PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A
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Experiences with a Coupled Global Model [and Discussion]

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Experiences with a coupled global model

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General circulation models of the atmosphere have been used to investigate the climate response to factors such as the changing concentration of CO_2. Their usefulness is restricted by the need to specify the sea surface temperature. Partial solutions to this problem exist, such as adding a model of the ocean mixed layer to the atmosphere model, but these cannot simulate the response of the ocean heat transport to changes in the atmospheric circulation. Only a coupled atmosphere–ocean–sea-ice model can represent the mechanisms that determine the climate on time scales of decades.

A coupled atmosphere–ocean–sea-ice model has been developed at the Meteorological Office. This paper describes the ocean and sea-ice components of that model and some of the characteristics of the ocean model when driven by observed fluxes of heat, fresh water, and momentum during a long spin-up experiment. Aspects of a four-year integration of the coupled model are discussed.

Many factors contribute to the simulation of the coupled model. Not only are the characteristics of the component models present, but the additional degrees of freedom introduced by the removal of fixed boundary conditions at the ocean surface also introduce new features into the simulation. Particular features that result from the interaction of the models used in the simulations described in this paper include a feedback between the sea-ice model and the simulations of the atmosphere model at high latitudes, and a warming of the tropical Pacific.

1. Introduction

Changes in climate on both long and short timescales are of social and economic importance. The El Niño of 1982–83 has highlighted the role that the ocean (in this case the tropical Pacific) plays in determining the weather patterns on seasonal timescales. Prolonged periods of drought, such as that experienced in the Sahel region during the 1980s (Owen & Folland 1988), can result in the loss of agricultural land essential to the existence of whole communities. Over longer timescales the Earth has undergone the great changes in climate of the Ice Ages. Observations of atmospheric carbon dioxide show a steady increase from pre-industrial levels as a result of increased use of fossil fuels and the removal of the tropical forests (see, for example, Sundquist & Broecker 1985). The impact of this and of changes in other radiatively active gases on climate must be understood. Long-term economic planning demands that such basic factors as the regional conditions influencing crop growth must be forecast. On the timescales over which such changes may occur the capability of the ocean to store, release and transport heat (Bryan & Spelman 1985) means that the oceans may be expected to strongly modify the climate response to such influences.

General circulation models of the atmosphere have been under development for many years. By the late 1970s they were able to reproduce the present-day climate with sufficient realism to enable experiments to be done that were aimed at understanding the factors involved in climate change. Initially such experiments were restricted by the need to determine the sea surface temperatures that would correspond to the perturbed climate being modelled. Gradually this restriction was removed. For example, Wilson & Mitchell (1987) used a 'slab ocean model' to simulate the response of the sea surface temperature to changes in the surface forcing: it was later refined to permit the parametrization of lateral heat transport within the ocean. Because the large-scale ocean circulation is expected to change as the atmospheric forcing evolves, a few centres have developed models in which an atmosphere general circulation model is coupled to a dynamic ocean model (see Bryan & Spelman 1985; Han *et al.* 1985; Schlesinger & Jiang 1988). Several idealizations have been introduced into these models: that of Bryan & Spelman represented only a sector of the globe, so that realistic geography was not used. That of Han *et al.* used a two-layer atmosphere. As computing power increases these restrictions are gradually being relaxed.

Development of a coupled model at the Meteorological Office required ocean and ice models to be integrated alongside the existing atmosphere model. It was decided to use an established ocean model for this purpose: that of Bryan (1969) as modified by Cox (1984) was chosen. An ice model based on the 'zero layer' model of Semtner (1976) with the addition of a representation of leads was adopted. A major requirement for integrations of the coupled model to be successful was that the surface fluxes of heat, momentum and fresh water generated by the atmosphere model should be realistic. These were assessed at an early stage, and have been kept under review as the atmospheric model has undergone developments. To date only 'synchronous' integrations of the coupled model have been performed, in which the atmosphere and ocean models are run at the same rate, so that a day is simulated by the atmosphere model for each day simulated by the ocean. Such an integration also forms part of an 'asynchronous' scheme, in which the ocean model is also integrated alone for long periods of time driven by fluxes calculated by a previous 'synchronous' integration: such experiments would be the only practicable method for deriving the equilibrium climate of the coupled system.

After a brief description of the models used, this paper outlines aspects of integrations of the global ocean model when integrated with observed surface fluxes of heat, fresh water, and momentum. This is followed by a discussion of a four-year integration of the coupled model. Finally, plans for future developments of the coupled model are summarized.

2. Model descriptions

(*a*) *Ocean model*

The global ocean model is based on that of Cox (1984). Values of temperature, salinity and currents are held on regular 2.5° latitude by 3.75° longitude grids for 17 unequally spaced layers in the vertical. In the upper ocean layer thicknesses of the order 10 m enable the seasonal evolution of the mixed layer to be represented. A variable vertical diffusion coefficient is used to represent the mixed layer, following Pacanowski & Philander (1981). Lateral diffusion is required to parametrize the small-scale mixing in the ocean. Redi (1982) described a form for the calculation of the diffusion that represents the strong dynamic constraints that confine the mixing of tracers (temperature and salinity in this version of the model) to isopycnal surfaces. This form has been adopted in the ocean model, and in later experiments has been modified further to revert to horizontal mixing within the mixed layer, so that the Pacanowski & Philander formulation can dominate there.

