A Reexamination of the Polar Halocline Catastrophe and Implications for Coupled Ocean–Atmosphere Modeling

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ABSTRACT

In this paper, the physical mechanism of the polar halocline catastrophe (PHC) is reexamined with emphasis on the role played by the surface heat flux. It is argued that, in a coupled ocean–atmosphere system, thermal changes in the atmospheric state in response to changes in heat flux from the ocean weaken the feedback responsible for the PHC.

So far, the PHC has been observed in models that use mixed boundary conditions; that is, the freshwater flux is specified, but the surface temperature is relaxed to a specified value. Previous explanations of the PHC have focused on the role of the freshwater flux in establishing a freshwater cap and shutting off the deep convection. However, the establishment of a freshwater cap reduces the depth of the water column that is cooled by surface heat loss. As a consequence, the surface temperature is reduced. Since the difference between this and atmospheric restoring temperature is now less, there is a corresponding reduction in the surface heat loss to the atmosphere, and this acts to further stabilize the water column. We examine the importance of this reduction in surface heat loss by considering two numerical experiments that are identical except that one is run under mixed boundary conditions and the other under a flux boundary condition applied to temperature as well as salinity. In each case, the surface fluxes are diagnosed from an experiment run to equilibrium using restoring boundary conditions on both fields. This also provides the initial state for both experiments. A PHC is easily induced in the mixed boundary condition case but not in the case using flux boundary conditions. By reducing the magnitude of the heat flux but not its sign, a pool of fresh water appears at the surface, but its effect is weaker than that under mixed boundary conditions and, in particular, there is no collapse of the meridional overturning circulation. A pool of fresh water also appears in an experiment in which a small, positive heat flux is added at all latitudes, a situation of relevance to global warming. This leads to an initial cooling in a shallow layer at the surface of the polar oceans, before heating at lower latitudes leads to a collapse of this state.

These experiments show that the reduction in the surface heat flux that occurs when the PHC develops under mixed boundary conditions is an essential feature of the PHC. The use of mixed boundary conditions assumes that the atmospheric state is fixed and does not respond to changes in heat flux from the ocean. If the atmosphere were allowed to adjust to changes in this heat flux, then a PHC would be less likely to occur. This has been demonstrated by coupling the ocean model to the zero-heat-capacity atmospheric model used by Schopf. This is justified, following Bretherton, because of the large horizontal scale of the sea surface temperature (SST) anomalies in the experiments. The authors were unable to induce a PHC with this model. In reality, the atmospheric boundary condition seen by the ocean lies somewhere between the two extremes of mixed boundary conditions, on the one hand, and Schopf's model on the other. We have investigated this intermediate region by conducting experiments in which SST anomalies are damped on successively shorter time scales. These show that if the damping time is reduced sufficiently, a PHC can again be induced.

1. Introduction

Following the work of Bryan (1986a, 1986b) many authors have used mixed boundary conditions (i.e., a restoring boundary condition on temperature and a flux boundary condition on salinity) to drive ocean models (e.g., Marotzke et al. 1988; Marotzke 1989; Weaver et al. 1991a,b,c, 1992; Wright and Stocker 1991; Stocker and Wright 1991a,b; Marotzke 1991; Marotzke and Willebrand 1991). The usual procedure is to spin up a model to equilibrium using restoring boundary conditions for both temperature and salinity and then diagnose the corresponding surface freshwater flux in this state. The boundary condition on salinity is then changed to this freshwater flux and the resulting model behavior examined. If the final state of the restoring boundary-condition case were stable under a switch to mixed boundary conditions, then no change in the model-computed solution should occur. Instead,

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many different kinds of behavior have been found, ranging from the polar halocline catastrophe (Bryan, 1986a, 1986b), to "flushes" (Marotzke 1989; Weaver and Sarachik 1991b), low-frequency oscillations (Marotzke 1989), and decadal time-scale oscillations (Weaver and Sarachik 1991a). Bryan (1986a) describes the polar halocline catastrophe (PHC) as a strengthening and spreading of the polar halocline. He noted that as the halocline develops, deep convection is interrupted and the meridional overturning circulation weakens dramatically while at the same time becoming increasingly confined to the upper layers of the ocean. In this way, the dense bottom water is no longer ventilated from the surface and is gradually warmed diffusively. Eventually, the bottom water becomes sufficiently warm that it is less dense than the fresh water overlying it, and a dramatic overturning event, that is, a flush, as described by Marotzke (1989), will result.

The great interest in these phenomena arises because of their implications for climate variability. This interest goes back to Stommel (1961). He demonstrated the existence of thermohaline convection with two stable regimes of flow, depending on whether temperature or salinity dominates the density differences in the system. He recognized the possibilities implied by his result for speculation about climate change. In a recent review, Broecker and Denton (1990) discuss the possibility that transitions in and out of periods of glaciation were accompanied by major reorganizations of the atmosphere–ocean system and of the ocean circulation in particular. The reason the ocean is important is because of its ability to store and transport heat. For example, a PHC is associated with a period often lasting many hundreds of years, during which the oceanic poleward heat transport is much reduced while at the same time the model deep ocean absorbs an enormous amount of heat. This heat is released to the atmosphere in only a few years during the subsequent "flush," sometimes at a rate of up to 100 W m^{-2} when averaged over the surface of the entire model domain, as noted by Weaver et al. (1992). Recent work has been directed to examining the robustness of this behavior as the surface boundary conditions on the models are made more realistic. For example, Weaver et al. (1992) have shown that including a stochastic component to the surface freshwater flux increases the frequency of "flushes" and decreases their intensity. These authors have also shown that the structure of the surface freshwater flux is very important for determining the behavior of a model upon a switch to mixed boundary conditions. In particular, strong high-latitude freshening seems to be important for a PHC to occur.

The annual heat budget at the top of the atmosphere for the current climate implies an excess of about 50–100 W m^{-2} at the equator and a deficit of the same order at the pole (Peixoto and Oort 1984). This in turn requires a meridional transport of heat to maintain energy balance. The heat transport is accomplished by the atmosphere and the ocean in approximately equal proportions (see, for example, Gill 1982). It follows that there is a zonally averaged net heat budget at the air–sea interface of magnitude 25–50 W m^{-2}. We shall show that under mixed boundary conditions, the onset of a PHC requires a much-reduced air–sea heat flux. As the restoring atmospheric temperature is held constant, there can be no change in the meridional heat transport by the atmosphere. It follows that the heat budget of the atmosphere is not satisfied under these conditions. An alternative to mixed boundary conditions is the use of a flux condition on temperature as well as salinity. This will allow for the heat budget to be satisfied but does not allow for any change in this budget. Indeed, as with mixed boundary conditions, flux conditions for both temperature and salinity only represent approximations to the coupled atmosphere–ocean system.

The use of mixed boundary conditions implies an unchanging atmosphere, that is, an atmosphere of infinite heat capacity. In reality, there is a large difference in the heat capacities of the atmosphere and the upper ocean, with the heat capacity of the atmosphere being roughly a factor 20 less (Dickinson 1981). This suggests an alternative to mixed boundary conditions; that is, to couple an ocean model to an atmosphere that has zero heat capacity (note that in the model we shall describe, the assumption of zero heat capacity applies as far as local changes in heat content are concerned and does not prevent the atmosphere from transporting heat). This approach has been used previously by Schopf (1983). His was an early attempt to show the influence of an interactive atmosphere on the development of sea surface temperature (SST) anomalies in an equatorial ocean model. By assuming that the transport of heat by the atmosphere is fixed in time, the zero-heat-capacity model implies that SST anomalies are damped through radiative relaxation from the atmosphere, giving a damping time roughly twenty times longer than in a model with a specified atmospheric temperature (as with mixed boundary conditions). The appropriateness of this model at middle and high latitudes rests on the assumption that there is no change in the atmospheric heat transport. This is because advection of heat by the atmosphere can play an important role in removing SST anomalies at these latitudes, as noted by Bretherton (1982). Bretherton points out that the importance of this effect depends on the horizontal scale of the anomaly. SST anomalies on the scale of a few tens of kilometers are rapidly removed by the atmosphere due to transport of heat by the wind; whereas global scale anomalies can only be removed by radiative relaxation (as in the model of Schopf) and so have a much longer lifetime. He quotes results from atmospheric general circulation models in support of this view.

