A Global Coupled Atmosphere--Ocean Model [and Discussion]

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A global coupled atmosphere–ocean model

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A low-resolution version of the European Centre for Medium Range Weather Forecasts global atmosphere model has been coupled to a global ocean model developed at the Max Planck Institut in Hamburg. The atmosphere model is driven by the sea surface temperature and the ice thickness calculated by the ocean model, which, in turn, is driven by the wind stress, the heat flux and the fresh-water flux diagnosed by the atmosphere model. Even though each model reaches stationarity when integrated on its own, the coupling of both creates problems, because the fields calculated by each model are not consistent with those the other model has to have to stay stationary, as some of the fluxes are not balanced. In the coupled experiment the combined ocean–atmosphere system drifts towards a colder state. To counteract this problem a flux correction has been applied, which balances the mean biases of each model. This method makes the climate drift of the coupled model smaller, but additional work has to be done to perfect this method.

1. Introduction

The confidence placed in estimates and simulations of the greenhouse effect by carbon dioxide or other trace gases is limited, because most of the models suffer from severe simplifications. Mitchell *et al.* (1987) obtained a rise in the atmospheric temperature, which they claim to be comparable to the greenhouse effect, but only after raising the sea surface temperature, their lower boundary condition, by 2 K. Other global models use a mixed-layer ocean and therefore neglect the redistribution of heat by currents (Washington & Meehl 1983, 1984; Hansen *et al.* 1988) or restrict the ocean and atmosphere to a fraction of the globe (Spelman & Manabe 1984; Bryan *et al.* 1988), thereby neglecting the interoceanic heat exchange, which might be crucial for climate catastrophes like the ice ages (Broecker *et al.* 1985). The only published global coupled atmosphere–ocean model (Schlesinger & Jiang 1988) has a very coarse vertical resolution (two levels). Its ability to represent the atmospheric circulation is therefore restricted.

In this paper, the development of a comprehensive global coupled atmosphere–ocean model from 'state of the art' atmosphere and ocean models will be described.

2. The coupled model

2.1. *The atmosphere model*

The atmosphere model has been developed as a medium-range forecasting model at the European Centre for Medium Range Weather Forecasts (ECMWF) (Louis 1986). For climate studies its resolution has been decreased to a spectral resolution of T21 and 16 vertical levels. Its physical parameterization includes turbulent vertical diffusion, shallow convection, large-scale condensation and cumulus convection, an interactive radiation scheme and a full hydrological cycle. In addition to the operational ECMWF model, the temperature over sea ice is calculated via a balance equation, and the deep-soil temperature and deep-soil wetness are prescribed by their climatological value every fourth day.

2.2. *The ocean model*

The ocean model has been developed by Maier-Reimer *et al.* (1982). Its fully implicit formulation allows a time step of over 30 days and makes it very suitable for climate experiments. In the coupled version it runs with a horizontal resolution of approximately 4° × 4° and 10 levels in the vertical. It includes a thermodynamic ice model and a realistic bottom topography.

2.3. *The coupling*

The ocean model provides the atmosphere model with the sea surface temperature (SST) and the ice thickness; the atmosphere model in turn calculates the total heat flux (i.e. latent, sensible, short- and long-wave radiative fluxes), the wind stress and the fresh-water flux for the ocean model (figure 1). Because in the ECMWF model the continental runoff is estimated as a global mean only, it has not yet been incorporated in the calculation of the fresh-water flux. As each of the models, even when driven by realistic boundary forcing, tends to generate its own climate, which is different from observation, a coupling creates a nonlinear feedback, which forces some coupled models to drift far from the observed state (Han *et al.* 1984). Sausen *et al.* (1988) propose a flux correction to compensate for the individual model errors, which

