On periodic solutions found in simple ocean models with fixed surface fluxes

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Abstract

In a recent paper Greatbatch and Zhang reported the finding of interdecadal oscillations in an idealized ocean basin forced by constant heat flux. This oscillation has many similarities to that found by Delworth et al. in a coupled ocean-atmosphere model. We have used three simple models with fixed surface fluxes, a box model, a Welander-like loop model and a three-level three-dimensional ocean basin model, in order to compare mechanisms associated with interdecadal oscillations in these models. Our experiments with the basin model revealed the three-dimensional character of the oscillations in three-dimensional models. Self-sustained oscillations are associated with a reduction of the propagation speed of cold anomalies in the north-western corner of the model because of anomalous strong convection in that region. Then anomalous
gyre circulation leads to the development of strong temperature anomalies. The extension of the period of strong (low) overturning in the three-dimensional (loop) model by some overcompensation mechanism is crucial for the occurrence of self-sustained oscillations. In conclusion we confirm Wintons view that fixed flux variability is not a meridional plane phenomenon related to loop oscillators.

1 Introduction

Decadal and interdecadal variability is an important phenomenon of the climate system. It is necessary to understand the relevant processes, which can lead to long-time variability in the system ocean-atmosphere to properly access the rate of global warming and to distinguish between long-time oscillations and the climate signal (see Ghil and Vautard 1991). There are different possible causes for long-term oscillations: (1) random radiative forcing of the climate system (see Schlesinger and Ramankutty 1995), (2) oscillatory forcing external to the climate system, for instance, by a variation in solar irradiance, (3) an oscillation internal to the climate system, produced by the atmosphere (James and James 1989) or the ocean (Delworth et al. 1993, Greatbatch and Zhang 1995) or by a coupled mechanism between the ocean and the atmosphere, perhaps not unlike that of the El Niño-Southern Oscillation, but with a longer time scale.

In this paper we consider internal oscillations in the ocean-only with fixed surface flux fields as boundary conditions. This is quite unusual, because most ocean models use restoring boundary conditions or mixed boundary conditions (restoring boundary conditions on the surface temperature and constant flux conditions on the surface salinity). In addition, some of these models use new parameterizations for atmospheric transports (Rahmstorf and Willebrand 1995). However, the boundary conditions are essential for the stability of the thermohaline circulation (THC) and the occurrence of “catastrophes” and oscillations in the THC. In a number of ocean-only models under mixed boundary conditions (e.g. Weaver and Sarachik 1991a,b) and also in a run of the coupled ocean-atmosphere model from the Geophysical Fluid Dynamics Laboratory (GFDL) (Delworth et al.1993) interdecadal oscillations are present. In the paper mentioned last, the authors found irregular oscillations of the THC in the North Atlantic with a time scale of 40-50 years. Greatbatch and Zhang (1995) have shown that similar oscillations can be produced by an ocean-only model under constant flux conditions. They argue that changes in the surface boundary fluxes play only a weak role in the Atlantic oscillations of the ocean-atmosphere model. This is confirmed by Figures 14 and 15 in Delworth et al. (1993). In addition Chen and Gill (1995) argue, that a fixed heat flux is consistent with an realistic atmosphere with zero heat capacity which can respond instantaneously to SST changes. However, it is interesting that, even with zero salinity flux and no windstress, one can generate oscillations in sea surface temperature (SST) and the THC with quite similar structures like in the complex GFDL-model. On the other hand Delworth et al. (1993) need advected salinity anomalies in their explanation for the oscillation in the North Atlantic.
The North Atlantic seems to be an essential part of the global climate system. This is not only obvious because of various model results, but also on account of data analysis. For instance, Schlesinger and Ramankutty (1995) reported that the 65-70 year oscillation in the observed global-mean surface temperature are mainly caused by oscillations over the North Atlantic ocean and its bordering continental regions. Kushnir (1994) and Deser and Blackmon (1993) rather think that the observed variations of SST in the North Atlantic with a time scale of 30-40 years were caused by changes in the ocean circulation. Obviously there is similarity between the SST-anomaly pattern described by Kushnir (1994) and that assigned to the oscillation in the GFDL-model (Delworth et al. 1993). Also, of course, this fact connects the oscillations described by Greatbatch and Zhang (1995) to the observations of Kushnir (1994).

