Abrupt Climate Changes: How Freshening of the Northern Atlantic Affects the Thermohaline and Wind-Driven Oceanic Circulations

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Key Words

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Abstract

Leading hypotheses for abrupt climate changes are focused on the ocean response to a freshening of surface waters in the north Atlantic. The degree to which such a freshening affects the deep, slow thermohaline, rather than the shallow, swift, wind-driven circulations of the ocean, and hence the degree to which that freshening affects climate in high rather than low latitudes, differ from model to model, depending on factors such as the treatment of diffusive processes in the oceans. Many comprehensive climate models are biased and confine the influence mainly to the thermohaline circulation and northern climates. Simulations of paleoclimates can provide valuable tests for the models, but only some of those climates provide sufficiently stringent tests to determine which models are realistic.

1. INTRODUCTION

Over the past few decades, the surface waters of the northern Atlantic have freshened at a relatively rapid pace (Dickson et al. 2002, Curry & Mauritzen 2005) and the heat transported northward across 25°N by the thermohaline ocean circulation[1] has decreased (Bryden et al. 2005). The freshening could increase further, should global warming increase high-latitude precipitation, accelerate the melting of glaciers, or suddenly release icebergs into the ocean (Ekstrom et al. 2006). This is of major concern as it is believed that freshwater discharges in the North Atlantic have played a significant role in past abrupt climate changes by reorganizing the thermohaline circulation (e.g., Broecker 1990, Rahmstorf 2002, Clarke et al. 2003, Alley et al. 2003). Are abrupt climate changes associated with a freshening of the northern Atlantic imminent? What form will they take? What additional parameters should be monitored to anticipate possible changes?

Answers to these questions are presumably available in the huge number of modeling studies devoted to the consequences of a freshening of the northern Atlantic (Rahmstorf 1995; Manabe & Stouffer 1995, 1999; Rind et al. 2001; Stouffer et al. 2006; and references therein). However, different models give different results, even in response to the same freshening. For example, some models have only a very modest tropical response to a freshening of the northern Atlantic (see Stouffer et al. 2006). The complex coupled ocean-atmosphere climate models that are used to anticipate the climate changes associated with future global warming, have a significant tropical response, as seen in **Figure 1a**. In this particular model (Stouffer et al. 2006), hosing 1.0 Sv (1 Sv = 1×10^6 m³/s) of freshwater uniformly over the North Atlantic between 50°N and 70°N caused a shutdown of the thermohaline circulation and induced the sea surface temperature shown in the figure. This broad hosing region is chosen to assure that the location of deep convection in different models lies below the freshening, thus allowing model intercomparison (Stouffer et al. 2006). (The initial equilibrium state of the model corresponded to a simulation of the climate of today. The freshening was maintained for 100 years. **Figure 1a** shows averaged conditions over the last 20 years of the simulation.) The lower sea surface temperatures in the high latitudes of the northern Atlantic accompanied a significant cooling over northern Europe. The warm surface waters in the equatorial and southern tropical Atlantic is, in this and a few other models, associated with a global displacement of rainfall patterns in low latitudes (Vellinga & Woods 2002, Zhang & Delworth 2005). Are the oceanic trends observed in the North Atlantic over the past few decades a prelude to such abrupt climate changes?

For an answer to this question, we need a realistic climate model, but, given that different models give different results, how do we decide which is the realistic model? Simulations of observed phenomena are tests for the models, but the interpretation of

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[1]There is much controversy as to the exact definition of the thermohaline circulation. (See Wunsch 2002 and Rahmstorf 2006). Here we distinguish between the wind-driven circulation, which is confined to the upper ocean, in and above the thermocline, and the thermohaline circulation, which extends from the surface to the ocean floor and is driven by the surface fluxes of heat and freshwater.
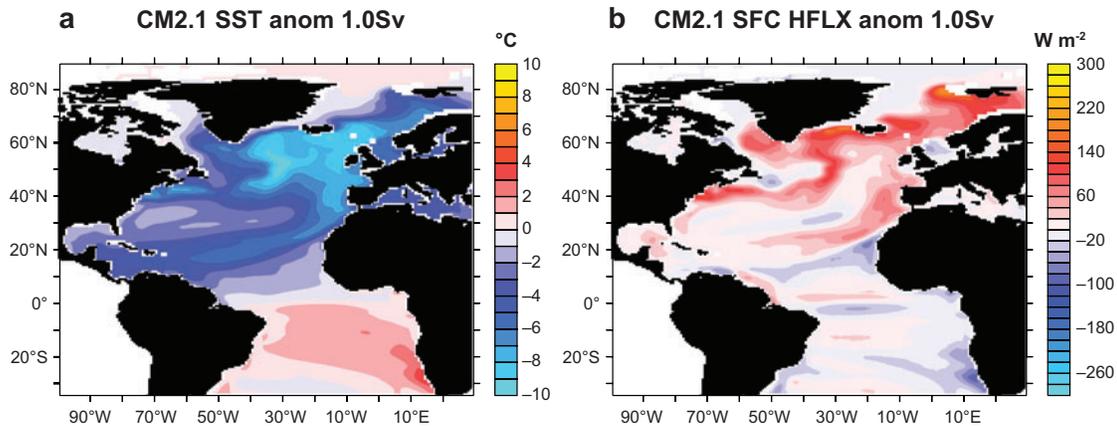
Figure 1

Experiment where 1.0 Sv of freshwater is added uniformly in the North Atlantic between
50°N–70°N in the CM2.1 coupled model. (*a*) Changes in sea surface temperature, and
(*b*) changes in the net surface heat flux. The positive anomalies in the high latitudes represent a
large decrease in the heat loss. The freshwater was added for 100 years, which represents a
global increase of sea level height of about 9 m. This water flux is within the range believed to
have occurred in events of meltwater release during the last glacial period and the deglaciation,
though for shorter periods (Clarke et al. 2003).

the observations can generate controversies. Consider, for example, climate records
for the past 50000 years extracted from the glaciers over Antarctica and Greenland.
Some scientists claim that the climate fluctuations in the two polar regions are linked
by the oceanic thermohaline circulation, with changes in Greenland lagging be-
hind those in Antarctica (EPICA Community Members 2006). This could be a test
for oceanic models that simulate the thermohaline circulation, provided the forcing
functions that cause changes in that circulation are known. A possible forcing func-
tion is a freshening of the northern Atlantic, but, unfortunately, critical information
concerning the location, timing, and intensity of the freshening is lacking. Hence,
a stringent test for the models is unavailable. To complicate matters, the statement
that the fluctuations in the two polar regions are related is questionable on statistical
grounds (Wunsch 2006).

The dispute over possible connections between climate fluctuations in Greenland
and Antarctica will remain unresolved until we have realistic models. In the (tempo-
rary) absence of suitable tests for the models, progress is possible by improving certain
features of the models, for example, increasing their spatial resolution, and enhancing
their parameterization of subgrid processes, such as turbulent mixing. These are the
factors that cause different models to simulate the thermohaline circulation differ-
ently and that cause different models to give different results in response to the same
freshening of the northern Atlantic.

The thermohaline circulation receives enormous attention, but several questions
concerning that circulation are still unanswered. Its most puzzling aspect is the fate
of the surface waters that sink into the deep ocean in high latitudes. Where does that
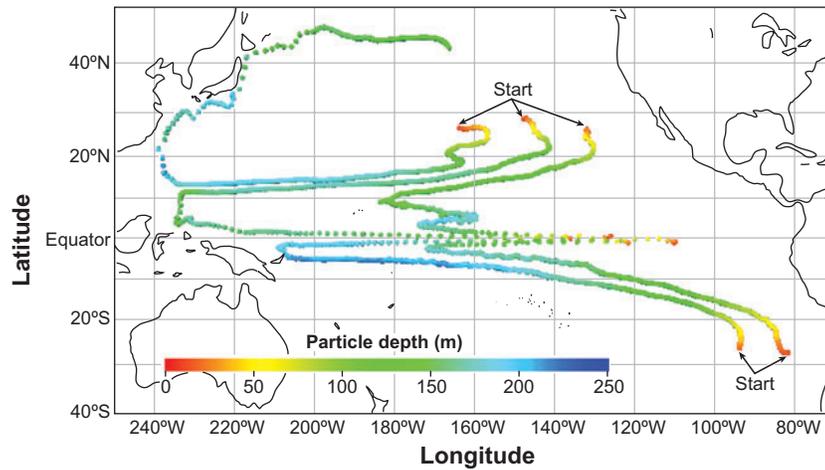
Figure 2

The tropical wind-driven circulation. The paths of water parcels over a period of 16 years after subduction off the coasts of California and Peru as simulated by means of a realistic oceanic general circulation model forced with the observed climatological winds. From the colors, which indicate the depths of parcels, it is evident that parcels move downward, westward, and equatorward unless they start too far west off California, in which case they join the Kuroshio current (off Japan). Along the equator they rise to the surface while being carried eastward by the swift Equatorial Undercurrent. Adapted from Gu & Philander (1997). Reprinted with permission from AAAS.

water return to the upper ocean? In many models, the cold, deep water rises uniformly throughout the mid- and low latitudes so that the maintenance of the thermocline depends on a balance between the downward diffusion of heat and the upwelling of cold water. The validity of this explanation for the thermocline is doubtful because the downward diffusion of heat necessary to warm up the cold water as it rises is far greater than measurements of oceanic mixing processes indicate. Apparently strong oceanic mixing is confined to a few small regions, for example, over ridges on the ocean floor (Ledwell et al. 1993, 2000). This result prompted some scientists to develop theories for the (ventilated) thermocline that neglect mixing processes. These are models of the wind-driven circulation[2] that build on Stommel's original two-dimensional model for the Gulf Stream by introducing variations in the third (vertical) dimension. Surface water subducts in certain subtropical regions and then moves on surfaces of constant density without any changes in temperature and salinity. The water rises back to the ocean surface, either in high latitudes or in an upwelling zone in low latitudes. **Figure 2** shows computer simulation results that plot the trajectories of water parcels, and the changes in their depths, over a period of 16 years after subduction.