In this paper we reexamine the physical mechanism responsible for the PHC. Previous authors have emphasized the role of the freshwater flux from the atmosphere. In this paper, we concentrate instead on the surface heat flux and then go on to examine the robustness of the PHC under changing atmospheric thermal conditions. The use of mixed boundary conditions assumes that the atmospheric state is fixed and unaffected by the changes in the heat flux from the ocean. Under mixed boundary conditions, the high latitude heat loss from the ocean to the atmosphere is greatly reduced when a PHC occurs. This reduction in heat loss acts to stabilize the water column and reenforce the effect of the developing freshwater cap. We shall show that this reduction in heat loss is an essential feature of the PHC. Without it, a PHC is very difficult to induce. We argue that if the atmosphere were allowed to respond thermally to changes in the ocean, as in a coupled ocean–atmosphere system, then the reduction in heat loss would be less than is found under mixed boundary conditions and that, as a result, a PHC is less likely, or at least will be less severe in its effect. We demonstrate this by coupling our ocean model to the zero-heat-capacity atmospheric model used by Schopf (1983). As noted above, this has the effect of damping SST anomalies by radiative relaxation from the atmosphere. We argue that this model is relevant because of the large horizontal scale of the SST anomalies associated with a PHC. We also discuss other experiments in which SST anomalies are damped on successively shorter time scales, in order to bridge the gap between Schopf's model and mixed boundary conditions.

The plan of this paper is as follows. Section 2 briefly describes the ocean model we use, and in section 3 we present the results of some model experiments. We find that on a switch to flux boundary conditions on both temperature and salinity, a PHC is not easily induced. We explain this by reexamining the PHC and, in particular, the role played by the surface heat flux. We illustrate our conclusions by describing other experiments in which the heat flux is varied and also, in section 4, cases in which the ocean is coupled to an atmospheric model that can respond thermally to changes in the ocean; for example, the zero-heat-capacity atmospheric model of Schopf (1983). Section 5 provides a summary and discussion.

2. The numerical model

The model used is the planetary-geostrophic (PG), ocean general circulation model described by Zhang et al. (1992, hereafter ZLG). It is a generalization to many levels of Killworth's (1985) two-level model. The governing equations consist of the full prognostic temperature and salinity equations and diagnostic momentum equations. The latter consist of the geostrophic balance, a linear friction in the vertically averaged part in order to provide a western boundary current, and wind-stress forcing as a body force acting over the top level of the model. The equation of state is quadratic in temperature and linear in salinity and is given by

$$\rho(T, S) = 3.0 + 0.77S - 0.072T(1 + 0.072T). \quad (1)$$

Here T is temperature in celsius, S is salinity in practical salinity units and ρ is (density − 1000) in units of kg m^{-3}. The coefficients are similar to those of Bryan and Cox (1972) but with compressibility effects ignored. In addition, the convective overturning algorithm ensures a completely stable profile after mixing (Marotzke 1991). The full model equations and their physical justification can be found in ZLG. All the experiments described in this paper are run using "no slip" conditions on the coastal boundaries. As found by Killworth (1985), the "no slip" condition can be substituted for one of "no normal flow" without significantly changing the results. Zhang et al. (1992) noted that additional physics must be added to the model to allow either of these conditions to be satisfied in the limit that the grid spacing goes to zero. In particular, horizontal Laplacian mixing is required in the case of the no slip condition (Colin de Verdiere 1988, 1989) and Rayleigh friction in the vertical momentum equation for the case of the "no normal flow" condition (Salmon 1986, 1990). To include either would involve a substantial increase in computational effort and is not done either here or in ZLG. Zhang et al. showed that despite these difficulties with the boundaries, the model yields very similar results to those of the well-known Bryan–Cox model (Bryan 1969; Cox 1984). In particular, ZLG compared results of experiments using the two models with the same model geometry and surface forcing and found their final equilibrium states to be very similar. Additional experiments have shown that the model also exhibits the same behavior under mixed boundary conditions as the Bryan–Cox model, for example, the PHC, flushes, and decadal and longer time-scale oscillations (see the Introduction). The advantage of the present model is its computational efficiency. As such, the model is an attractive alternative to using the distorted physics technique for accelerating the convergence to equilibrium of primitive-equation models (Bryan 1984). Also, the model can be used to study transient climate problems, for which the use of the distorted physics technique is not strictly justified, while at the same time using only modest computer resources. An example is the work presented in this paper.

The model uses spherical coordinates. For this study, the model domain consists of a flat-bottomed, 60° square box in latitude–longitude space extending between 5° and 65°N. The horizontal resolution is the same as in ZLG, that is, 2° in both latitude and longitude; 14 levels are used in the vertical. Table 1 gives the depth at the center of each level. The surface wind-stress forcing is zonally uniform. The eastward component is shown in Fig. 1, the northward component is zero. Restoring boundary conditions on either or

TABLE 1. The depths of the center of each model level.

Layer number	Depth
1	23 m
2	75
3	140
4	223
5	327
6	458
7	623
8	831
9	1093
10	1423
11	1838
12	2362
13	2990
14	3664

both of the temperature (T) and salinity (S) fields refers to the use of a Haney-type condition to specify the flux through the surface (Haney 1971). The "equivalent atmospheric" values used are shown in Fig. 1. Values of T and S in the top level of the model are relaxed to these values on a time scale of 30 days (this choice of 30 days is dependent on the depth of the surface level in the model, in this case 46 m). In all the experiments, the vertical and horizontal diffusivity used in the T and S equations are kept fixed at the values 0.63×10^{-4} and 2×10^3 m^2 s^{-1}, respectively. The numerical experiments are listed in Table 2. The control run (experiment 1) is obtained by spinning up the model under restoring boundary conditions on both T and S. The initial state is one of rest with uniform values for T (4°C) and S (35 psu). The model is run until a statistically steady state is reached with no trends in basin mean quantities. Model variables at the end of this calculation are shown in Fig. 2. Since in the vertical, a tick mark denotes the center of each level, the figures are distorted in the vertical direction. This has the advantage of emphasizing that part of the water column occupied by the thermocline. The meridional overturning streamfunction is dominated by an overturning cell with sinking near the northern boundary and smaller cells near the surface that are due to the wind-driven Ekman pumping. This structure, and also that of the pycnocline as revealed by the other fields, is very similar to that found in other studies (for example, see Fig. 8 in ZLG).

3. Model results

In all experiments described in this section, the model runs use as their initial condition the final state of the control run shown in Fig. 2. We begin with experiment 2 (see Table 2). This uses mixed boundary conditions; that is, the restoring boundary condition is applied to temperature but the surface boundary condition on salinity is replaced by the flux diagnosed from the final state of the control run and is the average over the last three months of integration (this is shown in Fig. 3b and is not significantly changed by averaging over a longer period). A negative salinity anomaly of −0.04 psu is added to the initial state in each of the top three levels north of 50°N. As found by previous authors (e.g., Bryan 1986a; Weaver and Sarachik 1991b), a PHC ensues. This is illustrated in Fig. 4, which shows model variables 20 years into the integration. Comparing with Fig. 2, we see that there has been a dramatic freshening of the surface layer in high latitudes, sufficient to cause a reversal in the north–south density gradient at the surface in the northern part of the basin. At this time, the overturning circulation has collapsed with the meridional overturning streamfunction now resembling that in the wind-driven-only case shown in Fig. 8b of ZLG. At this time a freshwater cap covers the high latitudes in the model.

