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Surface heat flux parameterization and the response of ocean general circulation models to high-latitude freshening

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ABSTRACT

A global ocean general circulation model (OGCM) is forced using *mixed* boundary conditions (i.e., a restoring condition on the upper level temperature but using a fixed, specified surface salt flux). A freshwater flux anomaly is then applied over the western half of the sub-polar gyre in the northern North Atlantic. The response of the model is found to be dependent upon the details of the parameterization of the surface heat flux: In particular the "*coupling strength*" or Haney relaxation time is crucial. Responses range from a *halocline catastrophe* at short relaxation times (strong coupling) to a very modest perturbation at longer relaxation times (weaker coupling). An accurate parameterization is therefore required to properly model the evolution of the response. It is uncertain that the restorative condition is sufficiently realistic, especially in cases where a significantly different climatology is obtained. It is possible, for example, that the evolution could move from an unstable trajectory to a stable one if the parameters in the heat flux formulation are also allowed to evolve. This might help to explain why OGCMs under mixed boundary conditions are more sensitive than the observations suggest they should be. When a recovery does occur it does so on decadal time scales. It is therefore tempting to speculate that the positive feedback on the initial perturbation provided by the heat flux response plays a central role in the dynamics of North Atlantic variability, in a manner that is analogous to the wind-stress feedback in the El Niño, Southern Oscillation.

1. Introduction

A significant contribution to North Atlantic Deep Water (NADW) is made by Gulf Stream water which has been diverted northwards via the North Atlantic Drift into the sub-polar seas beyond the Faeroe-Iceland Ridge. This water is much warmer (and more saline) than other waters at these latitudes and so it provides an important moderating influence on the regional climate and further east over western Europe (e.g., Broecker, 1991).

After further cooling, sometimes to the point where sea-ice is formed and brine is rejected, this water convects and mixes with Arctic and polar water, filling the sub-polar basins until the dense water overflows the ridge primarily through Denmark Strait between Greenland and Iceland (Swift, 1984). It then advects south along the western flank of the Mid-Atlantic Ridge at depth forming a western boundary undercurrent (Broecker et al., 1991), beginning what is believed to be a globally connected circulation (Gordon, 1986; Semtner and Chervin, 1988, 1992). The sub-polar seas are therefore considered to play an important role in determining the nature of the global thermohaline circulation (Aagaard et al., 1985) and climate in general.

The equation of state for sea water is such that at the salinities and temperatures typically encountered in these seas the addition of relatively small amounts of fresh water to the surface can stabilize

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the water column to the extent that convection can be prevented from occurring (Aagaard and Carmack, 1989). This sensitivity has helped to motivate several studies which attempted to determine the stability of NADW formation (NADWF) to freshwater anomalies imposed at high northern latitudes (Bryan, 1986; Maier-Reimer and Mikolajewicz, 1989; Marotzke, 1989; Wright and Stocker, 1991; Stocker and Wright, 1991). In these studies *mixed* boundary conditions (BCs), in which the salt flux is fixed and the heat flux is crudely parameterized (after Haney, 1971) in terms of deviations away from current climatic values of an apparent air temperature, T^*, (or it's proxy, sea-surface temperature) were employed. Under these conditions the meridional overturning associated with NADWF was found to be sensitive to such perturbations, with seemingly small anomalies leading to fundamentally different states in which NADWF has collapsed. In this state convection at high latitudes is subdued and less heat is transported northwards. As a result, the moderating influence of the ocean upon the regional climate is reduced.

These results fuelled an existing concern that NADWF could be halted if it became capped by low salinity water, thereby leading to a *halocline catastrophe*, a dramatic alteration to the global ocean's thermohaline circulation (ICSUSCO, 1991).

Other studies have highlighted the importance of the surface heat flux parameterization at high latitudes in determining the ocean model climate through its impact upon DWF (Maier-Reimer et al., 1991). More recently, Power et al. (1993) have shown that a collapse of the overturning associated with NADWF could also occur in a global ocean general circulation model (OGCM) with as realistic a representation of bathymetry and continental geometry as their coarse grid allowed, if the freshwater anomaly exceeded a certain critical value. While this value is probably a function of the horizontal diffusion and the precise location over which the anomaly is imposed (amongst other things), of more importance in the current context is the fact that the collapse was dependent upon the restorative condition on temperature. If the condition was replaced by a fixed heat flux condition then no collapse ensued. This result is consistent with the finding reported by Welander (1986) that only one solution can exist in a simple three-box model under fixed fluxes of both salt and heat.

Restoring the upper level temperature to currently observed estimates of T^* is strictly only a valid approximation for the surface heat flux when modelling the current climate (Power and Kleeman, 1993). This is because the values of parameters in the restoration formula are based upon observations made in the current climate and there is no guarantee that these values will remain appropriate in an evolving or different climate.

Another concern with the use of the restorative condition lies in the fact that the values usually chosen for the parameters in the heat flux formulation are based upon observations made at low and middle latitudes (Haney, 1971; Han, 1984; Oberhuber, 1988), yet the same values are applied at high latitudes by ocean modellers.

These two concerns beg the question: is the collapse significantly dependent upon the details of the heat flux parameterization at high latitudes? If we can show here that the details are important then perhaps we need to defer our assessment of the stability of the current climate until a more sophisticated treatment of the surface heat flux is employed. To answer the question we will determine how the response to a high latitude freshwater anomaly varies as a function of the relaxation time constant, τ, used in the Haney (1971) formulation for the surface heat flux. This constant can be considered as specifying how strongly the ocean is coupled to the crude atmosphere implied by the formulation. A shorter time constant, for example, implies a more rapid relaxation back to T^* and hence stronger coupling strength.

The OGCM used in this study is described in Section 2 and the results are presented in Section 3. The key results are summarized and discussed in Section 4.

2. Description of the Ocean Model

The OGCM used here is the latest version of the Geophysical Fluid Dynamics Laboratory (Princeton) code (Pacanowski et al., 1991) based on the work of Bryan (1969) and Cox (1984) and given the acronym MOM (Modular Ocean Model).

The horizontal grid is compatible with an atmospheric GCM R21 Gaussian grid, which has a longitudinal spacing of 5.625°, and a latitudinal spacing of approximately 3.2°. There are 12 vertical levels ranging in thickness from 25 m at the surface to 900 m in the deep ocean. The horizontal eddy viscosity is artificially large (9×10^5 m^2 s^{-1}) to ensure that the western boundary currents have horizontal scales that are resolved by the coarse grid. The horizontal eddy diffusivity is 2.5×10^3 m^2 s^{-1}. The vertical eddy viscosity and diffusivity take the same numerical values of 20, 1.5 and 1×10^{-4} m^2 s^{-1} in the first, second and subsequent levels, respectively, in order to crudely simulate a surface mixed layer. A fully non-linear equation of state is used (UNESCO, 1981; Gill, 1982), the topography has been smoothed to avoid topographic instabilities and convection is modelled as enhanced vertical diffusion. Further details regarding the configuration of this model have been given by Moore and Reason (1992).

3. Results

3.1. *The preliminary experiments*

The preliminary experiments conducted are represented in Fig. 1 and differ only in the specification of the surface fluxes of heat and salt. First the OGCM is integrated while restoring the upper level temperature, T, and salinity, S, to annually averaged climatological values based on the data compiled by Levitus (1982), using a relaxation time constant of 20 days. This leads to the solution R (the restoring solution).

The equations were integrated for 200000 tracer time steps. The acceleration techniques of Bryan (1984) and Bryan and Lewis (1979) were employed during the preliminary experiments: the time step was increased from 2 days at the surface to 16 days at the bottom. The barotropic and baroclinic equations were integrated with a time step of 1200 s. The globally averaged surface heat flux at the end of this stage is approximately 0.06 W m^{-2}, with a corresponding trend in the globally averaged temperature of about 0.01°C *century*$^{-1}$.

The salt flux, S_F, is then diagnosed over a period of about 50 surface years. The wind-stress is kept fixed to annually averaged observational estimates of Hellerman and Rosentein (1983) as they were during R.

Mixed BCs are then applied in which the salt flux is fixed to S_F but T is restored with a 20-day relaxation time again for 200 000 2-day time steps. This leads to the second solution M (mixed BC solution), in which the overturning has a maximum magnitude of about 22 Sv. Further details on these preliminary experiments are given by Power et al. (1993).

3.2. *The perturbation*

The acceleration techniques are then discontinued, in contrast to Power et al. (1993), as they were designed for accelerating the convergence towards the equilibrium solution (Bryan, 1984) but are not to be recommended if the transient behaviour is important.

A salt flux anomaly which reduces the surface salinity is then applied for 1500 days to the northern North Atlantic as shown in Fig. 2. The salt

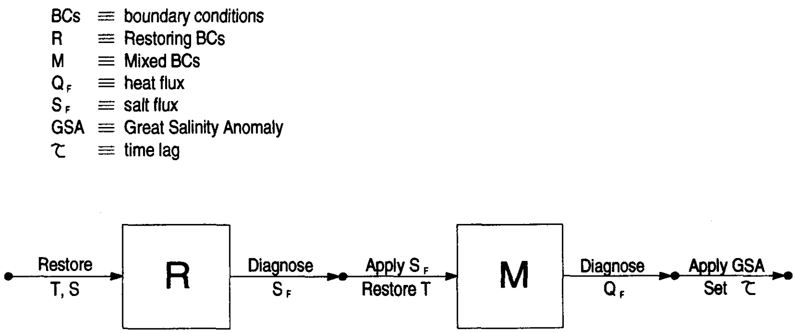

Fig. 1. A schematic representation of the experiments conducted.

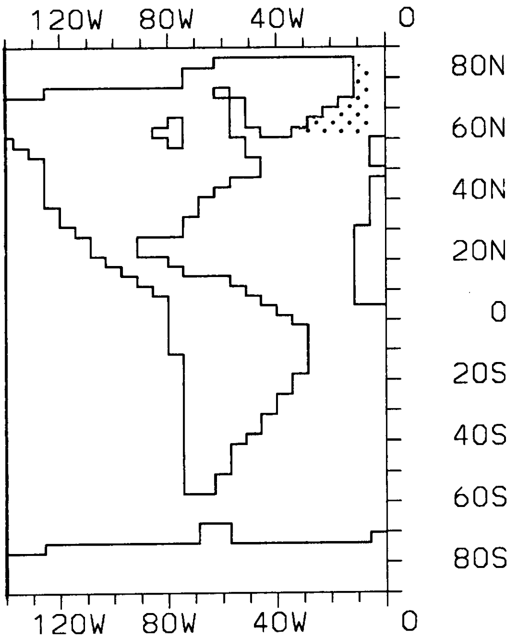

Fig. 2. The region off the south-east coast of Greenland over which the freshwater anomaly is dumped.

deficit of the anomaly is equal to about eight times that of the value estimated for the 1968–1982 "*Great Salinity Anomaly*" (GSA), i.e., 7.2×10^{13} kg (Dickson et al., 1988). We choose a duration of several years primarily because it is of the same order as the observed residence time of 10 years for the GSA in the Iceland Sea (Malmberg, 1973). The shorter value is chosen to reduce the cost of the experiments.

Power et al., (1993) found this to be a critical value: anomalies with salt deficits equal to or greater than this amount led to a collapse, whereas the overturning recovered if deficits less than this figure were applied. This result was obtained using a relaxation time constant of 20 days. If a constant heat flux was imposed instead, then no collapse ensued even for anomalies with salt deficits approximately 28 times that given for the GSA above.

The choice of such a large salt deficit (i.e., 8 GSAs) and relatively short duration is made because it provides the basis for a convenient framework in which to study existing methodologies employed to assess the stability of the current climate.

In order to understand how we will generalize these results consider the equation representing the parameterization of the heat flux, Q, over the North Atlantic between 10° and 80° N:

$$Q = Q_F - \alpha \, \Delta T / \tau, \tag{1}$$

where Q_F is the surface heat flux diagnosed from M (the mixed BC solution above), $\alpha = \rho C_p / h$, ρ the density, C_p the specific heat capacity, h the upper level depth, ΔT is the deviation of the upper level temperature away from the mean temperature diagnosed from M, and τ the relaxation time. Thus α/τ represents an inverse coupling strength. Note that eq. (1) reduces to the restoring condition if the time constant is not changed after the preliminary experiments have been completed.

The value of τ is subject to considerable uncertainty. In addition to the problems associated with the restoring BC already discussed in Section 1, use of estimates from Haney (1971) or Oberhuber (1988) assume that the air temperature does not depend on T, i.e., that the atmosphere has infinite heat capacity. Estimates we derived by regression from a ten year run of the BMRC atmosphere GCM (documented in Hart et al., 1990) indicate values for τ varying spatially from close to zero up to values of perhaps 50% of the estimates of

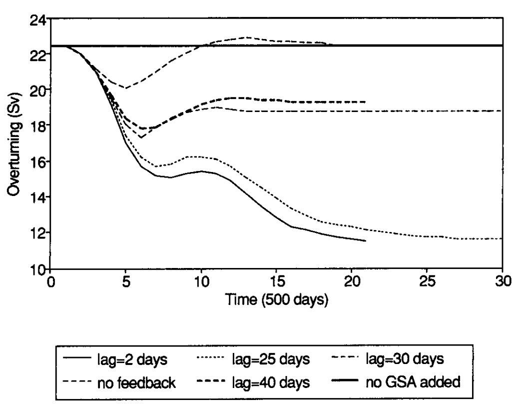

Fig. 3. The maximum zonally averaged meridional overturning associated with North Atlantic Deep Water Formation as a function of time during and after the addition of the freshwater anomaly. The trajectories for time lags, $\tau = 2$, 25, 30, 40 and as $\tau \rightarrow \infty$ (i.e., with a fixed heat flux diagnosed from the solution obtained under mixed boundary conditions, M) are presented. The control experiment in which no anomaly was added is also represented.

Oberhuber (1988). An adequate model of at least the air temperature is therefore required to accurately estimate τ. Such a model is currently under development and will be coupled to the OGCM employed here. In this study we merely wish to exploit the uncertainty in τ by conducting senstitivity tests as a prelude to work with the coupled model.

We exploit these uncertainties by choosing values of $\tau = 2$, 20, 25, 30 and 40 days during and after the addition of the salinity anomaly. The fixed flux condition (which corresponds to the case $\tau \rightarrow \infty$) is also considered. Away from the North Atlantic the upper level temperature is restored as usual (i.e., with a time constant of 20 days), as we are primarily interested in the coupling in the northern North Atlantic.

The zonally averaged overturning was diagnosed and the results for a representative subset of these cases is presented in Fig. 3. For $\tau \leqslant 25$ days a substantial collapse ensues, whereas a more modest collapse occurs for larger constants. The magnitude of the collapse tends to zero as $\tau \rightarrow \infty$. The cases $\tau = 25$ days and $\tau = 30$ days were then further integrated using the acceleration techniques described earlier to obtain the new climatologies. This leg of the integration continued for 150,000 tracer time steps, each 2 days long, with an acceleration factor of 8 between the surface and lowest levels.

The overturning remained subdued in the first case ($\tau = 25$ days) but recovered to about 18 Sv in the second. The difference between these two equilibria (i.e., $\tau = 25$ days, $\tau = 30$ days) is presented in Fig. 4. Associated with the reduced overturning are reduced surface temperatures (Fig. 4a) and salinities (Fig. 4b) in the northern North Atlantic. There are also significant differences in S and T (even if differences are based on 50-year averages) in parts of the Southern Ocean, highlighting the fact that NADWF is part of a globally connected thermohaline circulation (Gordon, 1986; Semtner and Chervin, 1988, 1992; Power and Kleeman, 1993).

In Fig. 4c, we see that the net horizontal transport in the North Atlantic's subpolar and sub-

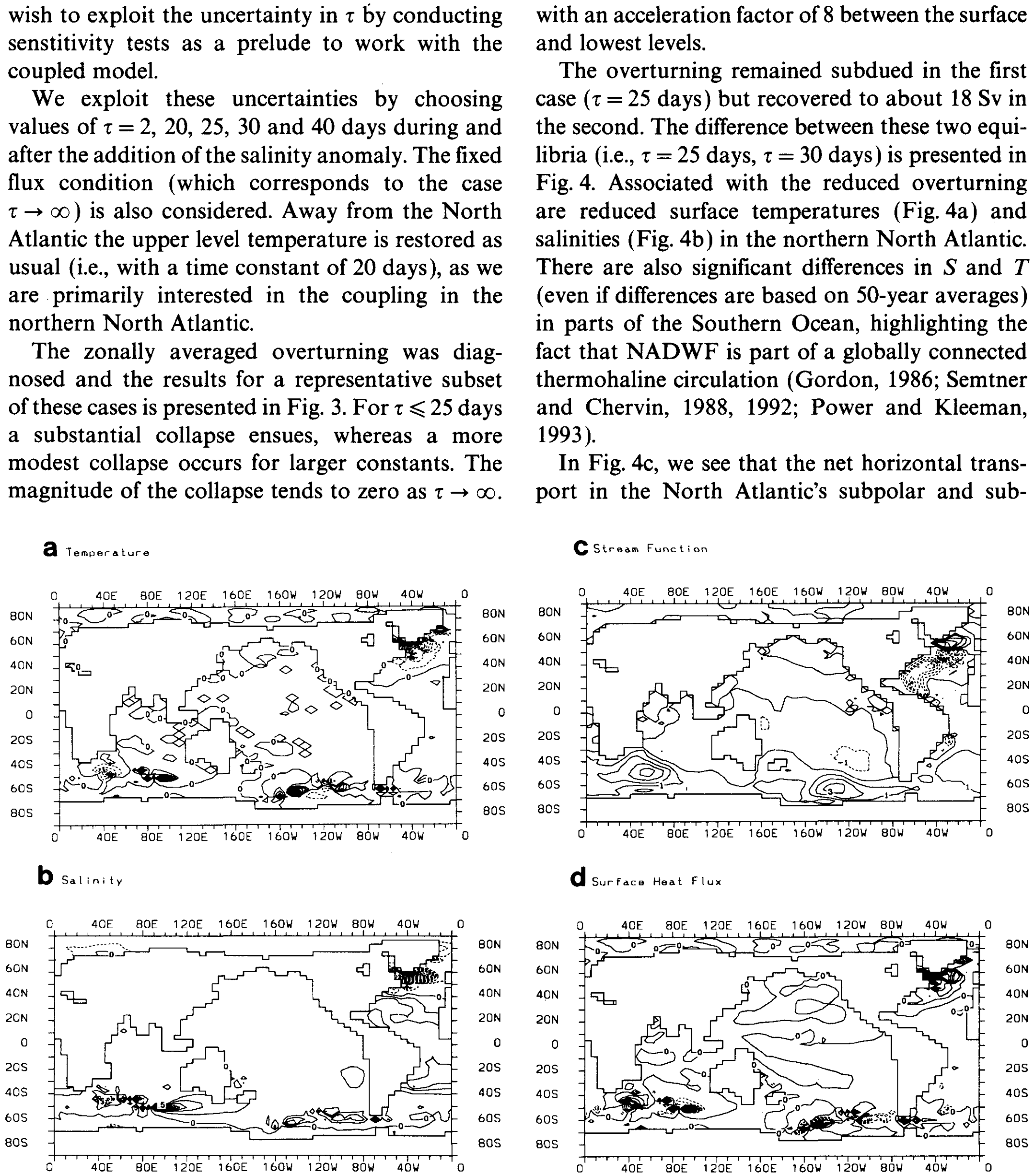

Fig. 4. The difference between the equilibrium states obtained after applying the freshwater anomaly first with $\tau = 25$ days and then with $\tau = 30$ days. Upper level surface (a) temperature (°C) and (b) salinity (ppt), (c) barotropic transport (Sv) and (d) surface heat flux (W m^{-2}) are presented. Contour intervals of (a) 0.25°C, (b) 0.1 ppt, (c) 1 Sv and (d) 10 W m^{-2} are used, respectively.

tropical gyres have been reduced and the Antarctic Circumpolar Current has anomalous meanderings. These changes can be attributed to the change in the thermohaline circulation since this circulation is coupled to the barotropic flow in an OGCM, especially one with bathymetry. To test this, the OGCM was driven with thermohaline forcing only. The subsequent equilibrium solution did indeed have a significant barotropic flow. In the Southern Ocean, for example, there is an Antarctic Circumpolar Current exceeding 10 Sv. Thus the barotropic flow can be altered if the buoyancy forcing is altered. The dependence of this interaction upon various model parameters is currently being investigated.

In previous studies solutions like those contrasted in Fig. 4 have been referred to as multiple equilibria under the *same boundary conditions* (for example, Marotzke and Willebrand, 1991). This statement needs clarification: The parametrization of the heat flux has the same functional form but the fluxes have *different* numerical values (Fig. 4d) and so the boundary conditions are not the same. This point is of more than just academic interest as the distinction has a very important implication: The solutions should be regarded as multiple equilibria of a *coupled* model and not of the ocean alone as they have been in the past. Previously, the uncertainty we might have associated with the existence of the multiple equilibria in our geological past (or perhaps our not-too-distant future) was bound up in how well we considered the OGCM to model the current ocean climate. The fact that we are instead considering multiple equilibria of a coupled model substantially increases this uncertainty because we are essentially dealing with a model in which all non-oceanic components (e.g., the atmosphere, cryosphere, etc.) are modelled by one or two simple formulae. So while we might be doing a fairly reasonable job of modelling the ocean component the same cannot be said for the other components of the climate system. See Power and Kleeman (1993) for further details.

4. Discussion

A coarse grid version of the GFDL global OGCM (Pacanowski et al., 1991) was integrated under mixed BCs (i.e., a restoring condition on the upper level temperature but a fixed, specified salt flux).

A freshwater anomaly with a salt deficit equal to about 8 times that estimated for the 1968–1982 "*Great Salinity Anomaly*" (or GSA) was imposed over the North Atlantic's sub-polar gyre, to the south-east of Greenland. This led to reduced sea-surface temperatures as a result of weakened convection and a weakened barotropic circulation in both the sub-polar and sub-tropical gyres. The evolution of the response was found to be critically dependent upon the relaxation time constant, τ, used in the equation parameterizing the surface heat flux (i.e., eq. (1)). For relaxation time constants less than about 25 days, a collapse occurs to a new state in which the overturning associated with North Atlantic Deep Water Formation is halved and the sea-surface temperature drops by over 2°C in the northern North Atlantic. If the time constant exceeds about 30 days, on the other hand, no significant collapse occurs. In fact no collapse occurred if a constant heat flux was imposed even for much larger anomalies than deployed here. In other words, as the coupling strength between the ocean and the crude atmosphere is increased the behaviour of the coupled system undergoes a fundamental change (i.e., exhibits a 'polar halocline catastrophe').

In the present context, we have seen that the ocean model climate can only be destabilized if the coupling strength implicit in the restorative condition on the surface temperature is sufficiently large. It is this condition, therefore, that provides the source of positive feedback on the initial perturbation provided by the high latitude freshening for the OGCM under mixed BCs: the freshwater anomaly stabilizes the water column and so convection is reduced. In this region the modelled temperature is actually unstable with respect to temperature (i.e., the temperature increases with depth) and so the reduction in convective activity means that less heat is brought to the surface from below and so the surface temperature drops. As it drops, less heat is lost by the upper level via the restorative condition, which further enhances the stability of the water column but drives the ocean further away from its initial state. It is this conspiracy between the initial perturbation and the restoring BC that we refer to as the *positive feedback*.

This positive feedback mechanism should be contrasted with the two discussed by Bryan (1986): The first mechanism was used by Warren (1983) to account for the absence of DWF in the North Pacific. It occurs if there is net precipitation over evaporation under fixed flux conditions, if the cessation of DWF is accompanied by a cessation of the advection of salty water into the surface of what was the convective region then the stability of the water column is further enhanced and DWF is even less likely to restart. The second mechanism (due to Rooth, 1982) occurs on a much longer time scale and is associated with salinity differences on a global scale and the strength of the large-scale circulation (Bryan, 1986). It is likely that both of these mechanisms *reinforce* the feedback due to the restoring BC.

It is worth pointing out that a positive feedback between the ocean and atmosphere is believed to be crucial in providing the *driving mechanism* for El Niño, Southern Oscillation tropical variability (Philander, 1984; Hirst, 1986; Battisti and Hirst, 1989). It is tempting to speculate that the coupled positive feedback shown to be important here, may be of importance in driving North Atlantic SST variability. Judging by the recovery time evident during the evolution depicted in Fig. 3 the pertinent variability is likely to be on decadal or longer time scales, time scales over which significant variability is known to occur (Bjerknes, 1964; Weaver et al., 1991, for example).

It is the ocean that provides most of the thermal inertia for our climate, thereby providing the stabilizing influence on the proxy values used for the apparent air temperatures, T^*, and the time constants used in the restoration of the upper level temperature. The ocean is, however, significantly altered in the collapsed state obtained here but this fact is not reflected in the heat flux parameterization. In more realistic climate models T^* and τ might also evolve (e.g., if storm tracks shifted from one region to another we might expect that the rate at which SST anomalies are eroded would differ because the mechanisms which give rise to the erosion would then differ, (Maier-Reimer et al., 1991)), and so the evolution might begin on one trajectory (depicted in Fig. 3) but subsequently move to another as the parameters change. As a result the response could be very different to that modelled using unique choices for these parameters.

It is possible, for example, that the substantial and sustained drop in the SST associated with the reduced convection and diminished northward heat transport might eventually lead to a drop in T^*. If this occurred then the heat flux anomaly will be reduced and so the evolution would then proceed on a more stable trajectory. This scenario would help to explain why OGCMs under mixed BCs are more sensitive than observations suggest they should be (Stocker and Broecker, 1992).

The only way of testing these hypotheses is, of course, to abandon the restoration altogether and develop realistic coupled atmosphere-ocean-ice models in which no ad hoc assumptions regarding the surface heat (salt and momentum) fluxes are required. Indeed the current generation of coupled models, from simplified, zonally averaged configurations (Stocker et al., 1992) to sophisticated atmosphere-ocean GCMs with sea-ice components (Manabe and Stouffer, 1988) do in fact exhibit a sensitivity to high latitude freshening. It is important to keep in mind, however, that flux correction techniques are required to stabilize their control climates, and so we should not be surprised to find that our ideas regarding the stability of NADWF are altered by future coupled models that provide more accurate estimates of fluxes of both heat and salt across the air/sea/ice boundary.

Two recent studies have helped to further clarify some of the issues raised here. In the first study by Zhang et al. (1993) a number of experiments similar to our own were independently conducted using a 2° by 2° geostrophic model (Zhang et al., 1992). The model consists of a flat 60° by 60° basin (bounded by longitude and latitude walls) between 5°–65°N, forced with zonally averaged, zonal wind-stress. They too exploited the uncertainty in the damping constant (or coupling strength). A freshwater anomaly was applied to all points north of 50°N and the response ranged from a 'polar halocline catastrophe' for damping constants ranging from 30 to 200 days, to no collapse for constants larger than 200 days. They argued (after Bretherton, 1982) that the longer time scales are appropriate for global scale SST anomalies because the 'damping' then occurs via long wave radiation to space. For smaller scale, anomalies atmospheric advection dominates the damping by shunting heat anomalies from one part of the ocean to another (Bretherton, 1982), a much more rapid process. Zhang et al. (1993) concluded that since the correct heat flux response probably lies between these extreme cases, the likelihood of a polar halocline catastrophe is reduced. They also highlighted the importance of the shallowing of the mixed layer as yet another source of positive feedback in the response and that the response is probably a strong function of the background salt flux field and the wind-stress applied (after Weaver et al., 1991).

In order to understand the large difference in the critical damping times given here (30 days) and by Zhang et al. (200 days) we need to understand some of the differences between the two models and the nature of the freshwater anomalies applied. First, however, a more meaningful comparison can be made between the damping rate/unit depth of the upper level, a 200 day lag in their model (which has an upper level depth of 46 m) corresponds to a lag of only 110 ($\approx 200 \times 25/46$) days in our model since it has an upper level depth of only 25 m. The remaining difference (i.e., 110 versus 30 days) is probably due to the various model differences (e.g., the version of the GFDL OGCM used here is global in extent and has a fairly realistic representation of bathymetry and continental geometry) and the fact that the magnitude, temporal and spatial character of the anomalies applied are different. The anomaly we applied was introduced to the upper level of a limited region to the south-east of Greenland over a period of about 5 years, whereas theirs was applied impulsively to the top 3 levels at all points north of 50° N.

In the second study, Power et al. (1993) showed that the response is indeed sensitive to both the magnitude and duration of the anomaly, the greater the anomaly or the more impulsively it is applied, the more likely a collapse. We would therefore expect that a shorter critical damping constant would have been obtained by Zhang et al. (1993) had they applied the anomaly more gradually. On the other hand, an even larger constant would have been obtained if a larger anomaly had been applied: less positive feedback is required for a larger perturbation (they applied the equivalent of about 3.5 GSAs whereas we applied 8). The first of these differences (i.e., the duration) would appear to dominate in this instance. While neither study has addressed the significance of the spatial character of the anomaly, it seems reasonable to suppose that this too will have an impact.

So far, the discussion has centred on the rôle of the heat flux feedback in the response, feedbacks on the salt and momentum fluxes have been ignored. Power et al. (1993) constructed a number of very crude coupled models (all incorporating the OGCM described here) in which salt and momentum flux anomalies were parameterized in terms of changes in the meridional SST gradient in an attempt to take account of increased storm activity that might accompany a polar halocline catastrophe. They found, however, that these additional feedbacks had no more than a minor modifying impact upon the response, the collapse still occurred. On the other hand, Zhang et al. (1993) reported that feedbacks due to anomalous ice formation can help to stabilize the system.

In summary, the important point to note from the results presented here, and in the earlier studies, is that despite the many differences in the models employed and the freshwater anomalies applied, the main conclusions are the same: The response of ocean models to high latitude freshenings depends critically upon the details of the restoring BC, and the use of a single damping constant (or apparent air temperature for that matter) is inappropriate in the context of climate change. This therefore highlights the need for a more sophisticated treatment of the surface heat flux BC in research aimed at assessing the stability of the current climate.

5. Acknowledgements

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