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[2]The textbooks Oce*an Circulation Theory* (Pedlosky 2004) and *Atmospheric and Oceanic Fluid Dynamics* (Vallis 2006) provide illuminating descriptions of the current understanding of the large-scale dynamics of the ocean, including the wind-driven and thermohaline circulations.
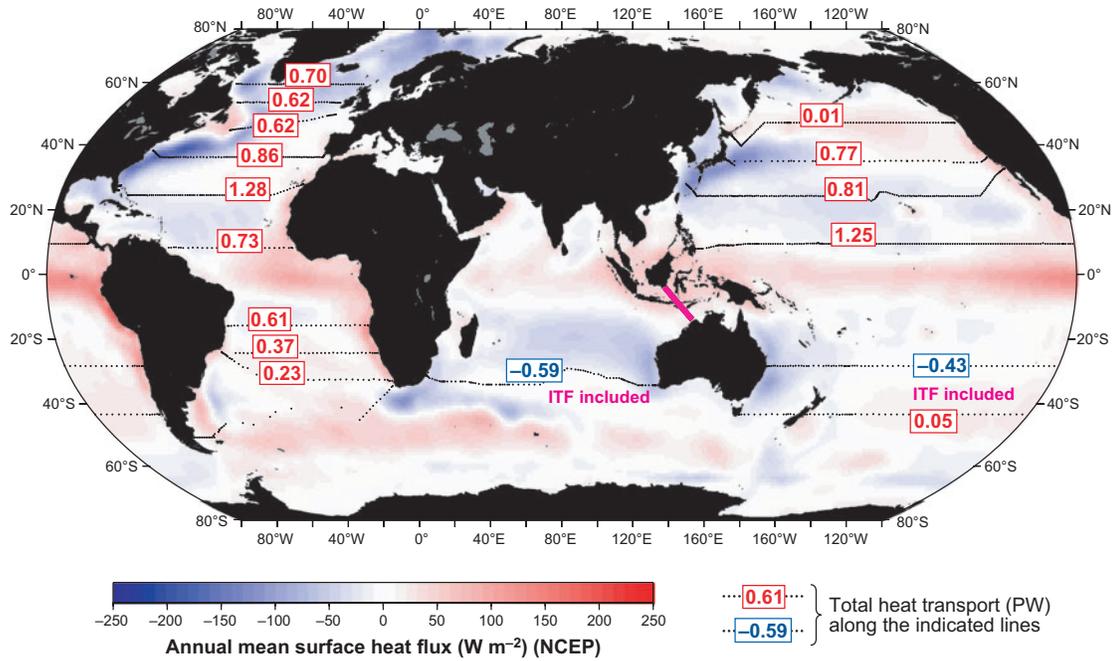
Figure 3

Annual mean net surface heat flux in W m^{-2} (calculated from NCEP reanalysis) and heat transports across the different latitudes. The ocean gains heat in the upwelling zones of low latitudes and loses heat in high latitudes. The Indonesian Throughflow (ITF) is indicated by a magenta line between Australia and Borneo. Adapted from Talley (2003).

In **Figure 3**, the ocean is seen to gain heat in the upwelling zones of low latitudes and to lose heat in higher latitudes, especially in regions where cold, continental air blows over warm ocean currents, such as the Kuroshio and Gulf Stream. The wind-driven gyres mainly effect the transport of heat from these regions of gain to the regions of loss. The common assumption (see Quadfasel 2005) that the thermohaline circulation is principally responsible for the oceanic heat transport overlooks the meridional overturning[3] component of the wind-driven circulation; that assumption is inconsistent with the finding that the oceanic heat transport depends mainly on currents in the upper ocean (Boccaletti et al. 2005).

The wind-driven circulation penetrates to a depth of a few hundred meters at most, as is evident in **Figure 2**. At greater depths, the oceanic thermal structure must therefore depend on processes separate from the directly wind-driven circulation,

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[3]A meridional overturning circulation (MOC) refers to motion in a north-south plane, downward in one location and upward in another. Most models of the thermohaline circulation involve sinking in the northern Atlantic, and rising motion everywhere else, but in reality it is not known where the cold water rises back to the surface. In the case of the wind-driven circulation, the water sinks in subtropical "subduction" zones, and rises in equatorial and coastal upwelling zones (see **Figure 4**).

processes associated with the deep thermohaline circulation. Hence the thermohaline and wind-driven circulations are complementary. Jointly they determine the oceanic thermal structure and satisfy the constraint of a balanced oceanic heat budget. That constraint determines the depth of the thermocline: The depth is such that the gain of heat in low latitudes balances the loss in higher latitudes (Tziperman 1986, Boccaletti et al. 2004).[4] This argument follows from **Figure 3**, which shows that, whereas the oceanic heat loss depends strongly on atmospheric conditions—cold air blowing over warm currents in winter—the heat gain depends on oceanic conditions, especially the depth of the thermocline in equatorial upwelling zones. In response to a change in the loss of heat, the ocean adjusts by changing the depth of the thermocline. When the loss is small, then, for the gain to be small, the thermocline has to be deep. The absence of cold surface waters from the eastern equatorial Pacific up to 3 mya implies a deep equatorial thermocline for a very prolonged period, and a very different oceanic heat budget than that of today (Fedorov et al. 2006b). What forcing at the ocean surface can induce such conditions?

Meridional overturning is a feature of both the thermohaline and wind-driven components of the oceanic circulation. Surface water sinks in high latitudes in one case, in subtropical subduction zones in the other case. A perturbation that increases the buoyancy of the surface waters inhibits the overturning in both cases. To shut down the meridional overturning component of the thermohaline circulation, the freshening of the surface waters has to exceed a certain threshold. There is similarly a threshold for shutting down the overturning of the wind-driven circulation, but the responses are strikingly different in the two cases: a cooling of high northern latitudes in the case of the thermohaline circulation and a deepening of the equatorial thermocline and a warming of low latitudes in the case of the wind-driven circulation (Fedorov et al. 2004, 2006a). The deep equatorial thermocline up to 3 mya could therefore have involved a freshening of the subtropics in the Pacific. Can a freshening of high northern latitudes in the Atlantic affect the equatorial Pacific?

Figure 3 shows that the Atlantic is a net importer of heat, whereas the Indo-Pacific is an exporter. The constraint of a balanced heat budget for the ocean, because it is global, implies that a perturbation in any one ocean can affect the others. Consider a freshening of the northern Atlantic that reduces its loss of heat to the atmosphere. (The sea surface temperature changes in **Figure 1***a* are associated with a similar reduction in **Figure 1***b*.) For a new equilibrium, that reduced loss in the northern Atlantic must be accompanied by a reduction either in the heat gained in the equatorial Atlantic or in the heat the Atlantic imports across 33°S. In the latter case, the heat budget of the Pacific is affected. That ocean, to export less heat, can either increase its loss in high northern latitudes or reduce its gain at the equator. (The Southern Ocean may also alter its heat transport in order to balance the changes in the Atlantic basin.)

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[4]Previous attempts to represent the thermohaline and wind-driven circulation in one coherent framework include that of Samelson & Vallis (1997) who propose a two-thermocline limit. In this picture, the adiabatic (diffusive) theories are appropriate in the upper (lower) part of the thermocline.
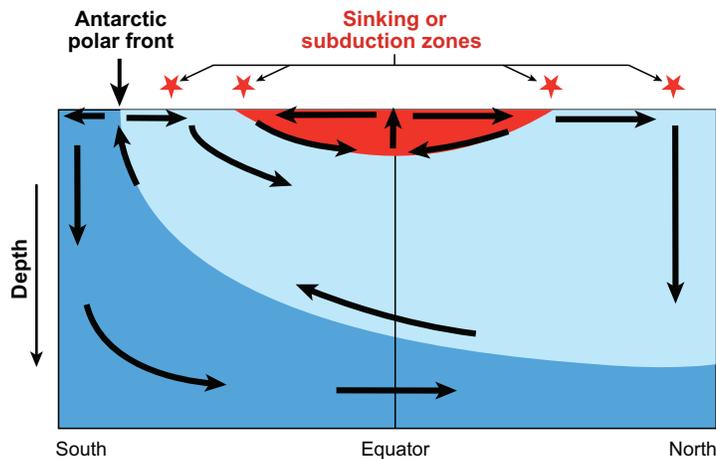
Figure 4

Schematic of the meridional ocean circulation. The two main components of the circulation are represented. The red region denotes the domain of the wind-driven circulation, which consists of upwelling at the equator, poleward Ekman drift, and subduction in the subtropics with a return flow in the upper 200 m. The thermohaline circulation has an upper branch where waters subduct in the North Atlantic, travel southward, and upwell in the Southern Ocean (*light blue*). The other branch consists of a bottom circulation where waters downwell in the southern polar region, travel northward, and join the deep waters to upwell in the Antarctic polar front (*dark blue*).

