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Dependence of multiple climate states on ocean mixing parameters

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Abstract. Multiple equilibria of the climate system, inferred from paleo reconstructions, have also been observed in both ocean-only and coupled ocean-atmosphere general circulation models. These multiple states are thought to be associated with different modes of operation of the meridional overturning circulation in the North Atlantic. It has recently been suggested that the stability of these states depends on the amount of vertical mixing in ocean models. Here we investigate the dependence of the hysteresis behaviour of the thermohaline circulation to sub-gridscale mixing processes in the ocean. Using a simplified coupled ocean-atmosphere model we find that both vertical and horizontal diffusivities have considerable influence on the stability of the different circulation modes. They also change the transition points from one circulation pattern to another. Larger vertical diffusivities lead to higher values of additional precipitation into the North Atlantic being necessary to stop the formation of deep water. However, for slightly increased evaporation, the state without deep water formation becomes increasingly unstable for stronger vertical diffusion. Larger values of horizontal mixing lead to a narrowing of the phase space for which two equilibria are stable. These results suggest that it is currently not possible, given the large uncertainty in ocean mixing, to quantitatively determine possible thresholds for the transition of the North Atlantic thermohaline circulation between *on* and *off* modes. This presents a policy predicament as it makes it extremely difficult to assign a probability for the future occurence of such nonlinear climate transition.

Introduction

The different nature of the observed heat and freshwater coupling between the atmosphere and the ocean lead *Stommel* [1961], through the use of a simple box model, to suggest that the oceanic thermohaline circulation (THC) could possess multiple equilibria. This fundamental property of ocean circulation was revisited by *Bryan* [1986] who demonstrated that an ocean general circulation model (GCM) also allowed for multiple equilibria of the THC, and that transitions between these states could be initiated through small perturbations to high latitude surface salinities. The existence of multiple equilibria of the THC has since been found in a hierarchy of ocean-only models [e.g. *Maier-Reimer and Mikolajewicz*, 1989; *Stocker and Wright*, 1991; *Marotzke and Willebrand*, 1991] and coupled ocean-atmosphere models [e.g. *Manabe and Stouffer*, 1988; *Stocker et al.*, 1992; *Fanning and Weaver*, 1997].

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At low temperatures the density of sea water is most sensitive to salinity changes. As such, small perturbations to the high latitude surface freshwater budget can lead to a stabilisation of the water column, suppression of deep convection and a transition from an active deep water formation mode to a passive mode of the North Atlantic THC. If, on the other hand, the system is already in the passive mode, enhanced evaporation at high latitudes can increase salinities to the point where the water column becomes unstable, switching the system to a state with active deep water formation.

Recently *Manabe and Stouffer* [1999] showed that in their coupled ocean-atmosphere GCM, the stability of the passive state without North Atlantic Deep Water (NADW) formation depends on the value of the vertical diffusivity. In particular, they noted that there appears to be a critical value of the vertical diffusivity, above which a stable passive mode of the THC did not exist.

The values of vertical and horizontal diffusivities used in ocean models are subject to considerable uncertainty. Measurements of tracer distributions in the open ocean [*Ledwell et al.*, 1998] yield much smaller values than those currently used in large scale ocean models [*Manabe and Stouffer*, 1999]. Diapycnal mixing is also not globally uniform, as assumed in most models, but enhanced in particular regions such as over continental slopes [*Munk and Wunsch*, 1998]. Therefore, it is difficult to infer global values of diffusivities needed in large scale models from local observations. In addition, *Schmittner and Stocker* [2000] showed that the effective vertical diffusion of tracers in their ocean model is also affected by the seasonality in wind forcing.

The purpose of this paper is to examine the dependence of the different equilibria of the THC on model parameters describing sub-gridscale ocean mixing. Our strategy is to use a simple coupled atmosphere-ocean model with which to explore a large range of parameter space.

Model and forcing

We use an idealised zonally-averaged ocean model [*Wright and Stocker*, 1991] in which the Pacific, Indian and Atlantic Oceans are individually resolved and connected through the Southern Ocean. The ocean component is coupled to a zonally- and vertically-averaged energy-moisture balance model with present day seasonal forcing [*Schmittner and Stocker*, 2000] and to a simple thermodynamic sea ice model [*Wright and Stocker*, 1993]. A standard model version with constant vertical ($k_v = 4 \times 10^{-5}\mathrm{m^2s^{-1}}$) and horizontal ($k_h = 1 \times 10^3\mathrm{m^2s^{-1}}$) ocean diffusivities is initialized by spinning up the ocean model under restoring surface boundary conditions from observations. After a steady state is reached, the atmospheric component is coupled to

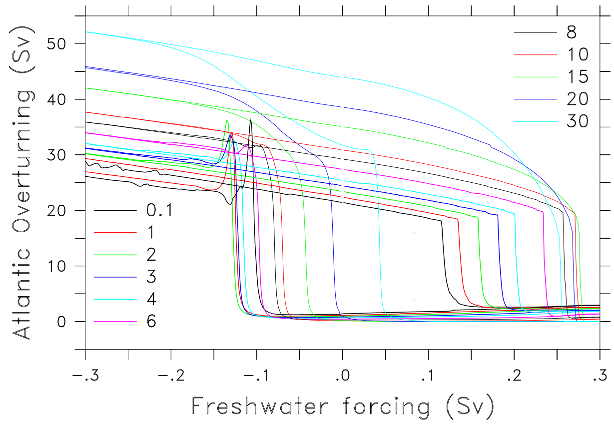

Figure 1. Evolution of the annual mean maximum overturning below 1000 m depth as a function of the freshwater forcing F for model versions with different vertical diffusivities k_v. The integrations start at the upper branch of the hysteresis curves with $F = 0$. F is then slowly increased for 3000 years until $F = 0.3$ Sv. The integration then proceeds on the lower branches, with F decreasing until $F = -0.3$ Sv is reached at year 9000. The upper branches for negative F are started at $F = 0$, and F is decreased until $F = -0.3$ Sv is reached. The different lines correspond to different values of the vertical diffusivity in units of $10^{-5}\mathrm{m}^2\mathrm{s}^{-1}$. $k_h = 1 \times 10^3\mathrm{m}^2\mathrm{s}^{-1}$ in all experiments.

the ocean. At the time of coupling several atmospheric parameters are determined involving the heat- and freshwater fluxes from the ocean spinup [*Schmittner and Stocker*, 2000]. After the coupled standard model has reached steady state, ocean mixing parameters are changed and for each combination of k_v and k_h the coupled system is integrated to a different steady state. This procedure ensures that all model parameters (except k_v and k_h) are the same for all model experiments discussed in this paper. The different responses of the coupled model are solely due to the differences in ocean diffusivities.

Two suites of experiments were conducted. In the first suite, equilibrium solutions were obtained for k_v varying from 0.1 to $30 \times 10^{-5}\mathrm{m}^2\mathrm{s}^{-1}$, and k_h constant ($1 \times 10^3\mathrm{m}^2\mathrm{s}^{-1}$). In the second suite of experiments, k_v was held constant ($4 \times 10^{-5}\mathrm{m}^2\mathrm{s}^{-1}$) and several equilibrium solutions were obtained for k_h varying from 0.1 to $10 \times 10^3\mathrm{m}^2\mathrm{s}^{-1}$. Note that the total freshwater budget in the Atlantic north of 47°S is very similar in all unperturbed steady states (0.30 - 0.31 Sv net evaporation).

In order to determine the transition points between the active (with NADW formation) and passive (no NADW formation) equilibria, we follow the methodology of *Mikolajewicz and Maier-Reimer* [1994] and apply a slowly-varying freshwater perturbation in the northernmost box of the North Atlantic. Starting from the present-day equilibrium with an active THC in the North Atlantic, the freshwater flux was changed by 0.1 Sv (1 Sv = $10^6\mathrm{m}^3\mathrm{s}^{-1}$) every 1000 years. Due to the slowly-varying nature of the surface forcing, the model is in quasi-equilibrium during the integration except during the mode transitions. Note that the freshwater forcing in the North Atlantic was not compensated for elsewhere. This leads to small changes in global mean salinity throughout our integrations.

Hysteresis for varying vertical diffusivities

Figure 1 shows the hysteresis curves for model versions with different vertical diffusivities. The unperturbed states ($F = 0$) with an active Atlantic THC (upper branches) show a stronger meridional overturning if the vertical diffusivities are larger, consistent with three-dimensional ocean GCM results [*Bryan*, 1987]. For small values of k_v and F, the response of the THC to the increasing freshwater input is almost linear. Slowly decreased surface densities in the North Atlantic lead to a slow advective spin-down of the THC. Once the forcing exceeds a certain threshold a rapid transition to the mode without NADW formation occurs, as found in different previous studies [*Stocker and Wright*, 1991; *Mikolajewicz and Maier-Reimer*, 1994; *Rahmstorf*, 1995]. This transition is associated with a cessation of convection in the North Atlantic. The critical value of F clearly depends on k_v, such that experiments with larger k_v exhibit higher threshold values. This behaviour can be understood by examining the different steady state initial condition strengths of the THC. The stronger the initial circulation (the larger the k_v) the stronger the northward near surface salt transport associated with the circulation. Therefore only a large perturbation can lead to the salinity reduction necessary for a circulation shut-down. Note that the critical overturning strength in our model (20 Sv) is nearly independent of k_v (Fig. 1).

Once the THC is in the *off* mode, the influence of the varying freshwater perturbations on the circulation is negligible. This is only true until increased evaporation in the North Atlantic exceeds a second threshold beyond which the circulation switches back to the state with active deep water formation. This second threshold also depends on the value of the vertical diffusivity, consistent with the result of *Manabe and Stouffer* [1999]. Higher diffusivities lead to less additional evaporation being necessary for a resumption of the THC. Globally-averaged stratification is reduced for model versions with larger vertical diffusivities due to increased downward heat diffusion in the tropics. Hence the stability of the water column is generally smaller for higher k_v. In order to destabilise the vertical density stratification in the North Atlantic, a smaller increase in sea surface salinities is

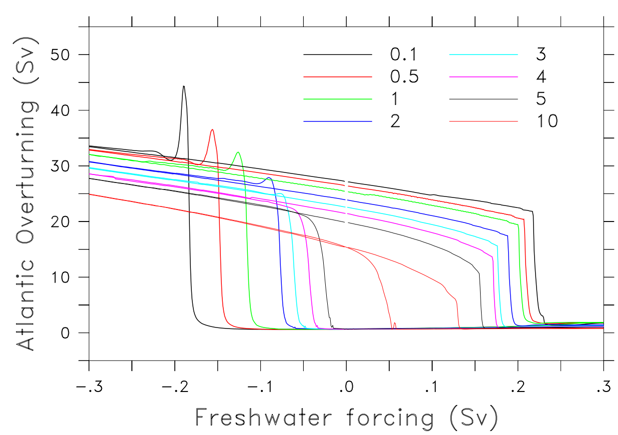

Figure 2. As figure 1 but for model versions with different horizontal diffusivities k_h in units of $10^3\mathrm{m}^2\mathrm{s}^{-1}$. $k_v = 4 \times 10^{-5}\mathrm{m}^2\mathrm{s}^{-1}$ in all experiments.

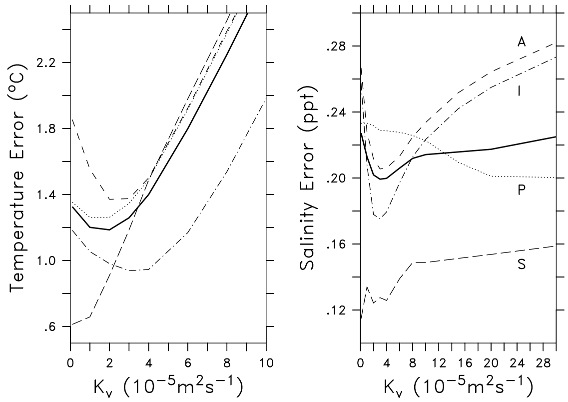

Figure 3. Mean deviation of modelled potential temperature (left) and salinity (right) from observations for experiments with different vertical diffusivities. In calculating the global mean rms error (solid line), we use area-averaged means from the individual basins (short dashed: Atlantic; dash dotted: Indian; dotted: Pacific; long dashed: Southern Ocean). Note that results for the potential temperature error are only given for values of $k_v < 10 \times 10^{-5} m^2 s^{-1}$ since the errors became very large for higher values.

therefore sufficient for model versions with higher vertical diffusivities.

From Fig. 1 it is obvious that the hysteresis curves are shifted towards values of increased precipitation for larger k_v. The width of the phase space in which two equilibria are possible, however, remains similar for the whole range of vertical diffusivities. This is in contrast to the results for different horizontal diffusivities as we will show in the following section.

Hysteresis for varying horizontal diffusivities

The hysteresis curves for model versions with different horizontal diffusivities are shown in Figure 2. The initial steady state overturning without forcing (upper branches with $F = 0$) depends on the value of the horizontal diffusivity. Model versions with larger k_h have a slower meridional overturning, due to a reduction in the meridional density gradients which drive the meridional overturning. The stronger the initial THC, the larger the threshold perturbations needed for a breakdown, similar to the results in the previous section. However, the circulation strength just before the rapid mode change depends on the horizontal diffusivity. It varies from 21 Sv for $k_h = 0.1 \times 10^3 \mathrm{m}^2\mathrm{s}^{-1}$ to 5 Sv for $k_h = 10 \times 10^3 \mathrm{m}^2\mathrm{s}^{-1}$. This is in contrast to the results in the previous section where the minimum sustainable strength of the THC without a collapse was around 20 Sv for all values of the vertical diffusivity.

The near surface poleward salt transport by the THC represents a positive feedback since it keeps salinities in the regions of deep water formation high. This positive feedback is also the reason for the existence of the hysteresis behaviour. Higher horizontal diffusivities lead to stronger diffusive near surface salt transport from the subtropics to high latitudes. Therefore with higher k_h a weaker overturning can still maintain the high salinities in the northern North Atlantic necessary to sustain convection and deep water formation.

The thresholds for a transition from the THC *off* state to the THC *on* state are shifted towards increased precipitation for higher values of the horizontal diffusivities. That is, model versions with larger horizontal diffusion need less additional evaporation to destabilise the water column in the North Atlantic than versions with smaller k_h. One reason for this behaviour lies in the development of the halocline in the North Atlantic in the states without NADW formation. In the cases of increased horizontal diffusion, the halocline is less pronounced and therefore it is easier to reach the critical salinity necessary for a switch to the active THC mode. The width of phase space in which two modes of the circulation are possible is therefore much smaller for model versions with larger horizontal diffusivities than for versions with small k_h.

Model evaluation with respect to diffusivities

In this section an attempt is made to evaluate those vertical and horizontal diffusivities that lead to the model simulation of the present day climate being in best possible agreement with observations. To accomplish this task we compute a global measure of the model performance. First we calculate the root-mean-square (rms) deviation of the annual mean coupled model results from the zonally-averaged observations of potential temperature [*Levitus et al.*, 1994] and salinity [*Levitus and Boyer*, 1994] in the individual basins weighted by the layer thickness. A weighted (with the surface area) average of the mean deviations in the basins yields the global mean model error.

Figure 3 shows the model errors for versions with different vertical diffusivities. Both the global errors in potential temperature and in salinity show absolute minima for $1 \times 10^{-5} m^2 s^{-1} \leq k_v \leq 4 \times 10^{-5} m^2 s^{-1}$. However, the minimum errors in the individual basins may deviate from the global value. The vertical diffusivities are strongly constrained to values smaller than $5 \times 10^{-5} m^2 s^{-1}$ by the potential temperatures in the deep ocean. For larger values of k_v, the deep ocean gets significantly warmer than the observations.

In Figure 4 the corresponding errors for the model versions with different k_h are given. The minimum in the global potential temperature error is much less pronounced than for the vertical diffusivities. From the global potential

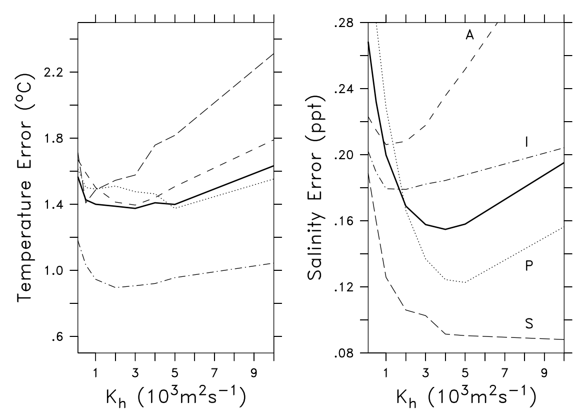

Figure 4. As Fig. 3 but for experiments with different horizontal diffusivities.

temperature error alone it is seemingly impossible to make a unequivocal choice for the optimum value of k_h. However, the global model errors in the salinity have a minimum at $4 \times 10^3 m^2 s^{-1}$. For the Atlantic alone a value of $k_h = 1 \times 10^3 m^2 s^{-1}$ leads to the best agreement with the observations. Mesurements in the open Atlantic [*Ledwell et al.*, 1998] are in excellent agreement with this value. The salinity error in the Atlantic also puts the strongest constraint on the horizontal diffusivities. For values larger than $5 \times 10^3 m^2 s^{-1}$ the error increases strongly because the distinct layers of Antarctic Intermediate Water, NADW and Antarctic Bottom Water are no longer present in the salinity distribution.

From Figs. 3 and 4 a realistic range of diffusivities for our model would be $1 \times 10^{-5} m^2 s^{-1} \leq k_v \leq 6 \times 10^{-5} m^2 s^{-1}$ and $1 \times 10^3 m^2 s^{-1} \leq k_h \leq 5 \times 10^3 m^2 s^{-1}$ with the globally-best estimates $k_v = 2 \times 10^{-5} m^2 s^{-1}$ and $k_h = 4 \times 10^3 m^2 s^{-1}$. Note that these ranges are only valid for the present model version. Other models, particularly GCMs, would most probably result in different parameter ranges. We are now in a position to draw a number of conclusions, with reference to Figs. 1 and 2, concerning the sensitivity of the threshold, between *on* and *off* states, to the magnitude of horizontal and vertical mixing. The threshold value for a breakdown of the North Atlantic THC increases by 75% for an increase of k_v in the above range, while the threshold for a transition from the *off* to *on* states, is less sensitive. For an increase of the horizontal diffusivities in the given range the threshold for a THC collapse decreases by about 40%, while the threshold for a restart decreases from around –0.1 Sv to –0.02 Sv. Over this same range, the value of the minimum sustainable overturning decreases by 30% from 20 Sv to 14 Sv. Taken together, these results strongly suggest that caution be taken when interpreting thresholds in any particular model given the great uncertainty in ocean mixing.

Discussion

Idealised models like the one we used in this study are not suitable for a realistic assessment of threshold values. However, our well tested model is useful for sensitivity studies like the one presented. We believe that the physical processes leading to the dependence of the stability of the different circulation modes and their transition points on sub-gridscale mixing parameters are a robust feature that is qualitatively valid in all large scale ocean models. While the study of *Manabe and Stouffer* [1999] is consistent with our results, a direct confirmation of our results with three dimensional GCMs is desirable.

We repeated the experiments described above with a version of the model [*Knutti et al.*, 2000] that uses isopycnal and diapycnal diffusion instead of lateral diffusion, as well as a parameterisation of eddy induced tracer advection. Varying isopycnal and diapycnal diffusivities led to very similar results to the ones presented here. Our study points to the central role of ocean sub-gridscale mixing parametrisations in simulations of future climate. Until uncertainties in these parameterisations are significantly reduced, it will remain impossible to provide a reliable estimate of the probability of rapid climate transitions associated with mode changes of the Atlantic thermohaline circulation.

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