Chapter 9

RAPID TRANSITIONS OF THE THERMOHALINE OCEAN CIRCULATION

A Modelling Perspective

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ABSTRACT

This chapter discusses the stability of the Atlantic thermohaline circulation with special emphasis on the critical thresholds and state transitions found in model experiments. The thermohaline ocean circulation is a major heat transport mechanism which causes the relatively mild climate in the North Atlantic region (including Europe) in the modern times. The formation of North Atlantic Deep Water and the associated large-scale meridional transports in the Atlantic are maintained by a positive salinity feedback (first identified by Stommel in 1961). A second positive feedback is responsible for the tendency of deep convection to reoccur in the same regions. These two feedbacks are the main reason for the non-linear behaviour of the thermohaline ocean circulation found in models; their characteristic processes, time and length scales are discussed. Simulations of plausible circulation changes during the last glacial maximum and due to future greenhouse warming are presented.

1. INTRODUCTION

The conditions that determine the climate of our planet are ever changing on all time scales. The output of the sun, the Earth's orbit, the distribution of continents, the chemical composition of the atmosphere, the elevation and vegetation cover of the land surfaces, the extent of ice cover and many other factors are variable. It is not possible to find a direct “analogue” for a future climate in the past; history never repeats itself. Therefore there is only one way in which the past can provide a window to the future: we have to dissect and understand the mechanisms of past climatic changes, and then put them together again in models. Reconstructions of past climatic changes, their spatial patterns and their timing provide hypotheses about possible mechanisms which can be examined in specific model experiments. Comprehensive climate models need to be tested on past climates; only if past climatic changes can be understood and simulated in models can we make confident projections into the future.

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This paper discusses some aspects of this process, focussing on the Atlantic thermohaline circulation. It summarises the understanding of mechanisms of circulation changes gained from a hierarchy of models, and it reports on an early attempt to test a coupled atmosphere-ocean-sea ice model on a past climate that is radically different from the present. The same model is then used for a glimpse into possible futures.

A highly simplified cartoon of the Atlantic circulation is shown in Fig. 1. The observed winter sea ice distribution is one indication of the heating effect of this circulation, visible directly on satellite images. Another indication is the strong warm anomaly centered over the Nordic Seas shown in Fig. 2. Simple heat budget calculations (based on heat transport values derived from hydrographic measurements, Roemmich and Wunsch 1985) as well as coupled climate models (Manabe and Stouffer 1988; Ganopolski et al. 1998b) show that this warm anomaly can indeed be explained by the heat transport of the thermohaline circulation.The magnitude of this anomaly gives an indication of the severe climatic impact that *changes* in the thermohaline circulation could have. Such circulation changes most likely play a role in the abrupt swings seen for example in the past climate records preserved in the Greenland Ice Sheet (Dansgaard et