We think there is not yet a complete and clear picture of the mechanism of these oscillations. But without this natural climate variability on decadal and longer timescales can not be understood. In addition to the arguments given above, using fixed surface flux conditions is attractive, because the situation in that case is less confusing than in the case of mixed boundary conditions. We consider interdecadal oscillations in three different simple ocean models with fixed surface flux conditions, a Stommel-like box model (Stommel 1961), a loop model similar to Welanders model (Welander 1986) and a three-dimensional model (3D model) similar to that of Killworth (1985). The concern of the present paper is two fold: (i), we investigate the ability of much simpler models than 3D ocean models to produce interdecadal oscillations and elucidate the mechanism responsible for these oscillations. (ii), we use a three-level basin ocean model, comparable to the model used by Greatbatch and Zhang (1995) (apart from the low vertical resolution) in order to reproduce the oscillations found by Greatbatch and Zhang (1995) and try to find common features between the low-dimensional models and the three-layer model. The motivation is to deepen the understanding of the oscillations found recently in more complex models by Delworth et al. (1993) and Greatbatch and Zhang (1995).

In this paper, the phase-relationship between the strength of the thermohaline overturning and the north-south temperature difference is discussed. We find transient solutions in the box model if we take into account the fact that in 3D models there is a phase lag between these two time-series. A deformed Welander-like loop model also shows a phase lag between these time-series, though different from the 3D and the box model. Although it is a new result that a non-inertial Welander-like loop model shows self-sustained oscillations when constant heat and no salinity fluxes are employed, the experiments confirm Winton's (1996) view that the oscillations in 3D models can not be simulated by two-dimensional or loop-like models. Our experiments with the 3D three-layer model reveal the subtle interplay between advection, convection and boundary waves which leads to self-sustained interdecadal oscillations. The discussion of one oscillation cycle at the end of the present paper reminds of the cycle described by Delworth et al. (1993) and is based on the mechanism discussed by Greatbatch and Peterson (1996). But in contrast to the last mentioned authors model results presented here show that anomalous convection in the north-western part of the model basin is a key process for the existence of self-sustained oscillations, whereas anomalous advection
is in the end responsible for the growth of temperature anomalies north of the center of the model basin.

In the present paper only the model description, the experiments and results of the 3D model are presented in detail. A perceptive discussion of the other two models used can be found in Harlander (1996). The structure of this paper is as follows. In the next section a brief description of the used 3D model is given, together with the assumptions for the used forcing terms. Also we discuss some experiments, done with different model parameters. The last section provides a summary and a conclusion.

2 Three-level model

2.1 Model description

In this section a simple three-level basin ocean model is used to examine the oscillatory behavior of a much more complicate system compared to a box or loop model under fixed heat and salt fluxes. The model is based on the Killworth (1985) model and is very similar to that used by Lenderink and Haarsma (1994). Further information about the model is given by Harlander (1996).

For the atmospheric forcing terms $F_T, F_S$ and the windstress $\tau$ we use the simple functions

$$F_S = -\alpha_S \cos(\pi \frac{y}{L})$$

$$F_T = \alpha_T \cos(\frac{y}{L})$$

$$\tau_x = -\alpha_W \cos(2\pi \frac{y}{L})$$

$$\tau_y = 0$$

$L$ is the width of the basin, $\alpha_S, \alpha_T, \alpha_W$ are constants given in Table 1.