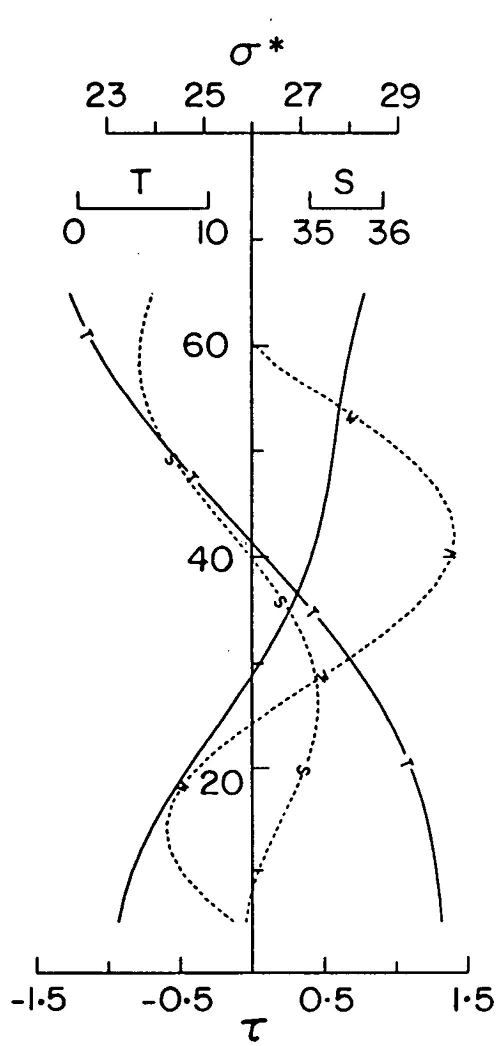

FIG. 1. The latitudinal (°N) distributions of the surface wind stress (labeled W), atmospheric "equivalent" temperature (T), and salinity (S). The corresponding "equivalent" atmospheric density, calculated using Eq. (5), is shown by the unlabeled solid curve.

TABLE 2. Summary of the numerical experiments.

		Surface boundary conditions				
		Temperature		Salinity		
	Experiment	Restoring	Flux	Restoring	Flux	Initial salinity anomaly (psu)
1	control	1	0	1	0	0
2	mixed	1	0	0	1	−0.04
3	both flux	0	1	0	1	−0.04
3.1	both flux	0	1	0	1	−1.00
4	90% heat flux	0	0.9	0	1	0
5	addition of 1.77 W m^{-2}	0	1 + 1.77	0	1	0
6	300-day mixed	0.1	0.9	0	1	−0.04
6.1	500-day mixed	0.06	0.94	0	1	−0.04
7	200-day mixed	0.15	0.85	0	1	−0.04

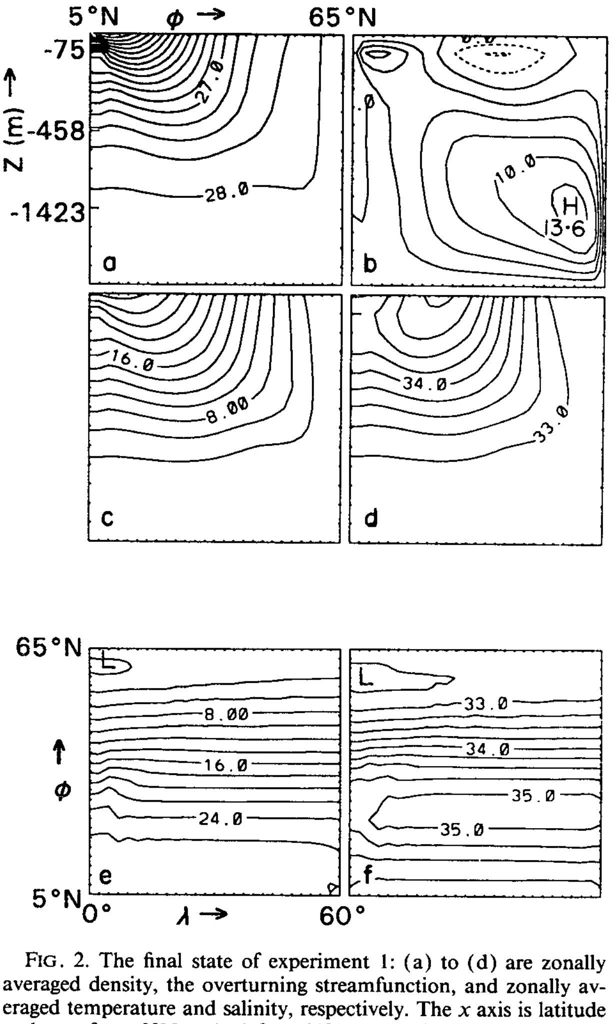

FIG. 2. The final state of experiment 1: (a) to (d) are zonally averaged density, the overturning streamfunction, and zonally averaged temperature and salinity, respectively. The x axis is latitude and runs from 5°N at the left to 65°N at the right. The y axis is in the vertical direction. A tick mark denotes the center of each model level, the depths of which are given in Table 1. This means that the figure is distorted in the vertical direction, with the upper part expanded. (e) and (f) are the top model-level temperature and salinity, respectively.

The absence of convective overturning can be seen by looking at the plots of zonally averaged density, temperature, and salinity (Figs. 4a,c,d). These show strong vertical structure at all latitudes—there is an absence of the vertically well-mixed region near the northern boundary in Fig. 2. What is particularly interesting is that the water is colder at the surface in the northern latitudes than it is underneath (Fig. 4c); the water column is, however, stabilized by salinity effects. This shows the insulating effect of the freshwater cap. That is, heat loss to the atmosphere is confined near the surface rather than being spread over the entire water column by the convective overturning. What is perhaps less obvious is that the heat loss to the atmosphere rapidly reduces as the freshwater cap builds up. This is because of the use of the restoring condition on temperature. Figure 4g shows the temperature difference between the top level of the ocean model shown in Fig. 4e and the "equivalent atmospheric temperature" shown in Fig. 1. This temperature difference is proportional to the heat flux from the atmosphere to the ocean. Figure 3a shows the corresponding temperature difference diagnosed from the last three months of the

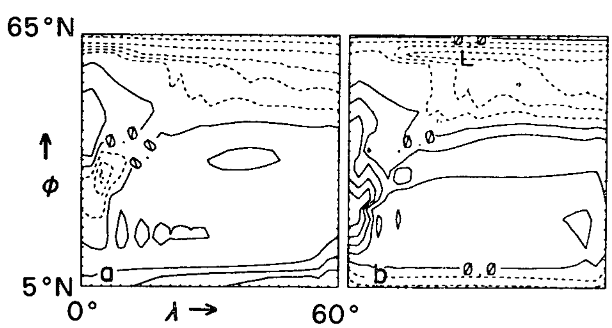

FIG. 3. (a) The heat flux from the atmosphere to the ocean averaged over the last three months of experiment 1. The contour interval is 0.5°C mo^{-1} for the 46-m top model level used in our experiments and (b) the corresponding freshwater flux contoured at an interval of 0.05 psu mo^{-1} for the 46-m top model level.

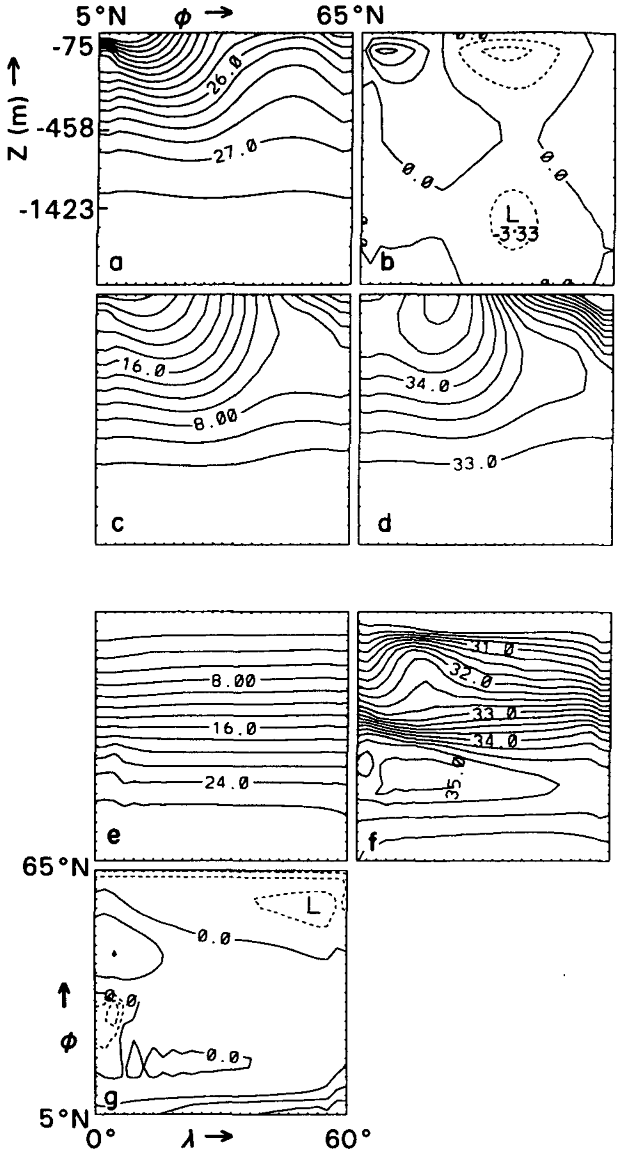

FIG. 4. Panels (a)–(f) as in Fig. 2 but for experiment 2 at 20 years. Panel (g) shows the heat flux from the atmosphere to the ocean at this time contoured at an interval of 0.5°C mo^{-1} for the 46-m top model level.

control run. Comparing these figures, it is clear that a reduction in the heat loss has occurred at high latitudes in experiment 2. As we now demonstrate, this reduction, which is associated with the use of mixed boundary conditions, is an essential feature of the PHC.