To explore these various possible responses requires a climate model that includes the Antarctic circumpolar current, an important connection between the Atlantic and Pacific oceans, and also an important link between the wind-driven and thermohaline circulations (Toggweiler & Samuels 1993, 1995). The winds over the Antarctic circumpolar current bring cold, deep water to the surface along the Antarctic polar front, whereafter some of the water flows southward and sinks near Antarctica, into the very deep ocean. The water that flows northward from the front subducts in the Polar Front Zone and Subantarctic Zone and subsequently determines the properties of water at intermediate depths, below the base of the directly ventilated thermocline, in the subtropics and tropics. This complex circulation, shown schematically in **Figure 4**, maintains a balanced heat budget for the ocean so that a change in the flux of heat across the surface in one region can affect the entire oceanic circulation.

A model appropriate for studies of the consequences of freshening the northern Atlantic has to reproduce the circulation in **Figure 4**. This is a challenge because of the vastly different adjustment times for the different components of the oceanic circulation. After an abrupt change in the surface forcing, the wind-driven circulation returns to a new state of equilibrium after a few decades, but the corresponding timescale for the global thermohaline circulation is on the order of a millennium. This means that the computer resources required to simulate changes in the thermohaline circulation are substantial. A compromise that permits simulations of the circulation over extended periods is to sacrifice spatial resolution in numerical models of the

ocean. In extreme cases, zonal (east-west) variations are neglected entirely, but more often the three-dimensional spatial grids are coarse. This brings us back to the oceanic diffusive processes because a model with a coarse numerical grid is necessarily highly diffusive and hence has unrealistic features.[5]

In any model, the relative importance of the thermohaline and wind-driven circulations in transporting heat poleward depends critically on the parameterization of diffusive processes in that model (Fedorov et al. 2006a.) The more diffusive a model—the coarser its resolution—the more important is the role of the thermohaline component, and the less important is that of the wind-driven circulation in transporting heat poleward. In such models, the gain of heat in equatorial upwelling zones is minimized—the oceans instead gain heat more uniformly throughout the tropics and subtropics—and perturbations that modify the heat budget have minimal impact on equatorial zones. In high-resolution models with low diffusion, on the other hand, changes near the equator can be far more pronounced. Simulations with such models can cover only limited periods of time, thus precluding a state of equilibrium for the thermohaline circulation. This trade-off must be kept in mind when evaluating the results from different models.

This review discusses the impact of a freshening of surface waters in different models. Section 2 shows how some of the results are qualitatively similar but quantitatively different. An analysis of the oceanic heat budget, in Section 3, provides new perspectives and a few surprises. The processes that determine the response of the tropics to different freshening patterns in the Atlantic and also the Pacific are discussed in Sections 4 and 5. The main results, and a discussion of issues that remain to be resolved, are the topics in Section 6.

2. THE MODELS

To study the consequences of a freshening of the northern Atlantic, scientists have used a broad range of tools, from realistic, coupled ocean-atmosphere, climate models to highly idealized models of the ocean forced at its surface with specified winds, heat, and freshwater fluxes. Because the available computer resources are limited, the realistic models can be used only for a small number of case studies. Fortunately, the idealized models are capable of reproducing some of the main features in the realistic climate models and can therefore be used to explore in detail the effects of changing certain parameters.

A model capable of realistic simulations of climate phenomena is the newly developed GFDL-coupled model CM2.1 (Delworth et al. 2006, Gnanadesikan et al. 2006). The oceanic component of this model is MOM4, which has 50 vertical levels and a horizontal resolution of $1.0° \times 1.0°$ longitude and latitude in the extratropics, increasing to $1.0° \times 0.3°$ degrees in the equatorial region. The atmospheric model

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[5]Heat and salt mixing in the ocean is driven by turbulence caused by tidal and wind forcing. In numerical models, finite differencing schemes that discretize the equations of motion can in effect contribute to mixing processes. Some schemes require physically unrealistic, explicit mixing to maintain numerical stability; others have implicit mixing.
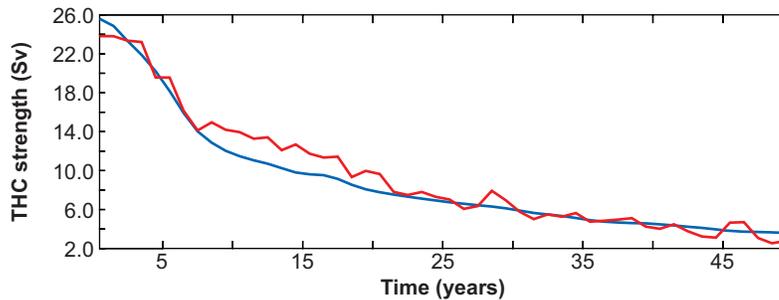
Figure 5

Strength of the North Atlantic Thermohaline Circulation (THC) in the experiments where 1.0 Sv of freshwater is uniformly added in the North Atlantic during 100 years. The red line shows the evolution in the CM2.1 coupled model and the blue line shows the evolution in the experiment using only the oceanic component of the coupled model. The strength of the THC is calculated as the maximum of the meridional volume streamfunction between 30°N-80°N.

has 24 levels in the vertical and a horizontal resolution of $2.0° \times 2.5°$. Because of the considerable computer demands of this model, it has been used in only two freshening experiments where the North Atlantic, between 50°N and 70°N, was freshened at rates of 0.1 Sv and 1.0 Sv, continuously for a period of 100 years (Stouffer et al. 2006). The results in **Figure 1** are for the higher value, which is above the threshold at which the meridional component of the thermohaline circulation shuts down. At what level is the threshold?

For an answer, we turn to a simplified model, the strictly oceanic component of the climate model, with specified surface forcing. The stand-alone ocean model is forced with observed climatological daily winds and monthly heat and freshwater fluxes. There is also a weak restoring with a 30-day timescale toward the monthly observed climatological sea surface temperature. **Figure 5** shows that this model captures changes in the intensity of the thermohaline circulation remarkably well. A comparison of **Figures 6** and **1** shows that the simulation of sea surface temperature amplitudes is less accurate, in part, because the ocean model has no sea ice, and the "atmospheric" temperature is fixed as a boundary condition. However, note that the strictly oceanic model reproduces the main features of the sea surface temperature patterns.

An oceanic model even simpler than the oceanic component of the climate model consists of two rectangular basins interconnected by a channel representing the Southern Ocean. The widths of the basins correspond to those of the real Pacific and Atlantic oceans. The resolution is $3.75° \times 3.75°$ in the horizontal, with 15 vertical levels. This model is forced with annual mean zonally uniform winds that vary in latitude and with mixed boundary conditions for temperature and salinity. **Figure 7** shows that this model reproduces an equatorial cold tongue, and an overturning circulation, that is reasonably realistic. On the other hand, the cold tongue is not nearly as intense or sharp as in reality, a consequence of the low resolution and associated high diffusion. In this model, the flux of heat across the ocean surface lacks the bull's-eye in the eastern equatorial Pacific seen in **Figure 3**. As a consequence the

Changes in sea surface
temperature in the
experiment where 1.0 Sv
of freshwater is added
uniformly in the North
Atlantic in the stand-alone
oceanic component of the
CM2.1.

OGCM SST anom 1.0Sv
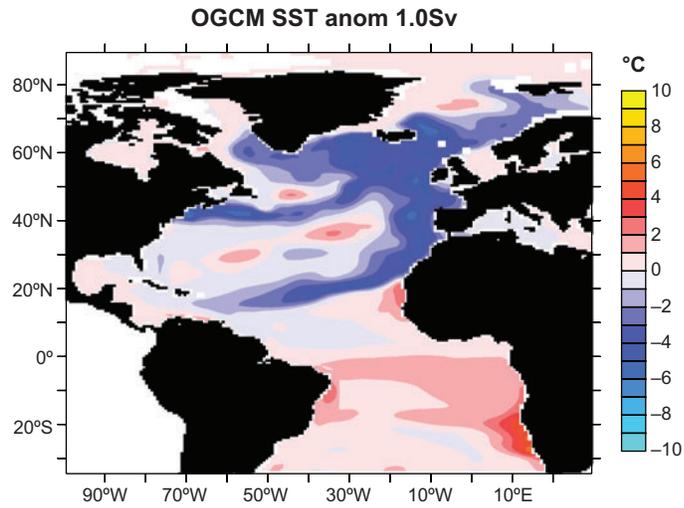
role of the thermohaline circulation, in transporting heat for example, is enhanced
relative to that of the wind-driven circulation.