Use of the acceleration technique of Bryan & Lewis (1979), in which different time steps are used for the integration of currents and tracers, enables long integrations to be performed at comparatively low cost. The technique is only suitable for slowly varying solutions, and has no impact on a stationary equilibrium state. Details of the transient response of the ocean to changes in surface forcing are distorted, making the technique unsuitable for the 'synchronous' phase of a coupled experiment in which high-frequency variations in the surface fluxes are generated. It is, however, of great utility in the determination of the equilibrium response of the ocean to surface forcing, or in the 'asynchronous' phase of a coupled experiment.

During stand-alone integrations of the ocean model the fluxes of heat and fresh water are modified by a relaxation to the observed temperature and salinity of the ocean surface by using the formula,

$$\text{flux applied} = \text{flux from external source} + \lambda(T_{\text{observed}} - T_{\text{simulated}}), \tag{1}$$

where λ is a relaxation coefficient (with a different value for each tracer, T, chosen so that, after the scalings used to convert the fluxes to rates of change of the tracer in the upper layer, the relaxation time for each tracer is the same). This allows the imposed heat flux to be modified as a crude representation of the response of the atmosphere to the sea surface temperatures of the simulation. Relaxation towards the observed salinity distribution is necessary to ensure that consistent boundary conditions are applied for density. This technique is equivalent to that of Haney (1971), but in the ocean-only experiments discussed in this paper the value of λ was spatially uniform.

(*b*) *Sea-ice model*

A modified version of the 'zero-layer' thermodynamic model of Semtner (1976) is used. Temperatures within the ice and snow are assumed to vary linearly, with a change of gradient at the ice–snow interface. In the coupled model, diurnal variations of the surface temperature are calculated within the atmosphere model, and to ensure that their amplitude is realistic a term representing the surface heat capacity is included. Although there is no representation of ice dynamics in the model, the importance of open stretches of water within a region of sea-ice is acknowledged by the inclusion of a model for leads. This is achieved through the introduction of the ice concentration, A, changes in which are determined by thermodynamic processes and the assumption of a distribution of ice depths within a grid box. Because completely unbroken ice is rare, the maximum value for A is 0.995 in the Arctic, and 0.98 in the Antarctic, following Parkinson & Washington (1979).

Presence of sea-ice affects the atmosphere and ocean models. In the atmosphere model, the albedo of the ice surface varies linearly with the surface temperature (between $\alpha = 0.8$ at 268 K and $\alpha = 0.5$ at 273 K) as a crude representation of the presence of melt ponds. Multiple reflections between cloud and ice are included through a reduction in the surface albedo. Leads are assumed to have a fixed albedo. The boundary-layer parametrization in the atmosphere model allows separate calculations of the turbulent fluxes over ice and leads.

Interaction of sea-ice with the ocean has three main effects. Momentum transfer between the atmosphere and ocean is, in reality, strongly modified by the presence of ice. In the model, because of the lack of ice dynamics, the wind stress is calculated by using a drag coefficient appropriate for sea-ice, but then passed directly to the ocean model. Heat fluxes to the ocean fall into two categories. Fluxes through the surface of leads directly heat or cool the ocean. In addition, there is a heat flux between the ice and ocean. This is calculated assuming that the exchange is proportional to the temperature difference between the base of the ice and the top layer in the ocean model. The constant of proportionality is 200 W m^{-2} K^{-1}. This technique was used by Pease (1975). Salinity changes in the ocean resulting from ice formation and melting play a significant role in the determination of the density of water at high latitudes. Sea-ice is assumed to have constant salinity (6 practical salinity units (psu)), and any excess is rejected into the upper layer of the ocean when ice forms. Ice melt results in the relatively fresh water being mixed with that in the upper layer of the ocean. Simplifications implied by this treatment are consistent with other assumptions made in the model.

(c) *Coupled model*

A coupled model of the atmosphere–ocean–sea-ice system has been constructed from the Meteorological Office eleven-layer atmosphere general circulation model (see, for example, Slingo *et al.* 1989), and the global ocean and sea-ice models described above. Integrations have been confined to the 'synchronous' mode of running in which the component models run at the same rate and exchange information frequently. In the form used for the experiment described in §4, surface fluxes of heat, water and momentum were accumulated by the atmosphere model for a period of five days and then passed to the ocean and sea-ice models. These were then integrated for five days (interacting at each half-hour time step) to produce new sea surface temperatures and sea-ice extents for the end of the period. These instantaneous values were then passed to the atmosphere model for use as its lower boundary condition for the following five days. This scheme is illustrated in figure 1.