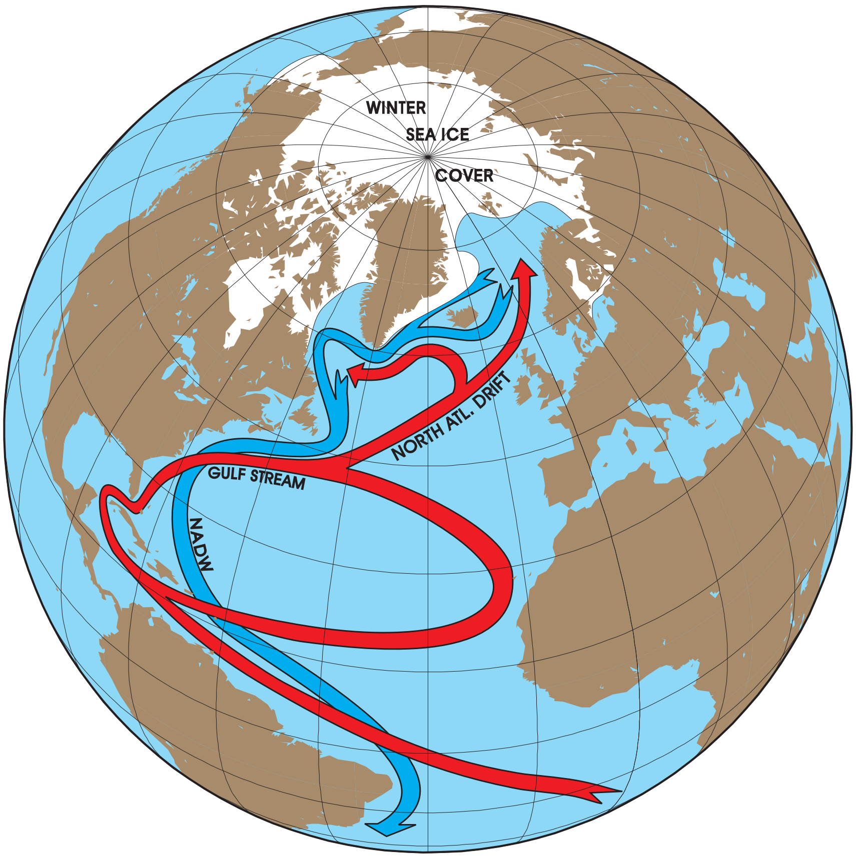

Figure 1. Highly simplified cartoon of the Atlantic circulation: darker shading shows surface currents, lighter grey North Atlantic Deep Water. The warm North Atlantic Current heats north-western Europe, pushing back the winter sea ice margin. From Rahmstorf 1997.

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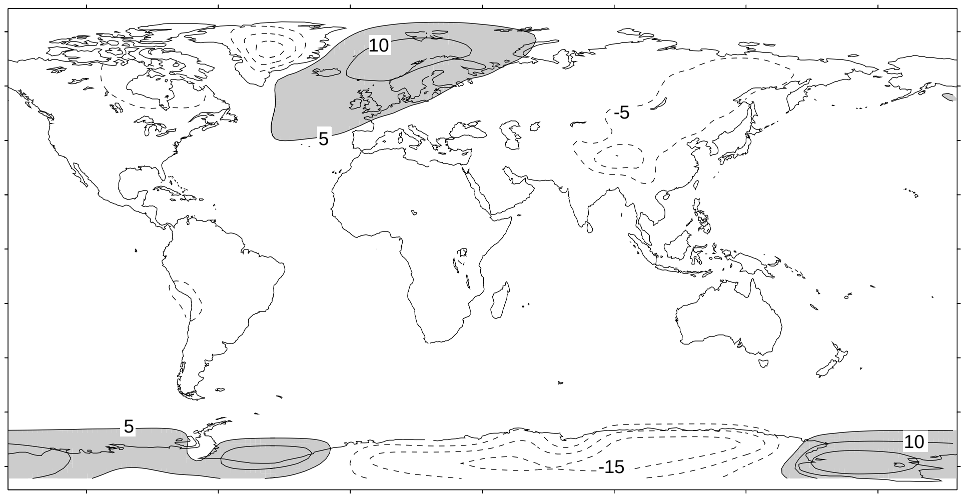

Figure 2. Deviation of the observed annual mean surface air temperature (NCAR air temperature climatology) from its zonal mean. Contour interval 5°C; regions more than 5°C warmer than the zonal average are shaded. From Rahmstorf and Ganopolski 1999.

al. 1993) and in oceanic sediments (Lehman and Keigwin 1992). The reason why the ocean is a prime suspect for some of the erratic behaviour of the climate system lies in the peculiar stability properties of the thermohaline circulation which are discussed in the next sections. For more comprehensive reviews of thermohaline circulation stability, see Weaver and Hughes 1992 and Rahmstorf, Marotzke, and Willebrand 1996.

2. POSITIVE FEEDBACK AND MULTIPLE EQUILIBRIA

The prime reason for the non-linear behaviour of the ocean circulation is the existence of positive feedback mechanisms. We know of two major positive feedbacks which affect the large-scale thermohaline circulation: an *ad*vective and a *con*vective feedback.

- The *advective feedback*: the thermohaline circulation advects salty water northward in the Atlantic, this enhances salinity and density in the north, which in turn keeps the thermohaline circulation going (Stommel 1961; Bryan 1986).
- The *convective feedback*: convective vertical mixing continually removes freshwater from the surface in areas of net precipitation; it thus prevents the formation of a fresh light surface layer which would inhibit convection (Welander 1982; Lenderink and Haarsma 1994).

Both feedbacks tend to reinforce an existing circulation pattern and help to maintain it once it is going. This makes it possible that several different circulation patterns are stable, i.e. multiple equilibrium states of the circulation can exist. The most well-known and dramatic example is that climatic states with and without deep water formation in the North Atlantic (sometimes called conveyor belt 'on' and 'off' states) are both found to be stable in models - e.g., in the coupled ocean-atmosphere circulation model of Manabe and Stouffer 1988. This is a consequence of the advective feedback,

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and the resulting two stable equilibria were first described in Stommel's (1961) simple box model. Some models further indicate that the convective feedback may lead to stable states with different convection patterns in the North Atlantic, e.g. with or without convection in the Labrador Sea (Rahmstorf 1995b; Rahmstorf 1995a). How relevant this is for the real ocean remains to be tested.