Experiment 3 is the same as experiment 2 except that the restoring boundary condition on temperature is now replaced by a flux boundary condition. As for salinity, this is diagnosed from the final state of experiment 1 and is an average over the last three months of integration; it is shown in Fig. 3a. As in experiment 2, a negative salinity anomaly of −0.04 psu is added in the top three levels to the initial condition north of 50°N. This time no PHC occurs, and there is no disruption of the meridional overturning circulation, although a small pool of cold, fresh water does become established in the northwestern part of the basin. Model variables 100 years after the start of the integration are shown in Fig. 5. The small fresh pool can be seen, but otherwise Fig. 5 is very similar to Fig. 2. Furthermore, the model remained in this state for the several thousand years of the integration, with no sign of any tendency to break down. The only difference between this and experiment 2 is that the surface heat flux has been kept constant; otherwise the model boundary and initial conditions are identical. It follows that the reduction in heat flux, which is apparent by comparing Figs. 4g and 3a, is *essential* for the PHC to occur. It should also be noted that the model failed to exhibit a PHC even when initialized with a negative salinity anomaly as large as −1.0 psu (experiment 3.1); although in this case, the meridional overturning circulation did weaken during the first 10 years, before becoming reestablished again. In this experiment, a very cold, fresh residual pool persists near the northern boundary.

In order to understand this behavior, we must examine more closely the physical mechanism of the

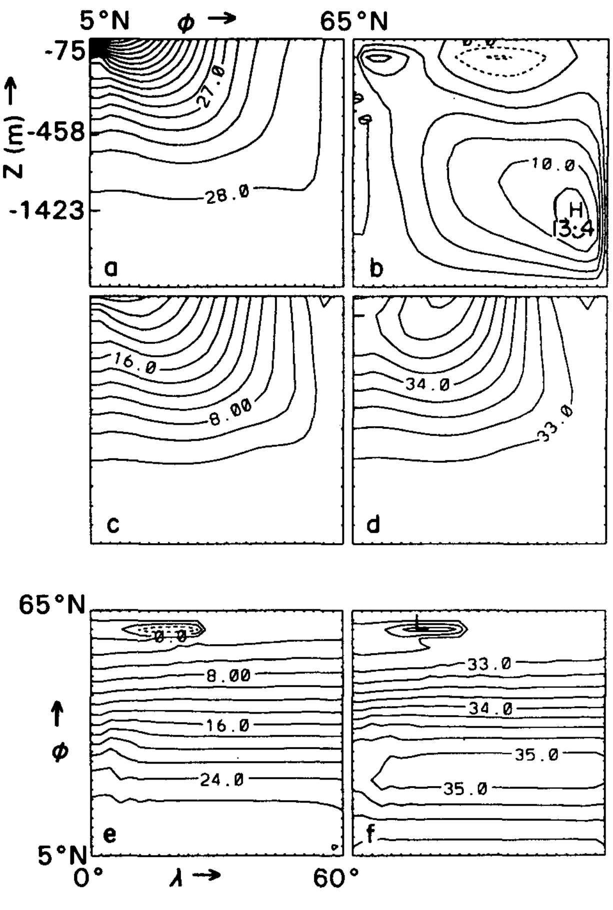

FIG. 5. As Fig. 2 but for experiment 3 at 100 years.

PHC. Weaver et al. (1992) have noted the importance of sufficiently strong high-latitude freshening. Here, our interest is in the role of the surface heat flux. As noted in the introduction, the occurrence of the PHC in an ocean general circulation model was first described by Bryan (1986a). He found that about 80 years after the change from restoring to mixed boundary conditions, the polar halocline strengthened and spread equatorward. At the same time, deep convection was interrupted and the meridional overturning circulation dramatically weakened. Marotzke (1991) has shown that the spontaneous occurrence of this behavior is related to the manner in which the standard Bryan–Cox code used by Bryan (1986a,b) handles convective overturning. However, even if a convective overturning routine is used that guarantees a completely stable water column (such as the one used in our model), the PHC can be easily induced by introducing a small freshwater anomaly at the surface in high latitudes, as we did in experiment 2 and as described by Weaver and Sarachik (1991b). The appearance of the freshwater anomaly acts to reduce the depth to which convective overturning penetrates. This, in turn, means that the flux of fresh water across the sea surface is mixed over a shallower depth than before, leading to an increase in the strength of the freshwater anomaly. At the same time, the heat lost through the surface is also removed from a shallower layer, leading to reduced surface temperatures in the model. When mixed boundary conditions are used, for which the surface heat loss is proportional to the difference between the temperature in the surface level of the model and the specified atmospheric temperature, this has the effect of greatly reducing the heat loss to the atmosphere (cf. Figs. 4g and 3a). This acts to further stabilize the water column. If a flux boundary condition is applied to temperature (as in experiment 3), the surface heat loss to the atmosphere remains the same. As the depth over which the water column is mixed is reduced, the temperature in that column is less than that of the mixed boundary condition case, thus, making the water column less stable. In experiment 3, this is sufficient to prevent the water in the region of the freshwater anomaly from remaining less dense than that underneath, and convective overturning soon penetrates again to the bottom. In this way, the PHC is prevented from occurring.

In order to further illustrate this effect, we now describe an experiment in which the heat flux is reduced in magnitude everywhere by 10%. This is experiment 4 in Table 2. Again, the initial condition is the final state of experiment 1, but this time without any salinity anomaly being added. Figure 6 shows model variables 10 years after the start of the integration. The reduced heat loss in high latitudes has allowed the development of a freshwater cap, due to the reduced convection and the fact that the freshwater flux is held constant. This time the freshwater cap does not extend all the way to the northern boundary, but exists as a pool of fresh, surface water just to the south. Indeed, convective overturning is still continuing at the northern boundary, as indicated by the vertical isopycnals in Fig. 6a. Furthermore, the meridional overturning streamfunction (Fig. 6b) shows only a slight weakening of the overturning circulation. Even after 50 years (not shown) it has only weakened to 11 Sv (Sv $\equiv 10^6$ m^3 s^{-1}) and indeed, at later times does not collapse as in the mixed boundary condition case (Fig. 4). A feature of this solution is the very cold water at the surface within the freshwater pool. This is because, although the heat loss to the atmosphere has been reduced by 10% from experiment 3, it has not been reduced as much as in the mixed boundary condition case (Fig. 4g). Heat is, therefore, removed from the freshwater cap even after it has formed. Since there is no convective overturning, this heat loss is distributed only over a small part of the water column and low temperatures can result. Clearly, if we were to include an ice model, the heat loss would result in ice formation and more realistic surface temperatures. However, the corresponding release of salt would further inhibit the collapse of the

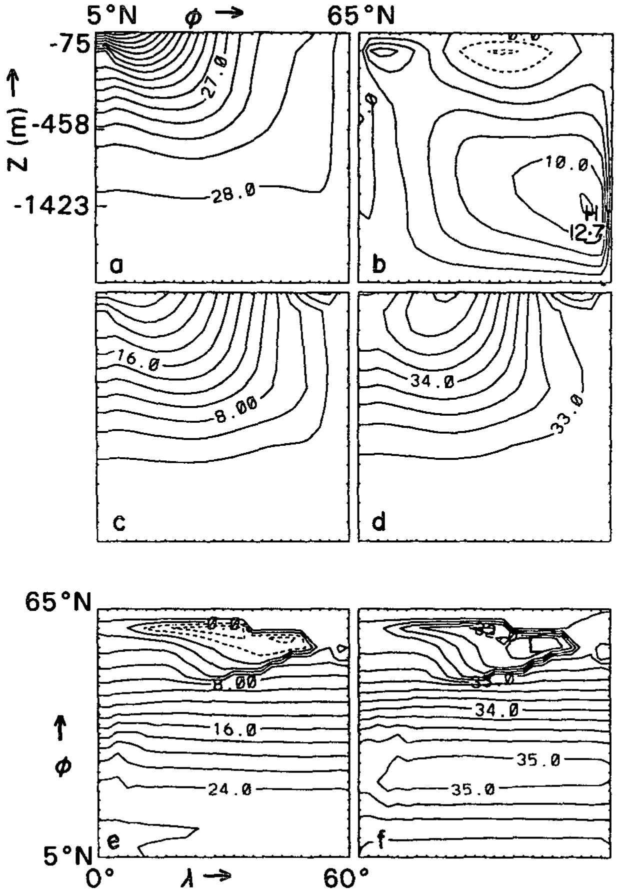

FIG. 6. As Fig. 2 but for experiment 4 at 10 years.

thermohaline circulation, as has been demonstrated using a simple ice model (Zhang et al., in preparation).

