NOTES AND CORRESPONDENCE

The Stabilization of the Thermohaline Circulation by the Temperature–Precipitation Feedback

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ABSTRACT

The meridional freshwater flux in the atmosphere, and hence the strength of the hydrological cycle, undergoes variations on glacial–interglacial as well as on some shorter timescales. A significant portion of these changes to the hydrological cycle are due to the temperature–precipitation feedback according to which there is more precipitation over the higher latitudes during warm periods when the moisture holding capacity of the atmosphere is higher.

It is proposed here that this feedback may play an important role in determining the stability of the thermohaline circulation (THC). The THC stability to different parameterizations of the meridional atmospheric freshwater flux is therefore investigated using a simple box model of the ocean, atmosphere, and sea ice. It is demonstrated that parameterizations that are consistent with the temperature–precipitation feedback, and hence with the observed variations of the hydrological cycle during glacial–interglacial cycles, stabilize the THC for a wide range of forcing parameters.

1. Introduction

The rate of snow accumulation on ice sheets during cold periods is significantly lower (by up to a factor of 4) than during relatively warm periods (Cuffey and Clow 1997; Alley et al. 1993; Lorius et al. 1979). This observation was termed the temperature–precipitation feedback (Ghil 1994; Ghil et al. 1987; Källén et al. 1979). Modeling studies of the response of the Antarctic ice sheet to greenhouse warming suggest that for a temperature increase of up to 5.3°C from present-day temperatures, the increased accumulation still dominates the increase in ablation (Huybrechts and Oerlemans 1990). The Greenland ice sheet seems more sensitive to global warming, but for climates colder than today's, the precipitation–temperature feedback is expected to work there as well [Huybrechts et al. 1991, Fig. 3]. Other studies suggest that this feedback may also work in the context of global future warming due to CO_2 increase (Miller and de Vernal 1992; Ledley and Chu 1994). The reduced precipitation rate during cold periods is not limited to precipitation on the ice sheets, but reflects a general reduction in the meridional atmospheric moisture flux during colder periods (Manabe and Broccoli 1985; Kitoh et al. 2001; Ganopolski et al. 1998).

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The purpose of the present study is to examine the effects of the temperature–precipitation feedback on the stability of the thermohaline circulation (THC). Initial studies of THC stability have considered the so called ''mixed'' boundary conditions in which the meridional moisture flux does not depend on the ocean state (Bryan 1986). Later studies, however, did consider the response of the meridional moisture flux to changes in the meridional gradient of the oceanic surface temperature and found this to introduce important new feedbacks (Marotzke 1996; Nakamura et al. 1994; Jayne and Marotzke 1999; Marotzke 2000).

There are two competing effects that determine the meridional atmospheric moisture flux. A colder period is normally characterized by a larger meridional temperature gradient in the atmosphere. This relation between the mean temperature and the meridional atmospheric temperature gradient is therefore implied throughout the discussions in the present work, and plays a major role in the followings. The increased meridional temperature gradient during colder periods may increase the intensity of atmospheric eddies, and thus of eddy meridional moisture flux (Manabe and Broccoli 1985; Hall et al. 1996). However, this increased moisture flux in colder periods seems to be limited to the margins of the ice sheets (Manabe and Broccoli 1985) and to a limited range of the atmospheric meridional temperature gradient (Nakamura 1992; Chang 2001).

On the other hand, a colder atmosphere has a reduced capacity for holding water vapor, and this tends to reduce the meridional moisture flux. Model studies using atmospheric GCMs (Manabe and Broccoli 1985; Krinner and Genthon 1998; Charles et al. 1994; Hall et al. 1996; Ganopolski et al. 1998; Kitoh et al. 2001) have shown that the accumulation rate over land glaciers is also affected by atmospheric circulation changes between cold and warm periods.

The parameterization of Marotzke (1996), Nakamura et al. (1994), and Jayne and Marotzke (1999) results in a reduction in the atmospheric meridional moisture flux when the mean atmospheric temperature is reduced, but in an increased moisture flux when the meridional atmospheric temperature gradient is increased. The dependence on the atmospheric temperature gradient seems dominant in these studies, and given that colder climates seem to be characterized by a larger meridional gradient in the atmospheric temperature, these studies therefore tend to find destabilizing THC feedbacks due to the increased meridional atmospheric moisture flux during such periods.