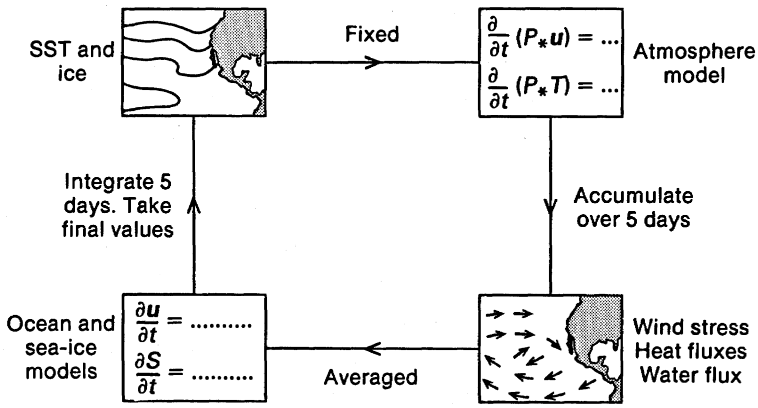

FIGURE 1. Schematic diagram of the system for integrating the coupled model.

3. OCEAN-ONLY EXPERIMENTS

(a) *Forcing and initial data*

Integrations of the ocean model driven by surface fluxes of heat, momentum, and fresh water derived from observed estimates and from the atmosphere model are described briefly in this section to indicate some of the characteristics of its climatology. Only those characteristics of the ocean circulation in these experiments that assist in the interpretation of the results of the coupled model presented later are discussed.

Sources of the observed surface flux data are given in table 1. The various data components were derived independently for assorted resolutions by different authors. Before being used the data were interpolated to the grid of the ocean model, and the monthly averages applied during the integration. These forced one integration (henceforth C). Surface fluxes were also

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

extracted from a four-year control experiment of the atmosphere model (Slingo *et al.* 1989) and used to force a second ocean model integration (M).

In addition to the externally derived surface fluxes, a relaxation towards the observed sea surface temperature and salinity was imposed, (1). The relaxation time associated with this was 57 days, assuming that the heat and salt fluxes generated by the relexation were distributed throughout a 50 m deep mixed layer.

Table 1. Sources of data for derivation of climatological fluxes

flux	source
wind stress	Hellerman & Rosenstein (1983)
heat	Esbensen & Kushnir (1981)
evaporation	Esbensen & Kushnir (1981)
precipitation	Jaeger (1976, 1983)
sea surface temperature	Bottomley *et al.* (1989)
salinity	Levitus (1982)

Initial data for the model integrations were derived by interpolating the values of temperature and salinity appropriate for March from the data set of Levitus (1982) onto the grid of the model. The currents were initially set to zero. The relatively short time step for currents required by the acceleration technique results in little time being available for the currents to adjust to the density field. To generate a balanced initial state, and thus speed convergence to quasi-equilibrium with the surface forcing, the model was first integrated for one month by using a half-hour time step for all variables and the observed fluxes described below. The resulting ocean state was used as the initial state for the spin-up experiment. Two hundred years (surface tracer timescale) of each spin-up experiment were performed. Only the averages of the final 15 years are discussed.

(*b*) *Sensitivity to fluxes*

Despite using the same version of the ocean model, and the same relaxation terms for the surface temperature and salinity, the two integrations differ in major respects. To illustrate the differences which may be found between the two simulations the flow in the North Atlantic at 96 m is considered; although there are major differences elsewhere this is the region which will be discussed for the coupled experiment. Figure 2 shows the difference between the currents simulated in each experiment. Differing paths of the North Atlantic Drift in the two simulations result in an anomalous cyclonic gyre in the difference chart near 45° N. In association with this, cold water extends further east in M resulting in the oval temperature anomaly of figure 3. Further, shorter, integrations have been performed to elucidate the cause of this and other differences. One run used observed wind stress and atmosphere model heat and water fluxes and the other used observed heat and fresh-water fluxes and atmosphere model wind stress. These enable the cause of the anomaly to be identified with the buoyancy forcing provided by the atmosphere model. This resulted in a region of cold water, which then generated a consistent velocity field.

Other differences (not shown) include a reduction in the strengths of the main gyres in M (the Gulf Stream transport is approximately $22\times10^6\ m^3\ s^{-1}$ in M and $32\times10^6\ m^3\ s^{-1}$ in C which represents flow in Sverdrup balance with the wind stress; Fofonoff 1981) and in the temperatures of the water at 1490 m. In much of the ocean at this level, M is warmer than C,

