NOTES AND CORRESPONDENCE

A Destabilizing Thermohaline Circulation–Atmosphere–Sea Ice Feedback

STEVEN R. JAYNE

MIT–WHOI Joint Program in Oceanography, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

JOCHEM MAROTZKE

Center for Global Change Science, Department of Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, Massachusetts

18 November 1996 and 9 March 1998

ABSTRACT

Some of the interactions and feedbacks between the atmosphere, thermohaline circulation, and sea ice are illustrated using a simple process model. A simplified version of the annual-mean coupled ocean–atmosphere box model of Nakamura, Stone, and Marotzke is modified to include a parameterization of sea ice. The model includes the thermodynamic effects of sea ice and allows for variable coverage. It is found that the addition of sea ice introduces feedbacks that have a destabilizing influence on the thermohaline circulation: Sea ice insulates the ocean from the atmosphere, creating colder air temperatures at high latitudes, which cause larger atmospheric eddy heat and moisture transports and weaker oceanic heat transports. These in turn lead to thicker ice coverage and hence establish a positive feedback. The results indicate that generally in colder climates, the presence of sea ice may lead to a significant destabilization of the thermohaline circulation. Brine rejection by sea ice plays no important role in this model's dynamics. The net destabilizing effect of sea ice in this model is the result of two positive feedbacks and one negative feedback and is shown to be model dependent. To date, the destabilizing feedback between atmospheric and oceanic heat fluxes, mediated by sea ice, has largely been neglected in conceptual studies of thermohaline circulation stability, but it warrants further investigation in more realistic models.

1. Introduction

The thermohaline circulation (THC) and its sensitivity to perturbations have become central points in the debate about climate variability. An understanding of the response of the system to changes in its forcing is essential for trustworthy climate predictions. However, the coupled ocean and atmosphere system is extremely complex with numerous nonlinear feedbacks that make data analysis and computer modeling difficult. State-of-the-art models of the fully coupled climate system are very complex and often do not allow a simple insight into the system's dynamics. Furthermore, they have often required order 1 adjustments to the modeled fluxes between the ocean and atmosphere to maintain a stable climate representative of the current conditions (e.g., Manabe and Stouffer 1988). Therefore, they are likely to distort the physics of the climate system, particularly in its sensitivity to changes in the forcing (e.g., Nakamura et al. 1994, hereafter NSM).

Since Stommel (1961) it has been known that models of the THC can exhibit at least two steady modes. One is expressed as sinking at high latitudes and upwelling at low latitudes and is qualitatively the regime that the North Atlantic exhibits today. But other modes exist without convection in the North Atlantic, and these modes compete to determine the overall pattern of the THC. For example, the THC in coupled general circulation models (Manabe and Stouffer 1988) can exhibit multiple equilibria either with sinking of salty water in the North Atlantic (the current mode) or with the sinking turned off. Manabe and Stouffer (1995) and Schiller et al. (1997) have also found that the sinking mode can be turned off temporarily by high meltwater runoff.

Many investigators have chosen to investigate the dynamics of the coupled system with simplified models to clarify the role of the feedbacks in the system. Box models, perhaps the simplest models of all, have been used to identify and analyze feedbacks between the atmospheric eddy transports and the THC (NSM; Marotzke and Stone 1995; Tang and Weaver 1995; Marotzke

Corresponding author address: Steven Jayne, Woods Hole Oceanographic Institution, Clark 3, MS 21, Woods Hole, MA 02543-1541.
E-mail: surje@alum.mit.edu

1996). The idealized models have not, however, included the effects of sea ice. Recently, Nakamura (1996) investigated the feedbacks owing to ice albedo and runoff from land on the stability of the THC and found the primary feedback, the storage of precipitation in the polar latitudes as ice, to be stabilizing for the climate system.

Since much of the direct forcing for the THC takes place at high latitudes, sea ice probably plays a significant role in the climate system; however, the exact nature of its interaction with the THC is unknown. Sea ice directly affects the density of the ocean in two ways. First, when sea ice forms, it excludes brine that can elevate the salinity and hence increase the density of the water beneath it. However, sea ice also decreases density because it insulates the warmer ocean from cooling by the overlying cold atmosphere. It is not obvious which of these two effects is dominant and how they interact with other parts of the system. The effects of sea ice coupled to the THC have been investigated by box models (Yang and Neelin 1997b), two-dimensional domains (Yang and Neelin 1993; Yang and Neelin 1997a), and idealized three-dimensional models (Zhang et al. 1995). Yang and Neelin (1993) found that brine rejection is an important dynamical effect and may lead to periodic behavior in the model. While Zhang et al. (1995) also found periodic behavior, they pointed to the insulating effect of ice on the water below as the source of the oscillations. However, none of these investigations have included a dynamic atmosphere in their models. Recently, Lohmann and Gerdes (1998) used an ocean general circulation model coupled to a one-dimensional atmospheric energy balance model and a thermodynamic sea-ice model to investigate the stability of the THC. Their main finding was that when sea ice was present, the response of atmospheric heat flux to sea surface temperature changes stabilized the THC, in contrast to the findings of ice-free models where atmospheric heat transport changes destabilize the THC (NSM; Marotzke 1996).

As our summary shows, the published literature on THC–sea ice interaction is small, and the only paper we are aware of that describes THC–atmosphere–sea ice interaction has just recently been published (Lohmann and Gerdes 1998). On the other hand, there have been numerous publications on THC–atmosphere interactions, as partially summarized in Marotzke (1996) and Rahmstorf et al. (1996). It is unclear which conclusions drawn from the many ''ice free'' models carry over to the case including sea ice. Moreover, the question arises whether the presence of sea ice would stabilize or destabilize the THC or, conversely, whether the absence of ice in models artificially destabilizes or stabilizes. Not only is this relevant for the interpretation of previous model results, but it also bears directly on ice effects on the THC's stability in colder or warmer climates. With the THC and meridional temperature gradients fixed, global cooling could lead from a warm climate without any sea ice to a threshold at which sea ice forms and temperature gradients must change. What then are the pure sea ice effects on the THC's stability under the two scenarios as opposed to those compounded by externally forced atmospheric cooling and reduction in moisture content?