It is difficult to estimate the quantitative contribution of each of the above two competing effects (mean temperature vs meridional temperature gradient) to the total meridional moisture flux. However, the observational and modeling studies mentioned above make it clear that for large-amplitude climate perturbations (such as glacial–interglacial cycles), the atmospheric meridional moisture flux and the high-latitude precipitation rate are reduced during cold periods (the temperature–precipitation feedback). It should be mentioned that the temperature–precipitation feedback does not necessarily work at all timescales and physical scenarios. For example, Schiller et al. (1997) discuss a meltwater experiment in a coupled GCM and find increased North Atlantic precipitation in a ''Younger Dryas'' state. This is consistent with the dependence of the meridional freshwater (FW) flux on the atmospheric temperature in the parameterization of Nakamura et al. (1994). We find here that, when the meridional atmospheric moisture flux parameterization reproduces the observed temperature–precipitation feedback, the result is a fairly dramatic stabilizing rather than a destabilizing THC feedback.

While the eddy moisture flux feedbacks considered by Nakamura et al. (1994) may be relevant to small amplitude climate variability about present-day climate, the temperature–precipitation feedback, and therefore also its role in THC stability, is relevant for large climate fluctuations such as glacial–interglacial transitions, large Heinrich events, or Dansgaard–Oeschger oscillations. Indeed, a number of studies have found that a positive correlation between the strength of the jet stream and eddy activity exists for wind speeds up to approximately 45 m s^{-1}, but for winds exceeding this value, the correlation becomes negative (Nakamura 1992; Chang 2001). Because the jet stream amplitude is related to the meridional temperature gradient, these studies indirectly support the notion that the meridional moisture flux may increase for a small increase in the meridional temperature gradient but decrease for larger increases of the meridional temperature gradient. The temperature–precipitation feedback and its effect on THC stability may also be relevant to greenhouse warming scenarios. Under increased atmospheric CO_2 scenarios, the high latitudes tend to warm more than lower latitudes, and the meridional atmospheric temperature gradient may therefore become weaker. In spite of this weakening, and consistent with the temperature–precipitation feedback, more precipitation occurs at high latitudes under greenhouse scenarios, leading to a shut down of the THC under a warmer climate (Manabe and Stouffer 1994).

Studies of proxy records have shown that large climate fluctuations may have occurred over a very short period of a few decades (Dansgaard et al. 1989; Taylor et al. 1993). Gildor and Tziperman (2001, 2000) showed that rapid changes of an extensive sea ice cover may play an important role in glacial–interglacial dynamics and speculated that they may also be crucial in shorter-term climate variability. Because large amplitude climate variability such as induced by large-scale sea ice changes may involve the temperature–precipitation feedback, the present study is also concerned with the effects of sea ice on THC stability. This issue has been examined in different contexts by several previous studies (Jayne and Marotzke 1999; Nakamura 1996; Kravtsov 2000; Yang and Neelin 1993; Grigg and Holbrook 2001). Large sea ice cover is expected to affect the meridional temperature gradient in the atmosphere. But it also shifts the storm track southward so that less precipitation falls on the high latitude ocean (Kapsner et al. 1995). Finally, sea ice cover prevents evaporation from the high-latitude ocean (Dansgaard et al. 1989; Charles et al. 1994; Hebbeln et al. 1994), which may again affect the THC stability.

To investigate the effect of the temperature–precipitation feedback and sea ice feedbacks on the THC stability, we use a THC box model (Stommel 1961), coupled to sea ice and atmospheric energy balance models. We do this by introducing two new elements into the analysis. First, we introduce a simple (and crude) meridional moisture flux parameterization, which represents the temperature–precipitation feedback. Second, we note that the albedo of sea ice is much higher than that of the ocean water, and this significantly changes the climate behavior. Apart from these new elements, we follow the analysis method of Jayne and Marotzke (1999) fairly closely. By considering the effects of sea ice and the temperature–precipitation feedback, we find that variable atmospheric moisture flux stabilizes the THC as opposed to the destabilizing feedbacks identified by Nakamura et al. (1994) and Jayne and Marotzke (1999). However, we see these results as complementary rather than contradictory to those of these previous stud-

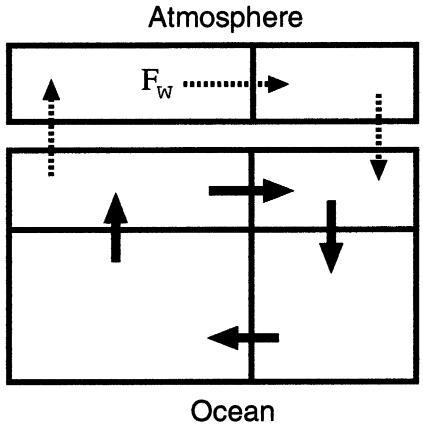

FIG. 1. Vertical cross section of the box model showing the air–sea freshwater flux (dashed arrows) and the THC (solid arrows).