Strictly oceanic models with specified forcing at the surface provide most of
the available information concerning the threshold that a freshening of high lat-
itudes has to exceed to shut down the thermohaline circulation. Do such models
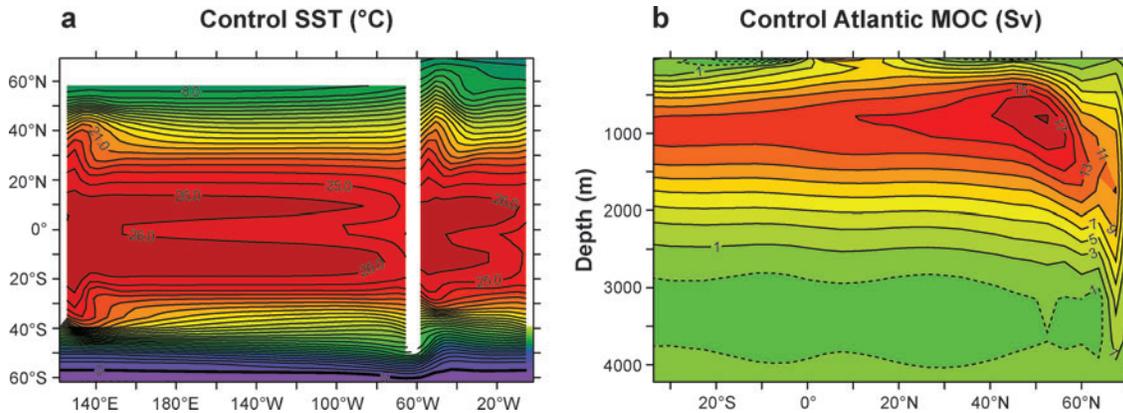
Figure 7

Control (unperturbed) state of the idealized ocean general circulation model: (*a*) SST (°C),
and (*b*) Atlantic meridional overturning circulation (MOC). Panel *a* shows that the model
reproduces equatorial cold tongues in the two basins, that is, regions of relatively cold waters
in the eastern equatorial oceans that extend westward. The contours in Panel *b* are the
streamfunction of the zonally averaged meridional flow (Sv). The positive values indicate that
the flow circulates clockwise, and the closer the isolines the faster the flow. The wind-driven
cells can be distinguished at the surface in the tropical region. There is also northward flow at
about 500 m that sinks in the North Atlantic and flows southward at about 1500–2000 m
(compare with **Figure 4**).

deny ocean-atmosphere interactions any role? One measure of the importance of those interactions is the difference between the results from strictly oceanic models (**Figure 6**), and results from the climate model (**Figure 1**, for example). For further exploration of the effects of ocean-atmosphere interactions, a useful model is the coupled atmosphere-ocean-sea ice model known as ECBILT-CLIO (version 3). The atmospheric component (ECBILT) is quasi-geostrophic, is global, has three vertical levels, and has a horizontal resolution corresponding to T21 (5.6° x 5.6°) (Opsteegh et al. 1998). ECBILT includes a simplified parameterization of radiation that uses prescribed climatological total cloud cover taken from the International Satellite Cloud Climatology Project. The hydrological cycle is closed over land by a land bucket model. The ocean–sea ice component (CLIO) is an ocean general circulation model with a free-surface, plus a thermodynamic/dynamic sea ice model (Goose & Fichefet 1999). CLIO has a horizontal resolution of $3° \times 3°$ and 20 vertical levels (13 levels in the upper 1000 m). The coupled model is run without heat flux correction, but includes a weak freshwater correction that reduces the precipitation over the northern Atlantic Ocean and redistributes the freshwater uniformly over the Pacific Ocean. In the ECBILT-CLIO, the response to an increase in atmospheric CO_2 and the response of the thermohaline circulation to a high-latitude freshening resemble that of more complex fully coupled atmosphere-ocean general circulation models (Stouffer et al. 2006, Gregory et al. 2005). On the other hand, this simplified climate model has such coarse resolution, especially in the ocean, that features of the wind-driven circulation, especially in low latitudes, are poorly represented. We return to this matter below.

3. SENSITIVITY TO THE RATE OF FRESHENING

3.1 Stratification of the Upper Ocean

Results from coupled ocean-atmosphere models show that, in response to a freshening of the surface, changes in the thermohaline circulation sometimes lead, and sometimes lag, those in sea surface temperature (Manabe & Stouffer 1995, Hall & Stouffer 2001). Under what conditions can the lead become a lag? What factors determine the response? These questions are important because the answer determines under what conditions we can predict abrupt climate changes by measuring the strength of the thermohaline circulation (Bryden et al. 2005). A factor that strongly affects the response is the rate of the freshening as is evident in **Figure 8**, which shows results for freshening at a rate of 1.0 Sv in **Figure 8*a***), and 0.1 Sv in **Figure 8*b***).

For large forcing, the sea surface temperature changes are large and are seen to lead the changes in the maximum intensity of the thermohaline circulation by several years. In the case of modest forcing, the changes in sea surface temperature are modest and practically in phase with the changes in the thermohaline circulation. The key difference is that hosing at a rapid rate stratifies the ocean to such a degree, and over such a large area, that it suppresses oceanic convection everywhere, thus decoupling the surface layers from those at greater depths. In the absence of convection, the surface layers are unable to gain heat from the deeper layers when cooled by the atmosphere, so that surface temperatures decrease sharply, as shown in **Figure 9**
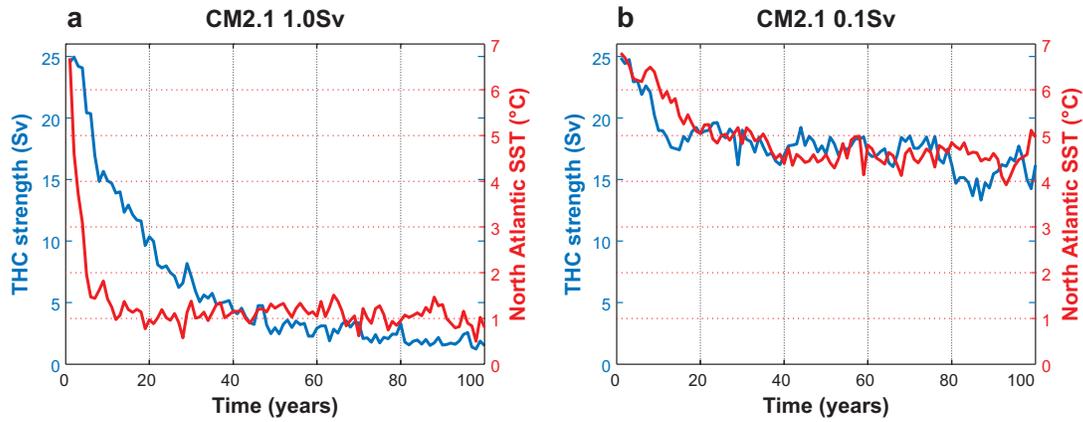
Figure 8

The panels show the time evolution of the strength of the thermohaline circulation in the North Atlantic (*blue*) and the average SST between 50°N–70°N (*red*) for (*a*) the 1.0 Sv freshening experiment and (*b*) the 0.1 Sv freshening experiment.

(see also Hall & Stouffer 2001). In the deeper layers, the northward transport of warm water at first continues in a narrow, geostrophically balanced jet (around 35°W in **Figure 9*a***). This jet, in the absence of convection, no longer loses heat to the surface layers; instead, the warm water is dispersed horizontally in the deeper layers.
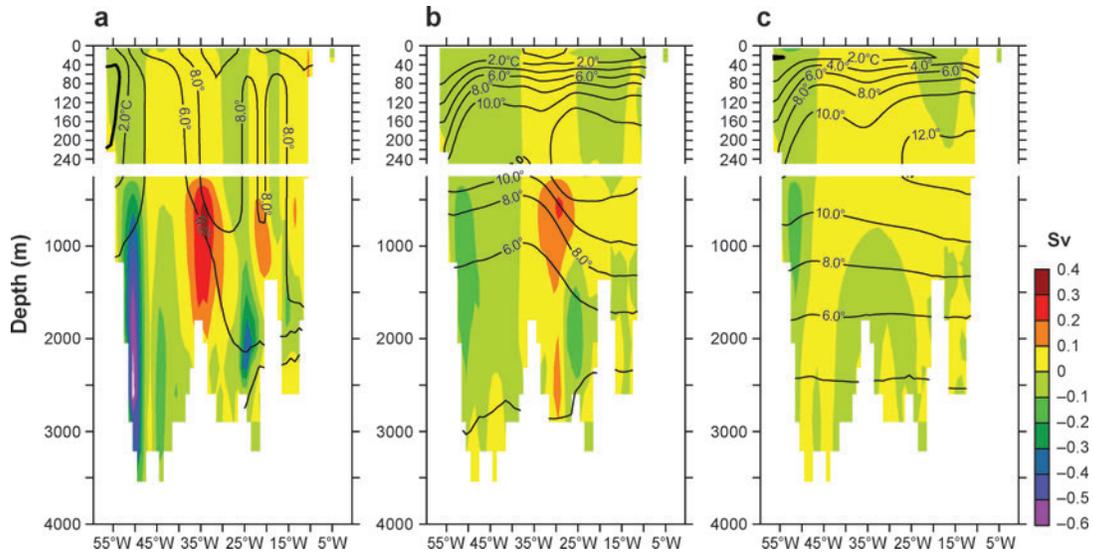
Figure 9

Volume transport (*colors*, Sv) and temperature (*contours*, C) at 55°N (*a*) at the start of and (*b*) 20 years into the 1.0 Sv hosing experiment. (*c*) The mean fields for the years 81–100. Note the different vertical scales for the upper ocean (0–250 m) and deep ocean (250–4000 m).