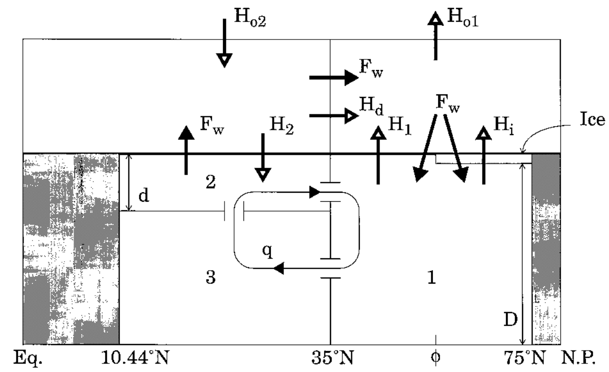

FIG. 1. Vertical cross section of the model, with flux arrows given in the positive direction.

We address this question in a similar manner as NSM, Marotzke (1996), and Rahmstorf (1995), who analyzed the impact of atmospheric feedbacks, by tuning an ice-free and an ice-admitting model to the same THC state. In contrast, Lohmann and Gerdes (1998) only investigated the impact of sea ice on THC–atmosphere interactions but not on the overall model stability. Hence, they were unable to say whether the presence of sea ice was or was not a stabilizing influence. Here we use the much cruder box model to focus on a complementary question: Does the presence of sea ice stabilize or destabilize the THC? More precisely, we try to mimic the comparison between a colder mean climate against one warm enough that, under the same meridional temperature gradient, sea ice would not appear. We investigate the behavior of the ocean–atmosphere–sea ice system by combining the coupled atmosphere–ocean model of NSM with a simple model of the two principal effects of sea ice: brine rejection and insulation. Our model is very simple, but serves to identify some of the fundamental feedbacks related to sea ice and the role of sea ice in the climate system.

2. Model

A thermodynamic parameterization of sea ice is added to a model with the same geometry and physics as that of NSM except that the parameterizations for the heat and moisture fluxes in the atmosphere have been simplified. The model represents an ocean basin with three boxes: the sea ice by a single box partially overlaying a portion of the northern ocean box and the atmosphere with two boxes (Fig. 1). The upper boxes represent the zonally averaged atmosphere between the equator and 35°N and 35°N and the North Pole. The ocean's three boxes span an area of 60° of longitude between 10.44°

and 35°N and 35° and 75°N. The surface areas of the boxes are equal and their depths are 400 m (d) for the shallow tropical box, 3600 m for the deep tropical box, and 4000 m (D) for the polar box. The volume of the polar box is equal to the combined volume of the tropical boxes.

The sea-ice layer models the thermodynamic effects of sea ice; however, the dynamics of sea ice, such as leads and unequal thickness, is not represented. There is a potentially important feedback, the ice–albedo feedback, that we have not included in our model for the purposes of simplification. As ice forms over the ocean it effectively increases the albedo of the ocean as it is highly reflective. Nakamura (1996) investigates the effects of the albedo feedback on the THC and finds it to be a stabilizing influence on the THC for temperature or salinity perturbations in a model without sea ice. Experiments in which the sea ice albedo is included, using the parameterization of North et al. (1981), indicate that it is of minor importance in our model and does not change the qualitative nature of the results. Since our model does not consider the dynamics of sea ice, the albedo of the polar box is uniform and unchanging in time. Also, since we examine only the steady states of the system and perturbations to them, we do not consider the temporal transition from the ice free to ice covered case (notice that this is an annual-mean model).

The water exchange between the boxes is representative of the thermohaline overturning circulation. Following Stommel (1961), the strength of the thermohaline cell in this type of model is simply related to the density gradient between the boxes, under the relation

$$q = k[\alpha(T_2 - T_1) - \beta(S_2 - S_1)], \tag{1}$$

where q is the volume flux in Sv (Sv $\equiv 10^6$ m^3 s^{-1}), $\alpha = 1.5 \times 10^{-4}$ (K)$^{-1}$, $\beta = 8 \times 10^{-4}$ (psu)$^{-1}$, and $k = 2.77 \times 10^9$ m^3 s^{-1}. The parameter k is set to reproduce the modern climate regime (NSM). For q positive, there is sinking in the polar box and upwelling in the Tropics and vice versa for negative q. The strength of the THC is a function not only of the surface density but also of diapycnal mixing and thermocline depth, but (1) reflects density differences as the fundamental driving force (Marotzke 1997).

It is assumed that the atmospheric boxes have no storage capacity for heat or water and therefore the flux between the ocean and atmosphere is the residual of the fluxes between the two atmospheric boxes and their external forcings. The parameterizations used by NSM for the fluxes of heat and freshwater between the atmospheric boxes are rather complicated nonlinear functions of the mean atmospheric temperature and the atmospheric temperature gradient. We use a simplification for the atmospheric fluxes to clarify the role of the fluxes in the model dynamics (Ahmad et al. 1997), the basis of which is that they are made to depend on the meridional temperature gradient to the third power.

Closely following the derivation of Ahmad et al. (1997), the integrated surface heat fluxes for boxes 1 and 2 are H_1 and H_2, respectively. When sea ice is not present, they are calculated as the sum of the incoming and outgoing fluxes, so that

$$H_1 = H_{o1} + H_d \tag{2}$$

$$H_2 = H_{o2} - H_d, \tag{3}$$

where H_{o1} and H_{o2} are the radiative fluxes at the top of the atmospheric boxes, and H_d is the meridional atmospheric heat flux across 35°N. Here, H_{o1} and H_{o2} consist of an incoming solar radiation component (held constant), a reflected shortwave component (proportional to the albedo of the box), and a dynamic outgoing longwave component taken to be a linear function of the meridional atmospheric temperature profile. The profile is found by fitting the second Legendre polynomial,

$$T_s(\phi) = T_{s0} + (T_{s2}/2) \times [3(\sin\phi)^2 - 1], \tag{4}$$

to the ocean surface temperature at two collocation latitudes ($\phi = 20°$ and 55°N). The total expressions for H_{o1} and H_{o2} derived from NSM can be written as functions of T_1 and T_2:

$$H_{o1} = 5.977 - 29.030\alpha_1 - 0.173T_1 - 0.0123T_2 \tag{5}$$

$$H_{o2} = 26.974 - 57.973\alpha_2 - 0.00328T_1 - 0.252T_2, \tag{6}$$

where H_{o1} and H_{o2} are in petawatts and T_1 and T_2 are in degrees Celsius. [see NSM Eqs. (8)–(11)]. The meridional atmospheric heat flux H_d in NSM can be written to an excellent approximation (Ahmad et al. 1997) as

$$H_d = C_2(\Delta T)^n + \frac{2L_v\rho_a}{R}F_w, \tag{7}$$

where ΔT is the atmospheric temperature gradient at 35°N, F_w is the moisture flux by the atmosphere, L_v is the latent heat of evaporation, ρ_a is the density of air, and R is the runoff factor, which controls the amount of water that goes into the ocean from river runoff. The first term represents the sensible heat flux and the second the latent heat flux. Here F_w is parameterized in the following way:

$$F_w = C_1R\frac{e^{-5240/T_M}}{T_M}(\Delta T)^n, \tag{8}$$

where T_M is the mean atmospheric temperature in kelvins. The exponential term divided by the mean temperature to the third power comes from the Clausius–Clapeyron equation for the saturation of air by water vapor; C_1 and C_2 are constants chosen to obtain the same values of the moisture and heat fluxes as a set reference climate. Varying the value of the power n is very instructive in determining the model's sensitivity as it directly controls the strength of the feedbacks due to atmospheric eddy transports (e.g., NSM). For this discussion, we limit ourselves to $n = 3$.

Sea ice is represented with an additional box of varying thickness, extent, and salinity over the northern ocean box. Sea ice is only present in the model when the polar ocean box is at the freezing point of seawater, here held to be constant at $-1.9°C$. Its extent (ϕ) is determined by the latitude where the atmospheric temperature profile equals the freezing point. If the temperature of the sea water in the polar box is greater than $-1.9°C$ (and correspondingly $\phi > 55°N$), sea ice is not present, the box has a thickness of zero, and the model simply reverts to the model of NSM, albeit with simplified atmospheric fluxes. Changes in ice volume due to freezing or melting are determined from energy balance at the ocean–ice interface: since the water must remain at freezing temperature, any excess heating (cooling) beyond $-1.9°C$ is used to melt (freeze) ice (e.g., Lemke 1993). The ice thickness is determined simply by dividing the ice volume by its extent. The freshwater flux that in the absence of the ice layer would have gone into the ocean is now assumed to fall in the form of snow on the ice. We make the simplification that the snow is instantly incorporated into the ice layer so that the mixing creates a uniform salinity within the ice. The equations for the sea ice volume are

$$\frac{dI}{dt} = \frac{d}{dt}(I_{\rm bot}) + \frac{d}{dt}(I_{\rm top}), \tag{9}$$

where I is the ice volume, and $dI_{\rm top}/dt$ is the change in ice volume due to snow on top of the sea ice given by

$$\frac{d}{dt}(I_{\rm top}) = \frac{\delta F_w}{\rho_0}, \tag{10}$$

where δ is the fraction of sea ice coverage, given by

$$\delta = \frac{\displaystyle\int_{\phi}^{75°} \cos\phi'\, d\phi'}{\displaystyle\int_{35°}^{75°} \cos\phi'\, d\phi'}. \tag{11}$$

The change in ice volume at the bottom of the ice layer $dI_{\rm bot}/dt$ due to direct melting or freezing of sea ice, which arises when the heat fluxes into ocean box 1 do not balance, is

$$\frac{d}{dt}(I_{\rm bot}) = -(H_1 + |q|(T_m - T_1)\rho_0 C_p)/(L_f \rho_0), \tag{12}$$

where $m = 2$ for $q > 0$ and $m = 3$ for $q < 0$ and L_f is the latent heat of fusion. The term that goes as H_1 represents the loss or gain of heat to the overlying atmosphere, while the term that behaves as $|q|(T_m - T_1)$ gives the amount of heat transported by the THC and represents a loss of ice due to melting. The equation for the change in the salinity of the sea ice, $S_{\rm ice}$, is given by

$$\frac{d}{dt}(S_{\rm ice}) = \frac{1}{2}\left[\frac{S_{\rm new}}{I}\left(\frac{d}{dt}(I_{\rm bot}) + \left|\frac{d}{dt}(I_{\rm bot})\right|\right) + \frac{S_i}{I}\left(\frac{d}{dt}(I_{\rm bot}) - \left|\frac{d}{dt}(I_{\rm bot})\right|\right)\right] - \frac{S_i}{I}\frac{dI}{dt}, \tag{13}$$

where $S_{\rm new}$ is the salinity of newly frozen seawater, set at 15 psu in this model.

The heat flux through the ice is specified using the parameterization of Yang and Neelin (1993). This causes the flux between the northern ocean and atmosphere to be determined by the temperature difference between the air and water as well as the ice thickness and extent, instead of only the residual of the atmospheric fluxes:

$$H_1 = (1 - \delta)(H_{o1} + H_d) + H_L, \tag{14}$$

where H_L is the heat released by snow freezing in the atmospheric box being given by

$$H_L = \delta F_w L_f. \tag{15}$$

The heat flux through the ice is governed by (16) relating it to the area of the sea ice and the heat conductivity of the sea ice divided by the ice thickness and multiplied by the area-weighted temperature difference between the ocean and atmosphere:

$$H_i = \delta A C_{\rm ice}\frac{\delta A}{I}\frac{\displaystyle\int_{\phi}^{75} \cos\phi'(T_s(\phi') - T_{\rm ocean})\, d\phi'}{\displaystyle\int_{\phi}^{75} \cos\phi'\, d\phi'}, \tag{16}$$

where A is the area of the ocean box between 35° and 75°N; $C_{\rm ice}$ is the heat conductivity of ice, taken to be 2 W m^{-2} K^{-1}; T_s is the air temperature profile given by the Legendre polynomial (4); and $T_{\rm ocean}$ is the ocean temperature, held in the presence of ice to be the freezing point ($-1.9°C$).