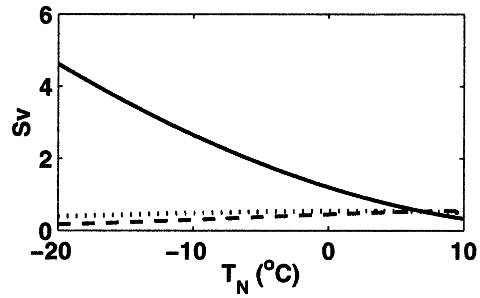

FIG. 2. Atmospheric meridional freshwater transport using the three parameterizations of equations (1: solid), (3: dashes), and (4: dots) as functions of the north box temperature T_N with the southern box temperature fixed to T_S = 25°C. The parameters in the above equations used for this figure are $C_1 = 60.e24$, $C_2 = 21.e7$, $C_3 = 50.e13$, L_N = 4150 km, L_S = 10 000 km, $r = 0.7$, $A = 2.53e11$ Pa, $B = 5.42e3$K.

ies, in the sense that our results here are more likely to be relevant for larger amplitude climate variability while the destabilizing moisture flux feedbacks of the aforementioned studies may be more relevant to the stability of the THC under small amplitude variability about present day climate. In any case, there is clearly a major uncertainty regarding the parameterization of the meridional moisture flux in the atmosphere in simple models such as used here, and one of the points of this work is the need to further develop more justifiable parameterizations that better fit the observed temperature–precipitation behavior.

The following sections describe the simple box model used here (section 2), the results of our stability analysis with and without the temperature–precipitation feedback (section 3), and we conclude in section 4.

2. The box model

The coupled ocean–atmosphere–sea ice model (Fig. 1) is based on that of Gildor and Tziperman (2000, 2001) except that it does not includes a prognostic land ice component. Only a brief model description is given here, and additional details may be found in the above papers. The ocean is represented in our model by two upper ocean boxes and two deep boxes. The meridional thermohaline circulation is calculated from the meridional density gradients (Stommel 1961) and is driven by air–sea fluxes of heat and freshwater. Sea ice forms when the ocean water temperature decreases below a critical freezing temperature and melts above that temperature. The sea ice cover is assumed to grow with an initial thickness of 2 m, and then to become thicker if the entire polar box is sea ice covered. The presence and evolution of sea ice affect the surface albedo, the salinity budget in the ocean, and also the air–sea fluxes (heat flux and evaporation) by partially insulating the ocean from the atmosphere. While sea ice tends to grow rapidly once the ocean reaches the freezing temperature, due to the ice-albedo feedback, it does not necessarily cover the entire polar box area in the present model, as will be explained in more detail below.

The atmospheric model (Fig. 1) roughly follows those of Marotzke and Stone (1995) and Rivin and Tziperman (1997) and is divided into two vertically averaged boxes, representing the same latitude bands as the upper ocean boxes. The lower surface of each atmospheric box is a combination of ocean, land, and sea ice, each with its specified albedo. The averaged potential temperature of each atmospheric box is calculated on the basis of the energy balance of the box, taking into account 1) incoming solar radiation using a box albedo calculated according to the relative fraction of each lower surface type (land, sea, or sea ice) in the box, 2) outgoing longwave radiation at the top of the atmosphere, 3) heat flux exchange with the ocean, and 4) meridional atmospheric heat transport.