The failure of the meridional overturning streamfunction to collapse in this experiment is interesting. Zhang et al. (1992) have argued that the sense of the overturning circulation is governed by the combination of a southward pressure gradient at the surface, driving eastward geostrophic flow, the presence of meridional boundaries that block this flow and, finally, a stably stratified pycnocline that is guaranteed by convective overturning. In experiment 4, the surface waters continue to become denser as one moves northwards, at least after zonally averaging, as can be seen from Fig. 6a. It follows that the north–south pressure gradient remains intact at the surface, driving eastward geostrophic flow and, hence, the meridional overturning. This contrasts with the mixed boundary condition case (Fig. 4a), in which the north–south density gradient reverses at high latitudes.

Next we consider experiment 5, which is the same as experiment 3 except that a positive heat flux of 1.77 W m^{-2} is added everywhere and, as in experiment 4, no salinity anomaly is added. This has the effect of reducing the heat loss at high latitudes, but increasing the heat gain by the ocean in low latitudes. Such a situation could conceivably occur in the event of global warming (although a more realistic model would have a gradual increase in the heat input from the atmosphere to the ocean, rather than the sudden "switch on" used here, and would also allow for latitudinal structure in this input). Model variables after 10 years are shown in Fig. 7. As in experiment 4, the reduction in heat loss at high latitudes leads to the development of a freshwater pool. Again, we find that the surface water has very low temperatures within this pool—an interesting effect in a model experiment in which heat is added at all latitudes! (In connection with this, some coupled ocean–atmosphere models exhibit cooling in the surface waters around Antarctica following a gradual increase in the CO_2 content of the model atmosphere; for example, Bryan et al. (1988) and Stouffer et al. (1989). It can be seen that the meridional overturning circulation in Fig. 7 is still intact and, in fact, has increased slightly from that in the initial state (Fig. 2). However, after 50 years (see Fig. 8) the circulation is collapsing, the freshwater pool has disappeared, and even the surface waters in high latitudes are up to 4°C warmer than in the initial state. The important point here is the appearance of the freshwater pool in response to a reduction in the heat loss at high latitudes. In the next section we examine the consequence of these results when our model is coupled to a very simple model of the atmosphere.

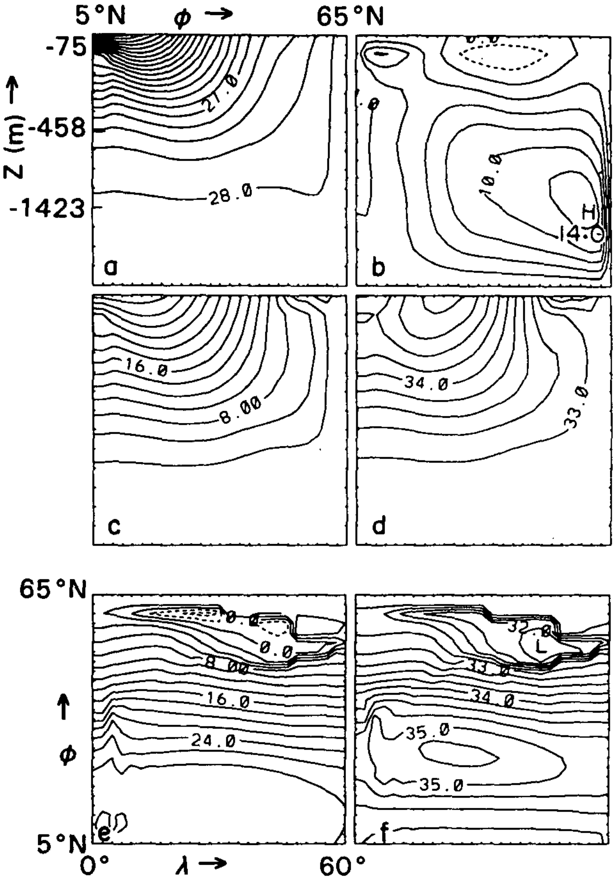

FIG. 7. As Fig. 2 but for experiment 5 at 10 years.

4. Coupling to a simple, thermally interactive atmosphere

The experiments described in the last section demonstrate that the reduction in heat loss at high latitudes is an essential feature of the PHC under mixed boundary conditions (at least that which occurred in experiment 2). Mixed boundary conditions assume that the atmospheric state is fixed and does not change in response to the changes in the heat flux from the ocean. In effect, the atmosphere is assumed to have infinite heat capacity. Dickinson (1981) points out, however, that the heat capacity of the atmosphere is much less than that of the ocean surface mixed layer and that as a result there is a strong tendency for the air temperature above the ocean to closely follow that of the ocean surface below. The exception to this is in regions of strong atmospheric transport of heat, such as along the eastern continental margins in winter during outbreaks of cold, continental air. As an example, Schopf (1983) has pointed out that the air–sea temperature difference in the tropics is rarely more than 1°C, even in areas where strong upwelling brings cold waters to the surface. He suggested that a better approach to model heat exchange between the atmosphere and the ocean (at least in the tropics) is to assume that the atmosphere has zero heat capacity.

Following Schopf, we, therefore, consider the following very simple model for the atmosphere and its coupling to the ocean:

$$C_a \frac{\partial T_a}{\partial t} = -K(T_a - T_o) - K'_r T_a + Q_a, \quad (2)$$

$$C_o \frac{\partial T_o}{\partial t} = K(T_a - T_o) + Q_o. \quad (3)$$

Here T_o is the sea surface temperature (taken to be the same as T in the top grid level of the model), and T_a is a representative atmospheric temperature; C_a and C_o are the heat capacities of the atmosphere and the ocean mixed layer; K'_r is the atmospheric radiative feedback constant, and Q_a and Q_o stand for the atmospheric and oceanic heat sources not included explicitly in the above equations (e.g., due to advection and transport by eddies). Since C_a/C_o is of the order of 0.045, we follow Schopf and set $C_a = 0$; that is, we assume that the heat capacity of the atmosphere is zero as far as local changes in T_a are concerned. We now define T_r by

$$T_r = Q_a/K'_r, \quad (4)$$

so that (2) becomes

$$0 = -K(T_a - T_o) - K'_r(T_a - T_r). \quad (5)$$

Rewriting (5), to express T_a in terms of T_r and T_o, and substituting for T_a in (3), now gives the following equation for T_o:

$$C_o \frac{\partial T_o}{\partial t} = K_r(T_r - T_o) + Q_o, \quad (6)$$

where

$$K_r = \frac{KK'_r}{(K + K'_r)}. \quad (7)$$

Typically, $K = 45$ W m^{-2} K^{-1} (Haney 1971) and $K'_r = 2.4$ W m^{-2} K^{-1} (Dickinson 1981), so that K_r is a factor of about 20 less than K. This means that instead of restoring sea surface temperature, T_o, to T_a, with a restoring time of 30 days, as indicated by Eq. (3), we can equivalently restore T_o to the temperature T_r, as indicated by Eq. (6), but using the much longer restoring time of close to 600 days.