Using the values of the sea ice thickness (I) and extent (δ) from the previous time step, the temperature of the atmosphere at the northern collocation point is solved for iteratively using the Newton–Raphson method, such that the sum of the fluxes into the atmospheric boxes equal zero, namely, that the following equation is satisfied:

$$H_1(T_s, I, \delta) + H_i(T_s, I, \delta) - H_{o1}(T_s) - H_d(T_s) = 0. \tag{17}$$

This is essentially a change from the ocean temperature being prognostic to the ice volume being prognostic, from which the air temperature over the ice is diagnosed. The air temperature profile is now decoupled from the ocean temperature in the polar box instead of being directly tied to it as it was before; this has important implications for the feedback mechanisms.

3. Results

The periodic behavior found by Yang and Neelin (1993) and Zhang et al. (1995) is not found in our model. This is most likely a result of using a box model for the ocean with steady forcing in our study. Ruddick and Zhang (1996) have shown that under steady forcing, the two box model of Stommel (1961) does not exhibit self-driven oscillations and using a three box model, Rivin and Tziperman (1997) found that the model's THC displayed damped oscillatory motions, which needed continuous variability in the external forcing to produce sustained variability. Instead, long-term steady-state solutions are found in which brine rejection plays no role in the equilibrium state and where only the insulating effect of the ice is important, consistent with the findings of Zhang et al. (1995) and Lohmann and Gerdes (1998). That brine rejection is inconsequential in the steady state is straightforward if one considers the dynamics of a layer of sea ice with a steady thickness. The snow falling from above will be constantly diluting the salt present in the ice layer, while at the same time, from the bottom of the layer, there must be melting to balance the incoming snow, which will remove some of the remaining salt with the meltwater. This leads to the long-term steady state where the salinity of the sea ice asymptotes to zero. Therefore, the ice layer in the steady state is a source of freshwater, at the same rate of snowfall on the top. However, at shorter timescales when the ice layer may be rapidly growing (e.g., during the seasonal cycle), brine rejection may play a more significant role. However, we examine only the steady states and perturbations to it.

To understand the role ice may play in the model, we explore the behavior of the model in a parameter space where we vary the albedo in the high-latitude box and the parameter controlling runoff. The high-latitude albedo effectively controls the amount of short wavelength incoming solar radiation the box receives. For low (high) values of the albedo, the box is very absorbing (reflective), which leads to the box being warmer (colder). The runoff factor is the ratio of ocean area to catchment area. This in turn controls the amount of river runoff that the ocean basin receives due to the atmospheric eddy moisture flux over land. The model was initialized with a positive overturning state. We show the steady-state overturning rates in Fig. 2a for the ice-free model and Fig. 2b for the model that includes ice with the associated ice extent and thickness shown in Figs. 2c and 2d, respectively. The principal effect to note in the case with sea ice is that for high albedos the direction of the thermohaline cell is reversed such that there is sinking at low latitudes and upwelling in the northern box under the ice. This is accompanied by relatively thicker ice as compared to the area of parameter space where there is sinking at high latitudes. It is consistent with the idea that the THC was weakened during glaciation and extended sea ice cover. When the model was initialized with a negative overturning, the entire phase space remained in the negative overturning state.

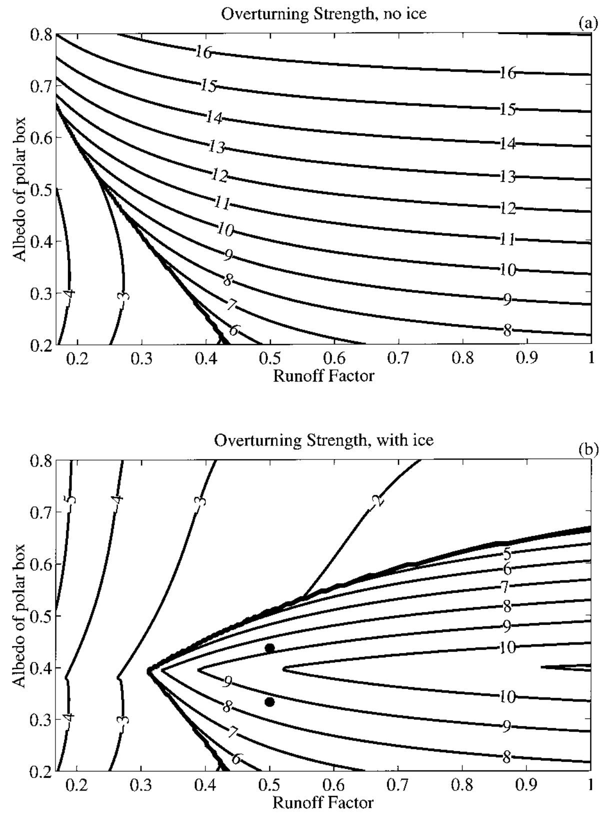

FIG. 2. (a) Phase space diagram of thermohaline circulation strength in Sv for model without ice for varying polar albedo and varying runoff factor. (b) As in (a) but for model with ice. Two points (solid circles) show positions in phase space used in perturbation experiments. (c) Phase space diagram of ice extent (latitude) associated with (b); the shaded area denotes where sea ice is not present. (d) Phase space diagram of ice thickness in (m) associated with (b). (e) Phase space diagram of atmospheric temperature gradient in degrees Celsius for model without ice. (f) As in (e) but for model with ice. (g) Phase space diagram of oceanic temperature gradient in degrees Celsius for model without ice. (h) As in (g) but for model with ice. (i) Phase space diagram of oceanic heat flux in petawatts for model without ice. (j) As in (i) but for model with ice.