Two parameterizations of meridional atmospheric moisture transport are used in this study. The first one is similar to that of Jayne and Marotzke (1999),

$$\begin{aligned} F_w &= -C_1|\nabla T|^2\nabla q \\ &= -C_1\frac{(T_N - T_S)^2 \times (q_N - q_S)}{((L_N + L_S)/2)^3}, \end{aligned} \tag{1}$$

where C_1 is a constant; T_S, T_N are the southern and northern atmospheric temperatures; L_S, L_N the meridional extents of the boxes; $q(T)$ is the humidity, calculated at each box assuming constant relative humidity and based on the atmospheric box temperature, using an approximated Clausius–Clapeyron equation,

$$q(T) = rA\exp(-B/T), \tag{2}$$

where r is a specified fixed relative humidity and A, B are constants. In this parameterization, the dominant effect on the atmospheric meridional moisture flux comes from the dependence of the flux on $|\nabla T|^2$ where ∇T is the meridional atmospheric temperature gradient. As can be seen in Fig. 2 (solid line), the meridional atmospheric freshwater flux in this case increases with the meridional atmospheric temperature gradient in contrast to the temperature–precipitation feedback. We show below that while our box model differs in details from Jayne and Marotzke (1999), the qualitative stability behavior of the two models is very similar when we use the above parameterization. The second parameterization used here is the one used by Gildor and Tziperman [(2001); see discussion around Eq. (14) there], which is able to reproduce the observed temperature–precipitation feedback, and may be written as

$$F_w = -C_2|\nabla T|q = -C_2\frac{(T_N - T_S) \times q_N}{(L_N + L_S)/2}, \qquad (3)$$

where q is the humidity of the box receiving the flux and C_2 is a constant. The meridional atmospheric moisture transport is proportional to both the meridional atmospheric temperature gradient (setting the strength of the synoptic eddies that carry the eddy meridional moisture flux) and to the humidity of the atmospheric box to which the flux is directed (representing the ability of the atmospheric box receiving the moisture flux to actually hold this moisture and carry it poleward from the common boundary with the midlatitude box toward the land glaciers and polar ocean). The (extremely crude) physical picture motivating this parameterization is that, if the polar box temperature is very low, the water evaporated in the midlatitude box is carried to the box boundary, and then condensates upon reaching the cold polar box before actually reaching the interior of this box. This is the reason for having the flux into the polar box depend on the polar box temperature. While obviously extremely crude and not well justified, this formulation results in an increased meridional transport of humidity during warmer periods, contributing to the temperature–precipitation feedback (Fig. 2, dashed line). We will discuss an alternative to this parameterization below and note that any parameterization that reproduces the temperature–precipitation feedback will have the same implications on the THC stability as (3), so the precise form and justification of the parameterization is not critical for our purposes here.

In order to study the effects of the temperature precipitation feedback on THC stability, we use the approach of Jayne and Marotzke (1999). We run the box model to a steady state for different choices of two parameters: the albedo of the northern land box (which may varies as a result of land ice accumulation during glaciations) and a freshwater transport coefficient K_w, which is a nondimensional number ranging from 0.75 to 1.55 and multiplying F_w. The outcome is therefore a plot of the THC amplitude as function of the freshwater forcing (freshwater transport coefficient, K_w) and the thermal forcing (northern land albedo).