Note how, in **Figures 9b,c** the weakening of the deep jet is associated with isotherms that become increasingly horizontal as the deep water warms.

In **Figure 8**, the response to a modest freshening of 0.1 Sv is modest, and the surface and deeper layers adjust at essentially the same rate. This happens because the freshening is so weak that the uncoupling of the surface layers from the deeper ones is confined to a few isolated regions. The change in oceanic stratification is nonuniform and convection shifts to regions where the stratification is weak. A shut down of the thermohaline circulation requires that the freshening be strong enough to stratify a huge area.

3.2 Atlantic Heat Budget

A freshening of the northern Atlantic at the rate of 1 Sv changes the heat fluxes across the ocean surface as shown in **Figure 1b**. To determine how this affects the heat budget of the Atlantic Ocean, it is useful to turn to the equation that expresses that budget:

$$dH/dt = IN + OUT + HT. \tag{1}$$

Here $H = \int \rho C_P T \, dv$ is the heat content of the Atlantic basin, which is taken to extend northward from 33°S; t is time; C_p is the heat capacity; T is temperature; ρ is density; IN is the net surface heat flux into the ocean; OUT is the net surface heat flux out of the ocean; and HT is the transport of heat across 33°S. (Additional small terms that are neglected include the export of heat through the Bering Strait and the energy change owing to the imposed freshwater flux, which raises sea level globally.) **Figure 10** shows how the various terms in Equation 1 change after the
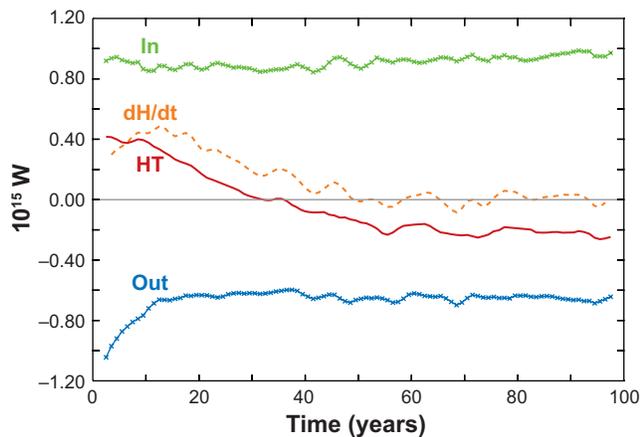
Figure 10

The oceanic heat budget of the Atlantic-Arctic basin for the 1.0 Sv hosing experiment using the CM2.1 model. Net surface heat flux into the basin (IN, *green*), net surface heat flux out of the basin (OUT, *blue*), ocean heat transport across 33°S (HT, *red*), and rate of change of oceanic heat content in the basin (dH/dt, *orange dashed*). Time series are smoothed with a 5-year low-pass (Parzen) filter.

hosing of freshwater starts. Before the freshening starts—in today's climate—the Atlantic Ocean imports heat across 33°S, gains heat in the equatorial region, and loses the sum in high latitudes. The freshening of the northern Atlantic alters this balance by rapidly cooling the high-latitude surface layers, thus reducing evaporation and hence the loss of heat to the atmosphere (**Figure 1***b*). In **Figure 10**, this is seen to happen rapidly, within a matter of a decade or so. The flux of heat into the ocean across its surface (the term IN in Equation 1) hardly changes because the increase in heat uptake in the northern region is balanced by reduced heat uptake in the equatorial region.

After the sudden onset of freshening at the ocean surface, the flux of heat across that surface reaches a new equilibrium within a matter of a decade. However, the adjustment of the thermohaline circulation, and of the transport of heat across 33°S, take far longer, as shown in **Figure 10**. After 60 years, the heat budget is close to a new balance, but the deep ocean continues to adjust very gradually over a much longer period.

3.3 The Threshold

The realistic climate model indicates that 0.1 Sv is below, and that 1.0 Sv is above, the threshold for shutting down the thermohaline circulation. A climate model of intermediate complexity, ECBILT-CLIO, permits a systematic exploration of the response to different rates of freshening and shows that the threshold is in the neighborhood of 0.3–0.4 Sv (**Figure 11**). For low values of the freshwater forcing (0.1–0.2 Sv), the cooling of the surface, the reduction in the intensity of the thermohaline circulation, and the decrease in the transport of heat across 33°S and across the equator are all modest.

As the rate of freshening increases, a threshold is reached at a value of approximately 0.4 Sv when the thermohaline circulation effectively shuts down. At this stage, the oceanic heat transport is poleward in both hemispheres (**Figure 11***c*), almost symmetrically about the equator.

For large values of the forcing (0.6–1.0 Sv), the high latitude surface cooling is independent of the freshening rate and the subsurface warms up, indicating a decoupling of the surface and deeper layers. This subsurface warming is associated with the collapse of the thermohaline circulation. The ocean recovers a balanced heat budget by reducing the heat uptake in the equatorial region (**Figure 11***d*), and by exporting heat (southward) across 33°S. Changes in the thermohaline circulation are of central importance to the associated changes in oceanic conditions in the far northern Atlantic Ocean, but what about the changes in lower latitudes?

4. THE LINKS BETWEEN HIGH AND LOW LATITUDES

A freshening of the northern Atlantic sufficiently large to shut down the thermohaline circulation also causes a warming of the surface waters at the equator (**Figure 1**), and a reversal of the oceanic transport of heat across 33°S (**Figure 11***c*). What are the links between the high and low latitudes? Clues are available in the band of cold,
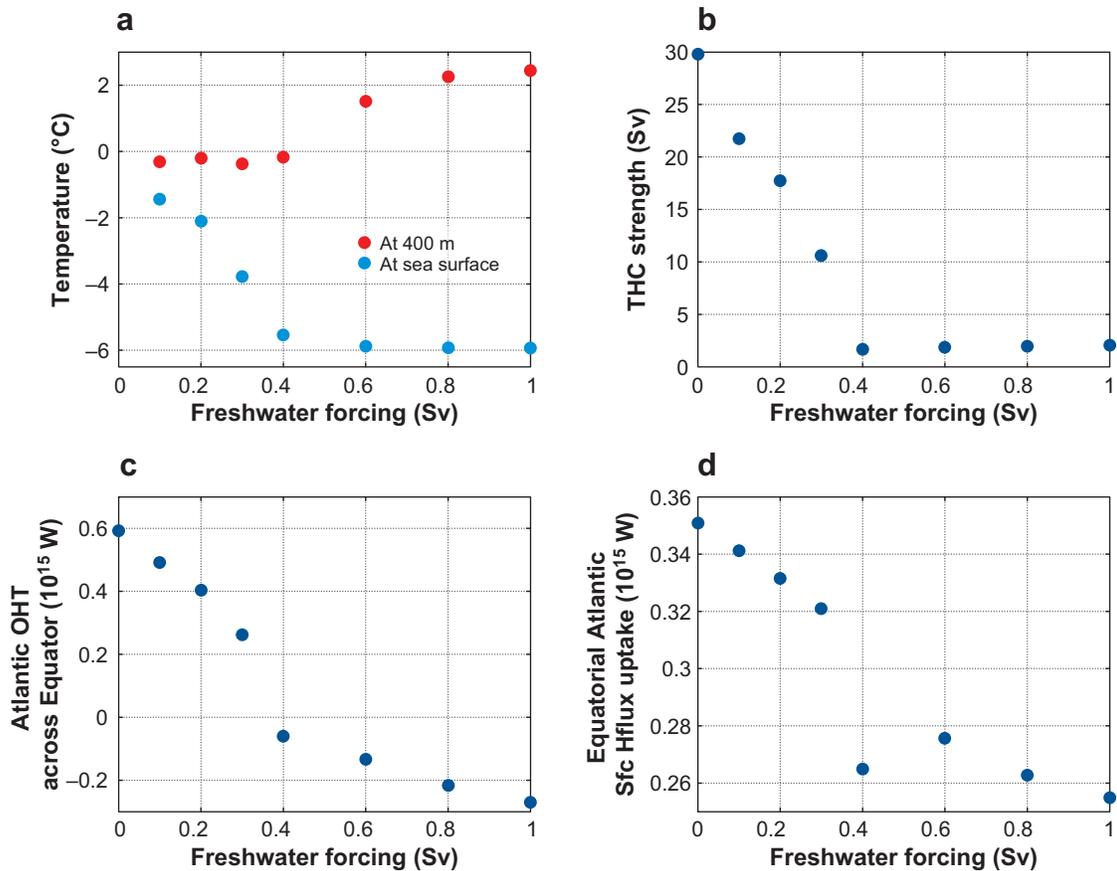
Figure 11

Experiments with ECBILT-CLIO for different values of freshwater forcing in the North Atlantic. (*a*) Temperatures at the surface (*blue*) and at 400 m (*red*) in the North Atlantic between 50°N and 70°N; (*b*) strength of the thermohaline circulation (Sv); (*c*) Oceanic heat transport across the Equator (10^{15} W); and (*d*) equatorial Atlantic heat uptake (10°S-10°N, 10^{15} W). Means are for the last 20 years of the runs. Hosing experiments were run 100 years for rates 0.1, 0.2, 0.6, 0.8, and 1.0 Sv, and 200 years for rates 0.3 and 0.4 Sv, which took longer to reach a state of quasi-equilibrium.

freshwater, which, in **Figures 1** and **6**, is seen to stretch from the northeast to the southwest across the Atlantic, reaching the American coast between 10°N–20°N. What mechanisms generate the currents that advect this cold water southwestward?

The differences between the sea surface temperature patterns in **Figures 1** and **6** indicate that ocean-atmosphere interactions have an amplifying effect, but that a strictly oceanic model captures the main features of the pattern. Hence some of the mechanisms that link high and low latitudes must be oceanic. Yang (1999), Knutti et al. (2004), and Cessi et al. (2004) explore those links and find that the freshwater discharge onto the northern Atlantic triggers Kelvin and coastally trapped waves

that travel along the western boundary and then along the equator. Once they reach Africa, the Kelvin waves split into a northern and southern branch. While moving poleward, they radiate Rossby waves, which readjust the ocean interior setting up a zonal density gradient that in turn induces a geostrophic flow, with a southward component, that advects cold and freshwater from the high latitudes.[6] These processes have an adjustment time of the order of one or two decades (Johnson & Marshall 2002).

A primarily atmospheric mechanism for linking high and low latitudes, one that involves limited interactions with the ocean, is a thermodynamic interplay among wind, evaporation, and sea surface temperature. To demonstrate how this mechanism can cause the equatorward propagation of the extratropical cooling, Chiang & Bitz (2005) use an atmospheric general circulation model coupled to a slab ocean.[7] This configuration is known to overestimate the importance of thermodynamic feedbacks (Barreiro et al. 2005), and denies the wind-driven oceanic circulation a role.

The mechanisms mentioned thus far all contribute to establishing links between high and low latitudes. In the connections between the subtropics and the equator, the wind-driven circulation, which includes the Gulf Stream, is dominant. A change in that circulation can have a significant effect on the equatorial upwelling and hence on tropical sea surface temperatures when a freshening of the subtropical surface waters suppresses the meridional overturning component of the wind-driven circulation by suppressing subduction in the subtropics (Fedorov et al. 2004, 2006a). The crucial factor is interference with the oceanic heat budget. A signal that induces a decrease in the loss of heat in the extratropics also sends adiabatic waves toward the equator where the ocean gains heat across its surface. The adjustment of that gain, so that it balances the reduced loss, includes a deepening of the thermocline. The process is diabatic (unlike the processes involved in interannual El Niño events) and has a decadal timescale comparable to that of the wind-driven circulation (Boccaletti 2005). The rapid adiabatic waves are but the first steps in the slower, diabatic, oceanic adjustment.

Different parameterizations of diffusive processes and of ocean-atmosphere interactions in models cause them to respond differently to the same freshening of the surface waters in the northern Atlantic. **Figure 12** illustrates this for the oceanic heat transport and for sea surface temperature patterns. The warming in low latitudes is seen to be weaker in the coarse resolution model, which instead has a more uniform warming in the southern hemisphere, in effect a "bipolar see-saw" (Knutti et al. 2004, Seidov & Maslin 2001). Different parameterizations for the ocean-atmosphere interactions could lead to even bigger signals in low latitudes. Paleoclimates that can help determine the appropriate parameterizations are those induced by the Milankovitch

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[6]Kelvin waves, oceanic gravity waves influenced by Earth's rotation, are trapped either along boundaries or along the equator. Rossby waves are planetary-scale waves that owe their existence to the latitudinal variation of the Coriolis parameter. A flow is geostrophic when the balance of forces in the horizontal is between the Coriolis and the pressure gradient force.

[7]Slab ocean: a simple ocean of fixed depth (~50 m) that interacts with the atmosphere only through local fluxes of heat and that allows no horizontal transport of heat.
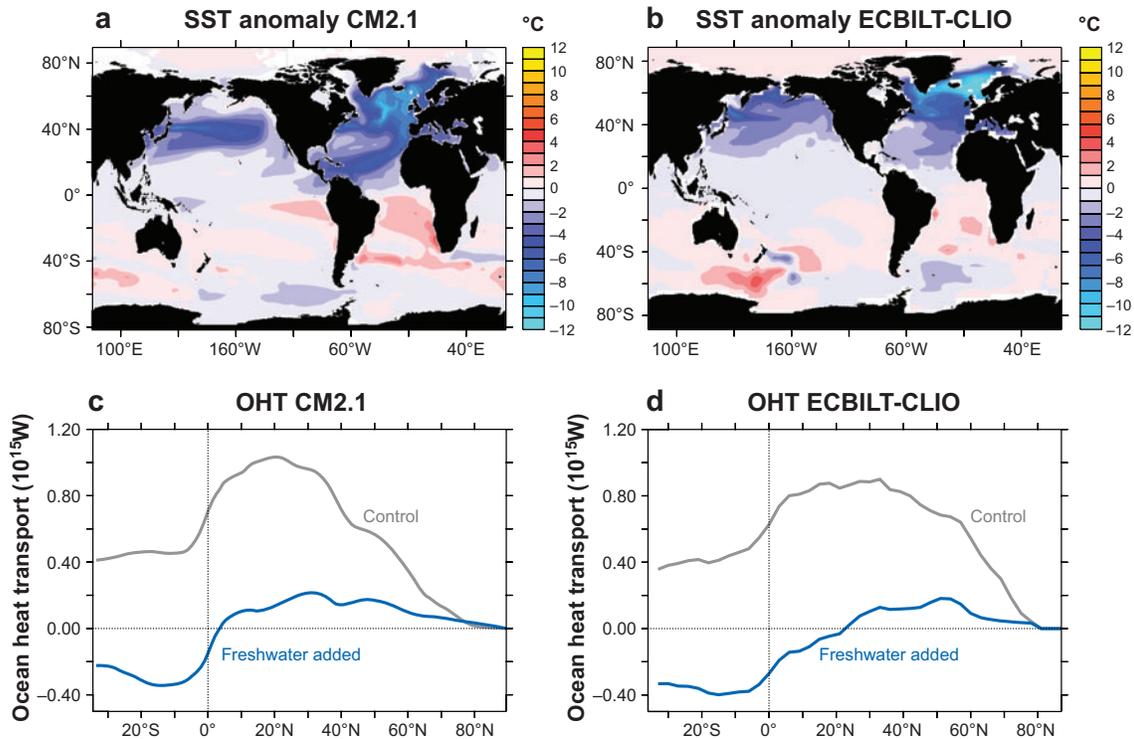
Figure 12

Comparison of changes in SST (*upper panels*) and in the oceanic heat transport (*lower panels*) between the high-resolution CM2.1 and low-resolution ECBILT-CLIO coupled models. In (*c*, *d*) the gray lines denote the ocean heat transport in the Control unperturbed experiment, and the blue lines indicate the ocean heat transport in the experiment where 1.0 Sv of freshwater are added uniformly in the North Atlantic. Experiments were run for 100 years and the plots represent the mean over the last 20 years of the runs.

forcing, which is known accurately. Simulations of so-called Dansgaard-Oeschger events[8] (Dansgaard et al. 1993) provide a less stringent test because information about the intensity and location of the forcing is very poor.

For meridional overturning of the wind-driven circulation to shut down, the freshening must exceed a certain threshold. In the case of the thermohaline circulation that threshold corresponds to a vanishing latitudinal surface density gradient, from the equator to polar regions (Zhang et al. 1999). What is the corresponding condition for the wind-driven circulation?

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[8] Dansgaard-Oeschger events, abrupt climate fluctuations during glacial periods, involve a rapid warming of the Northern Hemisphere in a matter of decades, followed by a gradual cooling over a long period. The geographical extent and the cause of these events are still unclear.