The reason for this switching of the THC mode over an area of the parameter space involves the following feedback: Thick ice over the ocean effectively insulates the polar atmosphere from the ocean, leading to colder atmospheric temperatures. This in turn increases the meridional atmospheric temperature gradient (cf. Figs. 2e and 2f), thus increasing both the atmospheric eddy sensible heat flux and the atmospheric eddy moisture flux. The increased moisture flux leads to an increase in the amount of snow over the ice and allows it to maintain a thick ice layer. Coupled with this, there is an upper limit placed on the ocean temperature gradient (cf. Figs. 2g and 2h) since in the ice free case the water temperature for the polar box becomes strongly negative in

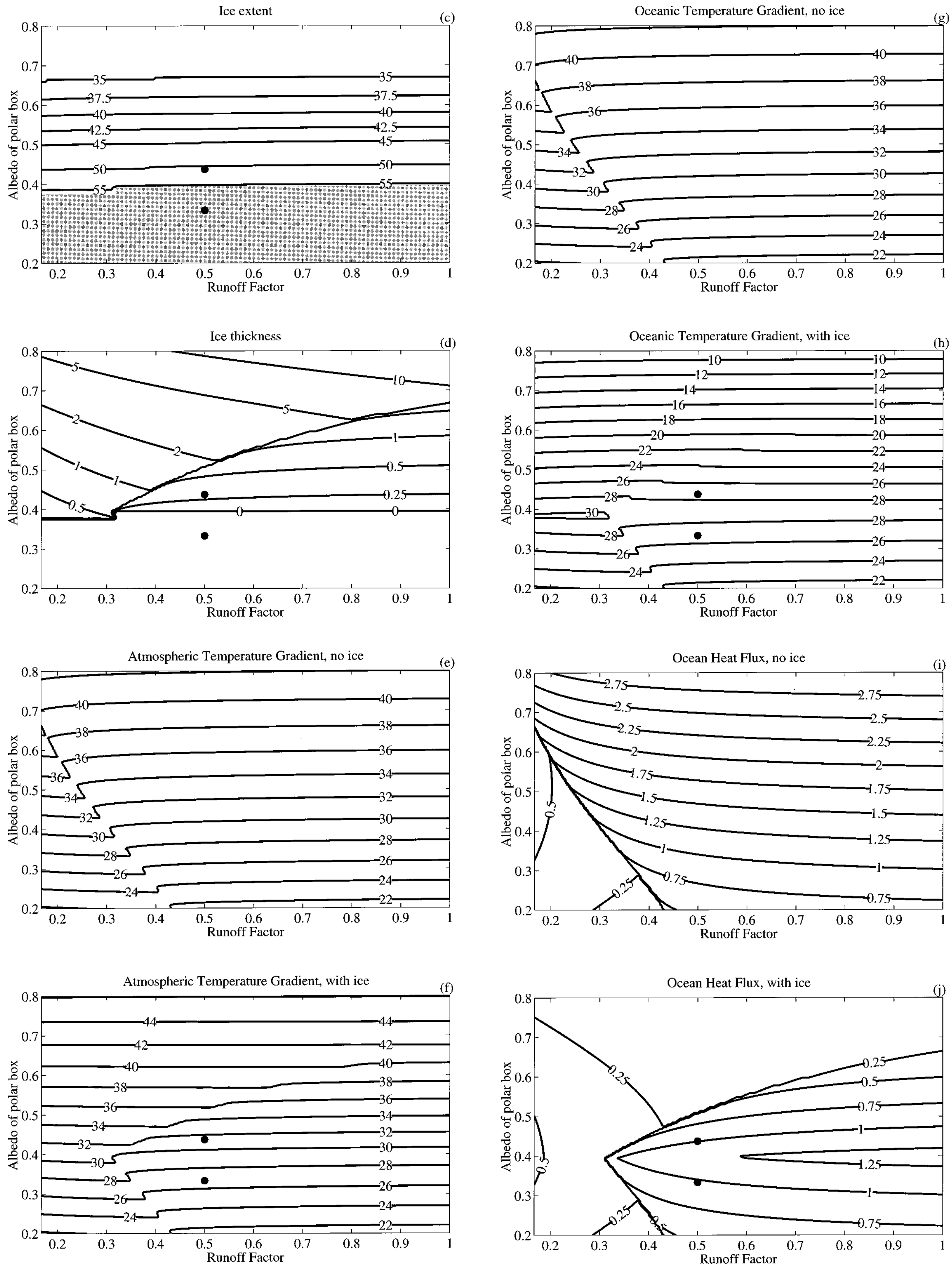

FIG. 2. (*Continued*)

sections of the parameter space, whereas in the ice-covered case it is bounded by the freezing point of seawater (held at $-1.9°C$). This capping of the ocean temperature gradient weakens the thermal forcing of the THC so that the haline component expressed as sinking of salty water in the Tropics dominates the circulation. This mode of the THC has very weak oceanic heat transports compared to the case with cold water sinking at high latitudes (cf. Figs. 2i and 2j). The decrease in heat transport by the ocean into the northern latitudes further supports the thickening of the ice layer. The heat flux required to balance the system is supplied by the increased sensible heat flux by the atmosphere due to the increased atmospheric temperature gradient. This is, in total, an interaction of the sea ice with the eddy flux feedback described in NSM.

To explore the finite amplitude stability of the THC under the influence of an ice layer to that of an ice-free ocean we examine the sensitivity of the system to finite amplitude perturbations. We choose two points in the polar albedo–runoff factor parameter space, having the same runoff factor but differing albedos: one with a steady state that produces sea ice and the other that does not. They have the same steady-state overturning strength and runoff factor but differing albedos (Figs. 2b,f,h, and j). In addition to having the same overturning, they also have the same oceanic temperature and salinity gradients, so the mean flow feedbacks are the same in these two cases. We integrate the two parameter regimes to steady state and then apply a finite amplitude salinity perturbation by instantaneously reducing the salinity of box 1 and increasing the salinity in boxes 2 and 3 by the same amount. A salinity perturbation of 0.76 psu is necessary to reverse the thermohaline cell in the model without ice, whereas it only takes a perturbation of 0.65 psu to reverse the circulation in the ice-covered case. This indicates that the ice-covered regime is somewhat less stable than the ice-free case; however, the effect is only a slight weakening of the stability but not by an order of magnitude change. How much this ultimately matters depends on how close the climate system is in phase space to the bifurcation of the two modes of the THC.