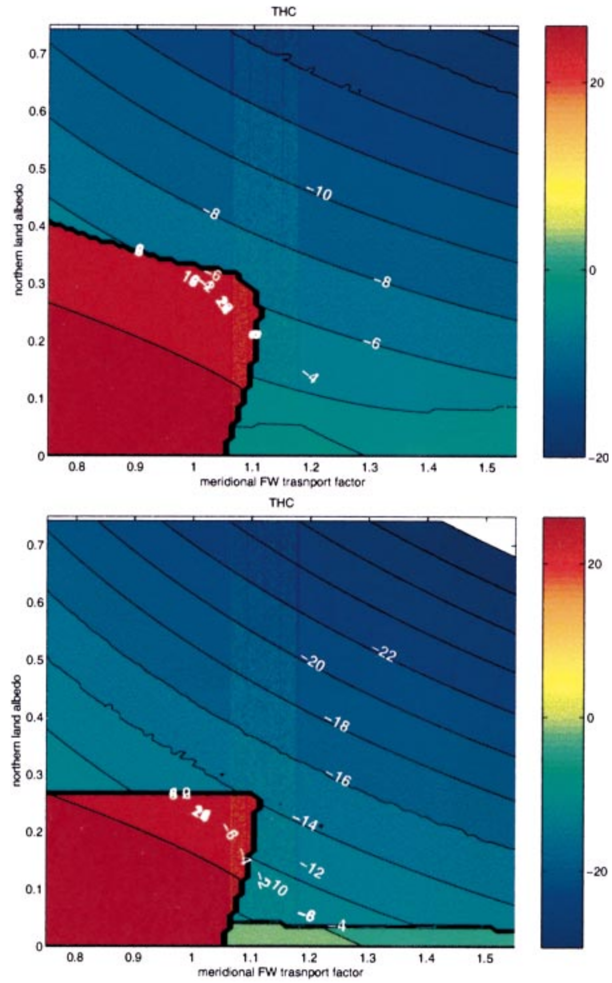

FIG. 3. THC stability using the parameterization (1) of the meridional freshwater atmospheric flux that is dominated by the meridional atmospheric temperature gradient. Shown is the steady state solution for the strength of the THC (in Sverdrups) as function of the albedo of the northern land box and of a nondimensional factor multiplying the meridional freshwater parameterization. (a) The results when sea ice albedo is assumed equal to that of open ocean and (b) when sea ice albedo is assumed more realistically to be higher than that of ocean water. Note that the thermally dominant THC is destabilized and leads to salinity dominated THC for high albedo and for large meridional freshwater factor.

3. THC stability with and without the temperature–precipitation feedback

The stability of the THC as function of the northern box albedo and the freshwater transport coefficient K_w, under the parameterization (1), is presented in Fig. 3. In the upper panel we can see the phase space diagram for the THC stability as function of the specified polar box land albedo and a nondimensional factor multiplying the atmospheric moisture flux parameterization, K_w. For each point in this diagram we initialize the box model with a thermally dominant state (i.e., initial conditions for temperature and salinity that correspond to sinking in the polar box), and we plot the steady state solution for the THC. The initial sea ice volume is set to zero. If the thermally dominant state at the specified values of the polar land albedo and freshwater factor is unstable, the model THC will reverse and a negative value appears in the diagram. By initializing the sea ice with zero we avoid the issue of multiple steady states with different values of the sea ice, which are expected to exist based on the work of Gildor and Tziperman (2000) yet are out of the scope of the present study. Sea ice albedo is set to 0.1 and is assumed in the calculation shown in the upper panel of Fig. 3 to be equal to that of the open ocean as in Jayne and Marotzke (1999). The qualitative behavior of the THC is very similar to that of Jayne and Marotzke (1999), confirming the robustness of these previously obtained results. The inclusion of a distinct and higher sea ice albedo (of 0.6, lower panel of Fig. 3) has a destabilizing effect on the THC since it further cools the atmosphere in the northern box, increases the atmospheric temperature gradient, and hence increases the meridional moisture flux (1). This results in a somewhat larger unstable regime than obtained by Jane and Marotzke (see area between land albedo values of 0.3 and 0.4 and between freshwater factor values of 0.75 and 1.1, which is unstable in the lower panel of Fig. 3 using the larger, and more realistic, sea ice albedo and stable in the upper panel using a sea ice albedo that is equal to the ocean albedo).

The instability of the thermally dominant THC occurs at two regimes of the polar box land albedo–freshwater factor diagram when employing the meridional moisture flux parameterization (1) and the higher sea ice albedo (lower panel in Fig. 3). When the land albedo is high, the atmospheric temperature tends to be colder, sea ice is formed, and the meridional atmospheric temperature gradient becomes large. This is the result both of the insulating effect of sea ice, which reduces the heat flux from the ocean to the atmosphere, and of the higher sea ice albedo. The large meridional temperature gradient results in a strong freshwater flux to the high latitudes, even when K_w is small (upper-left region of Fig. 3) and therefore in a reversed THC. The second regime of instability occurs when there is no sea ice (low land albedo regime in Fig. 3) yet the meridional moisture transport factor is large, which again causes an intense freshwater flux to the high latitude (lower-right region of Fig. 3).

As the increase of meridional moisture flux during cold periods induced by (1) does not fit what we know both from GCM studies and from the rate of snow accumulations reconstructed from proxy records, we test the sensitivity of the THC using the parameterization (3) and demonstrate that with this parameterization the thermally dominant mode of the THC is significantly more stable (Fig. 4). Before analyzing the stability behavior, we note that model parameters are chosen such that for polar box land albedo of 0.2 and for freshwater transport factor of $K_w = 1$, the two parameterizations produce a similar thermally dominant rate of THC.