An important question arises as to whether or not T_r in (6) is allowed to vary in response to changes in sea surface temperature. For example, we can mimic mixed boundary conditions with this model if we choose T_r so that T_a in (3) remains constant, independent of time. The required T_r is obtained by rearranging (5) with T_a fixed to give

$$T_r = (K(T_a - T_o) + K'_r T_a)/K'_r. \quad (8)$$

In this model, anomalies in sea surface temperature are removed on the "fast" time scale $1/K$ of 30 days. Clearly, however, varying T_r as required by (8) is equivalent to varying the atmospheric heat transport Q_a [this follows from (4)]. Since under mixed boundary conditions, the atmospheric structure is assumed to be fixed and independent of time, this represents a contradiction. In fact, as we noted in the Introduction, it is impossible to satisfy the atmospheric heat budget under mixed boundary conditions once changes in the ocean–atmosphere heat flux occur.

Bretherton (1982) has noted that the importance of atmospheric heat transport for damping sea surface temperature (SST) anomalies depends on the horizontal scale of the anomaly. For anomalies of scale a few tens of kilometers or less, atmospheric heat transport plays a major role. Indeed, the dominant mechanism for the removal of heat is advection by the wind, with most of the heat returning to the ocean elsewhere on the globe. Under this mechanism, anomalies are damped on the time scale of tens of days, as under mixed boundary conditions. On the other hand, he argues that global-scale anomalies will be damped by a quite different mechanism; that is, radiative relaxation from the atmosphere. This is because for global-scale anomalies, there is nowhere for heat to be advected and reabsorbed by the ocean. Rather, the heat associated with the anomaly must be lost to space. This process occurs on the much longer time scale of hundreds of days associated with $1/K_r$ in (6) and corresponds to relaxing the ocean surface temperature, T_o, to a fixed temperature T_r that is independent of time. He justifies his argument by appealing to results from atmospheric general circulation models. From this point of view, mixed boundary conditions are appropriate in the limit of very small-scale SST anomalies (which is why for finite-scale anomalies the atmospheric heat budget cannot be satisfied). As the horizontal scale of the anomaly increases, atmospheric heat transport becomes a less efficient process, and the damping will take place on a longer time scale, approaching that of radiative relaxation from the atmosphere in the limit of global-scale anomalies. In this limit, T_a, rather than T_r, varies in response to changes in SST. In fact, using (5) to solve for T_a with T_r fixed gives

$$T_a = (KT_o + K'_r T_r)/(K + K'_r). \quad (9)$$

Since $K'_r \ll K$, this reduces to $T_a = T_o + O(K'_r/K)$ showing that the representative atmospheric temperature is never far removed from the underlying SST. This indicates much smaller variations in the heat flux from the ocean to the atmosphere than would be the case under mixed boundary conditions. This is the zero-heat-capacity model adopted by Schopf (1983) and will subsequently be referred to as Schopf's model.

Since the changes in SST associated with a PHC (see Fig. 4e) have basin scale, it seems appropriate to investigate Schopf's model further by using it to provide the thermal boundary condition at the surface of our ocean model. Indeed, the above argument suggests that it is a plausible alternative to mixed boundary conditions (later we shall describe experiments that use a boundary condition that is intermediate between these two extremes). Since variations in the ocean–atmosphere heat flux will be less than under mixed boundary conditions, the results of section 3 suggest that the development of a PHC will be inhibited. To verify this, we have run experiments in which Eq. (6), with T_r fixed in time, is used to provide a restoring surface boundary condition on temperature, and a flux condition, shown in Fig. 3b, is applied to salinity, as in the experiments in section 3. To be consistent with the previous experiments, we wish to initialize with the final state of experiment 1. Letting T_1 be the surface temperature in this state, we can write the heat flux from the atmosphere to the ocean in Eq. (6) as

$$K_r(T_r - T_o) = K_r(T_r - T_1) + K_r(T_1 - T_o). \quad (10)$$

In this way, the flux is split into two parts. The term $K_r(T_r - T_1)$ is the flux from the atmosphere to the ocean in the final state of experiment 1 and is the diagnosed flux used to provide the thermal boundary condition in experiment 3 and shown in Fig. 3a. On the other hand, the $K_r(T_1 - T_o)$ term tries to remove changes in the ocean surface temperature, but only on the "slow" time scale C_o/K_r of several hundred days. What this term actually does is damp changes in sea surface temperature through radiative relaxation by the atmosphere. This is quite different from the mixed boundary condition case (experiment 2) in which changes in ocean surface temperature are removed on the "fast" time scale C_o/K of several tens of days.

We begin by showing results obtained with K_r chosen to correspond to a restoring time of 300 days in the surface level of the model (experiment 6 in Table 2). As in experiments 2 and 3, a negative salinity anomaly of 0.04 psu is added to the top three levels north of 50°N. As in experiment 3, there is no collapse of the meridional overturning circulation and no PHC such as occurred under mixed boundary conditions in experiment 2. However, a large pool of fresh water does develop at high latitudes and is associated with very cold surface temperatures. This can be seen in Fig. 9, which shows the model variables after 10 years of integration. In this respect, experiment 6 resembles experiment 4, in which the magnitude of the surface heat flux was reduced by 10%. This is not surprising, as can be seen by writing (10) in the form

$$K_r(T_r - T_o) = 0.9\text{FLUX} + 0.1\text{FLUX} + K_r(T_1 - T_o), \quad (11)$$

where FLUX is the heat flux from the atmosphere to the ocean at the end of experiment 1. This is given by

$$\text{FLUX} = K(T_a^* - T_1),$$

where T_a^* is the "equivalent" atmospheric temperature used in the spinup of experiment 1 and shown in Fig. 1. Noting that for this experiment $K_r = 0.1$ K, Eq. (11) can be rewritten

$$K_r(T_r - T_o) = 0.9\text{FLUX} + 0.1K(T_a^* - T_o), \quad (12)$$

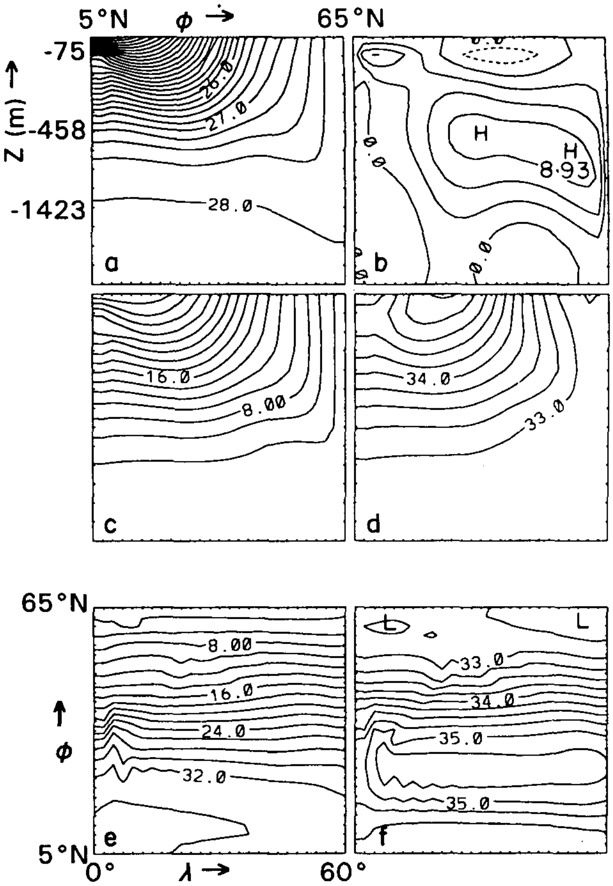

FIG. 8. As Fig. 2 but for experiment 5 at 50 years.

showing that this experiment differs from experiment 4 only by the addition of a slow relaxation (on a time scale of 300 days) of the ocean surface temperature to T_a^*. An interesting feature of this experiment is that the freshwater pool is advected towards the eastern boundary where it downwells, only to reappear again in the interior of the high latitude ocean (see Fig. 10) before being advected once more to the eastern boundary. This process occurs three times in the first 60 years of the integration and is revealed by a time series of the number of grid points at which convective overturning occurs, shown in Fig. 11. This is reminiscent of the decadal oscillations noted by Weaver and Sarachik (1991a) except that in this case they are not self-sustaining. After 70 years, the model settles down and remains in a state essentially the same as that of the initial condition shown in Fig. 2. An experiment with K_r, chosen to give a restoring time scale of 500 days, shows very similar behavior, with three oscillations of the 20-year time scale occurring before the model settles back into its initial state, but with a weaker freshwater pool. Figure 11 contrasts the behavior of experiment 6 with the mixed boundary condition case, experiment 2. Note how in the latter, the heat content of the ocean

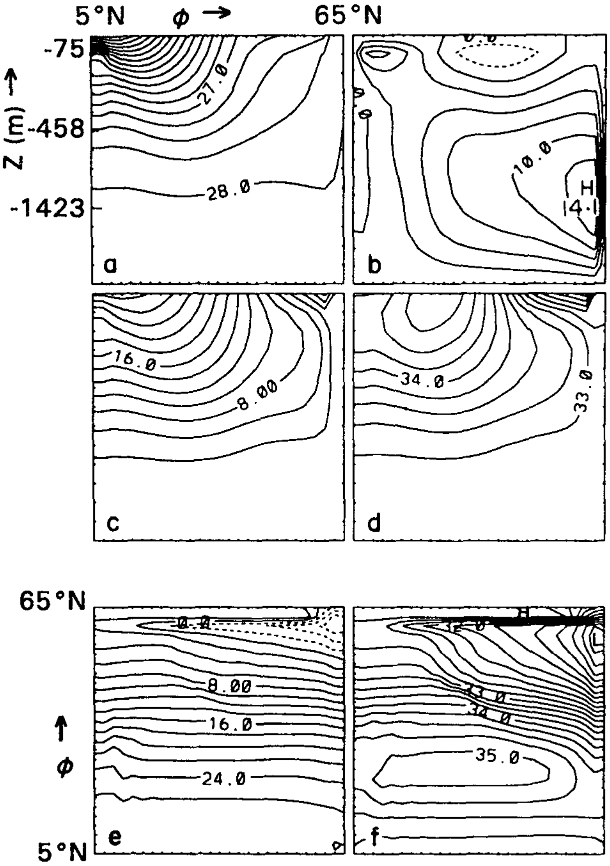

FIG. 9. As Fig. 2 but for experiment 6 at 10 years.

builds up and is eventually released in a "flush" a little over 100 years into the integration.

In reality, the atmospheric boundary condition seen by the ocean lies somewhere between the two extremes of Schopf's model on the one hand and mixed boundary conditions on the other. We have, therefore, run experiments identical to experiment 6 apart from the use of different time scales for the damping of SST anomalies. These range from the 30 days associated with mixed boundary conditions to the 300 days associated with experiment 6. We find that for a damping time of 250 days, the behavior is like that in experiment 6, with no PHC and, in particular, no collapse of the meridional overturning circulation. On the other hand, for 200 days or less a PHC ensues, and the overturning circulation collapses. The 200-day case (experiment 7 in Table 2) is particularly interesting. The initial behavior is not unlike that at 300 days, with the development of a freshwater cap that moves eastwards, with no reduction in the overturning circulation. However, after about 30 years, the freshwater cap spreads westward and eventually fills all the northern latitudes. At this time the meridional overturning circulation collapses. This state persists for about 100 years (as in experiment 2) before the subsequent "flush." This experiment shows that a PHC is possible in an ocean model coupled to a thermally interactive atmosphere [Manabe and Stouffer (1988) describe a quasi-stable state in their coupled general circulation model that resembles a PHC]. The critical damping time of 200 days probably depends on the strength of the freshwater flux and the strength of the wind forcing at high latitudes (Weaver et al. 1992). It would also be changed if we included ice in our model. One of the differences between the PHC that develops at 200 days and that under mixed boundary conditions is that the ocean temperature within the freshwater cap is much less (down to −9°C compared to near 0°C) because of the much greater heat loss that occurs from the ocean to the atmosphere. If we included ice in our model, the surface temperature in the cap would be more realistic, but the expulsion of salt in the freezing process would act to inhibit the development of the cap. Clearly, further research is required using both a more sophisticated model of the atmosphere and also incorporating ice into our ocean model.

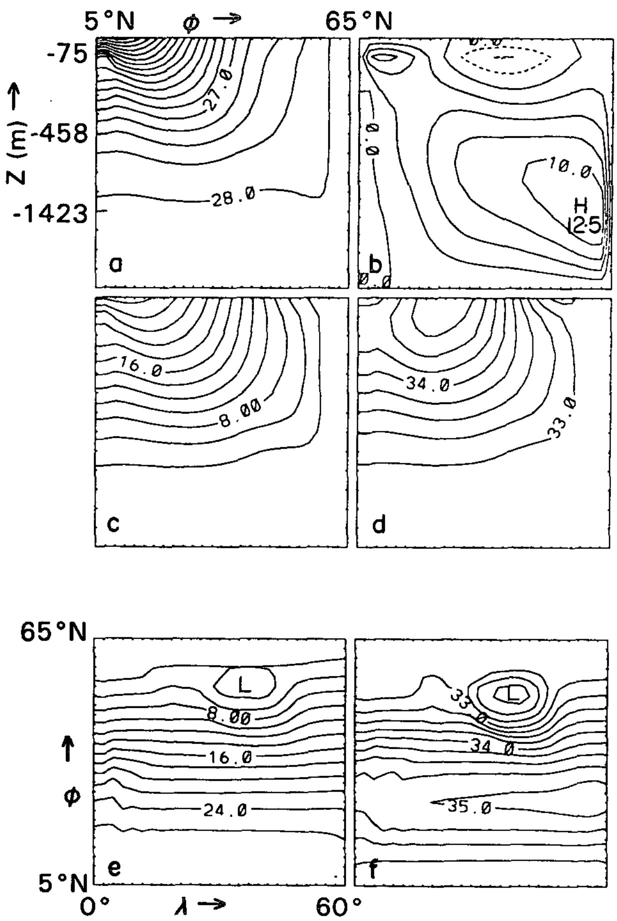

FIG. 10. As Fig. 2 but for experiment 6 at 50 years.

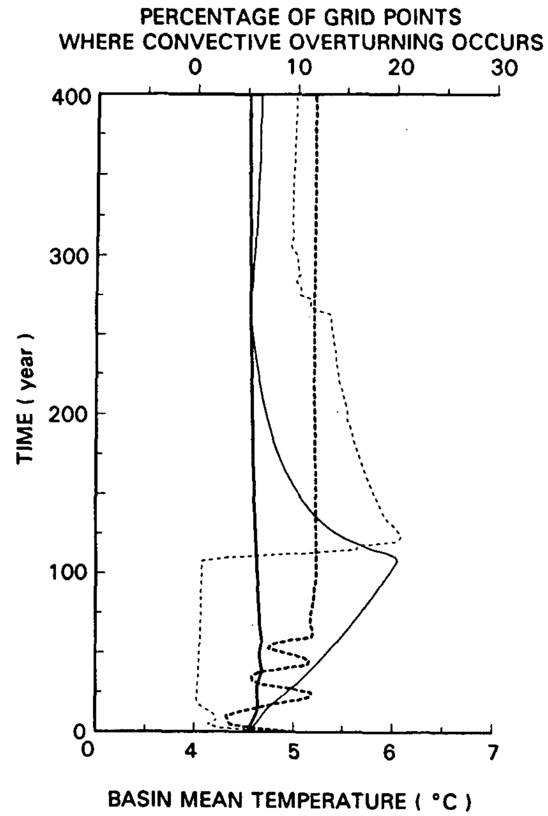

FIG. 11. Time series of basin mean temperature (solid line) and the percentage of grid points at which convective overturning occurs (dashed line). The thick lines are for experiment 6 and the thin lines for experiment 2.