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

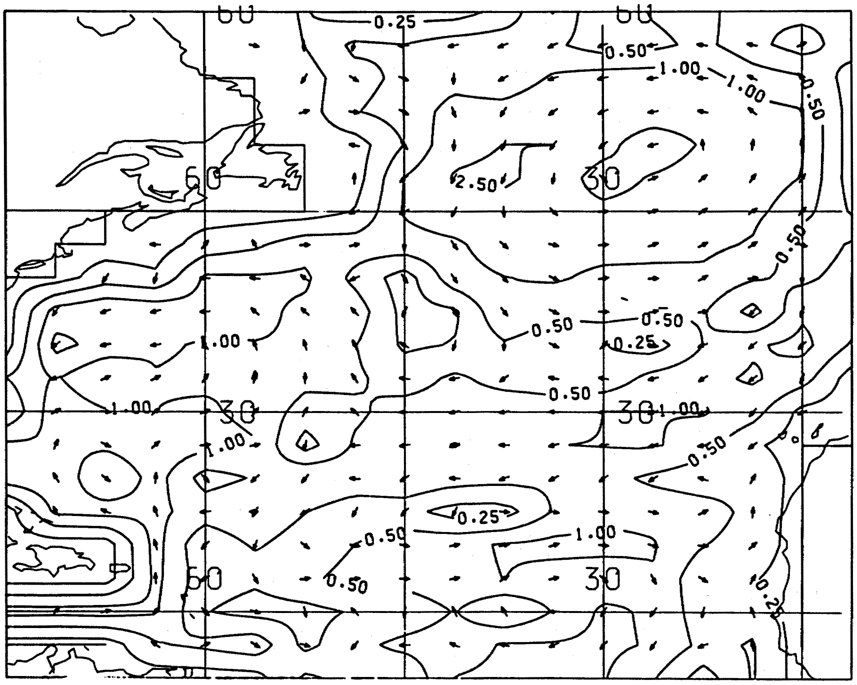

FIGURE 2. Differences between currents at 96 m in the N Atlantic simulated by the ocean model driven by surface fluxes calculated by the atmosphere general circulation model and those in the experiment driven by fluxes from observations. Contours at 0, 2.5, 5, 10 and 15 cm s^{-1}.

FIGURE 3. Differences between temperatures at 96 m in the N Atlantic simulated by the ocean model in the experiment driven by surface fluxes calculated by the atmosphere general circulation model and those in the experiment driven by observed fluxes. Contour interval 1 °C.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

typically by 0.5 °C. Southern Ocean temperature differences are, however, of opposite sign resulting in stronger meridional temperature gradients in M than C.

4. Coupled-model experiment

(a) *Initial data*

Primitive equation models of the atmosphere and ocean are capable of representing many types of motion. It is important to ensure that the initial data for model integrations are in an appropriate dynamical balance for the system under consideration. In a coupled model, each component model must be started from a state which is internally balanced. This is not, however, sufficient to ensure that the subsequent integration is representative of the long-term climate of the model. The boundary conditions for the atmospheric circulation are provided by the sea surface temperature and ice extent; the ocean is driven by the surface fluxes of heat, fresh water, and momentum. An additional constraint therefore applies: the initial states used for the model atmosphere and ocean must be chosen to be compatible. An example of the importance of this is the El Niño phenomenon, which is triggered by the ocean response to weakened wind stress in the tropical Pacific (Zebiak & Cane 1987). Poorly balanced initial states for the coupled system could cause the spontaneous and immediate generation of this phenomenon. A true equilibrum of the ocean with the atmosphere model surface forcing could only be attained after centuries of integration of the coupled model. It is possible, however, to obtain a more restricted balance in the upper ocean after only a few years. This compromise was used in deriving the initial state for the coupled model integration.

An individual day, corresponding to the end of June, was chosen from a climate experiment with the atmosphere model with observed sea surface temperatures (Slingo *et al.* 1989), and the simulated atmosphere for this day was used as the initial condition for the atmosphere model component of the coupled integration.

The sea-ice model was initialized by using extents derived from the climatology used in the atmosphere model control integration. For the purposes of initialization, sea-ice was classified as either seasonal or permanent. Ice thickness and leads fraction were then set to values representative of seasonal (0.5 m and 20% leads) and permanent (2.5 m and 2% leads) ice.

To provide a suitable initial state for the ocean model an ocean integration was first done, initialized with the temperature and salinity data for September taken from the digital atlas of Levitus (1982). Surface fluxes of heat, fresh water and momentum were extracted from the control integration of the atmosphere model. Monthly averages of these were calculated for each month leading up to the day chosen to initialize the atmosphere model. During this integration the fluxes of heat and fresh water were modified by a relaxation towards the observed surface values of sea surface temperature and salinity (1). The ocean model was then integrated by using these fluxes for a total of 21 model months. During the final six months the relaxation towards observed surface values was strengthed (that for temperature from 41 W m^{-2} K^{-1} to 100 W m^{-2} K^{-1}).

(b) *Simulation of sea surface temperatures*

Systematic trends in the coupled model simulation may be identified by the examination of the annual mean of the simulated fields. The difference between the annual mean simulated sea surface temperature during the final year and the observed climatology used for the control atmosphere experiment is shown in figure 4.