Several new elements can be observed now in the

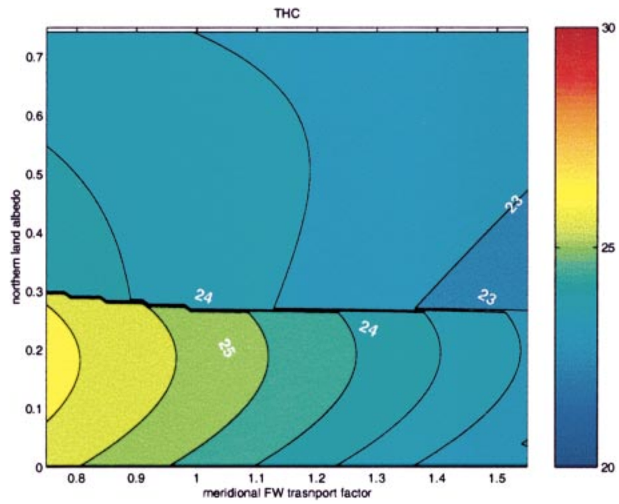

FIG. 4. As in Fig. 3 except for using the parameterization (3) of the meridional freshwater atmospheric flux that is consistent with the temperature–precipitation feedback. Note that the THC is stable for the entire parameters regime presented.

stability diagram (Fig. 4) using the parameterization (3). First and most obvious is that the thermally dominant THC is now stable for the entire parameters regime presented. This enhanced stability is entirely due to the meridional FW flux parameterization and the temperature–precipitation feedback: the larger land albedo parameter regime (colder states) induces less meridional atmospheric FW transport, and therefore stabilizes the THC. Second, the sea ice cover becomes smaller when present: for the parameterization (1), sea ice covers practically the entire ocean area in the northern box once the THC becomes salinity dominated. Using parameterization (3), the sea ice cover is smaller for the following reason. The steady-state ice cover is determined by the heat budget of the polar surface ocean box. The sea ice insulates the polar box from the cold atmosphere and reduces the atmospheric cooling so that it balances the northward heat transport by the THC. When using (3), sea ice is present together with a thermally dominant THC that transports plenty of heat northward and this implies a smaller area of insulating sea ice that is required to balance the northward heat transport by the THC. In contrast, when using (1), the sea ice cover exists when the THC is salinity dominant, does not carry heat northward, and therefore more insulating sea ice cover is required to reach a steady state. A third difference in the solution using the parameterization (3) is that a smaller area of parameter space now corresponds to a presence of a sea ice cover. In Fig. 3, sea ice cover exists when the THC is salinity dominant, which occurs for land albedo larger than 0.3 and for FW transport factor larger than 1.1. In Fig. 4, sea ice cover exists only above land albedo values of 0.3.

In the part of the parameter range when sea ice cover is present (land albedo values of more than about 0.3, upper part of Fig. 4), the meridional atmospheric temperature gradient is stronger and less freshwater is transported by the atmosphere to the northern box because of the temperature–precipitation feedback, so the THC remains stable. The THC weakens due to the insulating effect of sea ice, which reduces the cooling of the polar ocean by the atmosphere. With a better vertical model resolution, one might be able to observe a shift from deep-water formation to intermediate-water formation (Rahmstorf 1995) due to the presence of sea ice and the weakening of the THC. However, even in parts of the parameter space that represent cold, ''glacial-like'' states, the THC is not destabilized and does not become salinity dominant.

The meridional atmospheric freshwater transport formulation (3) is clearly oversimplified and not rigorously justifiable. We cannot and do not try to justify details such as why the meridional dependence on the temperature gradient is to the first and not higher powers such as in Nakamura et al. (1994), for example. We do note, however, that any parameterization that reproduces the observed temperature–precipitation feedback will result in the same stability behavior seen in Fig. 4, and in this sense the details of the parameterization simply do not matter. One may prefer a parameterization that is proportional to the atmospheric moisture gradient rather than to the moisture of the atmospheric box to which the flux is directed. An example could be a parameterization that relates the atmospheric meridional moisture to the humidity gradient at the boundary between the two atmospheric boxes:

$$
\begin{aligned}
F_w &= -C_3 \left.\frac{dq}{dy}\right|_{\text{box boundary}} \\
&= -C_3 \left.\frac{dq}{dT}\frac{dT}{dy}\right|_{\text{box boundary}} \\
&= -C_3 \frac{B}{T^2_{\text{boundary}}} rA \exp\left(-\frac{B}{T_{\text{boundary}}}\right) \times \frac{T_N - T_S}{(L_N + L_S)/2}.
\end{aligned}
\tag{4}
$$

