Climate Dynamics (1995) 11:447–458

Climate Dynamics

Climate drift in an ocean model coupled to a simple, perfectly matched atmosphere

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Received: 25 August 1994 / Accepted: 30 May 1995

Abstract. A very simple, diffusive energy balance atmosphere is coupled to the GFDL ocean circulation model. This provides a useful tool for analyzing climate drift in the ocean model after coupling, and may be used to assess various schemes for minimizing such drift. In the experiment reported here, the atmosphere is constructed in such a way that it provides the ocean model at the moment of coupling with the same fluxes as during spinup. The experiment is therefore equivalent to coupling a perfectly flux-corrected atmosphere model, and is used to investigate the response of the ocean model under these conditions. In spite of the steady, passive, flux-corrected atmosphere, the ocean model drifts to a new equilibrium state after coupling. The transition takes about 2000 years; the new state is characterized by different sites of deep convection and resulting changes in high-latitude SST and global deep temperatures. The mechanism for the transition is an instability of the oceanic convection patterns under the new feedback, felt after coupling. A similar state transition of the ocean model may be triggered by the coupling shock in fully coupled GCMs. If this is so, the transition would contaminate the results of climate scenario experiments, and it would explain part of the residual drift observed in coupled models in spite of the use of flux corrections.

1 Introduction

Coupled general circulation models of the ocean and atmosphere are an increasingly important tool used to study the climate of our home planet. They have even gained political relevance at a time when humanity is interfering with the climatic balance in numerous ways, and when the capability to understand and forecast the long-term effects of this interference has become a pressing need (Houghton et al. 1990). Due to the large computer power needed, it has only recently become feasible to run coupled circulation models for climate forecasts. While there is considerable experience with atmospheric circulation models and, to a lesser extent, with ocean circulation models, only little is known about their coupled behaviour.

It is still too expensive to perform routinely experiments of several thousand years duration with coupled GCMs. Such long model integrations would be required, however, for the coupled system to find a quasi-steady equilibrium state, since the response time of the deep ocean circulation is measured in thousands of years. To bypass the need for prohibitively expensive coupled "spin-up" runs, the ocean and atmosphere models are either spun up (i.e. integrated to equilibrium) using an asynchronous integration method (Manabe and Stouffer 1988), or they are spun up separately and then coupled (Cubasch et al. 1992; Manabe and Stouffer 1994). Both approaches have problems. A coupled spin-up from rest may lead to an unrealistic equilibrium state, e.g. without the oceanic conveyor belt circulation (Manabe and Stouffer 1988). A separate spin-up of ocean and atmosphere allows the preparation of realistic initial states for both subsystems, but the equilibria obtained separately do not necessarily match well, nor are they necessarily stable under coupled conditions.

The latter is particularly true for the ocean model. The ocean's deep circulation is a nonlinear system with complex behaviour; under realistic surface forcing it displays multiple equilibrium states, transitions between those and oscillations as well (see Weaver and Hughes 1992 for a review). The spin-up is normally performed with a physically unrealistic boundary condition, where surface temperature and salinity are strongly restored to some prescribed values (usually from observations). This ensures a stable equilibrium solution with surface temperature and salinity values close to the observed ones. This particular equilibrium state may be unstable when a different type of surface forcing is applied, for example by coupling an atmosphere model, leading to climate drift in the coupled model. Since there are different equilibria of the coupled system, the one that the model drifts towards may not be desirable; for example it may be one without

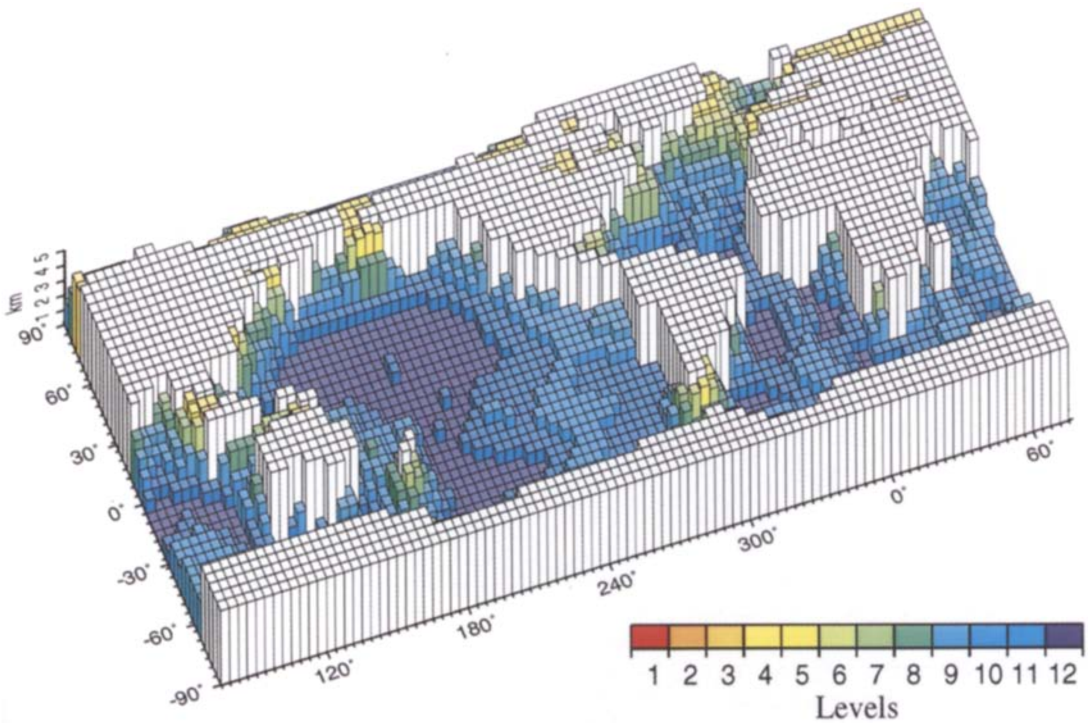


Fig. 1. Topography of the global ocean model. Vertical axis is depth (*one tick* is 1 km); *colour* gives the number of model levels

any North Atlantic deep water formation. And even if the new coupled equilibrium *is* a desirable, realistic climate, we cannot afford the integration time to reach it. Therefore, one usually attempts to suppress climate drift after coupling by introducing a "flux correction" (Sausen et al. 1988). This diagnoses the ocean-atmosphere fluxes computed from the ocean and the atmosphere model spin-up, calculates the difference between those and the fluxes arising under coupled conditions, and subsequently adds this difference as a fixed "correction" to the fluxes during the coupled experiment. In some cases the spin-up of both models is not done completely separately, but a restoring term is added during a coupled spin-up. The variable being restored to observed data is then essentially decoupled (e.g. salinity in Manabe and Stouffer 1988, or both salinity and temperature in Manabe and Stouffer 1994). The same stability considerations that apply for a separate spin-up at the moment of coupling apply also here, at the moment when the restoring term is switched off.

The use of flux "correction" is an ad-hoc measure not based on an understanding of the cause of the drift. Furthermore, in spite of the arguments presented in Sausen et al. (1988), flux corrections do not succeed in preventing all drift (Santer et al. 1994). A greater understanding of the behaviour of the ocean model under coupled conditions is urgently needed. Since full-blown atmospheric GCMs are so expensive to run, this knowledge can only come from coupled experiments with simple, 'cheap' atmosphere models, which approximate the behaviour of a real GCM (and the real atmosphere). Such a study is presented here. The 'atmosphere model' used is extremely crude and simple; nevertheless (or because of this) important insights can be gained. This study presents results with a perfectly matched, or "flux corrected", atmosphere, which at the moment of coupling offers the ocean model exactly the same fluxes as during spin-up, though with different feedback. Future work will examine experiments with different types of "mismatched" atmosphere.