5. DEEPENING OF THE EQUATORIAL THERMOCLINE

5.1 Tropical Thresholds

In the case of the thermohaline circulation, calculations with strictly oceanic models have shed valuable light on the processes that can cause a shut down of its meridional overturning. The same approach is adopted here to study similar aspects of the wind-driven circulation. The idealized oceanic model to be used, which consists of two basins interconnected by a channel representing the Southern Ocean, is described in Section 2. Conditions in the reference state, before the freshening starts, are those shown in **Figure 7** and correspond to a simulation of the conditions of today.

The hypothesis to be tested is that the shutdown of the meridional component of the wind-driven circulation requires only a vanishing of the latitudinal surface density gradient between the equator and the subtropics. We therefore run two freshening experiments. The first, FRESH-I, is very similar to the experiment that produces the results in **Figure 1** because 1.0 Sv of freshwater is added uniformly between 50°N and 70°N in the North Atlantic. This freshening is sufficient to make the latitudinal surface density gradient between the equator and 70°N to vanish. The second experiment, FRESH-II, has the same freshwater forcing as FRESH-I, but the sea surface salinity between 10°N and 30°N is strongly restored (with a 5-day timescale) to the values of the reference state. If the hypothesis is correct then, in the second experiment, the overturning component should shutdown in the case of the thermohaline circulation, but not in the case of the wind-driven circulation. Both experiments are run for 100 years, and anomalies are calculated as the average over the last 20 years.

When 1.0 Sv of freshwater is added uniformly in the North Atlantic (experiment FRESH-I) the meridional overturning circulation shuts down (**Figure 13c**). Moreover, the sea surface temperature anomalies are comparable in magnitude to the ones in the realistic climate model (compare **Figures 13a** and **1a**). The warming is largest on the equator and in the eastern side of the basin along the upwelling areas. Also noticeable is the presence of the anomalous geostrophic currents that flow along lines of constant density, cooling the eastern basin and advecting cold and freshwater southwestward (**Figure 13b**).

In experiment FRESH-II, as in FRESH-I, the high latitudes of the north Atlantic cool strongly (compare **Figures 13a,d**) and the thermohaline circulation collapses (**Figure 13f**). But, because surface salinity is restored between 10°N and 30°N, the density field in the subtropics cannot change significantly (**Figures 13b,e**). As a result, the extratropical cooling does not propagate south of 25°N. More importantly, the warming on the equator and upwelling areas is substantially smaller than that in FRESH-I, and the sea surface gradient along the equator is essentially unaffected.

The wind stress and the surface density gradient $\Delta\rho_{ST}$ between the equator and subtropics control the wind-driven circulation. In FRESH-I, the high-latitude freshening weakens the gradient from the equator to the polar region and, as a byproduct, weakens $\Delta\rho_{ST}$. In FRESH-II, only the gradient north of 30°N decreases so that the impact is mainly on the thermohaline (**Figures 13b,e**). In summary, shutting down the meridional components of the thermohaline and wind-driven circulations both
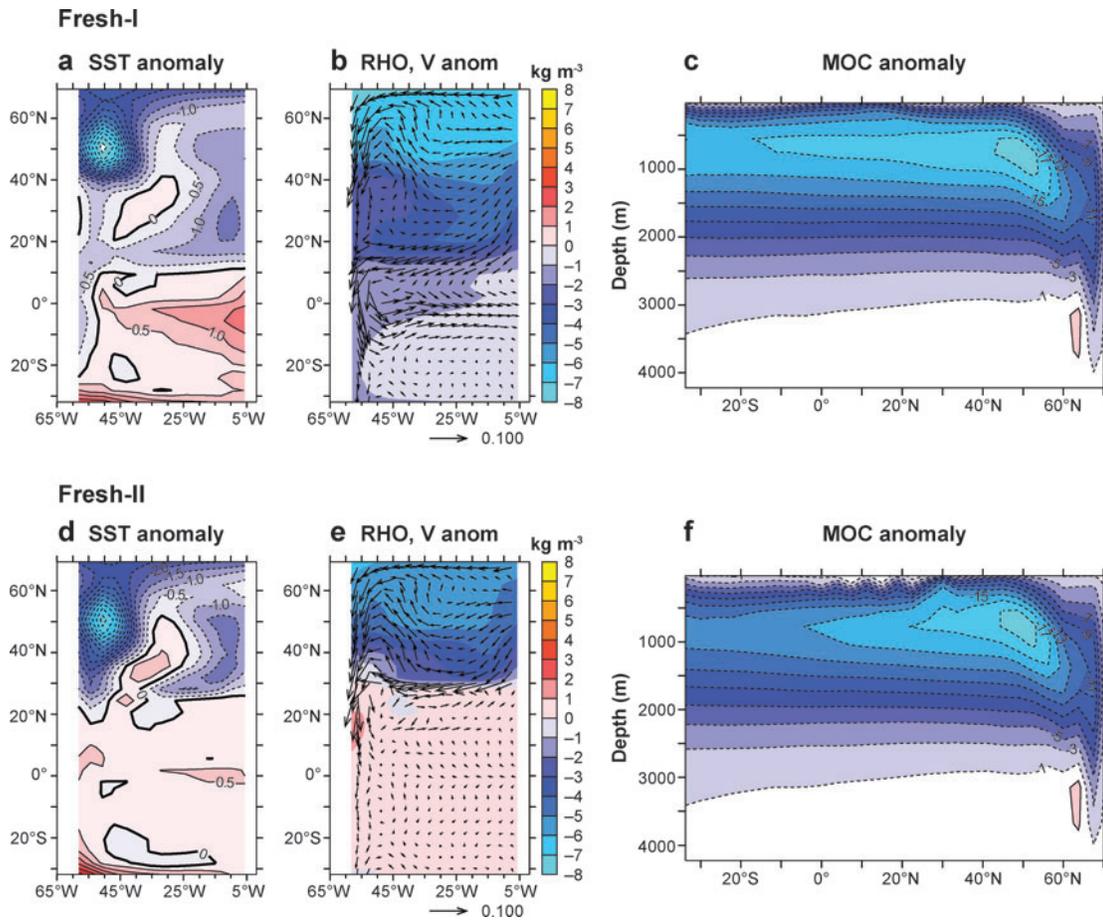
Figure 13

1.0 Sv freshening experiments with idealized ocean general circulation model. Upper panels show the results for FRESH-I: changes in (*a*) Atlantic sea surface temperature, (*b*) surface density (kg m^{-3}) and currents (m s^{-1}), and (*c*) meridional overturning circulation (Sv). Lower panels show the results for FRESH-II: changes in (*d*) Atlantic sea surface temperature, (*e*) surface density and currents, and (*f*) meridional overturning circulation.

require a freshening that exceeds a certain threshold, but the key differences are not in the magnitude of that threshold but in its spatial structure.

5.2 Differences between the Atlantic and Pacific Oceans

A key aspect of the thermohaline circulation, the sinking of surface water in the far northern Atlantic, has no counterpart in the Pacific. As a consequence, conditions in the upper ocean are essentially symmetric about the equator in the Pacific (except for phenomena associated with the northward displacement of the zone of maximum sea
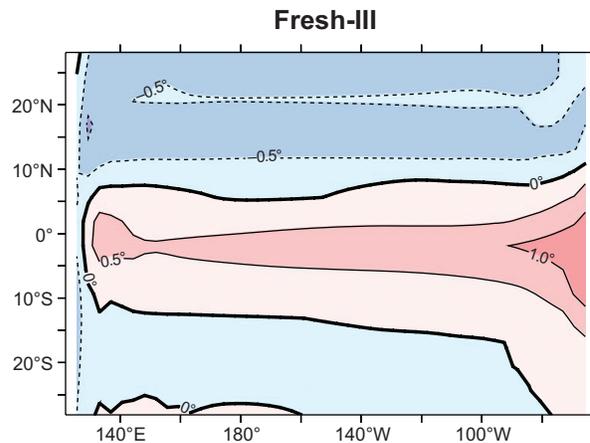
Figure 14

Changes in tropical Pacific SST (°C) in experiment FRESH-III.

surface temperature). The Atlantic, by contrast, has a net northward flow of warm surface waters across the equator (primarily in the North Brazilian coastal current) and a southward flow of cold water at depth. Hence we can expect that changes in the thermohaline circulation affect sea surface temperatures mainly in this Brazilian current (Yang 1999), whereas changes in the wind-driven circulation affect mainly sea surface temperatures in the equatorial upwelling zone.

To explore these differences between the Atlantic and Pacific further we complement the experiment FRESH-I with FRESH-III in which the Pacific is freshened between 10°N and 30°N at a rate of 2 Sv. (Because the Pacific basin is three times wider than the Atlantic, a large freshening is needed to modify the circulation.) The magnitudes of the induced decrease in the equator-to-subtropics density gradient are similar in FRESH-I and -III, with the latter experiment showing a clear equatorial warming with a strongly reduced zonal sea surface temperature gradient (**Figure 14**). The changes in the meridional surface currents that diverge from the equator are as expected. A freshening of the Pacific has a symmetrical effect and weakens the poleward flow in both hemispheres. A freshening of the far northern Atlantic, however, influences mainly the thermohaline circulation and reduces the northward flow across the equator in the surface layers.

The thermal changes associated with the changes in the circulation are shown in **Figure 15**. As expected from the discussion in Section 3.1, freshening of the northern Atlantic (FRESH-I) shows a large, local subsurface warming. At lower latitudes the northern subtropics cools while the tropical south Atlantic warms with a subsurface maximum at ∼300 m. A freshening of the tropical Pacific (FRESH-III) induces subsurface temperature changes with a similar pattern but with different amplitudes. Overall, the subsurface Atlantic is warmer than the Pacific in the tropics and subtropics (**Figures 15c,d**). The key difference is that the meridional overturning of the thermohaline circulation has been shut down in the Atlantic. Hence the downward
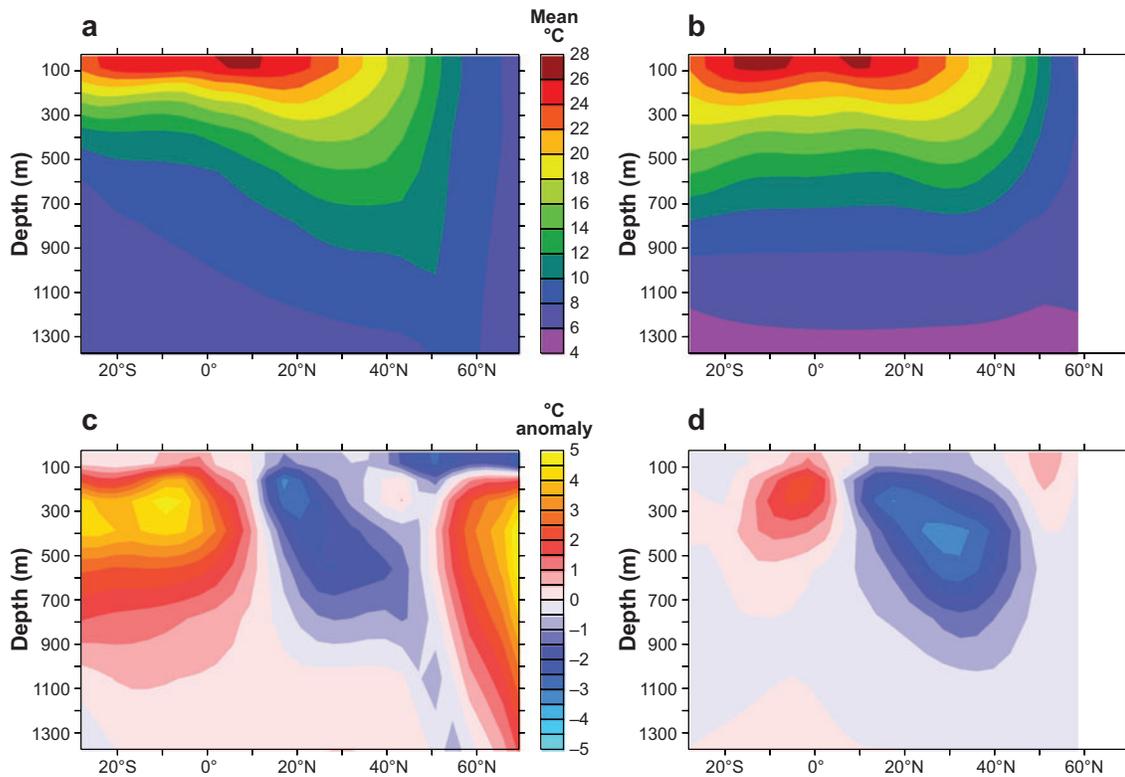
Figure 15

The zonally averaged mean temperature in the unperturbed Control experiment over the (*a*) Atlantic, and (*b*) Pacific basins in the simplified two-basin ocean model. Zonal mean temperature anomaly in (*c*) the tropical Atlantic in FRESH-I, and in (*d*) the tropical Pacific in FRESH-III.

diffusion of heat is no longer being balanced by the upwelling of cold water. Such a change is absent from the Pacific. In the northern subtropics, reduced subduction of anomalously fresh and cold water in the subtropical gyres cools the subsurface and changes the slopes of the isotherms in **Figures 15a,b** in experiment FRESH-I and -III.

6. SUMMARY

Whether a freshening of the northern Atlantic results in significant, abrupt climate changes depends critically on the magnitude and location of the freshening. The freshening must be sufficiently large to eliminate the north-south density gradient at the surface (Zhang et al. 1999). The results in **Figure 11** indicate that the necessary freshwater flux must exceed 0.3 Sv onto the zone between 50°N and 70°N, for at least a decade. This threshold is probably model dependent. Nevertheless, the current rate

of freshening is so small (Dickson et al. 2002, Curry & Mauritzen 2005) that it is probably well below the threshold. Even the great salinity anomaly[9] amounted to a freshening of only 0.07 Sv during a 5-year period (Curry & Mauritzen 2005). As reported in previous studies, and confirmed here, such rates will have a small effect on the thermohaline circulation, and will be associated with relatively small surface cooling in high latitudes.

If global warming should accelerate the melting of glaciers (Ekstrom et al. 2006) or the sudden release of icebergs into the ocean, or increase precipitation, thus rapidly freshening the surface layers of the ocean and increasing their buoyancy further by means of greenhouse warming, then it is conceivable that a shut down of the thermohaline circulation will occur, with far-reaching climate consequences. For example, model estimates of North Atlantic freshening due to melting, precipitation, and runoff in large CO_2-induced climate change scenarios can reach values of about 0.2 Sv (Stouffer et al. 2006). The first sign of such climate changes will be a sharp increase in the stratification of the upper North Atlantic Ocean, followed by changes in the deeper oceanic layers and in the thermohaline circulation. A significant reduction in the North Atlantic sea surface temperatures can actually precede the collapse of the thermohaline circulation. A monitoring strategy can therefore not be limited to that circulation but must include the density of the upper ocean over a large part of the northern Atlantic. This will require a combination of measurements from research vessels, ships of opportunity, and satellites, plus a modeling effort to synthesize the data.

A freshening of the northern Atlantic can affect the equatorial and southern Atlantic by means of several mechanisms. The results presented here indicate that the most important are oceanic processes whose impacts are amplified by the atmosphere through various feedbacks (for example, ice-albedo feedback in the high latitudes). Particularly important is the effect of high-latitude freshening on the latitudinal density gradient at the surface, especially the gradient between the equator and the subtropics. A decrease in that gradient can inhibit the meridional overturning component of the wind-driven circulation. The consequences can include a deepening of the thermocline in the upwelling zones at the equator and along coasts. A shutdown of the thermohaline circulation, if it fails to affect the surface densities in the subtropics because of changes in precipitation and evaporation patterns in low latitudes, will have only a small effect on the equatorial region and hence on the tropical climate. The northern Pacific Ocean has no deep meridional overturning associated with the thermohaline circulation, but it does have shallow meridional overturning, in the subtropics, associated with the wind-driven circulation. Hence, a decrease in the density of the surface waters in the high latitudes of the Pacific will leave the thermohaline circulation unaffected, but will influence the shallow wind-driven circulation, and hence can induce a deepening of the equatorial thermocline and a tendency toward El Niño conditions.

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[9]The Great Salinity Anomaly refers to a large near-surface pool of freshwater that appeared off the east coast of Greenland in the late 1960s (see Dickson et al. 1988).

In climate models, the impact of a freshening in high latitudes on low latitudes depends on the parameterizations of diffusive processes in the oceans and on ocean-atmosphere interactions. Different models will therefore give different answers to important questions, such as the impact of a change in the Pacific Ocean on the Atlantic and vice versa, and the types of disturbance that can alter the properties of El Niño in the Pacific and can change conditions in coastal upwelling zones. In addition to further modeling studies with different parameterizations, there is also a need for stringent tests for the models, tests in which the models simulate observed phenomena. Of special interest is the climatic response to the Milankovitch forcing, which is known precisely. (By contrast, very little is known about the forcing functions that led to abrupt climate changes over the past 50,000 years, the Dansgaard-Oeschger events for example.) The Milankovitch cycles that seem particularly suitable for simulations are those in response to obliquity variations. Although they affect sunlight in high latitudes the most, sea surface temperatures in the equatorial Pacific have a large response that lead the response in ice volume by several thousand years, which is out of phase with the local changes in sunlight, but is in phase with the sunlight variations in high latitudes (Fedorov et al. 2006b). If this response involves primarily the oceanic circulation, then simulations with models need cover only a few decades, not millennia. Solving the puzzle of the response to obliquity cycles will contribute to an improved ability to anticipate rapid climate changes.

DISCLOSURE STATEMENT

The authors are not aware of any factors that might be perceived as affecting the objectivity of this review.

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