In the last line of the above equation we have used the Clausius–Clapeyron relation at the linearly interpolated temperature at the box boundary

$$T_{\text{boundary}} = T_S + (T_N - T_S) \times \frac{L_S}{L_S + L_N},$$

and wrote the meridional temperature gradient explicitly in terms of the north and south box temperatures. Such a parameterization again reproduces the temperature–precipitation feedback because of the dependence of the moisture flux on the mean atmospheric temperature via the Clausius–Clapeyron relation (Fig. 2, dotted line). We have run the same set of experiments shown in Fig. 4 with this parameterization and find that it results in the same stability behavior as parameterization (3). We therefore conclude that the stabilizing effect of the temperature–precipitation feedback that was found in this work is robust and independent of the specific meridional atmospheric freshwater transport parameterization that is used here.

4. Conclusions

Simple box models (Stommel 1961) have been successfully used in the past for getting some surprisingly robust and useful insights on the sensitivity and behavior of the thermohaline circulation that have been confirmed by coupled ocean atmosphere GCMs (e.g., Tziperman 1997). In such simple models, though, there is a need to parameterized many processes that cannot be modeled explicitly. Such parameterizations should clearly be based on physical reasoning and/or observational evidence whenever possible.

The present study examined two parameterizations for the atmospheric meridional freshwater transport. We explained that a commonly used parameterization, which stresses the effect of meridional temperature gradient on the eddy moisture transport in the atmosphere, does not reproduce the observed reduction in the hydrological cycle during glacial time as reconstructed from both proxy records and GCMs. This parameterization results in an increased meridional atmospheric freshwater flux when the meridional atmospheric temperature gradient is larger, as in glacial times. Such a parameterization, which is likely a reasonable representation of the dependence of the atmospheric meridional moisture flux on the ocean state in small amplitude climate variability about present day climate, destabilizes the THC (Marotzke 1996; Nakamura et al. 1994; Jayne and Marotzke 1999). We have proposed here that different parameterizations of the atmospheric meridional moisture flux may be required for studying larger amplitude climate variability such as glacial–interglacial transitions, Heinrich events, Dansgaard–Oeschger oscillations, and possibly global warming scenarios. We presented two such alternative parameterizations that reproduce the observed temperature–precipitation feedback, meaning that they result in a smaller meridional freshwater transport during colder periods.

We have shown here that the variations of the atmospheric moisture flux in these alternative parameterizations actually stabilize the THC rather than destabilize it. This is our main result, and it may explain why the THC seems to have not collapsed during glacial periods. Recent interpretation of proxy records as well as recent modeling studies both suggest that during past colder periods (e.g., the Last Glacial Maximum), there was a similar rate of water formation in the North Atlantic to that of present day, possibly at different sinking sites and to a reduced depth (Legrand and Wunsch 1995; Bigg et al. 2000; Yu et al. 1996; Matsumoto and Lynch-Stieglitz 1999; Ganopolski et al. 1998; Weaver et al. 1998; Kitoh et al. 2001). This stability of the THC during colder periods in which the meridional temperature gradient in the atmosphere was much stronger than today is consistent with a stabilizing feedback as explored in this paper. Modeling results of global warming scenarios seem to support this conclusion as well. In these models, the outcome of global warming is a collapse or a significant weakening of the THC, in spite of the weakened meridional temperature gradient in the atmosphere (Kattenberg et al. 1996; Cubasch et al. 2001). There is clearly still a significant uncertainty regarding the behavior of the hydrological cycle and the meridional atmospheric temperature gradient during global warming scenarios.

We emphasized that it is far from clear at this stage exactly under what circumstances does the temperature–precipitation feedback act to control the meridional moisture flux, nor what precisely are the specific atmospheric dynamics behind this feedback. It is clearly important to improve our understanding of this feedback, and to derive appropriate parameterizations of the meridional atmospheric moisture flux that are more rigorously based on physical principles than those used in the present study. The temperature–precipitation feedback was shown here to have a potentially profound effect on the THC stability, and it would be interesting to see what will be the effect of such better parameterizations on the stability behavior of the THC.

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