2 An ocean model with simple atmospheric coupling

2.1 Model details and spin-up

In this study I have used the GFDL modular ocean model (Pacanowski et al. 1991, 1993) in a coarse resolution, global configuration with twelve levels in the vertical, very similar to the models of England (1993) and Manabe and Stouffer (1994). The topography of this model is shown in Fig. 1. The model was initially spun up from rest by applying wind stress from Hellerman and Rosenstein (1983) and restoring the surface level to annual mean temperature and salinity data from Levitus (1982) (averaged over the top 50 m and interpolated to the model grid). Very low salinity values from inside the Baltic as well as the fjords of Tierra del Fuego, which affected some grid points, were rejected and replaced by typical ocean values from the region (Rahmstorf 1995a). The restoring time scales were short for this spin-up (30 days for T and 50 days for S). The horizontal and vertical viscosities were $2.5 \cdot 10^5$ m^2/s and $5 \cdot 10^{-3}$ m^2/s respectively. Mixing was parameterized by depth-dependent diffusivities, varying for horizontal mixing from $1.0 \cdot 10^3$ m^2/s at the sur-

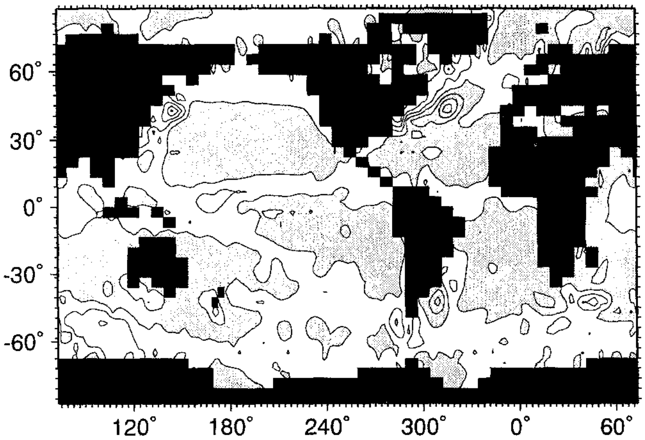


Fig. 2. Freshwater flux (P–E) field used to force the ocean model. Areas of net evaporation are *shaded.* Contour interval 2 m/y

face to $0.5 \cdot 10^3$ m^2/s in the deepest level, and for vertical mixing from $0.3 \cdot 10^{-4}$ m^2/s to $1.3 \cdot 10^{-4}$ m^2/s. All static instability was removed at each time step by a fast and complete convection scheme (Rahmstorf 1993). Split time stepping (Bryan 1984) was used for all the experiments (but one), with a tracer time step of one day and a momentum time step of one hour.

After 4800 years of integration the model was near equilibrium and the surface flux of freshwater was diagnosed. This flux field, shown in Fig. 2, was used as a fixed flux boundary condition in all subsequent experiments, after the small residual trend (less than 10^{-6} psu/y) had been subtracted. The rationale of using such 'mixed' boundary conditions (with fixed freshwater flux) has often been discussed before (e.g. in the reviews of Weaver and Hughes 1992; Marotzke 1994); they provide a physically more realistic response which is essential for time-dependent experiments and multiple steady states, while restoring is useful for producing a particular steady solution of the model. The diagnosed freshwater flux field shows characteristic regional features such as net rainfall along the equator and over the Antarctic Circumpolar Current, and evaporation in the subtropical gyres. Particularly relevant for the stability of North Atlantic Deep Water formation is the band of net precipitation across the northern North Atlantic at about 60°N. In its broad features this field is similar to the one diagnosed from the model of Maier-Reimer et al. (1993).

After the switch to mixed boundary conditions the model made a transition to a new equilibrium state; the vertical overturning rate in the Atlantic increased from 21 Sv (the value under restoring boundary conditions) to 24 Sv. A brief freshwater perturbation in the convection region off Greenland (an additional rainfall of 2 m/y was added for 10 years) caused another transition, this time to an equilibrium state with 18 Sv overturning. These multiple equilibria under traditional mixed boundary conditions are caused by changes in convection patterns (Rahmstorf 1995b), but are not the subject of this study. Rather, we will here use this final state as our starting point for experiments with a different atmospheric coupling, and will henceforth refer to it as the 'spin-up state'. Some values characterizing the barotropic flow field in this spin-up state are as follows: subtropical gyre in the North Pacific 58 Sv, in the South Pacific -49 Sv; in the North Atlantic 39 Sv and in the South Atlantic -40 Sv; subpolar gyres in the North Pacific -9 Sv and in the North Atlantic -40 Sv. The Drake Passage flow is 200 Sv, and the Indonesian throughflow 17 Sv. No depth-integrated flow through Bering Strait is allowed in this model. The zonally averaged meridional flow field has an NADW cell of 18 Sv and an AABW cell of 7 Sv in the Atlantic, and a deep inflow into the Indian and Pacific Oceans (combined) of 11 Sv.

2.2 *Thermal coupling to the atmosphere*

The restoring condition on temperature, as on salinity, is a useful tool for spinning up an ocean state with the desired SST distribution, but does not provide a physically realistic response to large-scale changes in circulation, as it assumes fixed atmospheric temperature (Haney 1971). An alternative approach, which is suitable for climate experiments, is to determine the atmospheric temperature from an atmospheric heat budget. This simple heat budget can be used to determine the response to SST anomalies. A warm anomaly, for example, will lead to increased heat flux from the ocean to the atmosphere, and the atmosphere will warm until it can balance this heat by increased longwave radiation to space and by transporting heat laterally; the heat capacity of the atmosphere is negligible in this context. How effective the lateral dispersion of anomalies is depends on the spatial scale of the anomaly. The atmosphere can remove small-scale anomalies more easily than large-scale anomalies; a global anomaly can only be damped by longwave radiation. Perhaps the most simple way to simulate this fundamental balance is a one-layer atmosphere model with horizontal heat diffusion and a linear parameterization of longwave radiation. Based on this approach, we have recently proposed the thermal boundary condition

$$Q = \gamma(T^* - T_O) - \mu\nabla^2(T^* - T_O) \qquad (1)$$

(Rahmstorf and Willebrand 1995). Q is the heat flux at the ocean surface, T_O the ocean surface temperature, γ a radiative relaxation constant (3 Wm^{-2}K^{-1}), μ a constant related to atmospheric heat diffusion, and T^* defines a climate without oceanic heat transport, around which Eq. 1 linearized. The atmospheric temperature is a diagnostic variable in this model; it is already eliminated from Eq. 1 by inserting the atmospheric heat budget equation. For any SST deviation from the 'background' climate T^*, caused by oceanic heat transport or storage, there is a balance between surface heat flux, longwave radiation to space (first term on the right) and lateral diffusion in the atmosphere (second term). This equation is in fact a truncated version of a full diffusive atmosphere model, using the additional assumption that air-sea coupling is infinitely strong. This truncation overestimates the damping of the smallest scale anomalies (which is limited by the effectiveness of air-sea exchange in the full model), but this poses no serious problem in coarse models, which only resolve large scales anyway. The damping of large-scale anomalies is limited by horizontal diffusion in the atmosphere and longwave radiation to space, and is approximated well by the truncated formula.

Equation 1 is sometimes mistaken for the traditional restoring formula with an added Laplacian term. While this may be formally correct, physically it is not. In traditional restoring (Haney 1971), the restoring term represents mainly air-sea coupling (for constant air temperature), and its sensitivity is of the order 50 $\mathrm{Wm}^{-2}\mathrm{K}^{-1}$. In Eq. 1, the term which has the form of a restoring term is in fact the linearized parameterization of longwave radiation, with a sensitivity of only 3 $\mathrm{Wm}^{-2}\mathrm{K}^{-1}$. The local air-sea coupling limit considered by Haney is assumed to be infinite in our truncated model, or 50 $\mathrm{Wm}^{-2}\mathrm{K}^{-1}$ in the full model, and affects only small-scale temperature anomalies.

Equation 1 provides a simple approximation to the kind of large-scale thermal feedback which the ocean model would feel if coupled to a 'real' atmosphere model. Studies with an idealized two-basin ocean model have already shown that this feedback is important and leads to rather different behaviour of the ocean model compared to classical mixed boundary conditions (Rahmstorf 1994, 1995b; Rahmstorf and Willebrand 1995). The main property of Eq. 1 is that it does not tie down the model SST as much as the traditional restoring approach, but allows SST to respond with a realistic amplitude to changes in oceanic heat transport. The effective coupling for the global mean temperature is 3 $\mathrm{Wm}^{-2}\mathrm{K}^{-1}$, for thermohaline circulation changes it is about 10 $\mathrm{Wm}^{-2}\mathrm{K}^{-1}$ (Rahmstorf and Willebrand 1995), and for grid scale anomalies it increases to 50 $\mathrm{Wm}^{-2}\mathrm{K}^{-1}$. The major restrictions are that atmospheric heat transport is parameterized as a simple diffusion law (a choice which is supported by experiments with a more elaborate energy balance model including advection, Kleeman and Power 1995) and that there is no heat exchange across land boundaries.

The constant μ was chosen to be $2\cdot 10^{13}$ W/K, tapered in zonal direction in the two northernmost grid rows for numerical stability reasons. This number is related to the diffusivity of the atmosphere as discussed by Rahmstorf and Willebrand (1995); our value corresponds to a diffusivity of the atmosphere of $2\cdot 10^{6}$ m^2/s, which is within the range deduced from Stone and Miller's (1980) empirical relationship between meridional temperature gradient and heat transport.

The forcing field $T^*(x, y)$ can be obtained in various ways, for example by running an atmosphere model with a 'swamp ocean' of zero heat capacity as lower boundary; T^* would be the temperature of the 'swamp'. For the present study, however, I constructed a T^* field which perfectly matches the 'spin-up state' of the ocean described, by inverting Eq. 1 using Q and T_O from the last time step of the spin-up. T^* was determined using a Gauss-Seidel iteration, to an accuracy in the heat flux Q of 10^{-6} $\mathrm{W/m}^2$ at each grid point. The resulting T^* field is shown in Fig. 3b; for comparison the Levitus SST field used for the spin-up is shown in Fig. 3a, and the final model SST field (see next section) in Fig. 3c.

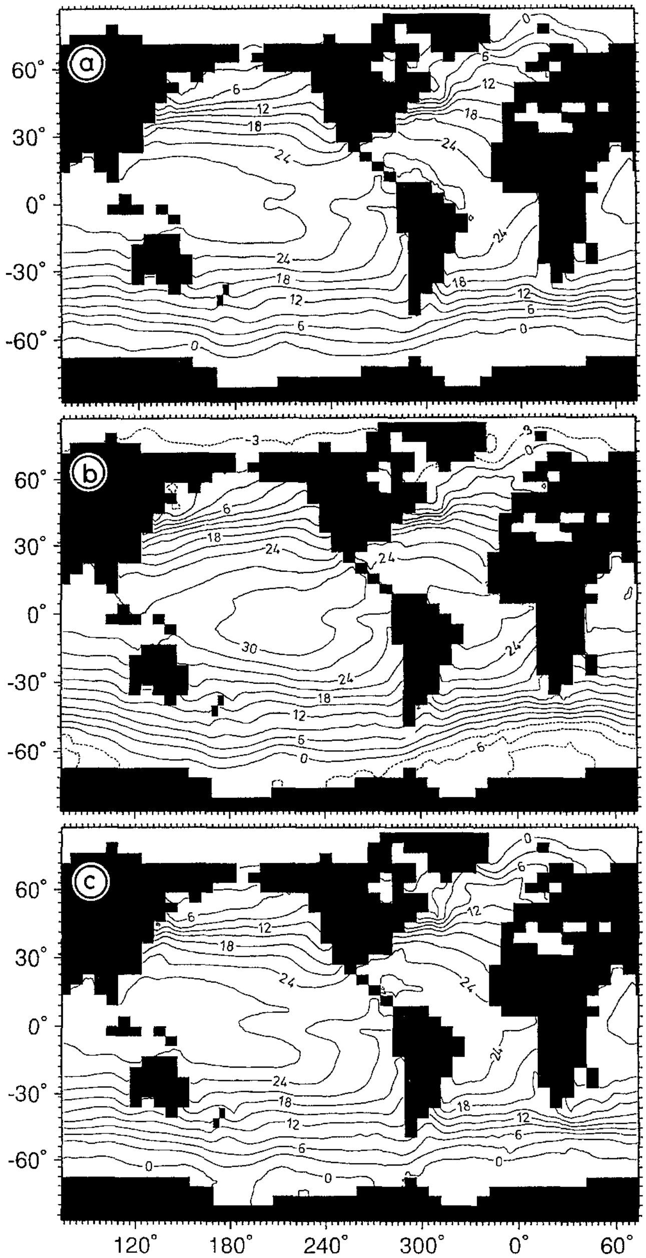


Fig. 3. a Temperature field from Levitus, upper 50 m of the ocean, interpolated to the model grid. This field was used as restoring field during spin-up of the ocean model; **b** field of T^* derived by inverting Eq. 1 with heat flux Q and surface temperature T_O from the spin-up state. This field was used in the experiment with the diffusive atmosphere; **c** final surface temperature after 5000 years of integration with the diffusive atmosphere; compare to **a**

The inverted T^* field is an interesting result in itself. If our model approach was perfect, then T^* should be free of all effects of oceanic heat transport, which are an important part of the observed (Levitus) or mod-

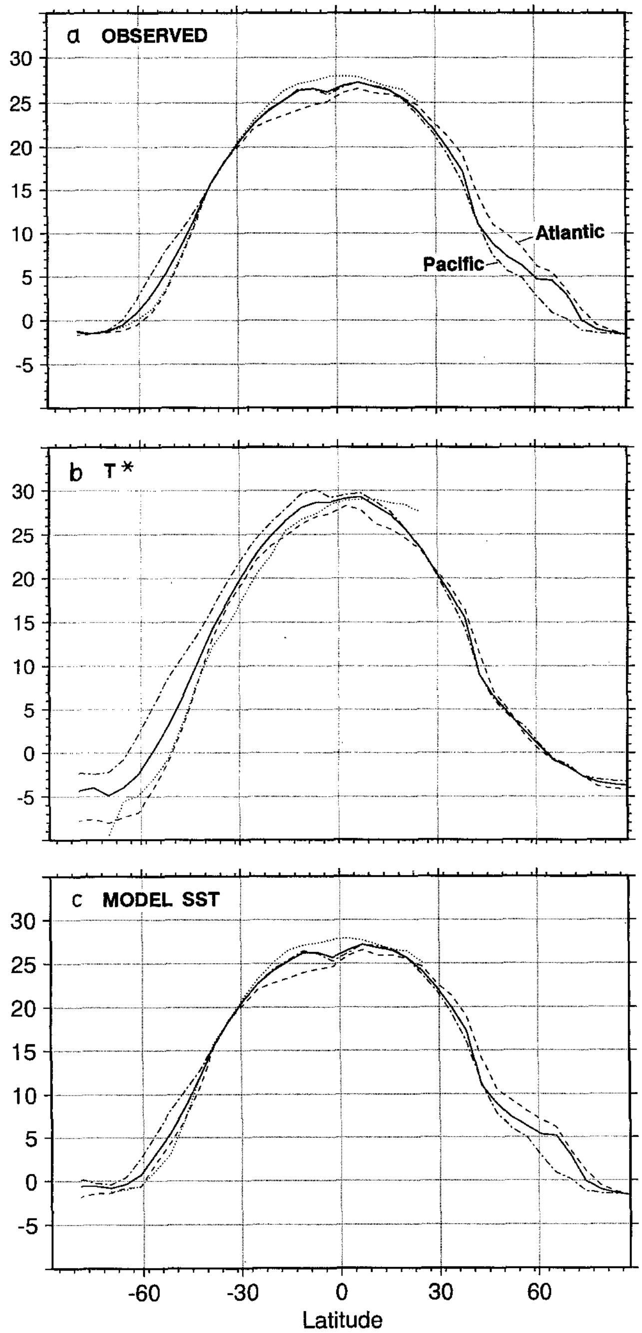


Fig. 4a–c. Zonal averages of the three temperature fields shown in Fig. 3. *Solid line:* zonal average of all oceans; *dashed line,* Atlantic ocean; *dash-dotted* line, Pacific ocean; *dotted line,* Indian Ocean. Note the 4° C temperature contrast between North Atlantic and North Pacific in the observations **a** and the model SST **c**, which is not prescribed in the forcing field T^* **b,** and the larger amplitude of the forcing

elled SST field. For example, the tongue of cold water spreading from Peru westward into the Pacific is probably an advective feature related to the Humboldt current; alternatively it could be created partly by stratus cloud off Peru. If the former explanation is true, the cold tongue should not be visible in the forcing field T^*, but rather be generated entirely by the ocean model. The fact that it is not then points to a weakness of coarse ocean models: swift currents are generally too sluggish. In contrast, the western Pacific warm pool *is* generated by the model by oceanic advection; in the forcing field the highest temperatures are in the *central* Pacific. The very cold forcing temperatures in the South Atlantic (below $-8°$ C) again point to a problem in the ocean model: large and probably unrealistic convective heat loss occurs there in the spin-up state. Of particular interest is the SST difference between North Pacific and North Atlantic, which is best seen in a zonally averaged picture (Fig. 4). If this temperature contrast is caused by the conveyor belt circulation of the ocean, then it should not exist in T^*. This is indeed the case in our model. Figure 4 also shows how the temperature contrast between low- and high-latitudes is larger in T^* than in the Levitus data or the final model SST (Fig. 4c), because meridional heat transport by the oceans ameliorates this contrast. The longitude-dependence of T^* in the Southern Ocean seen in Fig. 4b is a consequence of the distribution of convection mentioned.

3 Climate drift after coupling

3.1 Response of the ocean to coupling

We will now discuss how the ocean model responds to the switch from classical mixed boundary conditions (with restoring on temperature) to the new thermal coupling (Eq. 1). In some respects this corresponds to the coupling of an atmosphere model to an ocean model spun up with mixed boundary conditions. The atmosphere model is perfectly ‘flux corrected’ in this experiment; the freshwater flux remains the same as before, and T^* was constructed to provide exactly the heat flux which the spin-up state wants. It is important to realize that this is fully equivalent to starting from an arbitrary atmospheric model climate (i.e. an arbitrary T^* field) to which a fixed flux correction term is added, just as in ‘real’ coupling experiments and as described in Sausen et al. (1988). Any fixed additive flux correction field $\Delta Q(x, y)$ can be rewritten in the form $\gamma T'(x, y) - \mu\nabla^2 T'(x, y)$, so that the flux correction term can be absorbed in T^*.

All that changes for the ocean at the moment of coupling in this experiment is that it feels a different thermal *feedback.* The value of this experiment is that it isolates processes which occur in the ocean model upon coupling. In a fully coupled model, these are clouded by the variability of the atmosphere, and the reason for changes occurring after coupling cannot be attributed to a specific cause or process. In our case, the ‘atmosphere’ has no internal variability; it just responds to changes in the ocean in a simple and predictable manner. This experiment should reveal *part* of the variability which arises after coupling an ocean model to an atmosphere model; an important part, as we will see.

Figure 5 shows time series of three indicators of deep ocean circulation after coupling at $t=0$. It is im-

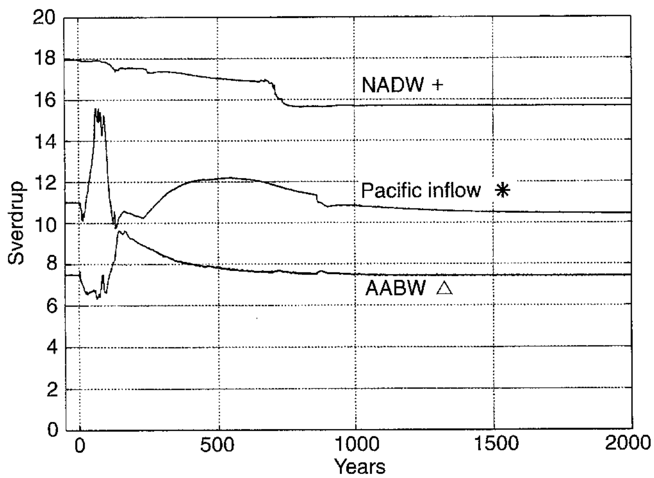


Fig. 5. Time series of deep circulation indicators just before and during the first 2000 years after coupling the diffusive atmosphere. The volume transport of the NADW and AABW cells in the Atlantic, and the deep inflow into the Pacific are shown (see also Fig. 6)

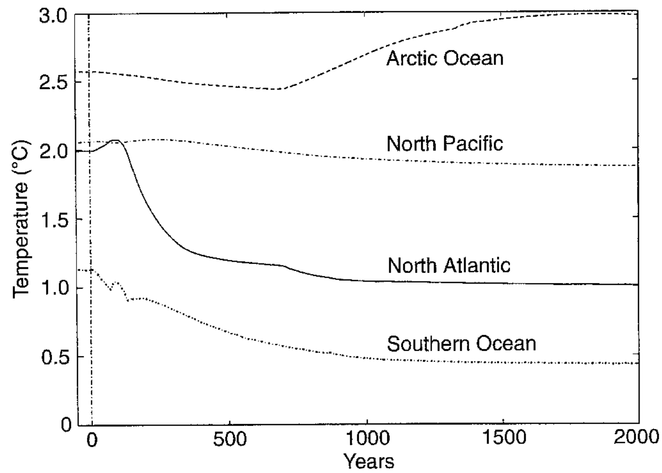


Fig. 7. Time series of basin-averaged temperatures at level 10 (3000 m) during the first 2000 years after coupling

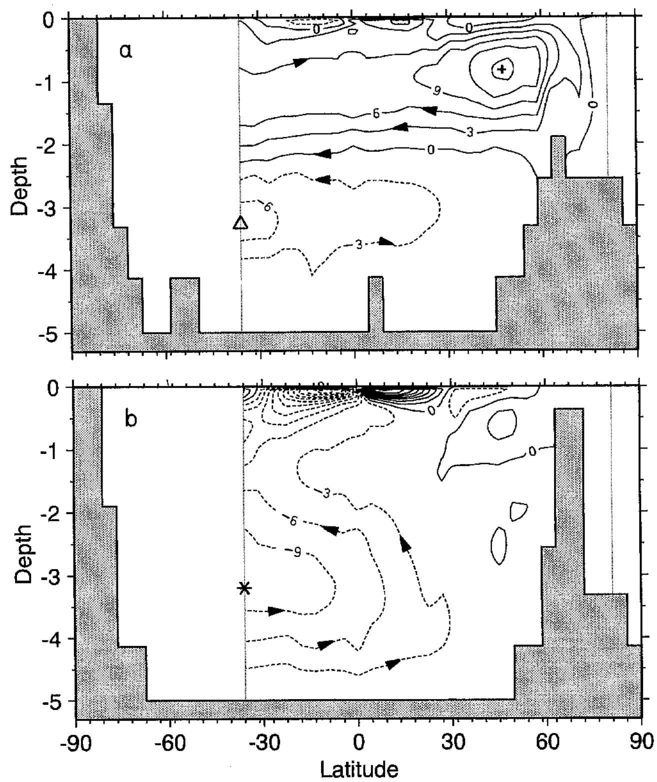


Fig. 6a, b. Meridional stream function of the final model equilibrium 5000 years after coupling. **a** Atlantic Ocean; **b** Pacific and Indian Ocean; maxima of NADW cell (+), AABW cell (△) and Indo-Pacific inflow (*) are marked; time series of these maxima are displayed in Fig. 5

mediately clear that the previous model equilibrium is not stable after the switch to a different thermal feedback at the surface. Major changes in the deep circulation occur; the deep inflow into the Indo-Pacific (see Fig. 6 for a definition) increases from 11 Sv to over 15 Sv within the first 60 years, then drops back after 100 years. Two thousand model years after the 'coupling shock' a new equilibrium is approached (see Fig. 6) with a circulation similar to the spin-up state except that NADW formation has dropped from 18 Sv to 16 Sv. Figure 7 shows, however, that large and permanent changes in deep ocean temperature have resulted from the coupling. The average temperature in the North Atlantic at 3000 m depth has dropped by 1° C.

How important are these changes for the 'climate' of a coupled model? The ocean model affects the atmosphere only through the sea surface temperature, so this is the most important climate variable. Snapshots of SST changes at four different times after coupling are shown in Fig. 8. We have faith that the magnitude of such changes, resulting from changes in oceanic heat transport, is realistic in the simple coupled model, because it produces the correct SST contrast between northern Atlantic and Pacific (see earlier).

The figure shows that regional SST changes of over 2° C occur after the 'coupling shock'. These changes are concentrated in the deep convection regions of the North Atlantic and Southem Ocean. The changes are no gradual drift or oscillation; rather they appear to be "switched on" at different times in different regions. The first major SST change is a 'dipole' in the South Atlantic, which is fully developed 10 years after coupling, and remains in all subsequent snapshots up to the end of the experiment after 5000 years. A sudden warming in the Ross Sea, in contrast, appears only after 250 years.

Figure 9 shows the corresponding temperature changes at 1600 m depth and Fig. 10 at 3700 m depth. At both levels there is widespread cooling of deep water at the end of the run compared to the spin-up (Figs. 9, 10b), but initially, during the first 100 years, some basins get warmer at depth (Figs. 9, 10a). Again it is clear that the changes originate in the convection regions.

Similar SST changes centred in high-latitude convection regions occur also in coupled GCMs after coupling (e.g. Santer et al. 1994). The Hamburg group has produced a 650 year (up to the moment of writing) 'control climate' experiment with a coupled GCM, which is comparable to my experiment except that a full atmospheric circulation model is used. This experiment exhibits variability of similar type, magnitude and time scale in the deep ocean circulation and in deep and surface temperatures after coupling (Jin-Song von Storch, personal communication). The coupled climate change scenario experiment of Cubasch et al. (1995) shows patches of warming and cooling of about 2° C amplitude in the Southern Ocean in the initial decades (from 1935 to 1995), which look very similar to the ones seen in Fig. 8. Such temperature changes are usually attributed to changes in sea ice cover in the Southern Ocean, which could be due to the sea ice model. The fact that similar changes also arise in the simplified model used here, which includes no sea ice, suggests an alternative explanation. Sea ice changes in the coupled experiments could be triggered by (and perhaps amplify) the kind of high-latitude SST changes analyzed in this paper.

For climate predictions with coupled models it is important to see whether the global mean surface temperature is also affected. Our simple atmosphere can produce no such changes in itself; the equilibrium global mean SST is fixed by Eq. 1 and is equal to the global mean of T^*. Nevertheless the ocean can cause temporary deviations of mean SST by releasing stored heat. The amplitude of such an SST change is governed by the global mean of Eq. 1, where the last term vanishes, and thus by the 'climate sensitivity' γ. This is the same for our simple model or a more sophisticated atmospheric model; our choice of γ is within the range given by the IPCC (Houghton et al. 1990; see also Rahmstorf and Willebrand 1995). The global mean SST changes resulting from the changes in deep circulation are shown in Fig. 11. The effect is not large (less than 0.1° C), since the deep temperature changes (Figs. 9, 10) occur slowly over a long time (Fig. 7).

3.2 The mechanism causing the drift

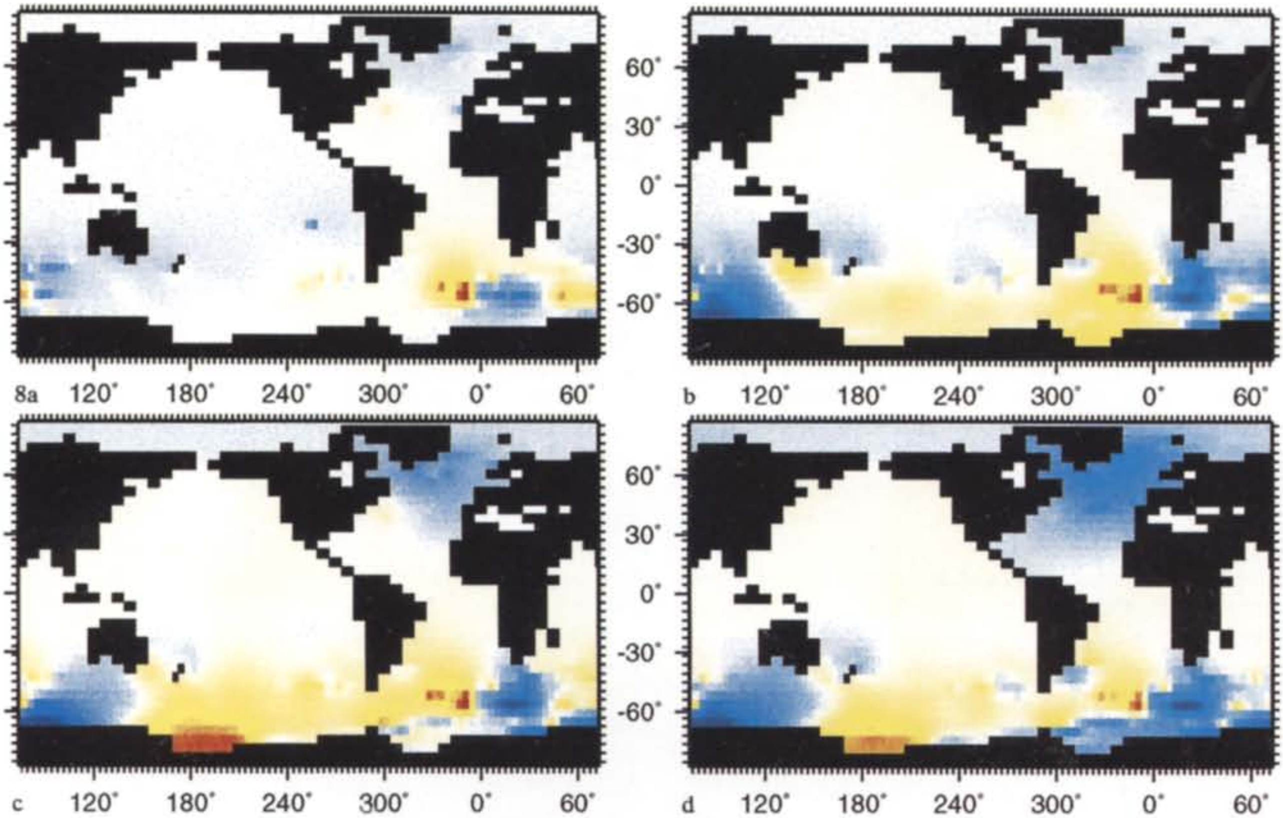
What is the mechanism of the observed transition to a new equilibrium, triggered by the act of coupling? The main cause is probably a change in convection patterns after the switch to the new forcing. Convection patterns are particularly sensitive to disturbance and can give rise to many different equilibrium states of an ocean model, because they are partly self-sustaining (Lenderink and Haarsma 1994). Using an idealized model, Rahmstorf (1995b) has analyzed in detail how changes in convection can be triggered by changes in surface forcing, and how they affect the circulation and deep water properties. Figure 12 shows the convection patterns in the model before and after coupling, i.e. for the spin-up state, 10 years after coupling and for the final state 5000 years after coupling. Large shifts in convection sites have occurred particularly in the Southern Ocean, which would explain the changes in the AABW cell and the deep Indo-Pacific inflow seen in Fig. 5. Time series of convection depth at representative points (not shown here) show that large changes in convection happen immediately after coupling, within the first year or two. This and the experiments presented in Rahmstorf (1995a, b) strongly point at the instability of convection as a *cause* for the observed state transition of the ocean model, not just as a side-effect of circulation changes caused by other feedbacks. Note that convection settles into a stable pattern again after the transition (except for minor oscillations at a few grid cells, which are also present in the spin-up state).

This interpretation is consistent with the sudden occurrence of regional SST changes seen in Fig. 8. The dipole arising in the South Atlantic is caused by a westward shift of convective heat release and represents the largest initial change in convection pattern. As the circulation slowly adjusts to the new situation, other convection sites become unstable, trigger a chain reaction as described by Lenderink and Haarsma (1994), and go through rapid transitions.

Further support suggesting this mechanism as the cause of the drift comes from a recent study of Power (1995), which analyzed climate drift in an OGCM not after a change in surface feedback, but after turning on noise in the surface forcing. A series of experiments revealed that the noise was only effective in causing climate drift if (a) it was added in high-latitudes, (b) mixed boundary conditions were used (and not restoring or flux boundary conditions) and (c) the noise was in the salinity forcing. These findings are consistent with a convective instability, triggered by freshwater anomalies, as the cause of the drift. Furthermore, Power (1995) showed that in a one-dimensional water column (like a single grid cell of the GCM) climate drift can be caused by noise in the forcing only if convective adjustment is present.

It is not surprising that the convection pattern of the spin-up state is not stable after switching to a new feedback. Some disturbances which are damped by the restoring boundary condition can grow under the new forcing. Convection becomes part of a new feedback: vigorous convection at one point tends to suppress convection in its vicinity through the diffusive term of Eq. 1. This is because convecting points mix up heat from below and are therefore warm at the surface; they heat neighbouring points at the surface, inhibiting convection there (see Rahmstorf and Willebrand 1995). Indeed we see that the total number of ocean points with deep convection (to below 2000 m) from the surface is reduced almost by half (it is 10 out of 2524 ocean points at the end of the experiment, compared to 18 in the spin-up state). As the deep ocean is now ventilated at fewer points, heat flux at these points is correspondingly larger, as seen in Fig. 12. This feedback alone is a reason to expect that the convection patterns of the spin-up state must change under the new forcing, or after coupling to a full atmosphere model.

In a paper introducing the method of flux correction, Sausen et al. (1988) have tested this method by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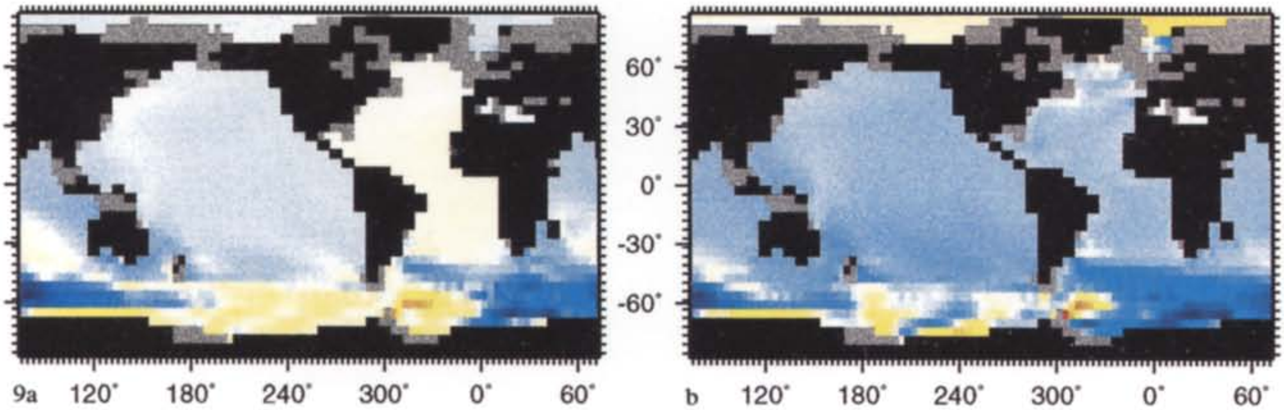
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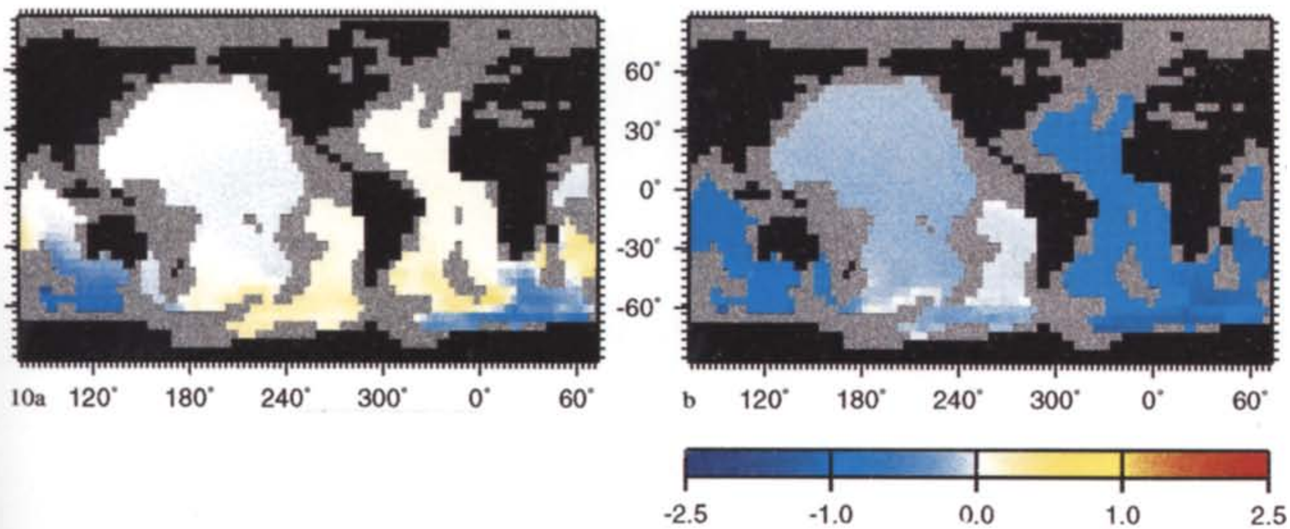
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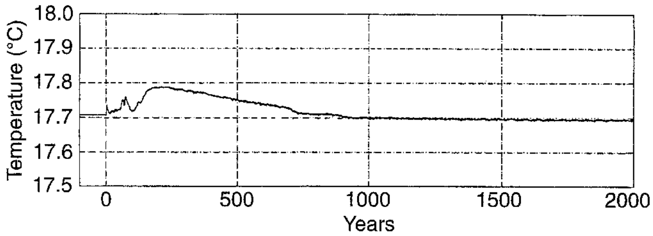


Fig. 11. Time series of global surface temperature change during the first 2000 years after coupling

performing a very similar experiment, namely coupling a spun-up ocean model to an energy balance atmosphere and observing the resulting drift. With flux correction, no drift was observed in the coupled model, in stark contrast to the results presented here. The crucial difference between these two experiments is in the salinity forcing. While Sausen et al. (1988) used a restoring boundary condition on salinity throughout the experiment, I have used a fixed freshwater flux condition. The positive feedback causing the convective instability, and thus the proposed mechanism for the drift, depends on this more realistic freshwater forcing (Lenderink and Haarsma 1994). With a salinity restoring boundary condition a strong negative feedback is introduced, which artificially stabilizes the ocean model.

It is often attempted to eliminate drift from the coupling shock effect by subtracting a control run from the climate scenario run, and then attributing the difference to the forcing scenario. This approach can only work if the response to coupling is deterministic, and similar in control and scenario run. This is most unlikely to be the case, as the transitions in convection patterns appear to be chaotic. Figure 13a shows a number of coupling experiments which differ only by a slight change in initial state (produced by integrating the spin-up state for another 10 years before each coupling, exploiting minor residual variability in the spin-up state). Each time the model takes a different trajectory. In two of the runs a step-like reduction in overturning (to about 15.5 Sv) occurs, in one case immediately after coupling, in the other case after 700 years. The short time scale of the overturning change again points to a convective instability being the cause (Rahmstorf 1995a), possibly affecting the same convection region in both cases.

Figure 13b shows the deep inflow into the Indo-Pacific from another set of runs starting from slightly different initial conditions. In one of these, it was not the initial condition that was varied, but the tracer time step: it was reduced from one day to one hour (making it synchronous with the momentum time step, i.e. dropping the acceleration technique). While trials show that this does not affect the transient behaviour of 'well-behaved' experiments, in this case it leads to a different model trajectory, highlighting again the chaotic nature of the reshuffling of convection. Note that the trajectory of this synchronous experiment is not different in character to the others; rather, it seems to be a different realization within the same ensemble. Except for one very stable experiment (dashed line), the curves all seem to reflect a similar type of convective transition, which differs between experiments in detail and timing.

Fig. 8a–d. Surface temperature change in °C compared to the spinup state at four different times, **a** 10 years after coupling; **b** 100 years after coupling; **c** 250 years after coupling; **d** 5000 years after coupling

Fig. 9a, b. Deep temperature changes in °C compared to the spin-up state at level 8 (1600 m) at two different times, **a** 100 years after coupling; **b** 5000 years after coupling

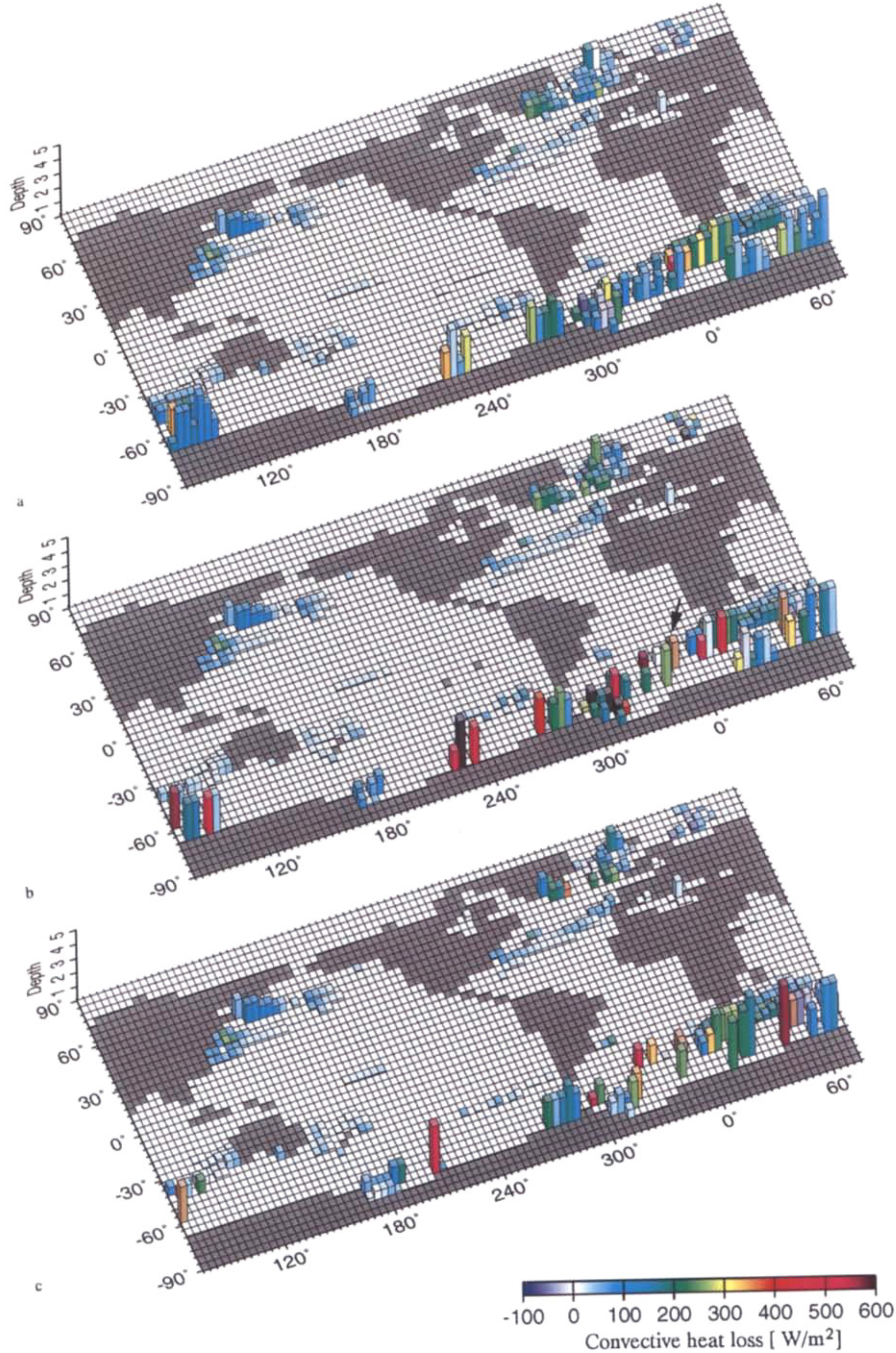
Fig. 10a, b. Deep temperature changes in °C compared to the spinup state at level 11 (3700 m) at two different times, **a** 100 years after coupling; **b** 5000 years after coupling

4 Discussion and conclusions

I have presented results of an experiment where an ocean model was coupled to an idealized simple atmosphere, which only has two properties: heat received from the ocean is radiated to space according to a parameterization of longwave radiation and heat is transported laterally according to a simple diffusion law. The atmosphere was constructed in such a way that it perfectly reproduces the heat flux of the final spin-up state of the model (obtained under mixed boundary conditions). Hence at the moment of coupling the surface fluxes do not change; only the thermal feedback changes. This is equivalent to coupling an atmosphere model with flux correction, and the experiment should produce a subset of the variability observed after coupling a full atmosphere model in this way. The advantage of this type of experiment is that it is cheap (a 2000 year experiment can be run overnight on one processor of a Cray C90, with minimal memory demands) and that it is simple enough to analyze the *cause* of the resulting variability.

The results show that the switch to the new feedback causes a transition of the model to a new equilibrium state. This is apparently caused by shifts in oceanic convection patterns, since the former convection patterns become unstable under the new feedback. The transition is accompanied by large global scale changes in deep water temperatures and regional changes in surface temperatures in the high-latitude convection areas. This state transition takes about 2000 years to complete.

A number of studies have recently demonstrated that the convective instability described by Lenderink and Haarsma (1994) can act as a trigger mechanism for different types of state transitions in ocean models, namely transitions following a high-latitude freshwater perturbation (Rahmstorf 1994, 1995b), transitions occurring after a change from restoring to mixed boundary conditions (Rahmstorf 1995a), or the transition after a change in thermal feedback discussed in this study. These studies have used different forms of steady forcing with no noise, leading to almost steady convection


Depth
1 2 3 4 5
90°
60°
30°
0°
-30°
-60°
-90°
120°
180°
240°
300°
0°
60°
a
b
c
-100 0 100 200 300 400 500 600
Convective heat loss [W/m²]

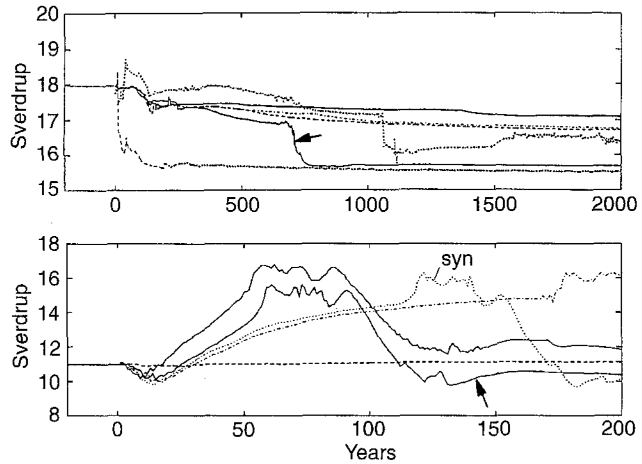


Fig. 13a, b. Time series of flow parameters for coupling experiments starting from slightly different initial conditions, **a** NADW cell during the first 2000 years after coupling; **b** Indo-Pacific inflow during the first 200 years after coupling. *Arrows* identify the standard experiment shown also in Fig. 5. The *dotted line* in **b** is a run with synchronous time stepping

patterns in equilibrium. It is thus natural to ask which effect atmospheric variability has on the oceanic convection patterns, and thus how relevant these results are for the fully coupled ocean-atmosphere system.

Stochastic variability resulting from atmospheric forcing may either be small compared to the basin of attraction of a particular equilibrium state in phase space, so that the model remains always near the same equilibrium with the same basic convection pattern. Or it may be so strong that convection patterns change all the time, and the underlying equilibria are hardly felt. A recent study by Weaver and Hughes (1994) with an ocean model and strong stochastic freshwater forcing (standard deviation = 32 mm/month), while not explicitly analyzing convection patterns, is apparently in an intermediate range where the model stays near a particular equilibrium for up to a thousand years before making a transition to a different state. If an atmosphere model has a weaker stochastic component than this experiment, convection patterns may be quite stable and the results of the steady experiments apply directly to the coupled model. If the stochastic component is much stronger, frequent transitions may mask the underlying multiple equilibria, but coupling an ocean model to such a noisy atmosphere, after spinning it up with steady forcing, would still mean a major transition in the ocean state (cf. Power 1995). If the ocean model was already spun up with noise, it would be interesting to check whether the new feedback acting after coupling leads to a similar reduction in deep convection points as found in this paper (see section 3.2). We can also expect that some convection states are close neighbours in phase space, e.g. the three different convection clusters southeast of Greenland in Fig. 12, while others are well separates, like states with or without Labrador Sea convection (the former did not arise in the experiments reported here, but in others conducted by the author with the same model).

The state transition caused by convective instability and analyzed in this paper has important implications for coupled climate models. From the time series shown in Figs. 5 and 7 it is clear that it is not a kind of 'natural variability' in the coupled system but an artefact; it is the response to the 'coupling shock', which takes several thousand years to die down. This coupling shock will probably contaminate the results of coupled climate change scenario experiments. Luckily the effect on global mean temperature is expected to be small; less than 0.1° C in this experiment. Regional effects can be large, however; over 2° C in this experiment, possibly more in fully coupled models if positive feedback effects with ice cover occur.

The state transition which the ocean model undergoes after coupling casts doubt on the very concept of flux correction. Flux correction is based on the idea that the atmosphere model and ocean model are mismatched, both being imperfect, and that this mismatch is 'corrected' by the flux correction. This implies that a flux correction can be derived for the *model*; in reality, the flux correction is only valid for the *spin-up state* of the ocean model. The 'coupled' equilibrium state of the ocean model is necessarily rather different, due to the different feedback, and would call for a different flux 'correction' (if any). For example, the very cold patches south of South Africa in the T^* field (Fig. 3b) are equivalent to a 'flux correction' which caters for the deep convection occurring there in the spin-up state, by allowing large heat losses. In spite of this, convection shifts away from these sites after coupling, and what used to be a flux 'correction' now becomes an artificially introduced error, leading to an unrealistic temperature drop in this region (Fig. 8). The basic premise of flux correction, namely that it maintains a realistic mean model climate while not affecting the perturbation response, may not be valid in the presence of the non-linear convective instability discussed in this paper.

I conclude that there are probably severe and until now unrecognized problems with the response of present ocean models after coupling to an atmosphere model, unless the state transition described here is somehow suppressed when a more active, variable atmosphere is used instead of the simple diffusive one tried here. These problems can affect at least the regional surface temperature response of coupled climate models, as well as the response of the global thermohaline circulation. Global mean temperature predictions are probably less affected.

The direct way to avoid these problems would be to spin up the coupled model together in fully coupled

Fig. 12a–c. Convection patterns of **a** the spinup state, **b** 10 years after coupling and **c** 5000 years after coupling. Depth of convection (in km) from the surface is plotted as a vertical bar. The bars are *colour coded* according to the convective heat flux across model level 1. A few bars shown in *brown colour* in **b** and **c** are far beyond the given scale; the point marked with an *arrow* has a convective heat loss of 1400 Wm^{-2}

mode for several thousand years, until the coupling shock has died down, before starting an experiment. This is costly, but may be feasible if the atmosphere model is not run in parallel during the whole time, but only updated at intervals, given the large difference in response times of atmosphere and ocean. Coupled spin-up may lead to different climatic states depending on initial conditions; to achieve a desirable state the ocean model can initially be 'nudged' towards a realistic circulation. When this nudging is suddenly removed the circulation might collapse; if it is replaced by a fixed flux correction a complete collapse may be avoided, but the 'coupling shock' problem discussed here still arises. Experience with many model experiments suggests that convection is easily interrupted and instabilities can be triggered by any sudden change in surface forcing; therefore a slow and gradual removal of the initial nudging may be a way to reach a realistic coupled climate state without flux correction, provided that the basic mismatch between ocean and atmosphere model is not too large.

Alternatively, ways to minimize the coupling shock may be found. One possibility may be to spin up the ocean model with a simple atmosphere like the one used here, or a cheap 'derivative' of the atmosphere model to be coupled. Given the chaotic nature of the transitions in convection, it is however uncertain whether this would bring significant improvement. Again, a gentle transition to the fully coupled mode may help to avoid disrupting the oceanic convection patterns.

On the other hand, a large part of the drift originates from the widespread convection regions in the Southern Ocean, which are clearly an unrealistic feature of the ocean model. In reality, convection appears to be confined to small regions near the Antarctic coast, particularly in the Ross and Wedell Seas. An improved representation of convection sites in ocean models, such as the one recently presented by Danabasoglu et al. (1994), may also lead to more stable convection patterns which cause less trouble in coupled model simulations. This is an open question for future research.

Acknowledgements. I wish to thank Matthew England for sharing his version of the GFDL global ocean model with me, and Suki Manabe for helpful comments on the first draft of this paper. I appreciate the support of Jürgen Willebrand for my research. Thanks to Dulcie Smart for editing the manuscript. This work was funded by the Deutsche Forschungsgemeinschaft and by the German Ministry for Research and Technology (BMFT).

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