2.2 Experiments

We carry out experiments with different parameters $\alpha_S, \alpha_T, \alpha_W$ (see Table 1). All the other model parameters are fixed (see appendix). Figure 1 displays the temporal evolution of the maximal meridional overturning in Sverdrup $(10^6 m^3 s^{-1})$, defined as

$$\omega_{Max}(t) = \max_{y,z} \left( \psi(y, z, t) \right) = \max_{y,z} \left( \int_0^L \left( \int_0^L v dx \right) dz \right)$$

of run 1 in Table 1 (no salinity flux, constant heat flux, no wind stress) and run 2 in Table 1 (constant salinity flux, constant heat flux, no wind stress)(black and gray solid line, respectively). Strong but damped oscillations can be observed. Besides the oscillations in the strength of the overturning we also find oscillations in the position of the maximal overturning, the area of deep and shallow convection, the heat and salt transport and the north-south temperature difference.
Table 1: Parameters in the atmospheric forcing terms. The units are $\alpha_S (10^{-8} \, \text{PSU s}^{-1})$, $\alpha_T (10^{-7} \, \text{K s}^{-1})$, $\alpha_W (\text{Nm}^{-2})$. The amplitude of the salinity flux is $0.7 \, \text{m} \, \text{Y}^{-1}$, the amplitude of the heat flux is $1.04 \, \text{°C} \, \text{M}^{-1}$ in run 1 and run 2.

<table>
<thead>
<tr>
<th>Run No.</th>
<th>$\alpha_S$</th>
<th>$\alpha_T$</th>
<th>$\alpha_W$</th>
</tr>
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</tr>
<tr>
<td>3</td>
<td>1.5</td>
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A full oscillation-cycle in SST and deep water temperature (DWT) anomaly is portrayed in Figure 3 (SST upper column, DWT lower column). The structure of the SST anomaly fields is comparable to Figure 6 of Greatbatch and Zhang (1995), but with larger amplitudes. In the state of growing overturning, we find anomalous cold water in the northern part of the model domain, in the SST as well as in the DWT (Fig. 3 (a) and (e)). After 19 years, the overturning is at its maximum (Fig. 3 (b) and (f)) and we find a dipole structure in the temperature anomaly fields in the northern part. While there is a strong meridional transport of warm waters to the north, the cold anomaly moves to the north-west corner of the model basin. There is also a weak cold anomaly in the southern surface level and a warm anomaly in the southern deep level. Figure 3 (c) and (g) show the anomalies in the phase of decreasing overturning. The warm anomalies are shifted to the north-west and, in the surface level, the strong warm anomaly is framed by two weaker cold anomalies. This structure is similar to Figure 18 (lag+10 years) of Delworth et al. (1993). After further 19 years, the overturning is at its minimum (Fig. 3 (d) and (h)). The warm anomaly drifted to the western boundary of the basin and a large cold anomaly developed in the surface level. Here again a weak warm anomaly exists in the southern surface level and a cold anomaly in DWT in the south-western part of the basin.

After 1650 years of integration the amplitude in run 1 is still 3.5 Sv with a period of 65 years. An addition of salinity flux (run 2 in Table 1) reduces the amplitude and lengthens the period (about 73 years) (gray solid line in Fig. 1). These results are in qualitative agreement with the results of Greatbatch and Zhang (1995), nevertheless Greatbatch and Zhang do not have damping in their experiment. In run 3 of Table 1 (dashed line in Fig. 1) we add a weak and temporal constant windstress and slightly reduce the amplitude of the thermal forcing. Undamped self-sustained oscillations with a much larger amplitude, even larger as in run 1 are observed under these conditions. An additional windstress leads to a more efficient northward heat transport at the western boundary of the 3D model. We have performed further experiments with a space and time variable windstress which show that in a particular parameter range windstress can support the interdecadal oscillations. On the other hand, if the windstress forcing is strong, the oscillations vanish. Hence, there seems to be a bifurcation point in the windstress parameter space. The connection between windstress and interdecadal
variability under constant fluxes is not the subject matter of the present paper and we hope to discuss this point more fully in a later investigation.

2.3 The oscillation mechanism

Before we give a quantitative explanation of the oscillations described above and shown in Figure 3 we first compare some phase-relations of the box and the ring model with the corresponding phase-relations of the three-level model. Afterwards we take a closer look at different processes which are important for the occurrence of oscillations in the basin model. Based on these preparations the course of events within the oscillation cycle shown in Figure 3 is evaluated.

2.3.1 Comparison of phase-relations

In Figure 4 the cross-correlation of overturning and north-south temperature difference of the box model run (solid gray line), the ring model run 6 (dashed line) and the three-level model run 1 (solid line) is shown. North-south temperature difference in the ring model is defined as \( T(\psi = 3\pi/2) - T(\psi = \pi/2) \) and in the basin model as the difference between the mean SST in the southern part of the model (grid point \( x = 1 \) to 22 and \( y = 1 \) to 6) and the northern part (grid point \( x = 1 \) to 22 and \( y = 17 \) to 22). In all models the correlation between these time-series is high and the phase-relation in the box model and the 3D model is comparable (a maximum in temperature difference is ahead a maximum in overturning). On the other hand, the phase-relation in the ring model shows that a maximal overturning is ahead a maximal temperature difference. So, similar to the box model, the problem arises that a loop-type model has no information about the east-west density gradient, which determines the overturning in the 3D model.

Next the phase-relations between the (maximal) overturning and the deep water temperature in the ascending part of the model circulation are compared. In that analysis the temperature time-series of the ring model is based on the mean temperature of the two grid points left from the connection point \( A \), where the non-curved part of the ring is connected to the curved part, whereas the time-series of the three-level model is obtained by averaging the temperature of an area in the lower left quadrant of the deep model layer (grid point 2 to 10 in x- and grid point 1 to 7 in y-direction). The cross-correlation for the non-inertial ring model and the three-level model is shown in Figure 5. The correlation between these two time-series is large in both models and the phase-relations are almost identical. A maximal overturning is followed by a minimal temperature in the deep southern part of the models. Based on these results it appears that the overcompensation described by Harlander (1996) for the deformed ring may also be relevant to the oscillations of the three-layer model.

To test this hypothesis we have performed a simulation equal to run 1 with the exception that any temperature anomalies are suppressed in the southern part of the model by a relaxation to the climatological mean temperature distribution in that area. Relaxation constants of \( 230^{-1} \text{days}^{-1} \) in the upper layers and \( 115^{-1} \text{days}^{-1} \) in the deep
layer were applied. As shown in Figure 2 (dashed line) we find now oscillations without damping and a slightly longer period. This evidences that temperature anomalies in the southern half of our model domain do not play an essential part for the occurrence of the observed oscillations, in contrast to the oscillations found in the ring-model.

Therefore we conclude that the two-dimensional view, e.g., with a conceptual model like the ring model, is not sufficient to describe oscillations occurring in the 3D model, even if we employ constant heat and no salt fluxes, which is probably the most simple forcing configuration. This statement is consistent with results of Winton (1996), who used a two-dimensional model version with a fixed forcing. However, in contrast to the two-dimensional model of Winton, the ring model does show self-sustained oscillations, even if the mechanisms responsible for these oscillations are different from that of the 3D model. In spite of these differences, we can draw an instructive analogy between the overcompensation in the loop and the 3D model, but this must be postponed until the discussion of the oscillations in the basin model is complete.

2.3.2 Boundary waves

As we have seen in the previous section, the temperature anomalies in the south are responsible for the damping in run 1 of Table 1. The reason for this is that traveling boundary waves disturb the development of temperature anomalies in the central part of the surface layer. These waves can be observed at the eastern boundary in Figure 3 A and D. This observation could give the impression that boundary waves are unimportant or even detrimental to interdecadal oscillations in coarse resolved ocean models, which is not the case. We turn our attention to the westward (southward) propagation of the temperature anomalies at the northern (western) boundary which is quite obvious in Figure 3. As pointed out by Winton (1996) and Greatbatch and Peterson (1996) this propagation is not associated with advection by the mean flow. Linear numerical boundary waves discussed by Killworth (1985), which adopt the role of coarsely-resolved, viscous, baroclinic Kelvin waves (cf. Winton 1996), are responsible for the observed propagation. An example of such a numerical wave propagation is shown in Figure 6. In this experiment any forcing is switched off. At time $t = 0$ the model was in a state of rest and the model-layer temperatures were constant $T_s = 15°C$, $T_1 = 7°C$, $T_2 = 5°C$. Then a Gaussian disturbance was introduced at the northern boundary of the model. This negative temperature anomaly propagates westward and also spreads southward at the western boundary. Besides that we find a prominent dispersion and diffusion of the wave packet. The propagation speed depends on the stratification of the model. Because of the existence of deep convection, usually a weak stratification occurs at the northern part of the model basin. Therefore, the propagation can have interdecadal time scales (see also Greatbatch and Peterson 1996).

2.3.3 Convection

As shown by Winton (1996) and Greatbatch and Peterson (1996) the propagation of boundary waves at the northern and western edge is essential for the occurrence of
interdecadal oscillations in coarse resoluted ocean models. Variations in deep convection seem also to be of great importance. In the model used here, convection is either switched on or off. If denser water is layered above less dense water, a complete mixing takes place. We define a corresponding convection parameter as follows: \( p^{ijk} = 1 \) in the case of convection, otherwise \( p^{ijk} = 0 \), where \( i, j \) and \( k \) denotes the grid point in x- and y-direction and \( k \) denotes the layer, respectively. Consequently it follows that the time averaged convection parameter is equivalent to the relative frequency of convection. Figure 7 shows snapshots of the deviation from the mean relative frequency of convection between the surface layer and layer 1 at the same time as the SST anomalies displayed in Figure 3. The convection pattern is clearly associated with the temperature anomalies and normally enhanced (reduced) convection occurs in regions with anomalous cold (warm) SST. If a temperature anomaly is propagating from east to west, it can also be seen in the convection pattern. Note that a switched off convection in the northern model domain leads to a decrease in the SST because of the forcing. To examine the role of convection for the occurrence of interdecadal oscillations we have repeated run 1 but after 200 years of integration either the area of convection is fixed between the surface and layer 1 or between layer 1 and layer 2. In both experiments the oscillations completely vanish after a few hundred years. So the propagating temperature anomalies may not be seen independently from convection anomalies.

2.3.4 One oscillation cycle

In this subsection we put together the single processes discussed so far to built a complete picture of the interdecadal oscillations in the three-level model with fixed surface fluxes. First of all, in Figure 8, the time evolution of the heat budget is considered for the last 180 years of the model integration (run 1 of Table 1 but with a relaxation in the southern part). For the grid point \( x = 13, y = 15 \), the change in SST due to advection (solid black line) and due to convection (dashed line) and, furthermore, the SST is shown. We find maximal (minimal) heat advection when a negative (positive) temperature anomaly is located west of the grid point and a positive (negative) anomaly east of it. In this period the SST increases (decreases) and convection is switched off (on). A striking feature of Figure 8 is that the period of convection at this particular grid point is much shorter than the period with no convection. That means that there is a relatively long (short) period with a cold (warm) anomaly in the north western part of the model domain. This is very obvious in Figure 9, where we have plotted the time evolution of the temperatures in all three layers at a grid point in the north-western corner \( (x = 4, y = 19) \). Most of the time convection occurs at that grid point together with equal temperatures in all layers. The cold anomaly in the north western part seems to be slowed down. Greatbatch and Peterson (1996) were the first who described the "held up" of disturbances at the north-western corner in their basin model. They believe the slow propagation along the northern boundary leads to a considerable increase in the amplitude of the wave, which is the fundament for the occurrence of self-sustained oscillations. Note that
Only cold anomalies are "held up". We believe the reason for this is anomalous strong convection in periods with anomalous cold surface water in the northwestern part of the model (see Fig.7 B). As mentioned by Killworth (1985) the phase velocity of numerical boundary waves in his two-level model is

\[ C_{ph} \approx \frac{\alpha S g H}{2 \rho_0 \Delta x}, \]  

where \( \Delta x \) is the grid point distance in x-direction, \( H \) stands for the depth of the fluid, \( g \) for the constant of gravity, \( f \) for the coriolis parameter, \( \rho_0 \) for the mean density, \( \alpha \) for the ratio of layer depth to fluid depth and \( \bar{S} \) stands for the density difference between the two layers. The density difference \( \bar{S} \) is almost zero in regions with convection, therefore the spatial mean of \( \bar{S} \) and the speed \( C_{ph} \) in the northwestern corner of the model domain is anomalous low in periods of anomalous strong convection. A very important point is that "new" temperature anomalies are produced in the center of the northern part of the model by anomalous advection or, in other words, by the secondary circulation of already exiting "old" temperature anomalies. If a cold anomaly is locked in the northwestern corner, positive heat advection has enough time to build up a large positive temperature anomaly. This seems to be the overcompensating mechanism in the 3D model which enables the oscillations to be self-sustained.

The results substantiate that the subtle interplay between boundary waves, anomalous convection and advection allows the non-stationary behavior of the model. Our experiments confirm this view: No oscillations will be found if we suppress boundary waves at the northern and western boundaries (see also Greatbatch and Peterson 1996) as well as if we suppress anomalous convection (see subsection 2.3.3).

Damped oscillations occurring in run 1 of Table 1 result from disturbing this interplay: Boundary waves, which are propagating to the north at the eastern boundary, modify the "slow-down" process of cold anomalies at the northern edge. By using a smaller heat flux \( \alpha = 2.5 \cdot 10^{-7} K s^{-1} \) in run 1 or by employing the parameter configuration of run 3 in Table 1, the system adjusts in a way that self-sustained oscillations occur. Therefore, in the three-dimensional \( \alpha_T - \alpha_S - \alpha_W \)-parameter space, self-sustained oscillations are not a "global" phenomenon but appear in particular regions of this space.

3 Summary and conclusion

We have studied interdecadal oscillations by three different simple models, a box model, a so-called ring model and a three-level ocean basin model, forced by constant heat and salt fluxes. The motivation was to deepen the understanding of the oscillations found recently in more complex models by Delworth et al. (1993) and Greatbatch and Zhang (1995) and to examine if lower dimensional models capture mechanisms which are also important for the occurrence of interdecadal oscillations in the 3D model presented here. Greatbatch and Zhang (1995) described the essential mechanism for interdecadal oscillations in their model as follows: "Since there is a constant rate of heat loss at
high latitudes, the surface residence time of a water particle determines the temperature anomaly at high latitudes. Strong overturning rates lead to strong meridional advection and to short residence times and therefore to warm surface anomalies at high latitudes. This weakens the circulation and anomalies with different signs may develop because of the forcing terms, and so on. The same mechanism was put forward by Huang and Chou (1994) and, quite similar, by Weaver et al. (1991) as an explanation for the oscillation they found under constant freshwater flux forcing. But from a naive point of view, one can not see why that mechanism alone can produce self-sustained oscillations in a model with constant forcing. It would be possible just as well that the system reaches a stable state after some initial oscillations. There needs to be some overcompensation mechanism, for instance inertia, to produce longer lasting oscillations.

Usually, in 3D basin models a maximum in north-south temperature difference is ahead a maximum in overturning. This out of phase relationship stands in contrast to the assumptions in box models of the "Stommel-type" (cf. Stommel 1961), where no phase lag between these two time-series exists. Our experiment with a box model with constant heat forcing showed that (damped) oscillations can be a solution of the model if a phase lag between the temperature difference of the two boxes and the box model "overturning" is artificially incorporated. A physical justification for the implementation of this phase relationship is that box models do consider neither the large distance of the two boxes nor the east-west pressure difference which determines the overturning in 3D models.

Our next step was to study the dynamics of a simple system which is able to produce a dynamically caused internal phase lag between overturning and north-south temperature difference. This system consists of a ring (or loop) of fluid which is forced by constant heat fluxes. A similar convective loop model was investigated earlier by Welander (1967), Malkus (1972) and Welander (1986). It can be shown that a circular symmetric ring can have non-steady solutions if inertia is incorporated (cf. Welander 1967) or if the relaxation constants of salinity and temperature are not equal (cf. Welander 1986). In addition, we have shown that even a non-inertial ring system with constant heat and no salinity fluxes can produce oscillatory solutions if the symmetry of the ring is broken by deforming its upper or lower side. The mechanism responsible for these oscillations in a ring with a flat bottom can be summarized in simplified form as follows. The ring overturning is at its maximum if a cold anomaly is located in the sinking part of the fluid and a warm anomaly is located at the rising part. After some time relative warm water is advected into the region of cooling and, in addition, the cold water in the sinking part of the ring enters the flat region of the ring and does not contribute to the momentum torque of the fluid loop anymore. Therefore, the overturning is strongly reduced. As a result of the slow circulation, the fluid particles in the forcing regions are exposed a relatively long period to the constant cooling and warming, respectively, and "new" temperature anomalies develop. A further decrease of the overturning rate, or at least a slowed down increase, takes place if the negative temperature anomaly leaves the non-curved part of the ring and starts to rise. This effect leads to an overcompensation and in the phase of minimal
overturning strong anomalies are produced in the forcing regions. These anomalies then produce an increase in overturning and so forth. Note that a Welander-like loop model can deal with the fact that the movements of waters between equatorial and polar regions are associated with large time-lag, but it has, of course, no information about zonal density gradients. The phase relationship between overturning and north-south temperature difference of the ring differs fundamental from that of our 3D model, but on the other hand the phase-lag between deep water temperature and overturning is remarkably alike in both models. In spite of this similarity, the experiments have shown that it is not straightforward to transfer aspects of the overcompensation in the loop model to the 3D case.

The interdecadal oscillation in the 3D basin model seems to be a fully 3D phenomenon and anomalous horizontal advection, numerical boundary waves and anomalous convection are indispensable ingredients for the occurrence of these oscillations. In the following one cycle of such an oscillation is summarized. We start from a state of anomalous weak meridional overturning. At this time a positive SST anomaly is observed in the north-western part of the basin whereas a negative anomaly is located north of the center of the model domain. This cold pool has an associated cyclonic circulation, therefore at the western flank of the cold anomaly there is anomalous southward advection and at the eastern side we find (weaker) anomalous northward advection. Because of boundary effects, the whole system is propagating westward. The propagation speed depends on the stratification, which is very low in areas with deep convection. At the western boundary the warm anomaly moves southward and weakens in intensity. The anomalous northward advection at the eastern side of the cold anomaly now leads to the development of a positive SST anomaly north of the center of the model basin. When the surface cold anomaly arrives at the north-western corner of the basin, the westward propagation of the dipole-like anomaly pattern slows down, because an anomalous large area of deep convection at the north-western part reduces the stratification. We believe that this retardation process is crucial for the maintenance of oscillations, because now the anomalous heat advection at the eastern flank of the cold anomaly has enough time to build up a large positive SST anomaly with an associated anticyclonal circulation. The cold anomaly is propagating to the south along the western boundary, weakens and the warm anomaly takes the place of the cold anomaly at the north-western corner of the model basin. This is the state of minimal meridional overturning, thus we have completed one cycle.

The described oscillation reminds one of the oscillations found by Delworth et al. (1993) in the coupled GFDL model. These authors also emphasize the close relationship between the anomalous gyre circulation and the THC. In contrast to the mechanism presented here, anomalous salinity transport into the sinking regions plays an important role in the mechanism described by Delworth et al. However, it is interesting that the authors referred to last found model evidence that the density induced gyre circulation propagates to the west with a speed on the order of $\approx 0.5-1\text{cm s}^{-1}$, which is consistent with the propagation speed of boundary waves in the Killworth-model and the model used by Greatbatch and Peterson (1996), respectively.

In conclusion it seems that the oscillations occurring in the basin model may be the
archetype of a variety of similar solutions in more complex 3D models. In the model experiments presented here the relationship between the oscillations in the ring model and the 3D model are rather weak. However, in both models the observed self-sustained oscillations are associated with a retardation process, which allows the development of strong temperature anomalies: In the asymmetric ring (with the flat bottom) the overturning is slowed down by cold anomalies which leave the flat part of the ring, in the 3D model boundary wave propagation is "held up" by anomalous strong convection and anomalous low stratification in the north-western part of the basin. Therefore, in the ring model the period of low overturning is extended, which in turn leads to strong temperature anomalies in the constant forcing regions. In the 3D case, the period of strong overturning is extended and with it the period of anomalous northward heat advection. The phase relationship between overturning and north-south temperature difference is a result of the overcompensating mechanisms discussed above. This completes the analogy recently drawn by Rahmstorf et al. (1996) between constant surface flux oscillations in 3D models and the thermal loop oscillator in the following way that inertia is not the only mechanism in low dimensional models standing in for the processes producing the oscillations in 3D models.

Acknowledgments. I thank Werner Metz and Yorck von Detten for stimulating discussions relating to this work. I am grateful to Joseph Egger for placing the main part of the three-level model at my disposal. I also thank Nicole Mölders for proofreading the manuscript. This work was partly supported by the BMFT-Projekt "Klimavariabilität und Signalanalyse".

APPENDIX
Model parameters

In all models the same timestepping scheme is used: A leapfrog scheme for the advective terms and a Euler forward scheme for the diffusive terms. The time-step is 3days for the three-level model. In all models $\rho_0 = 1001.3 \text{ kg m}^{-3}$, $k_1 = 0.7739 \text{ kg m}^{-3} \text{ PSU}^{-1}$, $T^* = -7.73^\circ C$ is applied and in model runs with constant salinity $S$ is set to 34.75 PSU. The spatial resolution of the three-level model is $\Delta x = \Delta y = 250km$. Furthermore the following model parameters are used: $f = f_0 + \beta y$, $f_0 = 1 \cdot 10^{-4} \text{s}^{-1}$, $\beta = 2 \cdot 10^{-11} \text{m}^{-1} \text{s}^{-1}$, the layer thicknesses $h_s = 50m$, $h_1 = 400m$, $h_2 = 4000m$, the horizontal diffusion parameter $\kappa_h = 1500m^2 \text{s}^{-1}$, the vertical diffusion parameter $\kappa_v = 1 \cdot 10^{-4} \text{m}^2 \text{s}^{-1}$ and the Stommel-friction $\kappa_s = 4 \cdot 10^{-6} \text{s}^{-1}$.

References


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Figure 1: Temporal evolution of the maximum of the meridional overturning in the basin model of run 1 (black solid line), run 2 (gray solid line) and run 3 (dashed line) in $Sv$ ($1 Sv = 10^6 m^3 s^{-1}$).

Figure 2: Temporal evolution of the maximum of the meridional overturning in the basin model of run 1 (solid line) and run 1 with relaxation (dashed line) (see text) in $Sv$ ($1 Sv = 10^6 m^3 s^{-1}$).
Figure 3: SST (left column) and deep level temperature anomalies (right column) during a full oscillation cycle in the basin model (run 1). Each of (a), (b), (c), (d) and (e), (f), (g), (h), respectively are 19 years apart, whereas (d), (a) and (h), (e) are 9 years apart, with the meridional overturning being a maximum at (b),(f) and a minimum at (d),(h). The contour intervals for SST and deep water anomalies are 1K and 0.1K, respectively. Negative anomalies are shown using dashed contours.
Figure 4: Cross-correlation between the time-series of meridional overturning and the time-series of north-south SST difference. Box model (gray line), ring model run 6 (dashed line) and three-level model run 1 (solid line).

Figure 5: Cross-correlation between the time-series of meridional overturning and the time-series of deep water temperature at the ascending part of the circulation (see text). Three-level model run 1 (solid line), non-inertial ring model run 6 (dashed line).
Figure 6: Surface boundary wave propagation at the northern boundary. The initial Gaussian SST perturbation of the steady state with $T_s = 15^\circ C$, $T_1 = 7^\circ C$, $T_2 = 5^\circ C$, a constant salinity and no forcing (a), at $t = 2.5$ years (b) and at $t = 4.1$ years (c). The contour intervals are 0.5 K.

Figure 7: Convection parameter anomalies (see text) between surface layer and layer 1 at the same time as shown in Figure 6. The dimensionless contour intervals are 0.1.
Figure 8: SST change at grid point $x = 13$, $y = 15$ associated with convection (dashed line), advection (black solid line) and SST (gray solid line) during the last 180 years of basin model integration (run 1 with relaxation) in $K \, s^{-1}$ and $10^7\, K$, respectively.

Figure 9: Temperature development at grid point $x = 4$, $y = 19$ in the north-western corner of the model basin during the last 180 years of model integration (run 1 with relaxation) in $K$. 