Sea ice insulates the high-latitude ocean and atmosphere from each other; furthermore, it sets an upper limit on the ocean temperature gradient because a lower bound is placed on the ocean temperature by the freezing point of seawater. To investigate the effect of the lower bound on ocean temperature in isolation, we perform a set of experiments in which the sea water temperature is not allowed to drop below $-1.9°C$, but high-latitude ocean and atmosphere are coupled as in the ice-free case. The energy to prevent cooling of the ocean beyond $-1.9°C$ comes from ice formation, that is, the latent heat of fusion. Since there is now no feedback (dynamical or through insulation) to balance the high-latitude heat budget, this interpretation implies that ice growth keeps on forever and a steady state cannot be reached. The alternative interpretation is that energy is not conserved. This procedure is not very satisfactory, but it separates the effect of the capping of the temperature gradient from the insulation effect of sea ice. Note that the most obvious way to eliminate the insulation effect, simply increasing the heat conductivity, is ineffectual since the increased heat flow through ice merely leads to greater ice thickness, such that the ratio of the two, and hence the steady-state heat loss, remains constant.

With T_1 limited to $-1.9°C$ and no sea-ice insulation effect, we find that the THC does not reverse from positive to negative sinking rates as before. Additionally, the finite-amplitude stability of the THC for the previously ice-covered point in phase space is higher, as it requires a salinity perturbation of 1.02 psu to reverse the circulation. Hence, we conclude that it is the insulating effect of sea ice and the resulting increase in atmospheric transports leading in turn to reduced tropical temperatures that is responsible for the destabilizing feedback.

There is a fundamental change in the model behavior when sea ice is present that explains the change in its stability. The atmospheric sensible and latent heat fluxes both increase with increasing atmospheric temperature gradient, regardless of the presence of sea ice. The oceanic heat flux also increases with increasing atmospheric temperature gradient when sea ice is not present. But, when sea ice is present, the oceanic heat flux decreases with increasing atmospheric temperature gradient. This is summarized in Figs. 3a and 3b, which plot the oceanic heat flux as a function of atmospheric temperature gradient for both the model with ice and the model without ice. Since the oceanic heat flux is not an obvious function of the atmospheric temperature gradient, we use the data from the phase-space plots (Figs. 2e,f,i, and j) to construct this plot and separate the data for positive overturning from negative overturning for clarity. In this data, other quantities that affect the oceanic heat flux are varying as well, explaining the scattering of the lines; however, the trend is obvious. When sea ice is not included, the oceanic temperature gradient increases with increasing radiative climate forcing and hence increasing atmospheric temperature gradient. In contrast, when sea ice is present, the oceanic temperature gradient becomes weaker as the atmospheric temperature gradient becomes larger. The weaker oceanic temperature gradient weakens the oceanic heat flux both by decreasing the thermohaline overturning strength and giving the overturning circulation a smaller gradient to transport. A weaker mean THC and the suppressed response of the temperature gradient to THC changes both lead to less stability (Marotzke 1996).

Even in the simple model used here, ocean–atmosphere interactions are considerably more complicated than in the corresponding case without sea ice. We therefore present the feedbacks in some detail, also because it will allow us to assess which of our conclusions are likely to carry over to more complex models such as

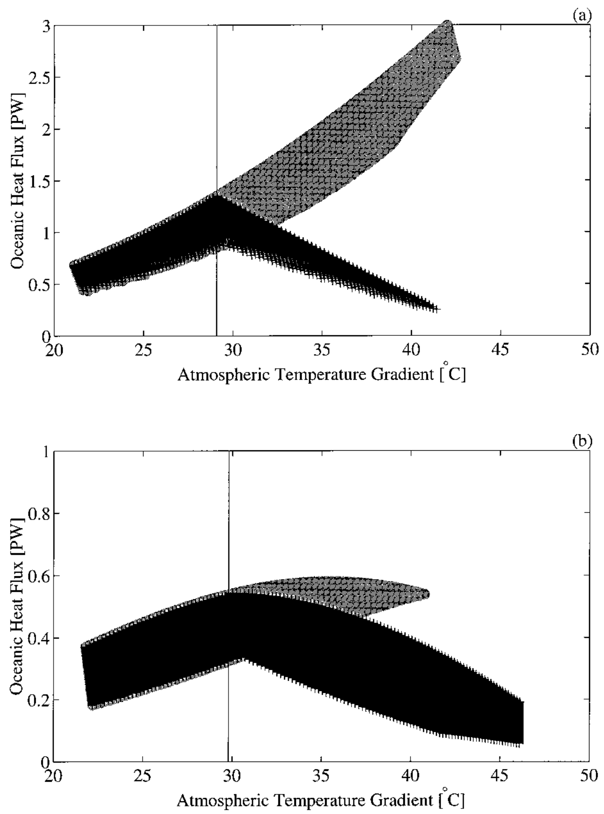

FIG. 3. (a) Oceanic heat flux in petawatts as a function of atmospheric temperature gradient over areas of phase space with positive overturning (Fig. 2) for model without ice (gray shading) and model with ice (black). The area to the left of the vertical line denotes the data where ice is not present in either model. (b) Same as (a) but for areas of phase space with negative overturning.

Lohmann and Gerdes (1998) or the real climate system. The same conceptual procedure as in Marotzke (1996) will be used, and we will point out the major differences from the ice-free case.

The starting point is an assumed small perturbation of an ice-covered, high-latitude sinking equilibrium. An increase in salinity gradient (low-latitude salinity minus high-latitude salinity) results in a weakened overturning circulation. The following effects ensue.

1) The mean flow advects the positive perturbation salinity gradient and reduces the initial perturbation (negative feedback).
2) The reduced overturning advects less mean salinity gradient and increases the initial perturbation (positive feedback).
3) The reduced overturning advects less mean temperature gradient, which increases the ocean temperature gradient and the overturning (negative feedback).

Feedbacks 1–3 are purely oceanic and occur independently of atmospheric or sea ice coupling. The reduced ocean heat transport (feedback 3) leads to thicker sea ice, less heat flux through the ice, and a decrease in atmospheric temperature over ice. Low-latitude oceanic and hence atmospheric temperature increases, and the atmospheric temperature gradient increases. This leads to three additional feedbacks (4–6).

4) The atmospheric moisture transport increases; increased evaporation at low latitudes increases salinity there, while increased snowfall at high latitudes first leads to thicker ice but ultimately, when ice thickness has equilibrated again, simply to a higher melting rate at the bottom of the ice and reduced salinity underneath. Hence, the salinity gradient is further increased and there is a positive feedback (the eddy moisture transport–thermohaline circulation (EMT) feedback of NSM, except that the modification in atmospheric temperature gradient might be stronger).
5) The atmospheric heat transport increases (in reality coupled to the moisture transport through the latent heat); and the air over ice gets warmer, which means that less heat goes from the ocean through the ice to the atmosphere; and ice thickness is reduced. By itself, this has no effect on the THC since the high-latitude ocean remains at freezing temperature. But, since the original sea ice thickening is counteracted, this is a negative feedback.
6) The atmospheric heat transport increases, which leads to a cooling of the low-latitude ocean and atmosphere. This means less thermal forcing of the THC leading to reduced flow (positive feedback).

Without ice, feedbacks 5 and 6 can simply be summarized by saying that the atmospheric heat transport reduces the atmospheric temperature gradient originally created by the reduction in ocean heat transport, hence weakens the (negative) ocean heat transport feedback and therefore is destabilizing. In particular, without ice, the net effect of all the above is an increase in ocean temperature gradient since the perturbation atmospheric heat transport cannot more than fully eliminate the temperature perturbation that caused it in the first place. With ice, however, the process cannot be couched in terms only of the oceanic temperature gradient. The large changes in atmospheric temperature over ice can lead to large changes in atmospheric heat transport, and it is possible (and in our model indeed the case) that the resulting reduction in low-latitude temperature outweighs the increase originally caused by the reduced THC.

We summarize that in our model, with its simplified physics and geometry, the presence of sea ice has the net effect of destabilizing the THC. However, this overall change in model stability is the sum of several feedbacks, some positive (destabilizing) and some negative (stabilizing). In our formulation, the additional negative feedback (feedback 5), the atmospheric heat transport acting on the sea ice, is outweighed by the positive feedbacks, the one involving the eddy moisture flux (feedback 4) and the feedback between atmospheric heat transport and the reduction in oceanic temperature gradient (feedback 6). The strength of these individual feedbacks is model dependent, as is their net effect. Hence, our model does not allow us to draw conclusions about whether sea ice stabilizes or destabilizes the THC in more complex models or the real climate system, and studies with more sophisticated models are warranted.

In this vein, the only other study of atmosphere–THC–sea-ice interactions we are aware of is that of Lohmann and Gerdes (1998) who use a three-dimensional ocean model coupled to an atmospheric energy balance model and a thermodynamic sea ice model. Since they do not directly contrast their model to an ice-free case, it cannot be said whether the combination of feedbacks involving sea ice overall stabilizes or destabilizes. Their main conclusion is that the presence of sea ice causes the atmospheric eddy heat transport to stabilize the THC, in contrast to the ice-free models. They emphasize a negative feedback that results from the insulating effect of more sea ice, as well as its higher albedo, causing the polar atmosphere to be colder, which enhances the atmospheric eddy heat flux that reduces sea-ice thickness and extent (similar to feedback 5). Their model also contains the destabilizing EMT feedback (feedback 4), which is very weak, however. Presumably, though they do not discuss it, their model also contains the interaction between the increased atmospheric eddy heat transport and the depletion of heat in the subtropical ocean and resulting reduction of thermal forcing of and heat transport by the THC (see feedback 6). But this positive feedback is weaker in their model than the negative feedback leading to less ice, which confirms our statement that the net effect of the feedbacks is model dependent.

The critical role of the atmospheric parameterizations in determining the effect of sea ice on the system is underscored by sensitivity experiments of Lohmann and Gerdes (1998) in which they change their atmospheric parameterizations. They switch from an energy balance model to 1) fixed atmospheric meridional heat transport or 2) completely prescribed surface temperatures. In both cases, the THC is much less robust to perturbations. In contrast, our model shows a monotonic behavior when the power n in (7) and (8) is varied between 0 and 10. Throughout, lower values of n result in weakening of the positive feedbacks (because of the weaker response of atmospheric eddies to temperature anomalies) and the system is more stable to perturbations. However, when we fix temperatures completely and thus eliminate the feedback described in feedback 6 the model lies in stability between the cases $n = 0$ and $n = 1$. Owing to the decoupling between atmospheric and oceanic temperatures in the presence of sea ice, the case of fixed temperatures (mixed boundary conditions) is not equivalent any more to very large n, as it is without ice (Marotzke 1996). Exactly why Lohmann and Gerdes (1998) find that the fixed meridional transport and fixed temperature cases have similar stability properties (although both less stable than the intermediate ones) is not clear to us, but it presumably has to do with a different sensitivity of feedbacks 4 and 6 to a different atmospheric heat transport parameterization.

4. Conclusions

We have used a simple model to investigate the role of sea ice in the sensitivity of the THC in the climate system. This simple model cannot represent some significant features of the real coupled ocean–atmosphere–sea ice system, such as the dynamic spatial structure of sea ice and the turbulent circulations of both the ocean and atmosphere. It has, however, allowed us to more clearly examine some of the feedbacks that are working in the full system, which would not be possible in a full climate model. To this end we have extended the coupled atmosphere–ocean box model of NSM to include a thermodynamic parameterization of sea ice. We emphasize that the model primarily serves to illustrate the resulting feedbacks, not to provide ''realistic'' sensitivities. Moreover, we emphasize that very little is known about large-scale THC–atmosphere–sea ice interaction, which is why we have started with very simple formulations and considerations.

We find that in our model sea ice acts as a destabilizing influence in conjunction with the atmospheric eddy feedbacks. The sea ice primarily causes the atmospheric meridional temperature gradient to be larger, as the air temperature over the sea ice is colder than the polar air temperature without sea ice. The increased atmospheric temperature gradient results in increased atmospheric heat and moisture fluxes, which weaken the ocean density gradient. The moisture transport essentially leads to the positive feedback discussed in NSM, and the heat transport reduces sea-ice thickness, as a direct effect, which is a negative feedback. Additionally, however, since the ocean cannot cool below the freezing point, the ocean temperature gradient is effectively capped, and the cooling of the low latitudes caused by the increased atmospheric heat flux reduces the ocean temperature gradient and hence the thermal forcing of the THC. Both effects act to reduce the ocean heat transport and hence enhance ice thickness, and thus help establish a positive feedback between atmospheric heat transport, sea ice, and the THC. Compared to a simulation without sea ice, a run with sea ice shows an increased response to finite amplitude perturbations, indicating that the positive feedbacks in the system have been strengthened by the presence of sea ice. The net effect is the sum of positive and negative feedbacks, and hence model dependent. Indeed, in the much more complex model of Lohmann and Gerdes (1998), the only study we know that addresses these questions, the presence of sea ice causes the atmospheric eddy heat transport to stabilize the THC when perturbations about an ice-admitting equilibrium are considered.

In contrast to Lohmann and Gerdes (1998) we have also investigated the overall effect of sea ice on the THC, mimicking the comparison of a colder mean climate against one warm enough that, under the same meridional temperature gradient, sea ice would not appear. Our results indicate that the effect of sea ice alone would make the THC less stable in a colder climate. Furthermore, our model results indicate that the many ‘‘ice free’’ studies considering THC–atmosphere interactions have left out a potentially important destabilizing feedback. In this note, we have used a very simple model to point out a number of fundamental processes and questions concerning the role of sea ice in the climate system; more realistic models are needed to investigate them further.

Acknowledgments. Support for this research came from a National Defense Science and Engineering Graduate Fellowship (SRJ), the Tokyo Electric Power Company through the TEPCO/MIT Environmental Research Program, funding from the U.S. Department of Energy (DOE) via the Program for Computer Hardware, Applied Mathematics, and Model Physics (CHAMMP), and 30% came from the National Institute for Global Environmental Change (NIGEC) Northeast Regional Center (JM). Comments from Jiayan Yang and anonymous reviewers were very helpful in improving the manuscript.

REFERENCES

Ahmad, J., G. Beg Paklar, F. Bonjean, G. Dal Bo’, T. Kaminski, F. Kauker, M. Monai, E. Pierazzo, P. Stone, and L. Zampato, 1997: The interaction between global warming and the thermohaline circulation. *Environmental Dynamics Series,* Vol. 4, Istituto Veneto di Scienze, Lettere ed Arti, 87–100 pp.

Lemke, P., 1993: Modelling sea ice–mixed-layer interactions. *Modelling Oceanic Climate Interactions,* J. Willebrand and D. L. T. Anderson, Eds., Springer-Verlag, 243–269.

Lohmann, G., and R. Gerdes, 1998: Sea ice effects on the sensitivity of the thermohaline circulation. *J. Climate,* **11,** 2789–2803.

Manabe, S., and R. J. Stouffer, 1988: Two stable equilibria of a couple ocean–atmosphere model. *J. Climate,* **1,** 841–866.

——, and —— 1995: Simulation of abrupt climate change induced by fresh water input to the North Atlantic Ocean. *Nature,* **378,** 165–167.

Marotzke, J., 1996: Analysis of thermohaline feedbacks. *Decadal Climate Variability: Dynamics and Predictability,* D. L. T. Anderson and J. Willebrand, Eds., Springer-Verlag, 333–378.

——, 1997: Boundary mixing and the dynamics of three-dimensional thermohaline circulations. *J. Phys. Oceanogr.,* **27,** 1713–1728.

——, and P. H. Stone, 1995: Atmospheric transports, the thermohaline circulation, and flux adjustments in a simple coupled model. *J. Phys. Oceanogr.,* **25,** 1350–1364.

Nakamura, M., 1996: Effects of the ice–albedo feedback on the thermohaline circulation. *J. Climate,* **9,** 1783–1794.

——, P. H. Stone, and J. Marotzke, 1994: Destabilization of the thermohaline circulation by atmospheric eddy transports. *J. Climate,* **7,** 1870–1882.

North, G. R., R. F. Cahalan, and J. A. Coakley Jr., 1981: Energy balance climate models. *Rev. Geophys. Space Phys.,* **19,** 91–121.

Rahmstorf, S., 1995: Climate drift in an ocean model coupled to a simple, perfectly matched atmosphere. *Climate Dyn.,* **11,** 447–458.

——, J. Marotzke, and J. Willebrand, 1996: Stability of the thermohaline circulation. *The Warmwatersphere of the North Atlantic Ocean,* W. Krauss, Ed., Gebruder Borntraeger, 129–157.

Rivin, I., and E. Tziperman, 1997: Linear versus self-sustained interdecadal thermohaline variability in a coupled box model. *J. Phys. Oceanogr.,* **27,** 1216–1232.

Ruddick, B., and L. Zhang, 1996: Qualitative behavior and nonoscillation of Stommel’s thermohaline box model. *J. Climate,* **9,** 2768–2777.

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

Stommel, H. M., 1961: Thermohaline convection with two stable regimes of flow. *Tellus,* **13,** 224–230.

Tang, B., and A. J. Weaver, 1995: Climate stability as deduced from an idealized coupled atmospheric–ocean model. *Climate Dyn.,* **11,** 141–150.

Yang, J., and J. D. Neelin, 1993: Sea-ice interaction with the thermohaline circulation. *Geophys. Res. Lett.,* **20,** 217–220.

——, and ——, 1997a: Sea-ice interaction and the stability of the thermohaline circulation. *Atmos.–Ocean,* **35,** 433–469.

——, and ——, 1997b: Decadal variability in coupled sea ice–thermohaline circulation systems. *J. Climate,* **10,** 3059–3076.

Zhang, S., C. A. Lin, and R. J. Greatbatch, 1995: A decadal oscillation due to coupling between an ocean circulation model and a thermodynamic sea-ice model. *J. Mar. Res.,* **53,** 79–106.