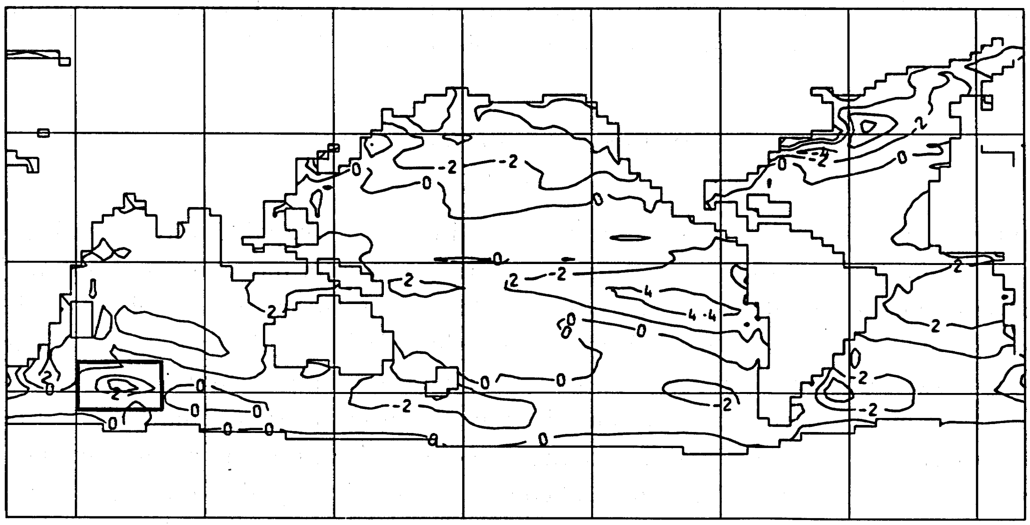

FIGURE 4. Annual mean sea surface temperature during the fourth year of the coupled model experiment less the climatology imposed during the control integration of the atmosphere model. Contour interval 2 °C. The boxed area is referred to in the text.

Over the whole of the tropical oceans the coupled model temperatures are higher than those observed, the smallest errors occurring on the Equator. Similar behaviour has been reported for a higher resolution tropical-ocean–global-atmosphere (TOGA) coupled model (Gordon & Corry 1989).

The major western boundary currents of the Northern Hemisphere, the Gulf Stream and the Kuroshio, are characterized by water cooler than observed along the coasts and warmer off the coast, consistent with the coarse resolution of the model that causes these current systems to be too broad and weak.

At middle latitudes in both hemispheres the simulated sea surface temperature is too cool. Although extrema occur in areas that are connected with the western boundary currents, such as the area to the south of Greenland, others occur in the open ocean. An example of the latter is near 45° S 50° E (the boxed area of figure 4) where the modelled temperature is 5 °C cooler than observed. This feature, and the warm water downstream to its east, may be caused by inadequate representation of the flow induced by a nearby submarine ridge, as it was also found in the integrations discussed in §3.

General cooling of the ocean at northern middle latitudes is, in part, a result of the underestimation of the solar heat flux entering the ocean from the atmosphere as a result of deficiencies of the cloud simulation (Slingo *et al.* 1989) (errors in excess of 20 W m^{-2} were found over large areas of the ocean, that would correspond to a 2.5 °C cooling of the top 100 m of the ocean during a three-year integration), but the ocean heat transport also has a role in producing this error, as discussed in the next section.

(*c*) *Simulated heat transport in the ocean*

A major role of the ocean in the climate system is to transport heat polewards from the tropics. Inadequate heat transport by the ocean may contribute to the cooling of the ocean at middle and high latitudes. It is instructive, therefore, to examine the latitudinal distribution of poleward heat flux in the ocean simulation even though the model is not in equilibrium. This

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

is shown in figure 5, which compares the total meridional heat flux in the ocean during the first and last years of the coupled integration with estimates derived from observations. Although the observed estimates differ greatly, outside the tropics the modelled heat flux is less than either, particularly in the Northern Hemisphere. Underestimation of meridional heat flux is common to many ocean models (Bryan & Lewis 1979; Meehl & Washington 1988). It is noteworthy that although the tropical heat transport decreases during the first four years of the experiment (associated with the warming of tropical waters) transports in middle latitudes increase with time.

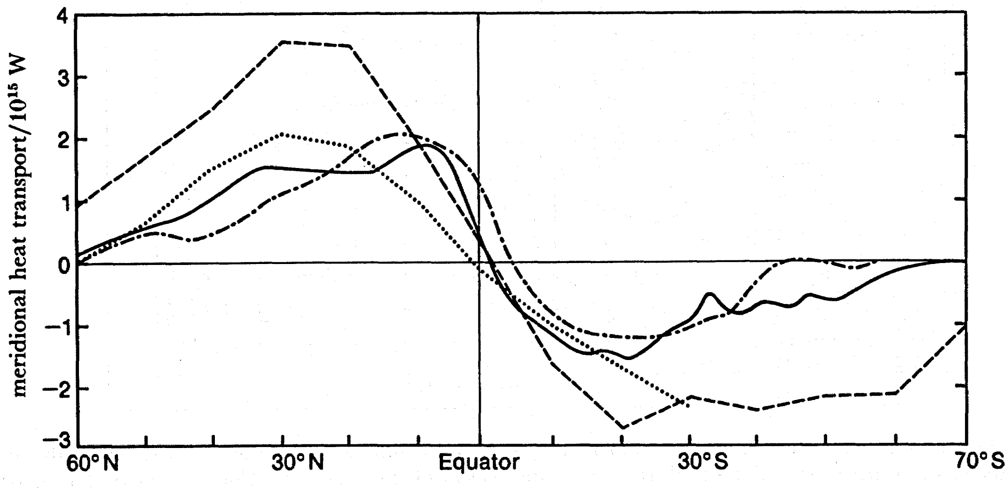

FIGURE 5. Meridional heat transport by the ocean. Two estimates derived from observations are shown (dotted and broken lines) together with the values for the first (dash-dot line) and final (solid line) years of the coupled integration.