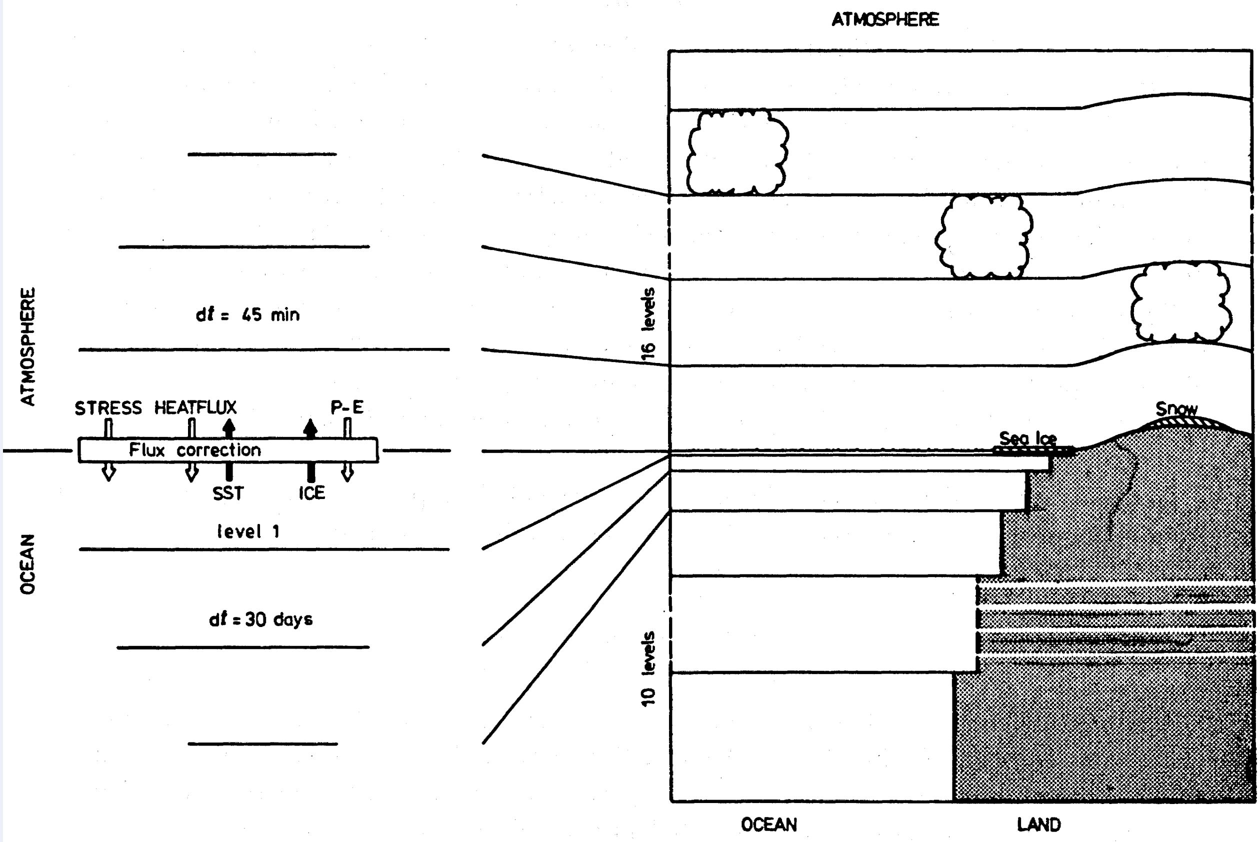

FIGURE 1. The coupling.

should stabilize the coupled model. This method will be tested in some of the experiments. The flux correction for temperature creates subfreezing temperatures in some regions without ice cover. In these regions a basic ice cover has been assumed.

2.4. *The initial data*

The initial data for the atmospheric part of the coupled model have been taken after integrating the atmosphere model alone (i.e. with climatological sea surface temperature after Alexander & Mobley 1976) for one year to let it find its own climate. From that date onwards it has been continued for another five years as a control experiment. This uncoupled run has been used to establish the flux correction fields. The coupled runs start at the same date as the control experiment at a January condition.

The initial data for the ocean have been derived from an integration in excess of 10000 years with the ocean model but at a slightly different resolution (Maier-Reimer *et al.* 1982). This data set has been interpolated to the resolution of the coupled model. From this data set a 1500 year integration has been done until all imbalances from the interpolation have faded away and a stationary state has been obtained. The ocean model in its uncoupled mode has been driven by the wind stress and a 'newtonian coupling' to salinity and temperature.

The wind stress has been taken from a publication by Hellermann & Rosenstein (1983). Because the observed wind stress is stronger than the one simulated by the atmosphere model (figure 2), particularly in the Southern Hemisphere and in the tropics, the coupling of this quantity creates some problems: a simulated ocean circulation initially forced by the strong winds of the observation reacts with a backlash when suddenly driven by the winds generated by the atmosphere model. There exist two possibilities to compensate for this effect: (*a*) again to introduce a bias correction, which will be added to the atmosphere model wind stress field to make it resemble the observed wind field in strength; or (*b*) to drive the ocean model with the atmosphere wind stress field derived from the control experiment until it reaches stationarity and to couple the models then directly. A bias correction in the Southern Hemisphere and in the tropics has the same magnitude as the model simulated forcing field.

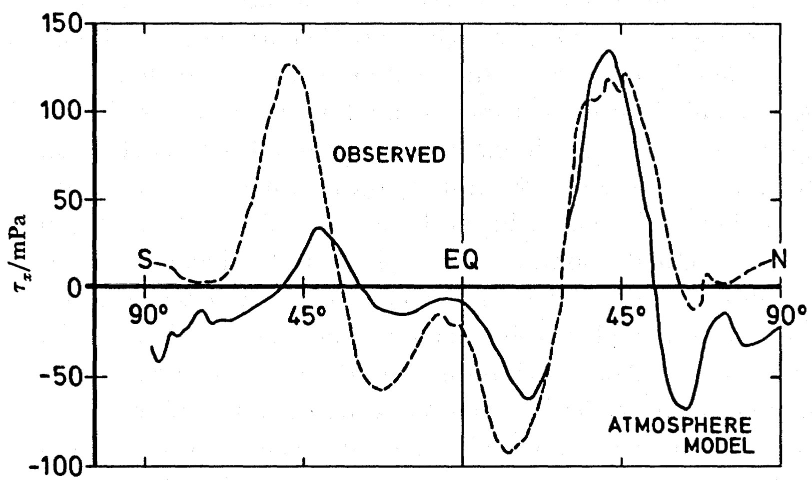

FIGURE 2. The zonal mean of the zonal wind stress for January as function of latitude.

To test the suitability of the model-generated wind stress to drive the ocean model circulation, a run of 1000 years with the ocean model driven by this wind stress has been performed. The oceanic surface currents simulated by the ocean model forced by the modelled wind stress are generally weaker than those forced by the observed wind stress. The circumpolar current around Antarctica is halved, but still reasonable, indicating that it is not only forced by the wind stress, but also maintained by the thermohaline structure of the ocean. The equatorial upwelling in the central Pacific has also become weaker, but also more detailed structures emerge in the Indonesian region in the ocean model forced by the modelled wind stress.

To generate a realistic temperature structure in the ocean the model has been driven by a 'newtonian coupling' to the 'equivalent' temperature and a variable coupling coefficient as defined by Oberhuber (1988). Figure 3 displays the temperature field as observed and as simulated by the ocean model driven by observed and driven by modelled wind stress. Both simulations show considerable differences from observation in the Gulf Stream and the Kuroshio region, where the gradient of the simulated sea surface temperatures is not strong enough. This is possibly connected with the too-coarse resolution of the ocean model. Large temperature differences also appear on the ice limits: the observed temperature contains also temperatures over ice, which can be quite low, whereas the ocean model never simulates temperatures lower than −1.9 °C, a threshold at which it forms ice. In the tropical Pacific the temperature is still about 2 °C colder than observed for the experiment driven by observed wind stress. This error decreases in the run driven by the simulated wind stress, because in this case the equatorial upwelling of cold water is smaller. The oceanic fields of the model driven by modelled wind stress are reasonable enough to use then as initial condition for most of the coupled experiment, thereby eliminating the need for a wind stress flux correction.

It is possible to estimate from the newtonian coupling the heat flux that would have been necessary to maintain the simulated circulation. The atmosphere model simulates a heat flux comparable to observation (Esbensen & Kushnir 1981), whereas the amplitude of the ocean model's heat flux is about 30% larger in the Kuroshio and Gulf Stream regions and about twice as large in the oceans of the Southern Hemisphere. In this hemisphere the flux correction is as large as the quantity it is correcting. In the monthly global mean of the heat flux over sea (figure 4) one finds a clear annual cycle in the model simulations and in the observation. None of the heat fluxes is totally balanced within a year (not even those observed). The heat flux simulated by the atmosphere model has a constant negative bias of about 7.3 W m^{-2}, which means it extracts heat from the ocean, and in an uncorrected run will cool the ocean, if no compensating feedbacks occur.

The fresh-water flux can be derived from the newtonian coupling to the salinity as observed by Levitus (1982). Figure 5 shows the fresh-water flux inversely modelled by the ocean model and the one obtained by the atmosphere model control experiment. The global water budget is balanced in the ocean model with a loss of 24 mm a^{-1}, in the atmosphere model with a loss of 46 mm a^{-1}, which in the latter accounts at least partly for the continental runoff, which has not been included into this calculation. The largest differences between the ocean model and the atmosphere model fresh-water flux appear in the Gulf Stream region, where the pattern of maximum evaporation is closely linked to Gulf Stream, which, as mentioned before is displaced in the ocean simulation, in the tropics, where the cumulus convection is underestimated in the atmosphere simulation, and along the Antarctic coast, where the melting and freezing of ice is not treated properly in the ocean model. It is interesting to see a lot of comparatively small-

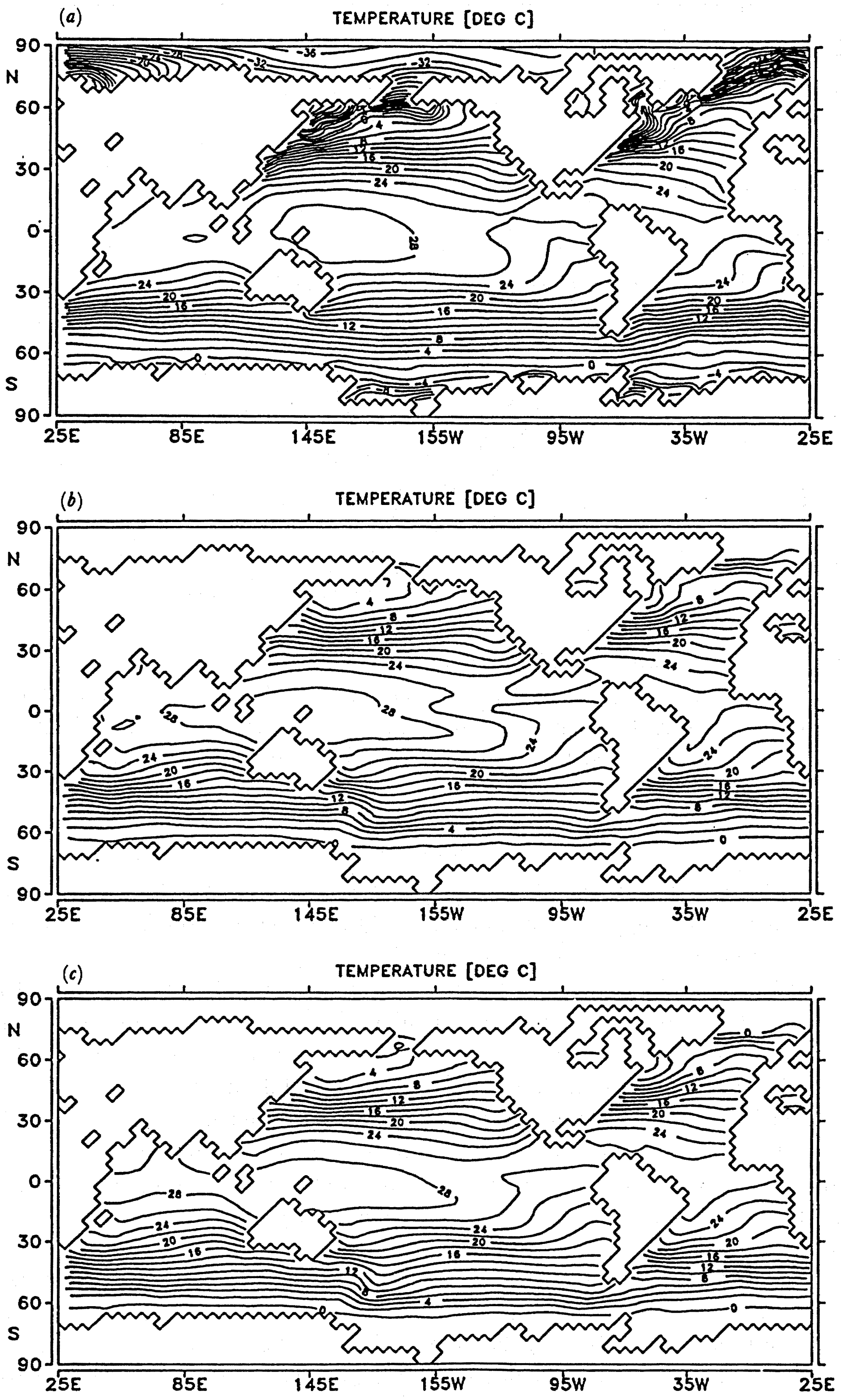

Figure 3. The sea surface temperature for January. (*a*) Observed (after Alexander & Mobley 1976); (*b*) ocean model driven by observed wind stress; (*c*) ocean model driven by simulated wind stress. Contour interval 2 °C.

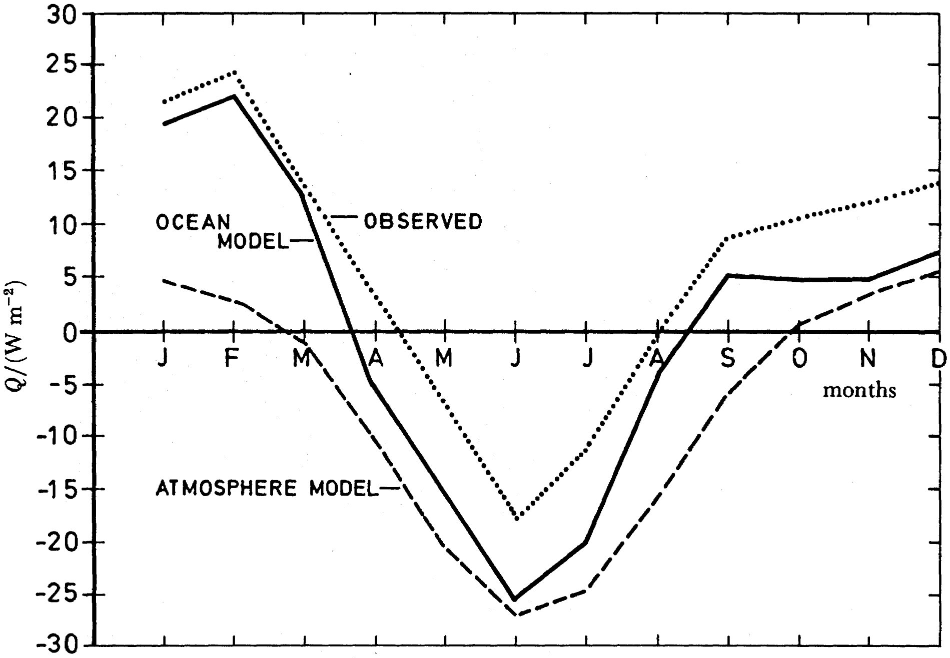

FIGURE 4. The monthly global mean heat flux balance over sea.

scale structures in the ocean simulation, making it difficult to apply a flux correction directly, because it would fix these patterns locally. A spatial filter has therefore been applied to the flux correction for this quantity. Compared with observation, both model simulations underestimate the amplitude of the fresh-water flux, particularly in the Indonesian region.

3. THE EXPERIMENTS

Up to now the only experiments that have been done were tests of the coupled model for computing errors and problems in the application of the flux correction. Three runs of 4 years will be described here.

(1) Coupled without flux correction, ocean start data generated by ocean model driven by modelled winds. This experiment gives a lower threshold, against which improvements on the models and correction methods have to be tested. The time step for the top level of the ocean model was 1.5 h, and for the remainder of the ocean model 30 days.

(2) Coupled with flux correction, ocean start data generated by ocean model driven by modelled winds, time step for the ocean model and the flux correction 30 days, only the heat flux and the sea surface temperature have been corrected. This run represents the most basic flux correction and was thought of as a test, to see whether the concept of flux correction is applicable to a comprehensive coupled ocean atmosphere model at all.

(3) Coupled with flux correction, ocean start data generated by ocean model driven by observed winds, flux correction applied to heat flux, wind stress, fresh-water flux and sea surface temperature. The top level of the ocean model has been calculated at a time step of 1.5 h, i.e. every other time step of the atmosphere model. This experiment was carried out to avoid the phase lag that the long 30 day time step entails. It uses the flux correction in the most comprehensive way.

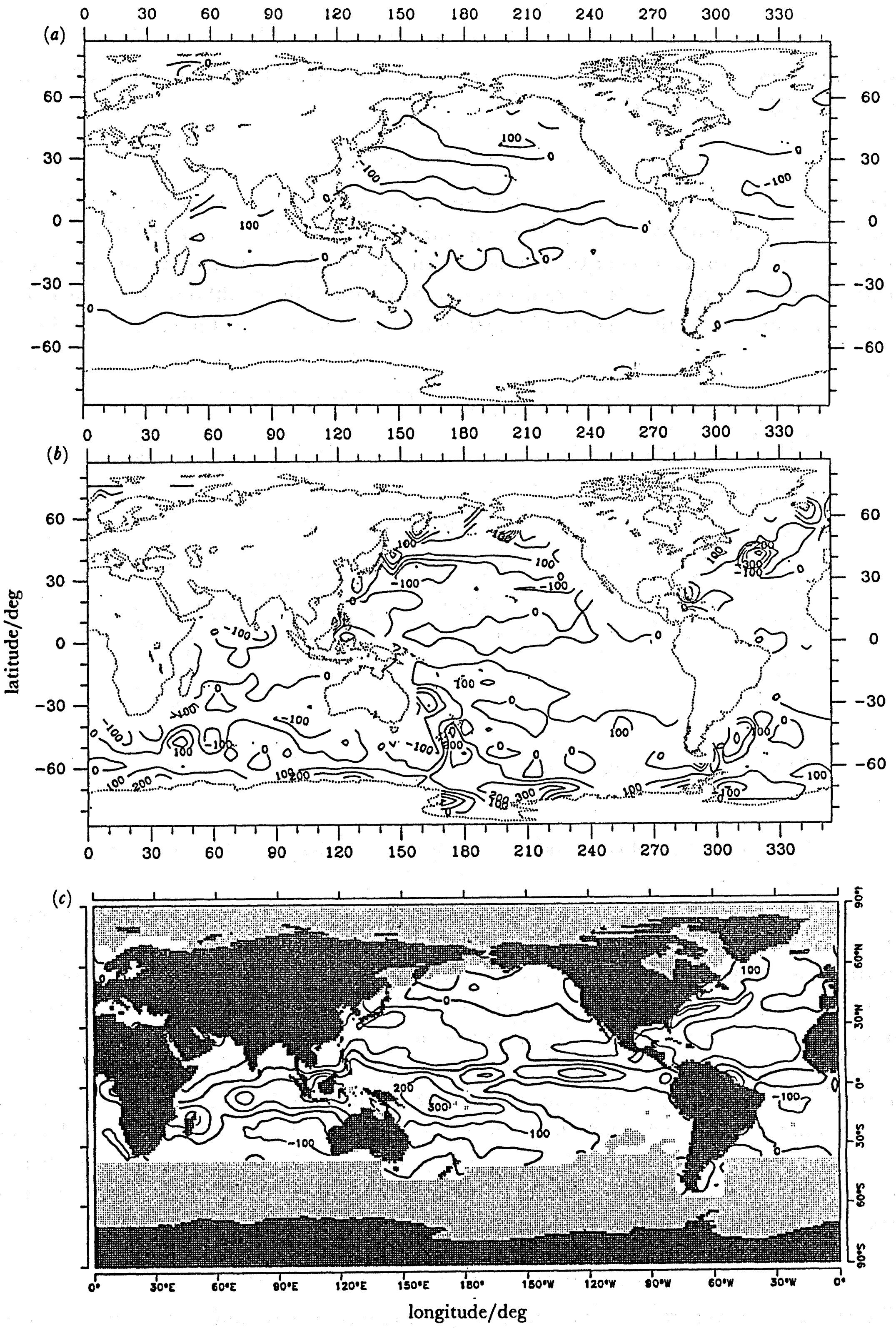

FIGURE 5. The fresh-water flux for January. (*a*) Simulated by the atmosphere model; (*b*) simulated by the ocean model; (*c*) observed after Oberhuber (1988). Units are millimetres per month.

4. Results

All experiments have been run for 4 years so far. In all of them a number of problems emerged, as will be described below.

4.1. *The atmospheric fields*

The global mean temperature of the atmosphere (figure 6) shows a distinct annual cycle in the control experiment caused by the asymmetry of the land–sea distribution in both hemispheres. This annual cycle can also be seen in the coupled experiments, but additionally a trend appears in experiments (1) and (3). Only the run with the 30 day ocean model time step and basic flux correction does not indicate a noticable trend. A primary cause of the trend in the uncorrected run seems to be a strong equatorial upwelling, which pumps comparatively cold water along the equatorial Pacific. It is generated by the Walker circulation, which has been triggered off by the too-cold SST in the initial conditions created by the uncoupled ocean model. This walker circulation becomes stronger as long as the temperature contrast along the Equator increases, thereby generating a positive feedback loop. This effect was so strong that experiments (1) and (3) have been abandoned, because the SST in the tropical Pacific eventually became unrealistically cold. In the experiment with the 30 day ocean time step and flux correction such a feedback loop does not emerge, as the SST deficiency has been compensated. In the run with the shorter time step in the top layers the inconsistency between flux correction time step and ocean model time step appears to be so large that this experiment enters into the Walker cell feedback loop even stronger as the uncorrected run.

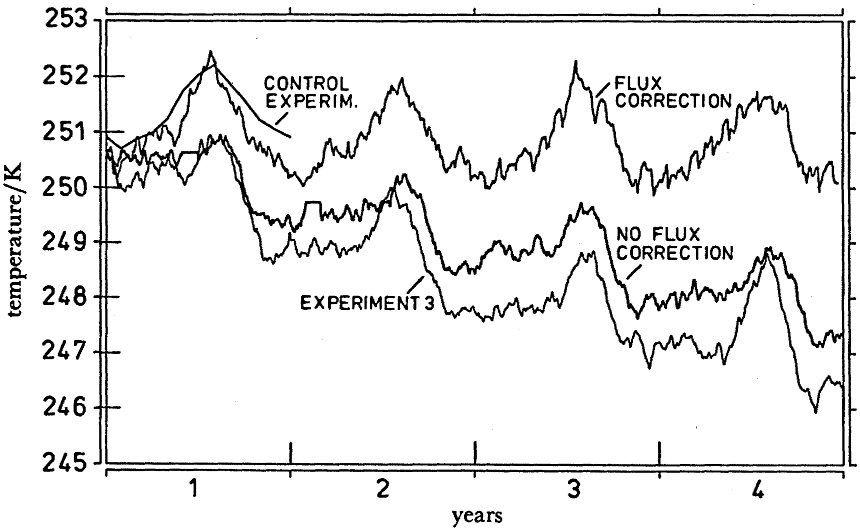

Figure 6. The global mean atmospheric temperature as function of integration time.

4.2. *The oceanic fields* (*see note added in proof*)

The sea surface temperature field (figure 7) has in all runs with and without flux correction a tendency to cool the equatorial Pacific and to warm the mid-latitudes. In a global mean, the SST cools with an annual rate of 0.35 °C for the experiment without flux correction, an annual

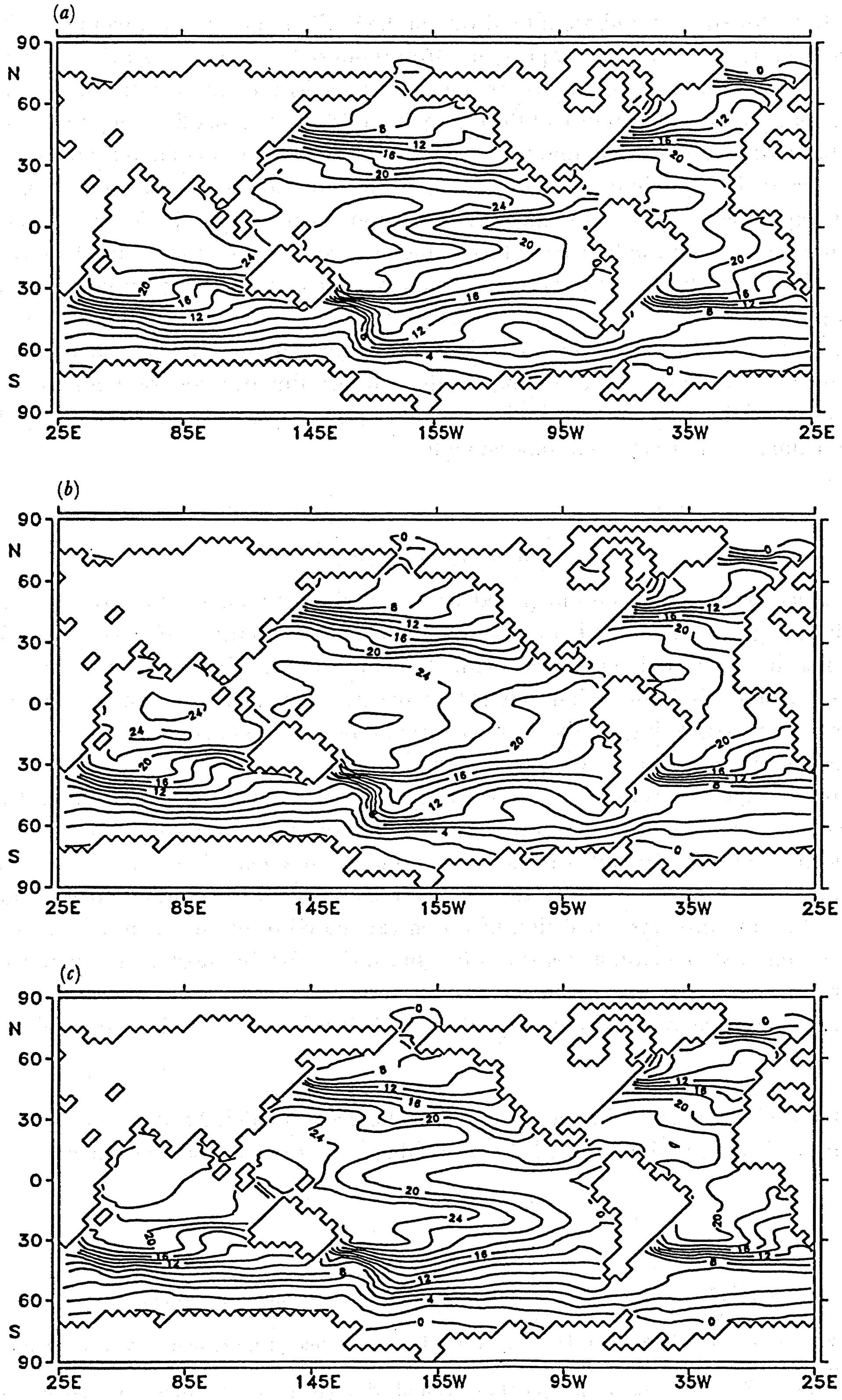

FIGURE 7. The ocean surface temperature after four years of integration time for January. (*a*) Experiment without flux correction; (*b*) with flux correction, ocean time step 30 days; (*c*) with flux correction, ocean top-level time step 1.5 h. Contour interval 2 °C.

rate of 0.37 °C for run (2) and an annual rate of 0.45 °C for run (3). A warming of the lower ocean layers indicates, that this temperature drop is not only caused by poorly simulated heat exchange between the models, but also by too-strong vertical mixing within the ocean model. The particularly poor performance of the ocean model with the short time step in the top layer can only be attributed to the fact that the flux correction has been calculated from a longer time step and is caused by nonlinear processes not transferrable to the shorter time step. It is not clear why experiment (2) shows such stable behaviour in the atmospheric part, whereas it is still cooling in the ocean-model part. The cooling as such might be caused by the unbalanced fresh-water flux, which alters the vertical exchange, but this should eventually feed back to the atmosphere. However, in this case the equatorial Pacific generally cools and does create a cold pool in the eastern Pacific, therefore it does not generate a stronger Walker circulation. The enhanced north–south contrast might still maintain the atmospheric temperature level. The curvature of the SST evolution indicates that it probably will reach stationarity, but the integration time has not yet been long enough.

5. Conclusions

The coupled model has been integrated with or without flux correction for 4 years so far. During this integration time neither of them has reached stationarity, but there are indications that the first flux-corrected run (experiment (2)) will eventually obtain stationarity.

The run without flux correction gets into a positive feedback situation in the equatorial Pacific region by triggering of the Walker circulation. This generates a strong equatorial upwelling and therefore colder SST in the Pacific.

The run with flux correction and an ocean time step of 30 days has also a small drift. This might be caused by inconsistencies, programming bugs or the fact that the fresh-water flux has not yet been corrected. Overall it is the most stable of three experiments.

The run with a short time step in the top layer shows the largest climate drift despite flux correction. Because this flux correction has been calculated from an ocean-model run with a longer time step, its application might not be appropriate for the shorter time step because of nonlinearities.

More work has to be done on the coupled model before it can be used to do sensible climate sensitivity studies.

I thank Dr Maier-Reimer for the provision of the ocean model, M. Böttinger and C. Rath for support with the graphics and M. Lautenschlager for the control experiment.

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Discussion

J. F. B. **Mitchell** (*Meteorological Office, Bracknell, Berkshire, U.K.*). In Dr Cubasch's coupled model, temperatures are too low in the tropics, and too high in high latitudes. In the Meterological Office model, the reverse is true (the model is too warm in the tropics). Does Dr Cubasch know why the meridional temperature gradient is smaller than observed in his model?

U. **Cubasch**. The coupled model gradually shifts towards this state. It is not yet clear whether it is caused by internal dynamics of the coupling, by the flux correction, or just by a programming error. It is, however, interesting to note that similar problems seem to arise in the coupled models of the University Corporation for Atmospheric Research and the Geophysical Fluid Dynamics Laboratory.

Note added in proof (*12 May 1989*). As development work on the coupled experiment continued it was found that the unusual temperature distribution in the ocean top layer was caused by a program error. After correction, the cooling vanished and the sea surface temperature stayed reasonable throughout a five year integration period for experiment 2. Experiments 1 and 3 have not yet been repeated.