5. Summary and discussion

In this paper, we have reexamined the polar halocline catastrophe of Bryan (1986a) from the point of view of the surface heat flux. We showed that in an experiment that develops a PHC under mixed boundary conditions, the reduction in the surface heat loss to the atmosphere during its development is an essential feature of the process. In particular, repeating the experiment but keeping the heat flux constant did not lead to development of a PHC. We found that by reducing the magnitude of the heat flux by 10%, a pool of very cold, fresh water appeared at the surface; but there was no collapse of the meridional overturning circulation. A freshwater pool also developed during the initial stages of an experiment in which a positive heat flux from the atmosphere to the ocean was added at all latitudes (this is of relevance to global warming, except that the positive heat flux would build up gradually rather than suddenly be applied, as in this experiment). The appearance of the freshwater pool is once again due to the reduction in heat loss to the atmosphere at high latitudes. We have also described experiments in which our ocean model has been coupled to the very simple atmospheric model of Schopf (1983). In these experiments, anomalies in sea surface temperature (SST) are damped through radiative relaxation from the atmosphere on a time scale of several hundred days. This is much longer than the damping time of tens of days associated with mixed boundary conditions. We believe this model to be relevant at middle and high latitudes because of the large horizontal scale of the SST anomalies that develop in our experiments, following an argument of Bretherton (1982). Since variations in the ocean–atmosphere heat flux are much less with this model, we expect the development of a PHC to be inhibited. Indeed, no collapse of the meridional overturning circulation was found, although during the first 60 years of integration, a pool of very cold, fresh water appears and is advected horizontally and vertically by the circulation in high latitudes, leading to decadal-type oscillations (Weaver and Sarachik 1991a). After 70 years, the model settles back to its initial state.

In reality, the ocean probably sees an atmospheric boundary condition somewhere between the two extremes of Schopf's model, on the one hand, and mixed boundary conditions on the other. We have, therefore, also described experiments in which SST anomalies are damped on a range of time scales between the 30 days associated with mixed boundary conditions and the 600 days associated with Schopf's model. We find that a PHC ensues for damping times of 200 days or less but not for times of 250 days or more. The critical time scale is likely to depend on the strength of the freshwater flux and the wind forcing at high latitudes (Weaver et al. 1992). It will also change if ice is incorporated into our ocean model.

These results suggest that thermal coupling with the atmosphere is likely to inhibit the development of a PHC and consequently reduce the likihood of flush events in which a large amount of heat is released to the atmosphere. Weaver et al. (1992) have already found that adding stochastic forcing to the freshwater flux at the surface increases the frequency of flushes under mixed boundary conditions and reduces their intensity. We are suggesting another mechanism for reducing the liklihood of a major flush event, namely the thermal response of the atmosphere to changing ocean conditions.

A rather unrealistic feature of our experiments are the low temperatures (down to −6°C) found in the freshwater pool in experiments 3, 4, 5, 6, and 7. This is because we have either removed the restoring boundary condition on ocean surface temperature (as in experiments 3, 4, and 5) or greatly weakened its effect (as in experiments 6 and 7). However, if our model also included the effect of sea ice, then the freezing that would occur as the ocean surface temperature drops would act to maintain the meridional overturning circulation and inhibit the occurrence of a PHC because of the salt released in the freezing process. This has been demonstrated using a very simple ice model described by Zhang et al. (in preparation). It is shown there that even under mixed boundary conditions, inclusion of the ice model prevents a collapse of the meridional overturning circulation. Experiments have also been conducted including a seasonal cycle, but this does not alter the conclusions we have drawn.

Note added in proof. We have since experimented using a realistic ice model. Preliminary results indicate a more complex behavior than suggested by the simple model. This will be discussed in detail in a future paper.

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REFERENCES

Bretherton, F. P., 1982: Ocean climate modeling. *Progress in Oceanography,* Vol. 11, Pergamon, 93–129.
Broecker, W. S., and G. H. Denton, 1990: What drives glacial cycles? *Sci. Amer.,* **262,** 49–56.
Bryan, F., 1986a: Maintenance and variability of the thermohaline circulation. Ph.D. thesis, Atmospheric and Oceanic Sciences Program, Princeton University, 254 pp.
——, 1986b: High-latitude salinity effects and interhemispheric thermohaline circulations. *Nature,* **323,** 301–304.
Bryan, K., 1969: A numerical method for the study of the circulation of the world ocean. *J. Comput. Phys.,* **4,** 347–376.
——, 1984: Accelerating the convergence to Equilibrium of Ocean climate model. *J. Phys. Oceanogr.,* **14,** 666–673.
——, and M. D. Cox, 1972: An approximate equation of state for numerical models of the ocean circulation. *J. Phys. Oceanogr.,* **2,** 510–514.
——, S. Manabe, and M. J. Spelman, 1988: Interhemispheric asymmetry in the transient response of a coupled ocean–atmosphere model to CO_2 forcing. *J. Phys. Oceanogr.,* **18,** 851–867.
Colin de Verdière, A., 1988: Buoyancy driven planatary flows. *J. Mar. Res.,* **46,** 215–265.
——, 1989: On the interaction of wind and buoyancy driven gyres. *J. Mar. Res.,* **47,** 595–633.
Cox, M. D., 1984: A primitive equation, 3-dimensional model of the ocean. GFDL Ocean Group Tech. Rep. No. 1, GFDL/Princeton University.
Dickinson, R. E., 1981: Convergence rate and stability of ocean–atmosphere coupling schemes with a zero-dimensional climate model. *J. Atmos. Sci.,* **38,** 2112–2120.
Gill, A. E., 1982: *Atmosphere-Ocean Dynamics,* Academic Press, 662 pp.
Haney, R. L., 1971: Surface thermal boundary condition for ocean circulation models. *J. Phys. Oceanogr.,* **1,** 241–248.
Killworth, P. D., 1985: A two-level wind and buoyancy-driven thermocline model. *J. Phys. Oceanogr.,* **15,** 1414–1432.
Manabe, S., and R. J. Stouffer, 1988: Two stable equilibria of a coupled ocean-atmosphere model. *J. Climate,* **1,** 841–866.
Marotzke, J., 1989: Instabilities and steady states of the thermohaline circulation. *Oceanic Circulation Models: Combining Data and Dynamics,* D. L. T. Anderson and J. Willebrand, Eds., Kluwer, 501–511.
——, 1991: Influence of convective adjustment of the stability of the thermohaline circulation. *J. Phys. Oceanogr.,* **21,** 903–907.
——, and J. Willebrand, 1991: Multiple equilibria of the global thermohaline circulation. *J. Phys. Oceanogr.,* **21,** 1372–1385.
——, P. Welander, and J. Willebrand, 1988: Instability and multiple steady states in a meridional-plane model of the thermohaline circulation. *Tellus,* **40A,** 162–172.
Peixoto, J. P., and A. H. Oort, 1984: Physics of climate. *Rev. Mod. Phys.,* **56,** 365–429.
Salmon, R., 1986: A simplified linear ocean circulation theory. *J. Mar. Res.,* **44,** 695–711.
——, 1990: The thermocline as an "internal boundary layer." *J. Mar. Res.,* **48,** 437–469.
Schopf, P. S., 1983: On equatorial waves and El Niño. II: Effects of air-sea thermal coupling. *J. Phys. Oceanogr.,* **13,** 1878–1893.
Stocker, T. F., and D. G. Wright, 1991a: A zonal averaged ocean model for the thermohaline circulation. *J. Phys. Oceanogr.,* **21,** 1725–1739.
——, and ——, 1991b: Rapid transitions of the ocean's deep circulation induced by changes in the surface water fluxes. *Nature,* **351,** 729–732.
Stommel, H., 1961: Thermohaline convection with two stable regimes of flow. *Tellus,* **13,** 224–230.
Stouffer, R. J., S. Manabe, and K. Bryan, 1989: Interhemispheric asymmetry in climate response to a gradual increase in atmospheric CO_2. *Nature,* **342,** 660–662.
Weaver, A. J., and E. S. Sarachik, 1991a: Evidence for Decadal Variability in an ocean general circulation model: An advective mechanism. *Atmos.-Ocean,* **29,** 197–231.
——, and ——, 1991b: The role of mixed boundary conditions in numerical models of the ocean's climate. *J. Phys. Oceanogr.,* **21,** 1470–1493.
——, ——, and J. Marotzke, 1991c: Fresh water flux forcing of decadal and interdecadal oceanic variability. *Nature,* **353,** 836–838.
——, J. Marotzke, P. F. Cummins, and E. S. Sarachik, 1993: Stability and variability of the thermohaline circulation. *J. Phys. Oceanogr.,* **23,** 39–60.
Wright, D. G., and T. F. Stocker, 1991: A zonally averaged ocean model for the thermohaline circulation. Part 1: Model development and flow dynamics, *J. Phys. Oceanogr.,* **21,** 1713–1724.
Zhang, S., C. A. Lin, and R. J. Greatbatch, 1992: A thermocline model for ocean climate studies. *J. Mar. Res.,* **50,** 99–124.