Anomalous heat fluxes generated by the relaxation term in the ocean-only integration driven by observed heat fluxes indicate that the ocean model requires an additional heat flux of 20 W m^{-2} into the ocean to maintain quasi-equilibrium in northern middle latitudes; the ocean and atmosphere models thus contribute similar errors at middle and high latitude oceans. Because the model is not in equilibrium the heat flux across the ocean surface need not balance that within the ocean, thus permitting trends of heating and cooling.

(*d*) *Ocean circulation*

All the major barotropic ocean gyres are represented in the simulation, although the Kuroshio and Gulf Stream transport less mass than observed (for the Gulf Stream the modelled value of 32×10^6 m^3 s^{-1} should be compared with the observed estimates of order $100–150 \times 10^6$ m^3 s^{-1} of which $30–35 \times 10^6$ m^3 s^{-1} is attributable to the wind stress and the remainder to recirculation (Fofonoff 1981) the dynamics of which are not adequately represented in a coarse-resolution model). This is consistent with the inadequate resolution of the ocean model. Examination of the currents in the upper ocean demonstrates the significance of these differences. Figure 6 shows the difference between the N Atlantic currents at 96 m near the start of the coupled integration and near the end. The change in circulation of the North Atlantic Drift is clearly seen as a gyre in the difference charts. This was first noticeable during the second winter of the integration at the surface and six months later was to be found at 96 m. Associated with the change in currents is a cooling in the region of the difference, illus-

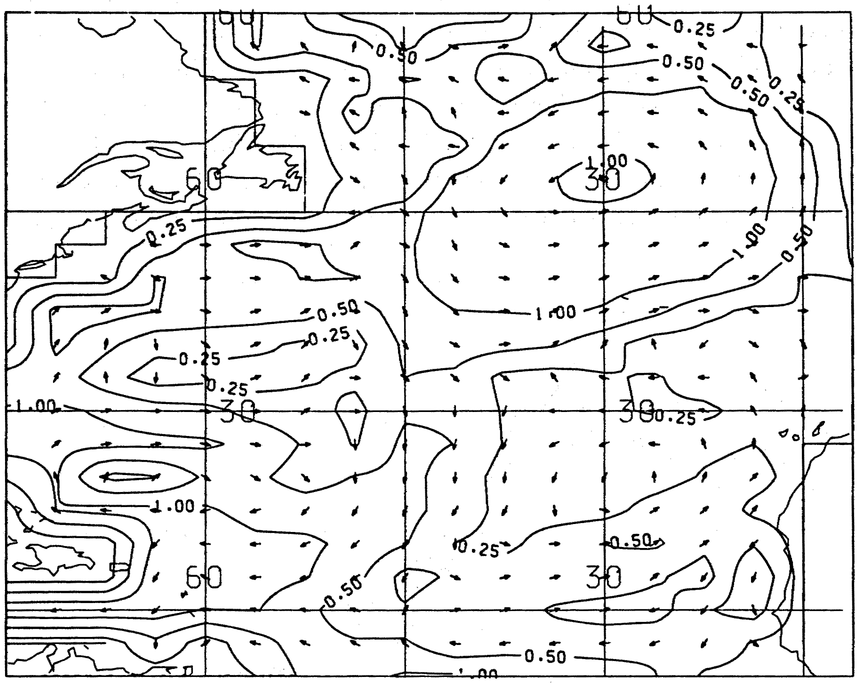

FIGURE 6. Differences between currents at 96 m in the N Atlantic simulated by the coupled model during the summer of the third year and those during the same season of the first year. Contours at 0, 2.5, 5, 10 and 15 cm s^{-1}.

