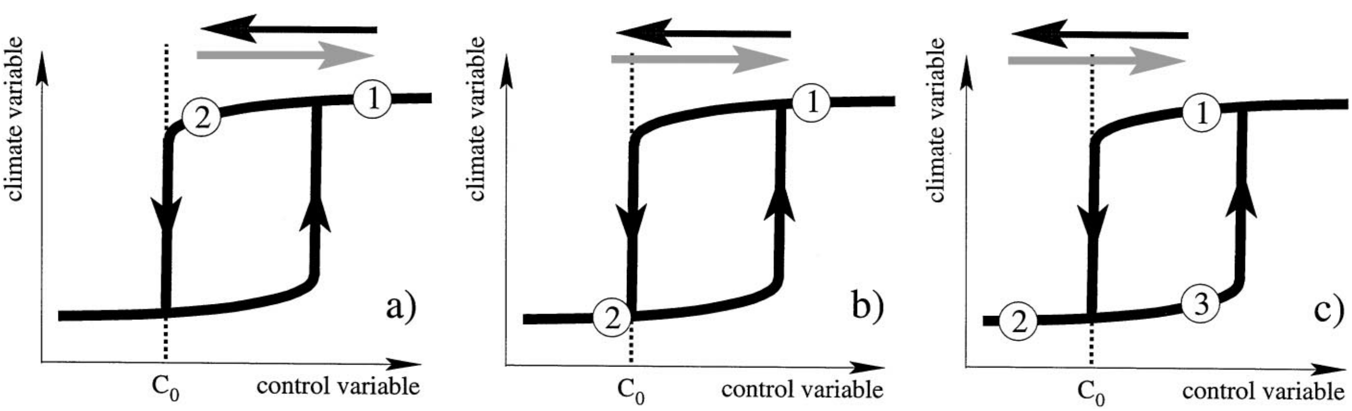

Fig. 2. The ocean–atmosphere system is a non-linear physical system that can exhibit hysteresis behaviour of the deep circulation in the ocean (Stocker and Wright, 1991). A given perturbation in the control variable causes three structurally different responses of the system (a) Linear, reversible response. (b) Non-linear, reversible response. (c) Non-linear, irreversible response.

perturbation may last for only a finite time (Fig. 2c). The response of the climate system to perturbations thus depends critically on the initial state. It is one of the important questions where exactly the present climate state is located on the hysteresis and where it might have been during the last glacial.

Multiple equilibria are intimately linked to the fact that sea surface temperature anomalies are strongly coupled to atmosphere–ocean heat flux anomalies, whereas sea surface salinity (SSS) anomalies are not influencing anomalies in atmosphere–ocean freshwater fluxes (evaporation–precipitation–runoff). An active THC is temperature-driven (Fig. 3a) – water sinks in the high latitudes where it is cooled — while the hydrological cycle acts as a brake, because excess precipitation reduces salinity, and hence density, at high latitudes (Schmitt et al., 1989). This is a direct circulation. In this circulation and with the corresponding mean vertical distribtutions of temperature and salinity, this implies northward meridonal fluxes of heat and freshwater. The latter is required to close the water balance in the presence of a northward freshwater flux in the atmosphere.

If the surface of the ocean freshens substantially due to some perturbation (e.g. increasing runoff or precipitation), deep water formation may be inhibited completely and the THC collapses. A polar halocline catastrophy occurs (Bryan, 1986). A new equilibrium state is achieved in which the circulation is shallow and reversed (Fig. 3b). Deep water is no longer cooled by sinking waters in the high latitudes and thus the vertical temperature gradients decrease. The meridional heat transport in the atmosphere is increased to compensate partly for the significantly reduced oceanic transport. Overall, the high latitudes experience a cooling due to the strong reduction of the oceanic heat transport. Since the circulation direction has changed, a reversal of vertical salinity gradients is required in order to close the freshwater cycle. This is only possible because there is no direct feedback between SSS anomalies and the atmospheric hydrological cycle. This behaviour is particularly present in ocean-only models that apply mixed boundary conditions but also occurs in fully coupled models (Manabe and Stouffer, 1988; Stocker et al., 1992b).

Thermohaline instabilities are associated with two different feedback mechanims: the advective and convective feedback (Table 1) (Rahmstorf et al., 1996). With regard to the advective feedback inducing instability, cooling leads to deep water formation only if the waters contain enough salt. In the Atlantic, the wind-driven circulation advects near-surface, high-salinity waters from the low latitudes to the high latitudes. If this supply of preconditioned waters is reduced, then the rate of deep water formation will decrease and, by mass continuity, meridional advection will become weaker. This advective feedback mechanism, with typical time scales of a few decades to a century, can lead to a complete collapse of deep water formation (Bryan, 1986). Instabilities can also be generated by the convective feedback mechanism (Welander, 1982). These are intially localised phenomena, which can spread quickly and lead to changes in the convective activity of larger areas (Lenderink and Haarsma, 1994). Such changes occur abruptly and within a few years. However, convective feedbacks and related processes are still poorly understood because of limited model resolution, missing processes in the models (e.g. formation of deep water on continental shelves) and serious gaps in the observational data (Marshall and Schott, 1999).

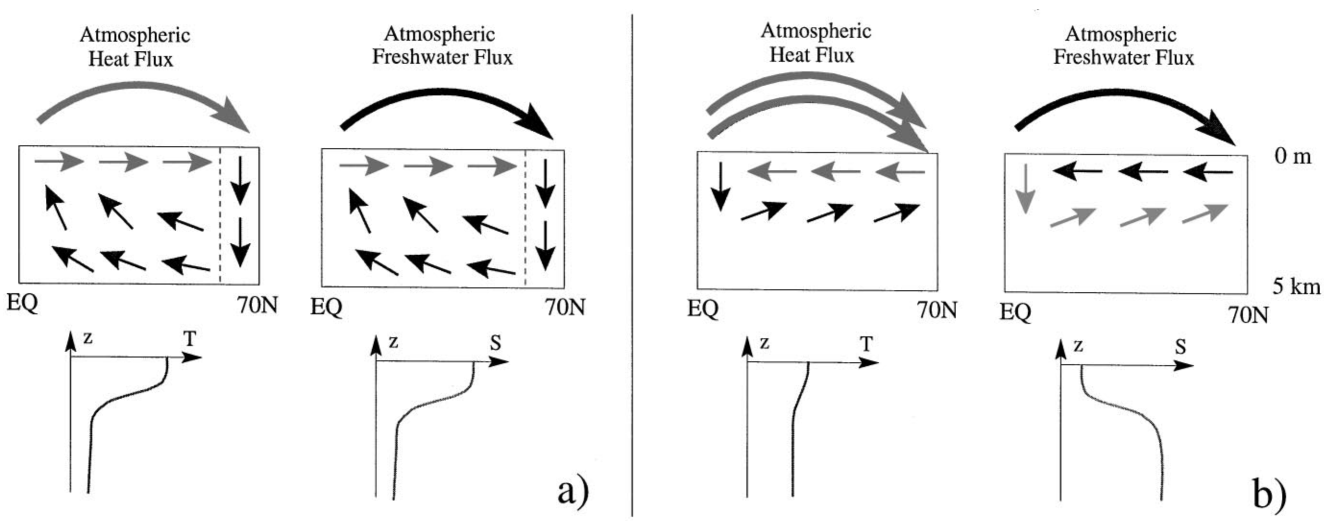

Fig. 3. Schematic representation of the basic mechanisms responsible for multiple equilibria of the THC in a simple hemispheric configuration (top row) and typical vertical profiles of temperature and salinity in the high latitudes (bottom row) for the corresponding circulation types. (a) Direct circulation. (b) Indirect circulation.

Table 1
Summary of the two principle feedback mechanisms that cause changes of the thermohaline circulation. The table is an extended version of Rahmstorf (1998); further feedback mechanisms are discussed in Marotzke (1996).

	Advective feedback	Convective feedback
Process	Advection of high salinity waters from the evaporative regions of the low latitudes to the deep water formation regions	Localized, upward mixing of warmer and more saline water into the surface layers of the high latitudes where cooling and excess precipitation prevails
Time scale	10–100 yr	1–10 yr
Trigger	Large-scale perturbation of atmosphere–ocean heat and freshwater fluxes, reduction of deep water formation at high latitudes	Localized input of freshwater, changes of sea ice cover, local changes in atmosphere–ocean heat and freshwater fluxes
Representation in models	Large-scale, robust process, fairly well represented in box models to coupled atmosphere–ocean general circulation models	Poorly parameterized in all coarse-resolution models. Only non-hydrostatic, high-resolution models simulate this process

4. Abrupt climate change documented in palaeoclimatic archives

4.1. Time scales and spatial distribution

The detection of abrupt climate changes on time scales of less than a few decades has only become possible with the exploitation of high-resolution palaeoclimatic archives which provide climate information that is nearly continuous in time. Palaeobotanical stratigraphic records were among the first to indicate that the last deglaciation was punctuated by climatic oscillations on time scales of a few thousand years (Mangerud et al., 1974; Alley and Clark, 1999). The most prominent of these changes is the Younger Dryas cooling event, dated at around 12,700–11,550 yr BP (Dansgaard et al., 1989; Taylor et al., 1993; Alley et al., 1997a), and which followed an earlier abrupt warming, the beginning of the Bølling phase (14,500 yr BP). Similar changes indicating rapid cooling are observed in lacustrine (Eicher and Siegenthaler, 1976) and marine (Ruddiman and McIntyre, 1981) records strongly suggesting a link with the ocean circulation (Oeschger et al., 1984; Broecker et al., 1985; Broecker and Denton, 1989). The analysis of the abundance of planktonic foraminifera in a high-accumulation core on the Norwegian shelf demonstrates that sea surface temperatures dropped by more than 5°C during Younger Dryas (Lehman and Keigwin, 1992b).

The most detailed and continuous information about climate variability comes from the Greenland ice cores (Fig. 4). Hans Oeschger (Oeschger et al., 1984), Willy Dansgaard (Dansgaard et al., 1984) and colleagues were the first to recognize the climatic significance of short interstadials (warming events) during the last glacial period; they were numbered consecutively (Dansgaard et al., 1993) and later named "Dansgaard/Oeschger Events" (Broecker and Denton, 1989) (see Fig. 5). All events exhibit a striking similarity in their temporal evolution: cooling extends generally over many centuries to about 3 kyr, while warming is abrupt and occurs within years or decades. This suggests that one common mechanism may be responsible for these climate swings. The recurrence time for the shorter D/O events is of the order of 1000 years.

Layers of coarse-grain lithic fragments are found in marine sediments (Heinrich, 1988), later termed "Heinrich Events" (Broecker et al., 1992). This is interpreted as material that was transported from the beds of the great ice sheets to the North Atlantic by calving icebergs from surging ice streams which then melted and provided an additional sediment layer. Heinrich layers have also low foraminifera counts indicating extended sea ice cover. However, phases of sea ice break-up in the nordic seas are also found at or shortly after Heinrich events (Dokken and Hald, 1996). Mineralogical analysis of these layers reveal that their origin is the Canadian shield (Bond et al., 1992), but discharge can also be traced to the Fennoscandian (Fronval et al., 1995) and British ice sheets (McCabe and Clark, 1998). Most of the debris material, however, originates from north of Hudson Bay (Gwiazda et al., 1996). The typical recurrence time is estimated at about 7 kyr, thus significantly longer than that for the D/O events found in the ice cores.

Correlation with events found in marine sediments from the Atlantic demonstrates the over-regional character of these records (Bond et al., 1993). This suggests that the stadials found in the ice cores and the layers of ice rafted debris may be causally linked. At first glance, D/O and H-events seem unrelated. However, Bond and Lotti (1995) shows in a high-resolution marine sediment core from the North Atlantic that smaller layers of ice rafted debris were buried between the prominent Heinrich layers, i.e. such layers also occur during or before D/O events. This, again, is strong indication of a common

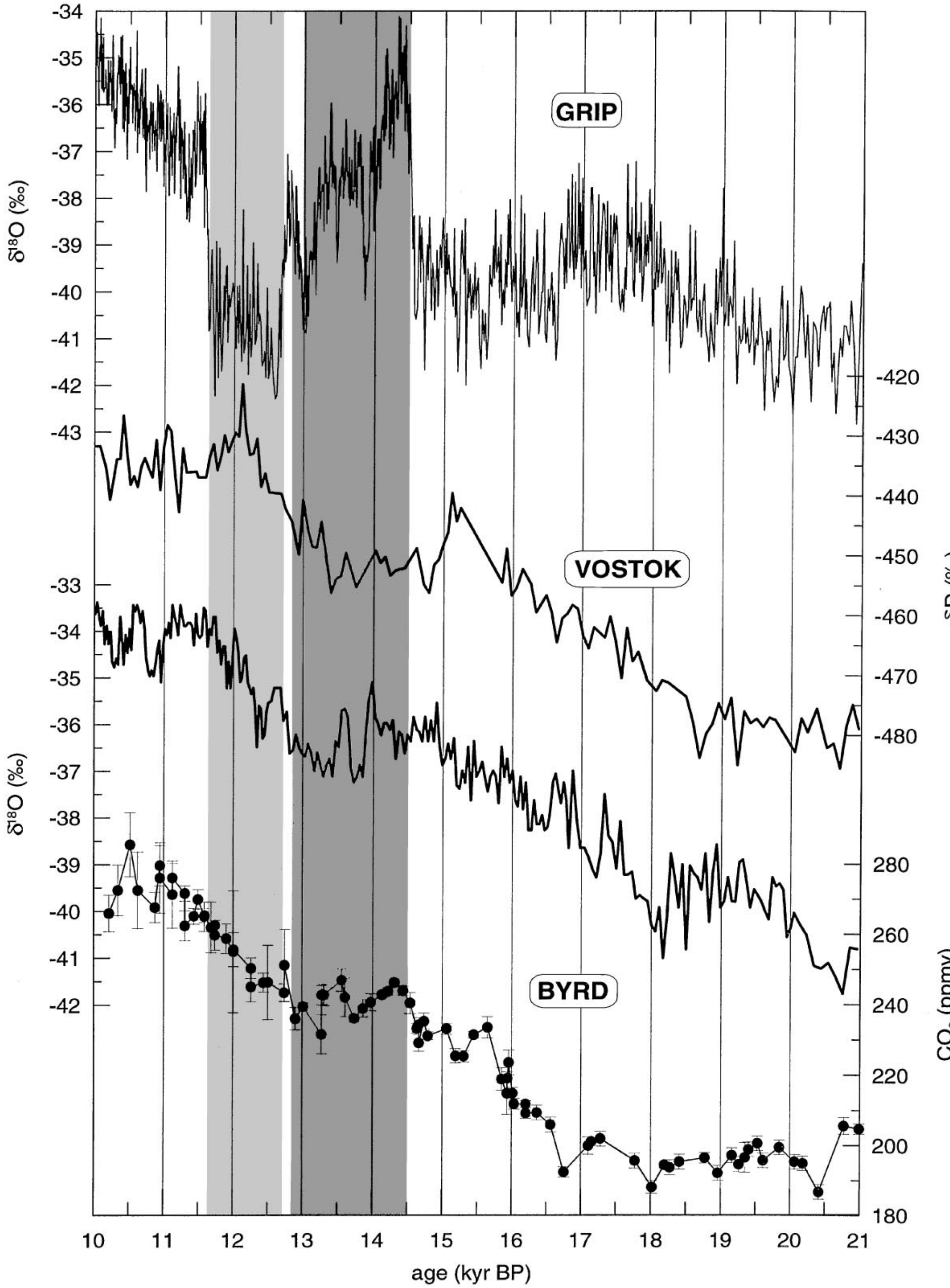

Fig. 4. High-resolution climate records based on polar ice cores from Greenland and Antarctica during the last deglaciation from 21 to 10 kyr BP including a sequence of abrupt climate changes in the northern hemisphere. Changes in the isotopic composition of the water molecule, and $\delta^{18}O$ and δD, indicate temperature variations. Atmospheric CO_2 is measured in bubbles enclosed in Antarctic ice from Byrd station (Blunier et al., 1997; Marchal et al., 1999b) and increases almost linearly during Younger Dryas (YD). The increase is interrupted during the Antarctic cold reversal (strong shading). Deglaciation begins at around 21 kyr BP in both hemispheres but is interrupted in the north, likely due to Heinrich event 1 (16.5 kyr BP). The transition into the Bølling (strong shading) warm phase is abrupt and probably initiates the cooling in Antarctica (Antarctic cold reversal). This slight cold phase is interrupted when YD begins in the northern hemisphere at 12.7 kyr BP (weak shading). The northern cold phase is coeval with the resumption of warming in the south which is, again, interrupted when YD terminates. This sequence strongly suggests an antiphase coupling of northern and southern hemispheres during this sequence of abrupt climate change. (Figure based on Blunier et al., 1997).

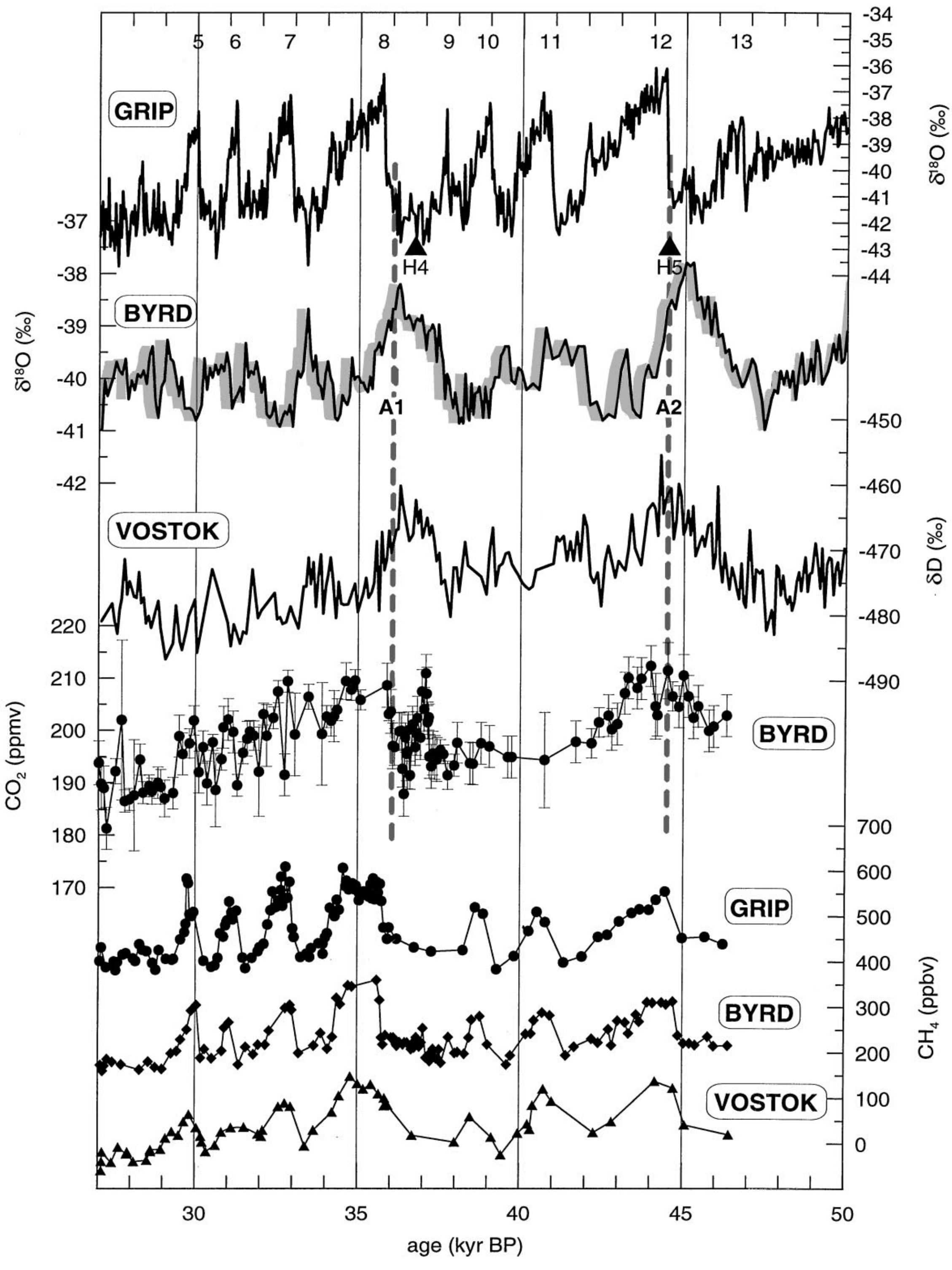

Fig. 5. Succession of D/O events 5–13 during 27 to 50 kyr BP in the last glacial as seen in Greenland (GRIP) and Antarctic (Byrd Station, Vostok) ice cores. The records are synchronised in time based on high-resolution records of methane (Stauffer et al., 1998) (bottom three curves). Events 8 and 12 are followed by longer cooling events and are accompanied by changes in Antarctica. For both of these, the slow warming in Antarctica begins about 2000 years before the abrupt warming in the north (events 8 and 12). At the time of the abrupt transition to the interstadial in the north, cooling starts in the south (events A1 and A2). No distinct changes in Antarctica are detected during the shorter D/O events 5, 6 and 9, 10. For events 7 and 11, there may be an indication of changes similar to A1 and A2 but smaller in amplitude. The D/O events (except 9) are coeval with increased values of atmospheric methane, but there is no correlation with events A1 and A2 suggesting that humid conditions in tropics may have been more strongly influenced or even caused by climate changes of the northern hemisphere. Changes in CO_2 of about 15 ppm (Stauffer et al., 1998) are registered only during the warming events in Antarctica (A1, and A2), i.e. the strong D/O events but not during the series of shorter D/O events following them (compiled by A. Dällenbach).

cause or trigger in the climate system responsible for these changes. A recent marine record suggests that such abrupt changes only occur between some limits of the global ice volume: for stronger glaciation, or interglacial states, the climate system appears to be more stable (McManus et al., 1999).

The duration of the abrupt warming at the end of Younger Dryas at 11,550 yr BP is estimated at several decades based on oxygen isotope records in Greenland ice cores (Dansgaard et al., 1989) (a proxy for temperature), isotopic composition of gases (Severinghaus et al., 1998), planktonic foraminifera assemblages in North Atlantic sediments (Lehman and Keigwin, 1992b) (a proxy for sea surface temperatures) and tree rings (Björck et al., 1996), to only a few years from electrical conductivity measurements based on a Greenland ice core (Taylor et al., 1993) (a proxy for atmospheric circulation), (Alley et al., 1993) and other indicators (Mayewski et al., 1993).

Signals of abrupt climate changes are not limited to the Atlantic region. Cold reversals during the transition from the last glacial period to the early Holocene are found in records from the Pacific west coast (Mathewes et al., 1993; Peteet, 1995) and the tropical Atlantic (Hughen et al., 1996). The entire sequence of Heinrich and Dansgaard/Oeschger events is also expressed in annually laminated deposits from the Santa Barbara basin (Pacific west coast), indicating the absence of bioturbation and hence reduced vertical mixing (Behl and Kennet, 1996). Pollen sequences from Florida appear correlated with Heinrich events (Grimm et al., 1993). The most convincing evidence for a global signal associated with these abrupt climate events is provided by the ice archive: the concentration of atmospheric methane exhibits significant variations for nearly each of the 13 Dansgaard/Oeschger events during the last 45,000 years (Chappellaz et al., 1993). This is direct evidence for changes in the hydrological cycle during each of these events, because atmospheric methane is primarily determined by the source strength of tropical and low-latitude wetlands.

But are there signals associated with these abrupt changes in palaeoclimatic archives from the southern hemisphere suggesting global synchroneity? Clear evidence from terrestrial palaeorecords does not yet exist (Peteet, 1995). In the Chilean Andes (Lowell et al., 1995) and the New Zealand Alps (Denton and Hendy, 1994) radiocarbon-dated moraine deposits indicate that mountain glacier advanced during the last deglaciation. However, the response of a glacier to climate changes strongly depends on a combination of temperature, moisture supply and local characteristics. Moreover, interpretation of radiocarbon dates can be difficult (Mabin, 1996), and precise dating to better than about 200 calendar years is required to link these glacier advances with cooling phases identified in isotopic records of Antarctic ice cores (Jouzel et al., 1995; Blunier et al., 1997).

4.2. The biogeochemical signal of abrupt changes

Changes in the deep circulation of the ocean, if real, must leave distinct signals in marine sediments. Several palaeoceanographic proxies are available to reconstruct circulation changes but only a few of them are suitable to be used to detect changes also on time scales shorter than a millennium. The series of H-events is found in the stable isotope records and grey scales of various North Atlantic sediment cores (Oppo and Lehman, 1995; Cortijo et al., 1995). Evidence for past changes in the deep circulation of the North Atlantic, which were coeval with changes at the surface, also comes from deep sea corals. Adkins et al. (1998) measure a reversal of radiocarbon ages at around 15.4 kyr BP on growth rings of deep sea corals living near Bermuda rise. This suggests that waters with a high radiocarbon content (young waters from the northern source) were rapidly replaced by waters with a lower ^{14}C content (old waters from southern origin) which makes younger growth rings appear older than they actually are. Moreover, the younger growth rings have higher Cd/Ca values typical of waters of southern origin suggesting changes in the deep water mass composition.

Stable isotopes of carbon also indicate rapid events such as YD and earlier abrupt events, but they give results that are apparently contradictory. In the mid-latitude Atlantic, a clear reduction of $\delta^{13}C$ in benthic foraminifera and an increase in Cd/Ca during YD is consistent with a replacement of NADW by a water mass of southern origin as would be expected following a break-down of the thermohaline circulation (Boyle and Keigwin, 1987; Keigwin et al., 1991). Such changes are registered also earlier during glacial time (Keigwin et al., 1994). Lower values of benthic $\delta^{13}C$ and surface $\delta^{18}O$ are found in North Atlantic sediments during Heinrich event 4 (35 kyr BP) suggesting a reduction of the THC caused by meltwater (Vidal et al., 1997; Zahn et al., 1997). However, $\delta^{13}C$ from other locations are seemingly inconsistent with a large-scale reduction of the THC during the YD: a site in the southern Atlantic (Charles and Fairbanks, 1992) and locations in the northern Atlantic (Jansen and Veum, 1990; Veum et al., 1992, Sarnthein et al., 1994) do not show negative anomalies of $\delta^{13}C$ during YD. This would suggest that there was no major reduction of the THC. Generally, identification of a relatively short event such as YD (duration about 1200 yr) is difficult in marine cores because of limitations in resolution and difficulties in sufficiently precise dating.

Polar ice cores permit direct reconstruction of the evolution of the three most important greenhouse gases after water vapor, CO_2, methane, and N_2O (Raynaud et al., 1993). CO_2 increases by about 15 ppm before the two large D/O events 8 and 12 in Fig. 5 (Stauffer et al., 1998), similarly in timing as the warming events A1 and A2 found in Antarctic ice cores (Blunier et al., 1998). Slow warming in the southern ocean starting about 1.5 kyr before the abrupt warming in the north would lead to a reduction CO_2-solubility and could explain the increase of 15 ppm. One may speculate that the slow warming in the south is initiated by the Heinrich events in the north that bring the THC to a full collapse (see below). CO_2 is apparently not influenced by the subsequent series of shorter D/O events, and no indication of warming is found in the Antarctic ice cores during those events.

CO_2 is also an indicator against a global extent of cooling during YD, because CO_2 is steadily increasing during the event (Fig. 4). Results of simulations with a prognostic physical-biogeochemical climate model (Marchal et al., 1999a) are consistent with this reconstruction and suggest that the strong cooling in the northern North Atlantic and, subdued, throughout the northern and tropical southern hemisphere (Lehman and Keigwin, 1992b; Kennet and Ingram, 1995; Thompson et al., 1995) was compensated by a warming in the Southern Ocean.

5. North–South Connections

5.1. The perspective of palaeoclimate archives

The best evidence for interhemispheric connections comes from the ice core records from Greenland and Antarctica which are put on a common timescale by methane synchronisation (Blunier et al., 1998). The longest D/O events (number 8 at 36 kyr and number 12 at 45 kyr BP) in the Greenland ice cores have a correlative in the Antarctic ice cores (Fig. 5). These events are characterized by millennial warming in the south lasting for about 2–3 kyr and are interrupted roughly at the time when the north switches from a cold to a warm climate on a time scale of only a few centuries or less. Cooling then starts in the south as well as in the north, although it is significantly slower in the north. A plausible scenario is that the south responds to a rapid resumption of the Atlantic THC. The warming in the north then initiates glacial melting or an increased meridional transport of freshwater through the atmosphere which again slowly reduces the THC. This hypothesis should now be checked with the earlier D/O events 14, 21 and possibly 23 (Dansgaard et al., 1993).

The ice core record shows strong north–south coupling also during the last deglaciation superimposed on the warming (Jouzel et al., 1995; Sowers and Bender, 1995; Blunier et al., 1997). Warming continued in the south since about 18 kyr BP while it was suppressed in the north, probably due to Heinrich event 1 and the associated meltwater discharge (McCabe and Clark, 1998) (Fig. 4). Note that the north-south pattern here is very similar to that of D/O events 8 and 12, where the south registers warming about 2 kyr before the abrupt warming in the north. At 14.5 kyr BP abrupt warming occurs in the north and the long-term trend reverses in the south (Antarctic cold reversal). At the time YD starts in the north, the Antarctic cold reversal stops and warming resumes in the south. The cold reversal in the south thus coincides with the gradual cooling during the Bølling/Allerød period. The seesaw behaviour manifests itself for a third time at the termination of YD: warming in the south stops when the north warms within a few decades (Fig. 4).

Climatic changes in these Antarctic ice cores evolve on a much longer timescale than those in the Greenland ice cores. There is, however, one important exception: the isotopic record of Taylor Dome, an ice core from the coastal zone of Antarctica, shows one abrupt warming at around the time of the Bølling warm transition at 14.5 kyr (Steig et al., 1998). The timescale chosen by Steig et al. (1998) implies that the complete warming, registered at Taylor Dome, took significantly longer than in the Greenland cores (about 1 kyr) and was completed only by 14.0 kyr BP, about 500 years later than in the north, if the time scale is correct. More high-resolution isotopic analyses and CH_4 measurements during the glacial part of the Taylor Dome ice core, and comparison with results from other Antarctic ice cores are required in order to obtain a clearer picture of possible regional expressions of abrupt climate change in Antarctica.

The atmospheric concentration of radiocarbon, $^{14}C^{atm}$, is another, indirect indicator for changes in the THC. High-resolution reconstructions across the YD based on varved marine sediments show an increase of $^{14}C^{atm}$ at the beginning of YD (Hughen et al., 1998) consistent with a complete shut-down of North Atlantic deep water formation as simulated by models (Stocker and Wright, 1996). Also, the end of YD is marked by a significant and rapid decrease of atmospheric ^{14}C which would be indicative of a rapid turn on of the ventilation of the deep ocean.

5.2. The model perspective

Several modes of the THC have been realised in models (Manabe and Stouffer, 1988; Stocker et al., 1992a; Rahmstorf, 1995; Schiller et al., 1997) and it has been shown that the THC is very sensitve to freshwater flux perturbations (Maier-Reimer and Mikolajewicz, 1989; Stocker and Wright, 1991; Wright and Stocker, 1993; Mikolajewicz and Maier-Reimer, 1994; Rahmstorf, 1994; Manabe and Stouffer, 1995; Stocker and Wright, 1996; Manabe and Stouffer, 1997; Fanning and Weaver, 1997). These models range from simplified, low-order models such as zonally averaged ocean circulation models to coupled, three-dimensional atmosphere–ocean models. But so far, attempts to simulate an entire series of abrupt climatic events have been rather limited (Paillard and Labeyrie, 1994; Stocker and Wright, 1998; Stocker, 1999). These circulation modes can be classified according to their different export rates of NADW into the Southern Ocean (Table 2).

The following discussion of the deep circulation, its changes and their effect on temperature and geochemical tracers is based on results we have obtained using a zonally averaged, three-basin ocean circulation model (Wright and Stocker, 1992), coupled to an atmospheric energy balance model (Stocker et al., 1992b) and a recently developed prognostic biogeochemical module that represents the global carbon cycle (Marchal et al., 1998b). The model is in close agreement with three-dimensional models both with regards to dependencies on fundamental parameters (Wright and Stocker, 1992) as well as the basic dynamical balances (Wright et al., 1998).

Strong overturning is generally associated with strong export of NADW (Fig. 6a) when overturning is shut down (Fig. 6b) there is no export of NADW into the Southern Ocean. The strength of NADW export is controlled by surface heat and freshwater fluxes (Stocker et al., 1992). Furthermore, there are modes with similar export of NADW but different primary convection sites such as the Greenland-Norwegian-Iceland (GIN) Sea and/or the Labrador Sea in the North Atlantic or locations further south in the North Atlantic (Lehman and Keigwin, 1992a; Weaver et al., 1994; Fichefet et al., 1994; Rahmstorf, 1995; Seidov et al., 1996). Because water mass and heat exchange with the Southern Ocean are linked, only those mode switches for which the export of NADW into the Southern Ocean changes, should affect the climate in the southern hemisphere. If, instead, changing convection sites in the North Atlantic has little impact on water export to the Southern Ocean, then the heat balance in the south is not expected to change much.

Temperature differences between strong THC and collapsed THC exhibit a clear north-south antiphase relation with strong cooling in the northern North Atlantic and compensatory warming in the south (Fig. 7). This signal is also imprinted in the atmosphere. An active THC in the Atlantic transports heat from the southern ocean northward (Crowley, 1992), today at the rate of about 0.2×10^{15}W. In consequence, a reduction of the THC leads to a cooling in the north due to a decrease in meridional heat transport and to a corresponding warming in the south. The amplitude of the warming in the south depends on the degree of reduction of the North Atlantic THC (Marchal et al., 1999a,b). Qualitative agreement is found with results from much more complex models Manabe and Stouffer (1988), but the amplitude of the temperature changes in the present model are larger by about a factor of two. The larger heat capacity of the Southern Ocean compared with that of the North Atlantic results in changes in the south that are of smaller amplitude and evolve on longer time scales. For a simple estimate, consider a volume of water of the size of the southern ocean (55–70°S, 4 km depth) with a heat capacity of about 5×10^{23}J/K. If the meridional heat flux out of the Southern Ocean (0.2×10^{15}W) stopped due to the collapse of the THC, a heating rate in the Southern Ocean of about 1.6°C per century (a zonal mean) would result (Stocker, 1998), if all else is unchanged. During abrupt events, in which the THC in the Atlantic is fully shut down, the climate system thus has the potential to operate like a "bipolar seesaw" (Broecker, 1998) (Fig. 8). If the THC remains essentially active and switches only between different convection sites in the North Atlantic (GIN Sea, Labrador Sea, LGM type), the

Table 2
Possible modes of the thermohaline circulation ordered according to their location of deep water formation in the North Atlantic (LGM denotes Last Glacial Maximum at around 21 kyr BP). It is argued that only switches to and from the mode in column four (no deep water formation) are registered in Antarctic climate records because of changes in the heat exchange between Southern Ocean (SO) the Atlantic Ocean. Switches between modes in the left three columns alone are only registered in the Atlantic region

	Location of deep water formation in the North Atlantic			
	GIN Sea	Labrador Sea	LGM type	None
Climatic phases	Holocene, Bølling phase, interstadials after H warm states of D/O events	8200 yr event (probable), cold states of D/O events	LGM circulation, cold state of D/O events, probably weakly stable	YD, during the strong H-events, before D/O events 12 and 8
N–S coupling	Strong heat water exchange between Atlantic SO, SO is cooled by THC		Weak coupling, intermediate Atlantic–SO heat water exchange	No heat and water exchange between Atlantic and SO, SO is not cooled by THC
SST northern North Atlantic	Warm	Intermediate	Intermediate	Cold
Models	Most ocean and climate models under modern conditions, Fig. 6a	Not well resolved in most models, transitions possible	Present under LGM conditions, reduced export of NADW to south, forced by weak freshwater pulse	Forced by strong meltwater pulse, natural mode, Fig. 6b

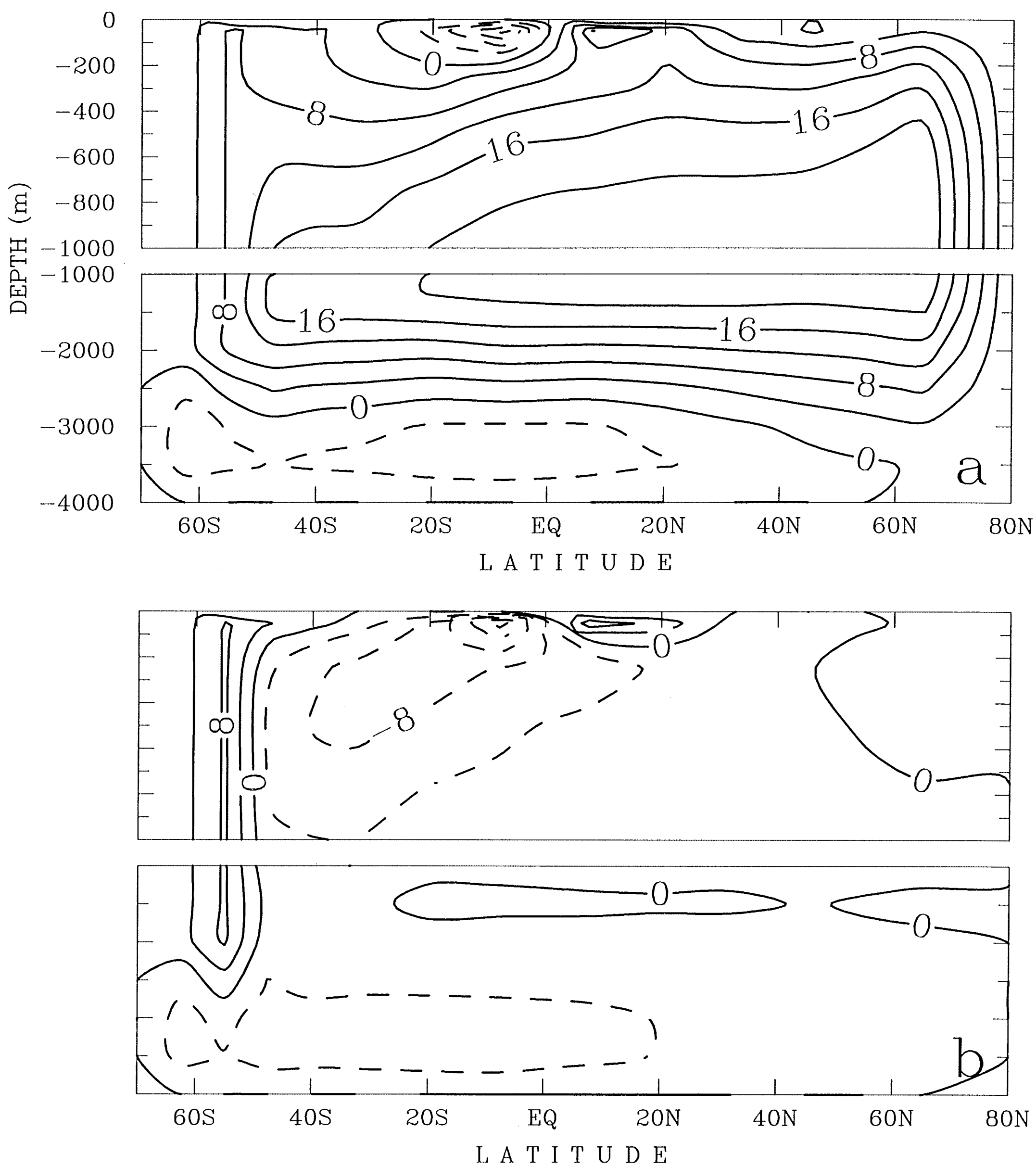

Fig. 6. Two different states of the meridional overturning streamfunction (in Sv, 1 Sv = $10^6 m^3 s^{-1}$) in the Atlantic (Stocker and Wright, 1996). (a) strong overturning is characteristic of the present-day deep circulation in the Atlantic. More than 10 Sv of North Atlantic Deep Water (NADW) are transported into the Southern Ocean and influence the water mass properties significantly on a global scale. (b) fully collapsed Atlantic THC following a polar halocline catastrophe. NADW formation is shut down, but the Atlantic is still weakly ventilated from the southern hemisphere.

south is not much affected and the two hemispheres are decoupled.

The simple mechanical analogy of a bipolar seesaw does not address the important question of the driver of the observed changes. There are two possibilities that, unfortunately, cannot yet be distinguished based on the evidence from the paleoclimatic record. According to the traditional interpretation of the paeloclimatic reconstructions, the center of action is in the high latitudes of the Atlantic Ocean (Fig. 8a). Perturbations (e.g. meltwater from the continental ice sheets) trigger transitions between different states of the THC. Through the effect of the meridonal heat flux in the Atlantic, the amplitudes of climate signals are large in the region around the North Atlantic and are weak and of opposite sign in the south. This is the typical inter-hemispheric seesaw forced in the northern North Atlantic region. There is a second possibility (Schmittner et al., 1999). Perturbations originate in the tropical region e.g. hydrological cycle feeding into the Hadley cell, heating of

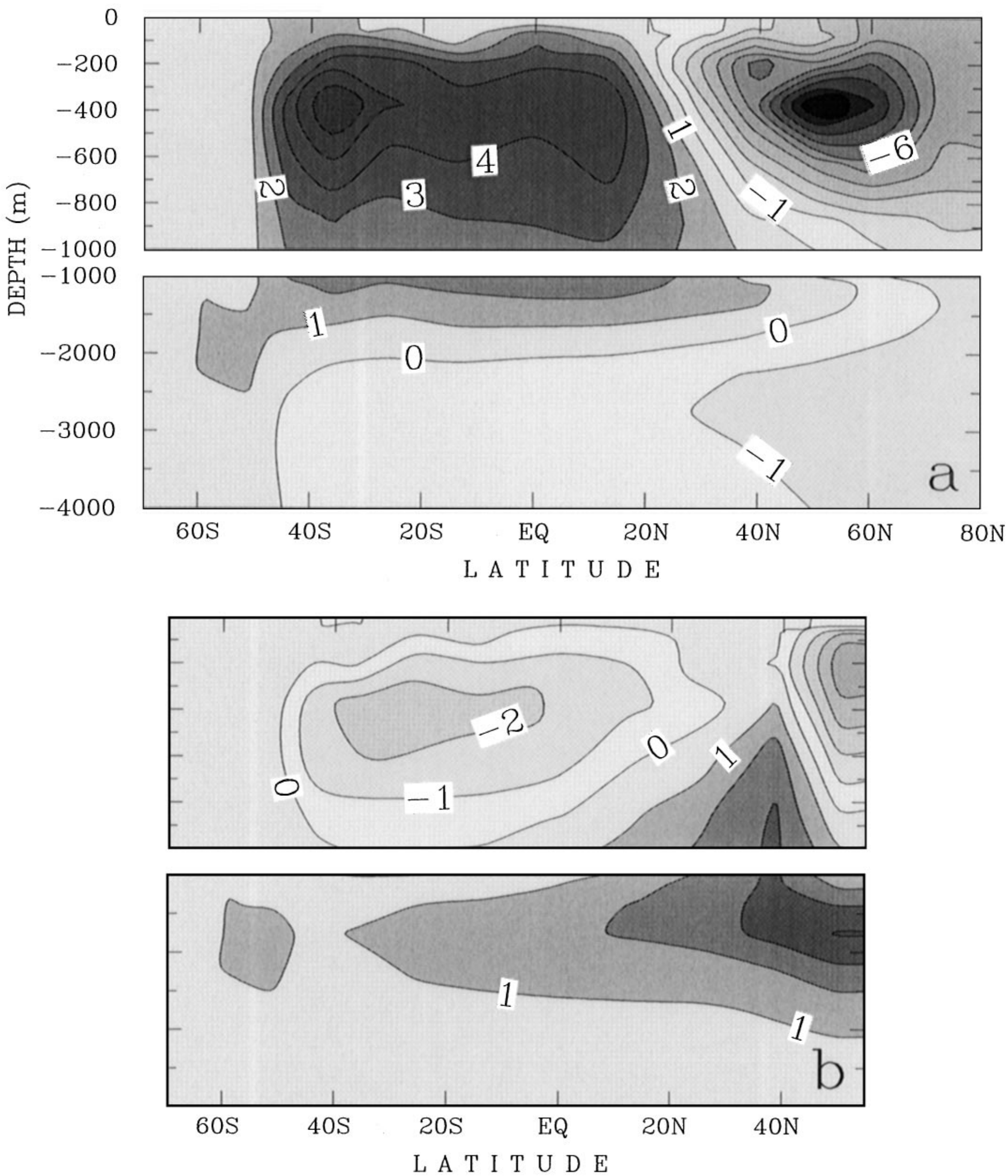

Fig. 7. Temperature differences for the transition between the two states of the Atlantic THC in Fig. 6. (a) In the Atlantic, the cooling is limited to the area north of 20°N, while a compensatory warming occurs south of it. There is a clear anti-phase relationship between the two hemispheres. (b) In the Pacific, changes are smaller in magnitude with a slight warming at depth and cooling in the intermediate waters. (Based on simulations by Stocker and Wright, 1996).

large areas of the equatorial ocean) and trigger transitions between different states of the thermohaline circulation in the North Atlantic (Fig. 8b). The effects would be essentially identical to (a). It is therefore rather difficult to distinguish between these two types based on a few paleoclimatic archives. The THC thus only responds and acts as an amplifier of climate changes. This is analogous to a seesaw, whose motion is forced by changes of the position of the hinge point. Finally, perturbations in the tropical region lead to parallel changes in northern and southern hemispheres. Here, the seesaw is locked and changes are synchronous and in phase everywhere (Fig. 8c). The results from ice cores rather exclude this last possibility for the abrupt events during Younger Dryas and D/O events 8 and 12 (Blunier et al., 1998).

Why did the seesaw operate only during D/O events 12 and 8 and during the Bølling/Allerød-YD transition? Before each of these phases, strong H-events occur (H5, H4, H1). Experience from model simulations suggests the following scenario. The associated discharge of icebergs into the North Atlantic may have shut down the THC, leading to a warming in the south and a moderate

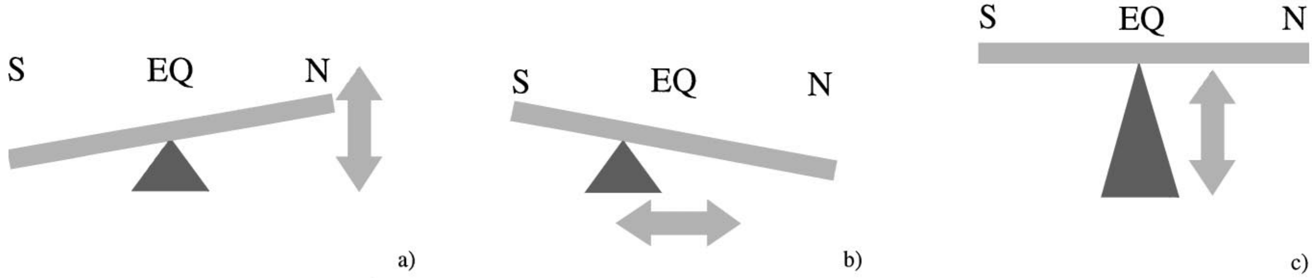

Fig. 8. Simple mechanical analogy of three different types of behaviour of the climate system as a response to perturbations (green arrows). (a) Perturbations in the northern high latitudes drive the seesaw. (b) Perturbations in the tropical region drive the seesaw. (c) Perturbations in the tropical region are synchronous, in-phase changes in northern and southern hemispheres. (From Stocker, 1998).

increase of atmospheric CO_2. Once the major iceberg discharge is over and the meltwater has been mixed away by near-surface processes (wind-driven circulation, air–sea fluxes), the THC resumes in the Atlantic. The south is now cooling again and CO_2 starts to decrease. The switch-on of the THC is rapid because of a positive feedback mechanism: starting convection brings to the surface saline and warmer water which cools rapidly when exposed to the surface re-enforcing convection (Wright and Stocker, 1991). The active THC warms the high latitudes and enhances the hydrological cycle bringing the precipitation, which is necessary to build up the ice sheets again. At the same time, the enhanced transport of moisture would also slow down the THC gradually. If the reduction of the THC is not complete, the climatic influence remains limited to the northern hemisphere: the series of shorter D/O-events is starting. After about 8 kyr the cycle repeats itself. This is probably the time required to condition the ice sheets for another major discharge resulting in an H-event. A further process that may be important is the interaction between changes in sea level and the stability of ice shelfs.

A full simulation of this hypothetical sequence of processes has not yet been presented. However, there are results from two recent three-dimensional, coupled atmosphere–ocean general circulation models that support this mechanism. In Manabe and Stouffer (1995) the circulation is perturbed with a short freshwater spike into the North Atlantic which leads to only a partial shut-down of the Atlantic THC. In the North Atlantic region, a pronounced cooling of more than 4°C is simulated. The export of NADW into the southern ocean decreases only by about 40%, and a temperature response in the southern high latitudes is not registered. This is an example of one of the shorter D/O events. In contrast, a stronger freshwater perturbation results in a full collapse of the THC (Schiller et al., 1997). The deep exchange flow with the southern ocean actually reverses, and the high southern latitudes are warming. This is an example of a long D/O event following an H-event. The model also indicates regional patterns of the climate response associated with changes in convection sites.

On the other hand, the model experiments of Manabe and Stouffer (1997), in which the THC does not fully collapse, cooling is simulated around Antarctica. Strong changes in convection and sea ice extent are associated with these changes. This result indicates, that conclusions based on models, as far as they concern areas distant from the centers of action (e.g. the North Atlantic) must still be interpreted with caution.

Further support for the seesaw hypothesis during abrupt change comes from recent physical-biogeochemical model experiments (Marchal et al., 1998a; Marchal et al., 1999a,b). Using a zonally averaged ocean circulation model, coupled to a prognostic carbon cycle component, it is possible to give a first estimate on the contribution of changes in the water mass distribution, biological production and gas exchange to the overall changes in $\delta^{13}C$ as a function of location (latitude, depth). Based solely on a water mass argument for the interpretation of $\delta^{13}C$ in the Atlantic as given by Fig. 9, it would be concluded that NADW decreases in the north Atlantic below 1 km depth ($\Delta\delta^{13}C < 0$), but that NADW increases in the south Atlantic ($\Delta\delta^{13}C > 0$). Clearly, this model is too coarse to resolve specific spatial responses of $\delta^{13}C$ to changes of the THC, but it does indicate that the simple water-mass interpretation can be problematic.

5.3. Open questions

Coupled atmosphere-ocean models respond to meltwater perturbations by a reduction or a complete collapse of the Atlantic THC. The response is immediate and depends strongly on the rate and amplitude of the perturbation (Stocker, 1999). This is apparently at odds with the paleoclimatic record: there is a lag of about 1 kyr between the first major rise of sea level during the Bølling/Allerød and the onset of YD (Bard et al., 1996). All current climate models would respond with an immediate shut-down of the THC to a freshwater flux of

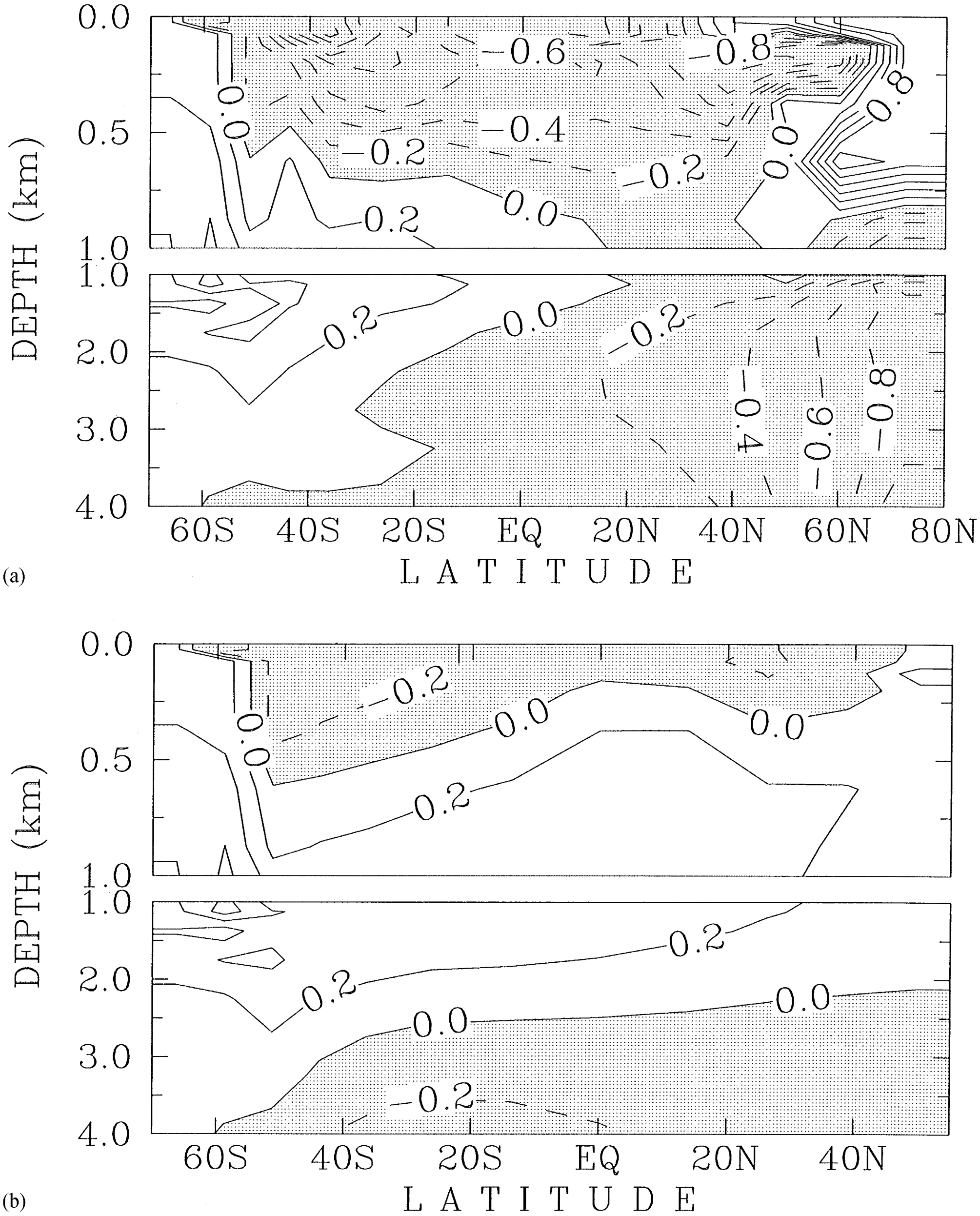

Fig. 9. Differences of $\delta^{13}C$ in the Atlantic (a) and Pacific (b) Oceans between the state with a collapsed (Fig. 6b) and with an active (Fig. 6a) THC as calculated with a zonally averaged physical–biogeochemical model including a prognostic carbon cycle. Negative values of $\Delta\delta^{13}C$ (shaded) would be classically interpreted as a reduction of NADW, while white areas would indicate an increase of the proportion of waters from southern origin. Note that if $\delta^{13}C$ is taken simply as a water mass indicator, statements regarding water mass changes may be misleading and inconclusive. (From Marchal et al., 1998a).

such magnitude if applied close to the convection regions in the North Atlantic. For the present generation of models, this seems a robust result which could be indirect support for the hypothesis that most of that sea level rise probably originated in Antarctica (Clark et al., 1996).

A second problem is associated with atmospheric radiocarbon. While model simulations indicate that $^{14}C^{atm}$ is highest just before the end of YD, the record shows a continuous decrease in $^{14}C^{atm}$ already a few centuries after the beginning of YD (Hughen et al., 1998) implying that ocean ventilation resumed at least partially or that production of $^{14}C^{atm}$ decreased significantly. This could be explained by intermediate water formation in the North Atlantic shortly after a full shut-down of the deep circulation at the beginning of YD or point at other deep water formation sites (Broecker, 1998).

Although the simple picture of a thermohaline seesaw is plausible and the paleoclimatic record is in many aspects consistent with this mechanism when the biggest abrupt changes are considered, there is no direct evidence available for a strong link between changes in either hemisphere and the ocean circulation (Boyle, 1999). There is still a lack to synchronise with sufficient accuracy marine sediment cores, terrestrial records and ice cores. Although the stable isotope of oxygen offers a possibility (Sowers et al., 1993), the long cycling time of this tracer (about 2000 years) precludes synchronisation on time scales shorter than a millennium, i.e. the time scales required to determine the phasing during D/O events.

For ice cores, the situation is better with regards to synchronisation. However, the locations of polar and tropical ice cores are almost singular points on the Earth's surface, and therefore information is very patchy. This is why a combination of circumstantial indicators and physical–biogeochemical models (e.g. Marchal et al., 1999a or Marchal et al., 1999b) is needed to complement the information from high-resolution paleoclimatic archives.

6. The past as a window to the future

The combination of evidence from the paleoclimatic record and a wide range of model simulations suggests that the coupled atmosphere–ocean system has thresholds beyond which large-scale reorganisations of the deep ocean circulation can occur. The system is driven across these thresholds either by slow change of the background state or perturbations. During the glacial, such perturbations were most likely due to meltwater from the last great ice sheets on the American and European continents, and possibly also from Antarctica. For a few centuries to millennia they modified the surface freshwater balance in the North Atlantic and forced shut-downs or shifts in the locations of deep water formation.

During the early Holocene, these ice sheets have disappeared and abrupt climate change of the magnitude typical for the glacial has no longer been observed. Nevertheless, changes in the surface freshwater balance are not impossible in a future climate that is warmed by increased levels of greenhouse gases, and there is reason for concern (Broecker, 1997). A warmer atmosphere carries more moisture: model simulations suggest that the meridional transport of moisture would increase (Manabe and Stouffer, 1994; Stocker and Schmittner, 1997). This could lead to an increase in precipitation thereby reducing sea surface salinities particularly in regions of deep water formation. The ocean would then respond with a reduction of the thermohaline circulation in the Atlantic. This process has been found in an entire suite of models (IPCC, 1995), all of which exhibit a decline in the overturning strength, although its magnitude is strongly model-dependent. There are first indications from the GIN Sea that surface-to-deep exchange of water has been interrupted since about 1980 (Bönisch et al., 1997). However, it is not clear whether this is a local phenomenon as part of long-term natural variability, for instance of the North Atlantic Oscillation (Dickson et al., 1996) or whether this has larger-scale implications.

Up to now, estimates of future uptake of CO_2 by the ocean have generally assumed that the ocean circulation does not respond to the warming. However, this assumption is no longer warranted for simulations extending over more than a few decades. A modification of the circulation and the associated vertical mixing rates will influence the uptake of anthropogenic carbon as well as the supply of bio-limiting nutrients to the surface layers. A second effect of ocean circulation changes is related to CO_2 solubility: warming of the surface ocean leads to a decrease in solubility and outgassing, thereby further increasing the atmospheric CO_2 concentration. This would be a positive feedback on the warming. Model simulations suggest that these feedbacks will become important when the THC will shut down completely (Joos et al., 1999).

It must be clearly noted that the present models are useful in furthering our understanding of non-linear processes in the climate system, but they are not predictions. First, the spread of results between different models is still too large to consider the amount of reduction of the THC for a doubling of a CO_2 a robust result — estimates of the decrease in overturning range from 10 to about 50% for a doubling of CO_2 (IPCC, 1995). Second, the prediction of threshold values is per se difficult, because it is an inherently non-linear problem which is linked to the small-scale processes of atmosphere–ocean exchange, deep water formation and convection. These are still poorly modelled or crudely parameterized in ocean models (Marshall and Schott, 1999), and serious uncertainties remain. An important lesson is that the non-linearities in the atmosphere–ocean system give rise to irreversible changes even though perturbations may not be permanent.

7. Conclusions

Abrupt climate change has been the rule rather than the exception during the last glacial period. The last abrupt event happened about 8200 years ago (Alley et al., 1997b) when still some ice melting on the American continent was underway (Licciardi et al., 1998). Ocean circulation, in particular, the thermohaline circulation of the Atlantic, and its changes have played a crucial role in amplifying these changes and transmitting them to other areas of the globe. A mechanism for interhemispheric coupling involves the export of NADW to the Southern Ocean which influences the heat balance in this region. Paleoclimatic records suggest that this mechanism operated during specific events of the last glacial period and deglaciation.

Many open questions remain. The proposed sequence of events needs to be investigated in more well-dated, high-resolution archives from the southern hemisphere, especially from the Southern Ocean. Temperatures at the sea surface and at intermediate depths would be the desired indicators (Labracherie et al., 1989; Vidal et al., 1999). Paleoclimatic records need to be synchronised, because a correct determination of leads and lags requires an accuracy of better than 500 years. Wiggle matching must no longer be used for paleoclimatic records from different geographical regions unless the proxy is global. Difficulties remain with radiocarbon dating because phases of abrupt change are likely to coincide with strong modifications of the radiocarbon clock and surface reservoir ages (Bard et al., 1994; Stocker and Wright, 1996).

Much progress has been made by combining palaeoclimatic data and models in order to quantiatively interpret the records. More improvement is needed on the formulation of boundary conditions and simplified atmospheric modules for ocean models such that they can be utilized for palaeoclimatic simulations over many 1000 years. A special effort is needed to better constrain the timing, the location and amount of freshwater discharge into the ocean, because it is the primary agent to trigger abrupt changes during the glacial and thus needed as a forcing function for these models. The goal is to simulate entire series of D/O and H events. Quantitative modelling of these events crucially depends on this information which is required before the modelling of entire sequences of D/O events can be undertaken.

The lessons from past changes are clear. As continuing global warming will change the hydrological cycle, a further mechanism emerges that modifies the surface freshwater balance and the thermohaline circulation. It should be a priority to better understand the factors and processes that influence the surface freshwater balance — evaporation minus precipitation minus runoff — in critical areas such as the deep water formation sites of the world ocean. Included in such an effort must be continued, long-term observation of the hydrological cycle on a global scale, intergovernmental commitments for long-term oceanographic observation networks, and observation and analysis of changes in the water mass structure of the intermediate and deep ocean, in order to detect early signs of sustained thermohaline circulation changes in the near future.

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