From Fig. 2 it is clear that a transition between different modes of ocean circulation can alter the heat transport and lead to major regional climate changes. It can even have a global effect, as a change in sea ice extent changes the planetary albedo. Because of the different feedbacks, different types of transition can be distinguished in model experiments.

3. ADVECTIVE SPINDOWN

The mode transition associated with the advective feedback is called an advective spindown and was found in the GCM experiments of Bryan 1986. For an explanation, Stommel's (1961) classic box model can be extended to cross-hemispheric flow to make it applicable to the Atlantic (Rahmstorf 1996), and it yields a simple stability diagram of the thermohaline circulation. This shows the equilibrium thermohaline flow rate m as function of the freshwater input F_1 into the North Atlantic (Fig. 3, dotted line), which is given by the quadratic equation:

$$m^2 + k\alpha(T_2 - T_1)m + k\beta S_0 F_1 = 0$$

Here $(T_2\text{-}T_1)$ is the temperature difference between the northern and southern Atlantic boxes, which is the prime driver of the flow; α and β are thermal and haline expansion coefficients, S_0 is a reference salinity and k an empirical constant. For a derivation and detailed discussion see Rahmstorf 1996. The circulation has a saddle-node bifurcation (S in Fig. 3), i.e. a critical threshold of how much freshwater input the circulation can sustain. The dashed line shows how the circulation in the box model responds to a slow, linear increase of the freshwater input F_1 with time. When the critical threshold is exceeded, an advective spindown of the circulation occurs on a time-scale of centuries. This stability behaviour agrees surprisingly well with state-of-the-art general circulation models (Rahmstorf 1995a; Rahmstorf 1996); the solid line was obtained when the same experiment as with the box model was performed with a global GCM. Note that the GCM locates the present climate in a region of the stability diagram where two stable equilibria exist, with NADW formation "on" or "off" (to the left of the origin only the "on" equilibrium exists). The reason for this bistable behaviour is that the circulation is driven by cooling while freshwater input acts as a brake. This contrasts with Broecker's (1991) concept of an evaporation-driven "conveyor belt", which balances the net evaporation from the Atlantic by importing freshwater (and which would be monostable, in the left half of Fig. 3). Table 1 summarises the two different views of the freshwater budget of the Atlantic. Recently, Weijer et al. 1999 have presented some observational support for a net southward freshwater export from the Atlantic by the thermohaline circulation.

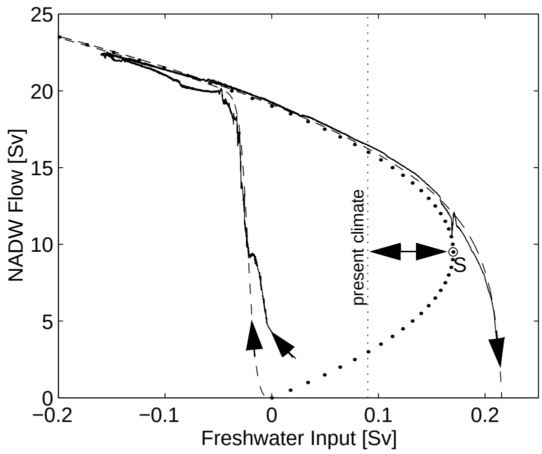

Figure 3. Hysteresis curve of a simple box model (dashed) and a global circulation model (solid). The dotted parabola is the *equilibrium* solution of the box model with the saddle-node bifurcation point *S*. The hysteresis curves were obtained in transient runs by slowly increasing and then decreasing the freshwater flux into the North Atlantic, at a rate of 0.05 Sv per 1,000 yr. The initial state of the circulation model, obtained by a spinup with present-day climatological forcing, is marked as 'present climate'. Arrows show how close to the critical threshold the present climate is, according to this model estimate. The location of the present climate on the stability diagram is model dependent; the diagram itself is more robust. (Figure from Rahmstorf 1996.)

4. CONVECTIVE INSTABILITY

The mode transition associated with the convective feedback is called a convective instability. In contrast to the advective spindown this is a very fast process, leading to circulation changes on a time scale of a decade or less. This is a mechanism which could explain some of the abrupt climate changes seen in the ice core record, e.g. the Younger Dryas event. There are two types of convective instability: a basin-wide ('polar halocline catastrophe') and a local one. A basin-wide convective instability interrupts all deep water formation in the ocean basin and leads to a rapid collapse of the thermohaline circulation. A localised convective instability shuts down convection just in one area and

Table 1: Proposed freshwater budgets of the Atlantic north of 30°S, ignoring the Bering Strait contribution. Freshwater input into the Atlantic is given a positive sign. The term wind-driven gyre refers to the subtropical gyre of the South Atlantic. Note that there is net evaporation from the Atlantic in both cases, but only in Broecker's concept does it drive (and is balanced by) the thermohaline circulation

Broecker's (1991) budget	
net evaporation	-0.35 Sv
thermohaline circulation	+0.35 Sv
Rahmstorf's (1996) budget	
net evaporation	-0.2 to -0.3 Sv
wind-driven gyre	+0.3 to +0.35 Sv
thermohaline circulation	-0.1 to -0.05 Sv

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leads to a rearrangement of convection patterns without shutting down the large-scale circulation; for example, it could cause a shift of convection from the Greenland Sea to a location south of Iceland. Although not as drastic as a complete shutdown of the circulation, shifts in convection pattern can also have a major effect on climate (Rahmstorf 1994).

Table 2: Overview over properties of the two instability mechanisms relevant to the Atlantic ocean circulation

	Advective Spindown	Convective Instability
Time Scale	gradual (~ 100 y)	rapid (~ 10 y)
Mechanism	large-scale advection	local convection physics
Cause (forcing)	basin-scale heat and freshwater budget	local forcing in convection region(s)
Effects	conveyor winds down	shift of convection locations or complete breakdown of conveyor
Equilibria	conveyor “on” or “off”	several equilibria with different convection patterns
Modelling	modelled quite well by climate models	large uncertainty in forcing and response

Table 2 summarises the properties of the two mechanisms of transition, and a schematic stability diagram including possible transitions is presented in Fig. 4. It should be noted that the circulation changes discussed do not have their ultimate cause in the ocean but occur in response to a change in external forcing. The positive oceanic feedbacks involved make this response highly non-linear and can strongly amplify the reaction of the climate system to gradual and subtle forcing changes. They thus provide a mechanism that can translate gradual forcing, such as the ‘orbital forcing’ due to the slow changes in the Earth’s orbit, into rapid and strong climatic swings when certain thresholds are crossed.

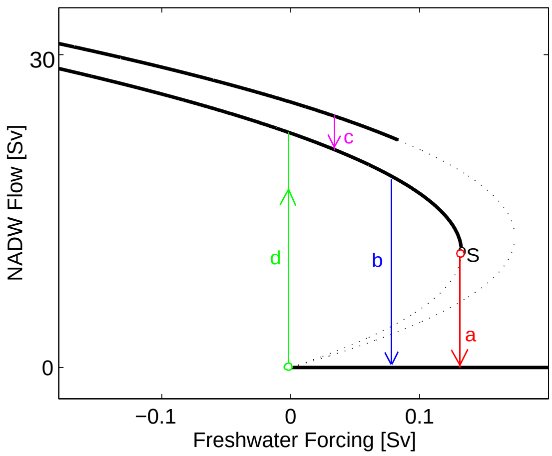

Figure 4. Schematic stability diagram of the Atlantic thermohaline circulation. The two upper heavy branches indicate the possibility of multiple states with different convection sites. Possible transitions indicated are: (a) advective spindown, (b) polar halocline catastrophe, (c) convective transition, (d) start-up of NADW formation. From Rahmstorf 1999; see Stocker and Wright 1991a for an earlier, similar diagram.

5. OCEAN CIRCULATION AND CLIMATE OF THE LAST GLACIAL

Recently it has become possible to simulate the climate and atmospheric and oceanic circulations of the last glacial maximum (LGM) with a numerical model (Ganopolski et al. 1998c). It takes several thousand years for the climate to equilibrate (due to the slow diffusive time scale of the thermohaline ocean circulation), and such a simulation has not yet been attempted with a coupled general circulation model (GCM) as this would require an excessively long computation time. Another problem is that flux adjustments at the air-sea interface (still used in many GCMs) cannot be justified for a climate very different from the present. Ganopolski et al. therefore used a coupled climate model of intermediate complexity, CLIMBER-2, which can be run for 5,000 model years within a few days and which does not require flux adjustments.

The atmospheric component is a dynamical-statistical atmosphere model with 10° latitudinal and 51° longitudinal resolution. It explicitly resolves the large-scale circulation but does not resolve individual synoptic weather systems. The vertical structure includes a planetary boundary layer, a free troposphere (including cumulus and stratiform clouds) and a stratosphere. The ocean component is a zonally averaged model with three separate basins (Atlantic, Indian and Pacific Oceans) similar to the one used by Stocker and Wright 1991b, with parameterisations of the vorticity balance and of Ekman transport. It includes a thermodynamic sea ice model which predicts the sea ice fraction and thickness for each grid cell, with simple treatment of advection and diffusion of sea ice.

In spite of its relative simplicity the model reproduces the large-scale features and seasonal cycle of the present climate quite well (Petoukhov et al. 1998). For the simulation of glacial climate, the annual cycle of insolation was changed (reflecting the change in the Earth's orbit), the atmospheric CO_2 level was lowered to 200 ppm, and continental ice sheets were prescribed (and consequently the sea level lowered by 105 m and global salinity increased by 1.0 ‰) to reflect conditions 21,000 y before the present. Ultimately one would like to simulate rather than prescribe CO_2 and ice sheets, as these are not external causes but part of the internal dynamics of climate change.

After the model reaches equilibrium with the LGM boundary conditions, a glacial climate is simulated which is on average 6.2°C colder (surface air temperature) than the modern (pre-industrial) climate. The strongest cooling (up to 30°C) is over the North Atlantic and Europe in winter; in summer there are two slightly weaker maxima over the ice sheets of the northern continents. The cooling over the continental ice sheets is a test mainly of ice-albedo feedback and agrees with the results from the atmospheric GCMs participating in the PMIP model intercomparison project (Joussaume and Taylor 1995), as well as being consistent with ice core data from Greenland. Sea surface temperatures predicted by the model are in good agreement with alkenone data (E. Bard, personal communication).

The most interesting aspect in this context is the change in ocean circulation simulated by the model (Fig. 5). In the Atlantic, deep water formation retreats to the south as sea ice advances, the outflow of deep water moves to a shallower depth and bottom water of Antarctic origin pushes northward and fills the deep Atlantic. Similar changes have been deduced for the LGM from sediment cores (e.g. Duplessy and Maier-Reimer 1993). This can be seen as a different convective state of the circulation, and one may speculate that interstadials could be caused by a switch to a circulation state more like the present, with convection north of Iceland.

To study the effect of the ocean circulation changes on the LGM climate, another LGM experiment was performed in which the present-day ocean heat transport was kept fixed in the model. Global cooling for the LGM in this experiment was only 4.7°C. This shows that the changes in the ocean circulation enhanced global cooling by 30% in the model; in the Northern Hemisphere the cooling was even enhanced by 50%. This was due

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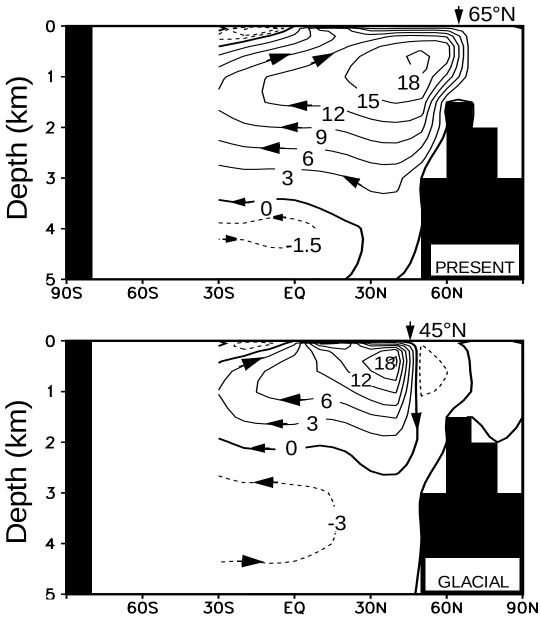

Figure 5. Meridional transport stream function in the Atlantic for the present climate (top) and the Last Glacial Maximum (bottom), as simulated by the coupled CLIMBER-2 model (Ganopolski et al. 1998c).

to the Atlantic ocean's heat transport reaching not as far north; the maximum of the heat transport, situated at 20°N, remained unchanged at 1.1 PW. Locally over the North Atlantic the difference between the runs with and without ocean heat transport changes exceeded 20°C, even though the *rate* of overturning hardly changed (Fig. 5). This illustrates the dramatic effect that shifts in ocean circulation can have on the surface climate.

6. OUTLOOK FOR THE FUTURE

An advective spindown of the Atlantic thermohaline circulation over the next centuries occurs in some global warming scenarios (Manabe and Stouffer 1993; Rahmstorf 1997; Stocker and Schmittner 1997). More recently, Wood et al. 1999 found a shut-down of Labrador Sea convection early in the next century in their simulation of greenhouse warming.

We have used the CLIMBER-2 model to study a climate change scenario (Fig. 6a) in which the CO_2 concentration of the atmosphere increases as observed until the present and then follows a high emission scenario (IPCC IS92e, Houghton et al. 1995) to the year 2100. Concentrations peak in the year 2150 at 3.3 times the present level and then decline as the fossil fuel era comes to an end. We studied several scenarios in which we artificially varied the "hydrological sensitivity" of the model, i.e., the amount of freshwater input into the northern North Atlantic resulting from the warming of the Northern Hemisphere (Rahmstorf and Ganopolski 1999). The intention is to account for the uncertainty in the hydrological cycle of climate models, both due to uncertainty in atmospheric vapour transport changes and in the amount of meltwater runoff.

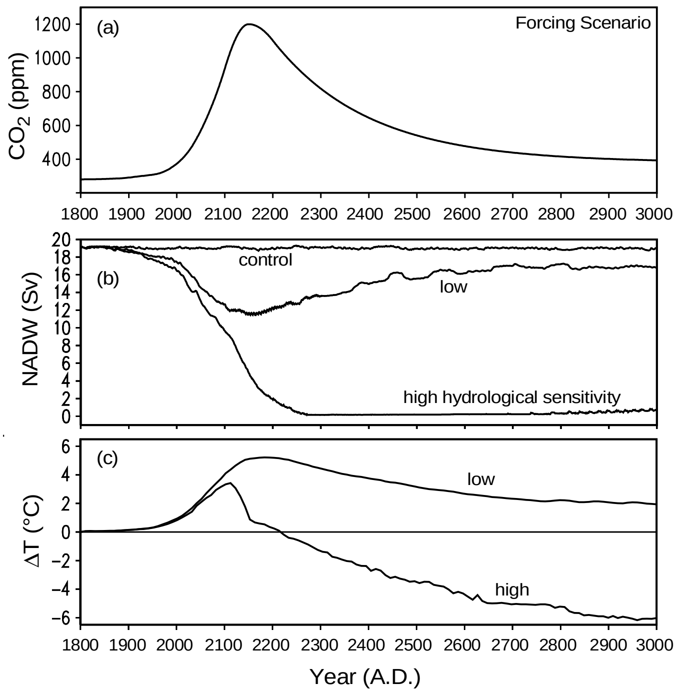

Figure 6. Long-term simulation of global warming scenarios with the CLIMBER-2 model (Rahmstorf and Ganopolski 1999), for low and high hydrological sensitivity. (a) Prescribed atmospheric CO_2 forcing. (b) Maximum of the Atlantic overturning stream function (i.e., formation rate of NADW). (c) Mean air temperature over the North Atlantic between 50°N and 60°N (deviation from the preindustrial climate).

Fig. 6b shows the response of the Atlantic thermohaline circulation for a “low” value (corresponding to the unchanged CLIMBER-2 model and no glacier runoff) and a “high” value (roughly corresponding to the Manabe and Stouffer 1994 GCM) of the hydrological sensitivity. For the low sensitivity the thermohaline circulation declines from 19 Sv to 12 Sv but then recovers. For the high sensitivity it collapses completely and does not recover for several thousand years; the anthropogenic peak in CO_2 concentration has shifted the climate system into a new stable state. While the global mean temperature is hardly affected by the change in ocean circulation, Fig. 6c shows that it leads to a major cooling over the northern North Atlantic in the simulation.

Future progress in modelling paleoclimates may develop in several directions. Further climatic processes will be included in climate system models, such as vegetation feedback (e.g. Ganopolski et al. 1998a), the dynamics of continental ice sheets and biogeochemical cycles. Specific climatic proxies, such as the oxygen isotope composition of snow, could be simulated by models. The regional resolution of the models will increase; enhanced computer power will mean that coupled ocean-atmosphere GCMs will increasingly be used for long-term paleoclimate simulations. Improved models will put increasing demands on the quality of the paleodata, in terms of accuracy, reliability and spatial and temporal coverage, so that the models can be validated and hypotheses about specific mechanisms can be tested.

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