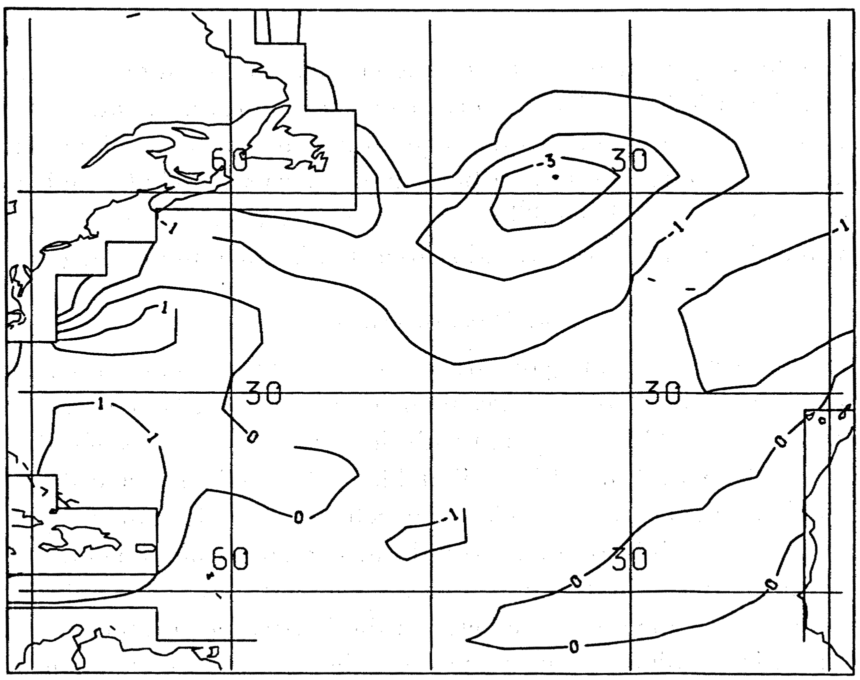

FIGURE 7. Differences between temperatures at 96 m in the N Atlantic simulated by the coupled model during summer of the third year of integration and those during the first year. Contour interval 1 °C.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

trated in figure 7. This difference in the pattern is qualitatively similar to that between the two ocean-only experiments, indicating that the atmosphere model buoyancy flux is the responsible agent, as discussed in §3. This feature persists until the end of the integration.

(*e*) *Simulation of sea-ice*

In the Arctic the seasonal variation of sea-ice compares favourably with observations (figure 8) in both amplitude and phase, although the actual extent is greater than that observed. Particular areas where this may be seen clearly are Baffin and Hudsons Bays, where ice erroneously persists throughout summer, and in winter the Gulf of Alaska is ice-bound in the model.

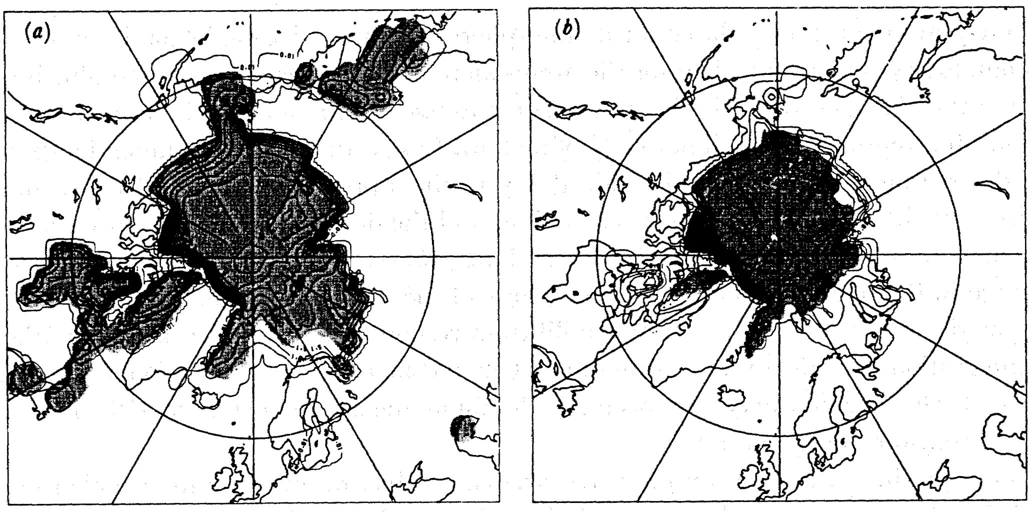

FIGURE 8. Ice depths calculated by the coupled model during the final year, for (*a*) March and (*b*) September. Contours are for 0.01 m and at multiples of 0.5 m. Shading represents the extent of ice used in the control integration of the atmosphere model.

Excessive Arctic ice might be expected if the oceans transported insufficient heat to high latitudes, a feature of the present integration (figure 5). This is not the sole cause of the excessive ice. When integrated by itself, the atmosphere model in the form used for the coupled experiment underestimates the surface temperature over land at high latitudes during summer by several degrees because of excessive low cloud in the simulation through a slowing of snow melting. In the coupled model, this led to a reduction of advection of warm, continental, air over the already too-cold oceans. The cold sea and ice surfaces resulted in excessive low cloud, the radiative properties of which are poorly represented at high latitudes in the model, encouraging further cooling. This mechanism resulted in the surface temperature over the northern Canada and the U.S.S.R. being too cool by over 10 °C, the atmosphere model directly contributing up to 5 °C of this.

Antarctic sea-ice in the model (not shown) does not display the strong annual cycle of the observations. This results in insufficient ice in winter and excessive ice in summer, although a systematic erosion of southern sea-ice extents during the integration modifies this.

Hibler (1984) discussed the role of ice dynamics in the Southern Ocean and found that seasonal variations were larger when the dynamics were represented. It is consistent that the present integration has insufficient seasonal variation. Systematic reduction in southern sea-ice cover was also a feature of the integrations of Gates *et al.* (1984) and Washington *et al.* (1980). Gates *et al.* attributed this trend to errors in the heat fluxes calculated by the atmosphere model; Washington *et al.* concluded that the high lateral diffusion required by the coarse-resolution ocean model used in their study was the primary cause. Although the present model has finer horizontal resolution than the earlier ones it still requires unrealistically large lateral diffusion, although figure 5 shows that there is too little transport of heat towards high latitudes by the ocean model. Heat fluxes at the ocean surface over the Southern Ocean are too poorly observed to enable conclusions to be drawn about their errors in the numerical model.

5. Conclusions and future developments

A four-year integration of the coupled atmosphere–ocean–sea-ice model has been performed. Although four years are sufficient for the atmosphere model to reach a quasi-equilibrium with the sea surface temperature (this is achieved on a timescale of months) it is barely long enough to establish a representative climatology of its flow. Ocean timescales are much longer; apart from the seasonal response of the mixed layer, the upper layers take decades to reach equilibrium with changes in the surface forcing, and the deep ocean may require centuries to attain an equilibrium state. The coupled model integrations done so far therefore represent the first stage in the investigation of the climatology of the coupled system.

Although not representative of the equilibrium response of the model, aspects of the four-year integration provide an indication of what should be expected from a longer experiment. A particular feature that may be expected to dominate the eventual climate of the model is the low heat transport of the ocean.

It is not possible, with the computer resources available, to integrate the coupled model to equilibrium by using the synchronous technique described in §4. Washington *et al.* (1980) used as 'asynchronous' technique in which the coupled model was integrated for short periods by using the synchronous scheme and the fluxes calculated by the atmosphere model stored. These were then applied to a longer integration of the ocean model. It is proposed to use a similar technique for longer integrations.

Further experiments are planned in which the ocean model will be integrated by using surface fluxes derived from different sources. By comparing integrations forced by different combinations of fluxes from the atmosphere model and observations (for example, modelled heat fluxes and observed wind stress) it will be possible to distinguish the roles of the different fluxes in the coupled integration.

Systematic errors in the simulations of the component models of the coupled system produce 'climate drift', the tendency of the model simulation to become removed from the present climate despite the ability of both models to produce an acceptable simulation independently. Experiments to investigate climate change require the climate of the model to be qualitatively the same as the present climate. Once a long integration of the coupled model has been performed it will be assessed and the processes responsible for errors in the simulated climate identified. Wherever possible the formulations of the component models will be improved to reduce the error, although it is expected that in the short term it may be necessary to use *ad hoc* corrections, such as the flux correction method of Sausen (1988).

Experiments are planned to determine the impact of increased resolution on the aspects of the ocean simulation of importance to the coupled model, such as the heat transport. Further development of the parametrizations in the ocean model is also being investigated. Bottom friction is being introduced into the ocean model to investigate its impact on the strength of the Antarctic Circumpolar Current and on the inter-basin exchanges of heat and salt. Pacanowski & Philander's (1981) formulation of the vertical mixing was calibrated to produce an acceptable representation of the equatorial undercurrent in their model. Its use in the extra-tropics needs justification, and it may be desirable to modify the form of the relation between the vertical mixing coefficients and Richardson number, which is used in the scheme to produce a parametrization that is more representative of middle and high latitudes or to adopt a bulk mixed layer formulation.

The eleven-layer atmosphere model is under constant development. Work is in hand to convert the coupled model to use the latest version of the atmosphere model that uses an improved representation of cloud and of the transport of moisture in the boundary layer. One of the major factors affecting the simulation of sea-ice in the coupled experiment was the unrealistically dense cloud at high latitudes that resulted in insufficient heating of the surface. Modifications to the radiation scheme of the atmosphere model have been included, and are designed to correct this error. A further experiment with the coupled model with this improved form of the atmosphere model is under preparation.

Future integrations of the coupled model will be of longer duration. This will enable the short-term climatology of the model to be determined, and provide a control integration for perturbation experiments to investigate climate change.

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Discussion

H. von Storch (*Max-Planck-Institut für Meteorologie, F.R.G.*). In his talk, Dr Foreman has shown that his coupled model is capable of reproducing the basic aspects of the annual cycle of sea surface temperature. Does this annual cycle repeat itself, apart from the trend that he mentioned, in the four-year run? Or does the model develop transient large-scale sea surface temperature anomalies, for example in the North Pacific?

S. J. Foreman. Firstly, I must emphasize that the model integration was only of four years' duration. This is not sufficient for the coupled system to attain a balance. It is not clear whether anomalies in the experiment are typical of those that would be present in a longer integration.

Despite this, the N Pacific does show interesting patterns within the general region of sea surface temperature cooling. The region of greatest anomaly appears to progress around the gyre, identifiable for two years. This may be the product of an anomalously intense atmospheric low during the second simulated winter, but the integration has not been analysed sufficiently to confirm this.

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES