Climate Calculations with a Combined Ocean-Atmosphere Model

SYUKURO MANABE AND KIRK BRYAN

Geophysical Fluid Dynamics Laboratory, ESSA, Princeton University, Princeton, N. J.

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Empirical evidence indicates that the poleward heat transport by ocean currents is of the same order of magnitude as the poleward transport of energy in the atmosphere (Sverdrup, 1957). A significant contribution to the heat exchange across latitude circles is also associated with polar pack ice. Thus, any serious attempt to calculate climate must take into account the entire fluid envelope of the earth, consisting of the atmosphere and the hydrosphere. Although the cryosphere, consisting of ice packs over the oceans and continental ice, is not a fluid in the usual sense, it must be included in a general climatic model because of its large reflectivity to the solar insolation and its ability to store and transport heat.

The object of this note is to report the completion of a calculation based on a combined numerical model of the atmosphere and the ocean carried out at the Geophysical Fluid Dynamics Laboratory, ESSA. The extended numerical integration of this model is successful in producing many realistic features of climate starting with quite arbitrary initial conditions. Only a brief description will be given here. For the full details the reader is referred to Manabe (1969) and Bryan (1969).

The numerical model of the atmosphere is very similar to that described by Manabe *et al.* (1965) except that the hydrology of the continental surface is taken into consideration. Velocity, temperature, water vapor and surface pressure are calculated at each of the grid points which are spaced approximately 500 km apart. Calculations are carried out at 9 levels which are chosen so that they resolve the structure of the lower stratosphere and the Eckman boundary layer. The radiation model is essentially that described by Manabe and Strickler (1964). For the sake of simplicity, the seasonal and diurnal variation of solar insolation are not taken into consideration; instead, annual mean insolation is assumed for this study. The depletion of solar radiation and the transfer of terrestial radiation is computed taking into consideration cloud and gaseous absorbers such as water vapor, carbon dioxide and ozone. The distribution of these absorbers is specified in advance to correspond to zonal averages taken from climatological data. The prognostic equation of water vapor involves the three-dimensional advection of water vapor, condensation and evaporation. Over continental surfaces, the depth of snow cover and the amount of soil moisture are predicted based upon detailed balance computations of snow and soil moisture, respectively.

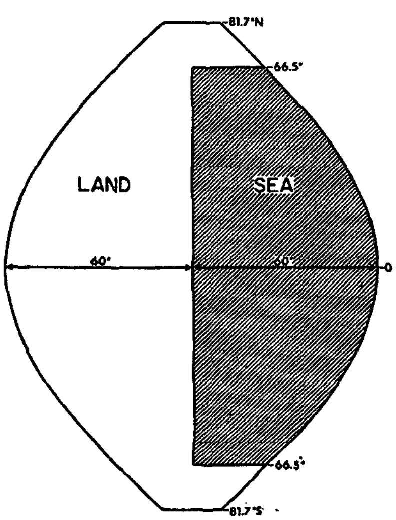

FIG. 1. Ocean-continent configuration of the model.

The ocean model is similar to that of Bryan and Cox (1968), except that the fields of temperature and salinity are calculated explicitly, and density is calculated from a realistic equation of state. Another new feature of the ocean model is a simplified method of calculating the growth and movement of pack ice in polar latitudes. Calculations are carried out for 5 different levels with respect to the vertical coordinate. The horizontal resolution is similar to the atmospheric model in the interior, with extra rows of grid points for greater resolution near the western boundary.

The calculations are carried for a region on a globe bounded by two meridians 120° of longitude apart, cyclic symmetry being assumed in the atmosphere at these meridional boundaries. The regions immediately adjacent to the poles are excluded by free slip, insulated walls at 81.7N and 81.7S. In the interval between 66.5N and 66.5S, half of the area is covered by ocean. Fig. 1 shows the ocean-continent distribution chosen for this study.

In the first stage of the calculation the effect of heat transfer in the ocean model is suppressed. The ocean surface is simply regarded as a wet surface without any heat capacity. An equilibrium state is reached by numerically integrating the atmospheric model starting from isothermal initial conditions. The final "climate" attained is used to specify fixed boundary conditions for the ocean model in the second stage of the calculations. These fixed surface boundary conditions consist of the wind stress pattern, the sea surface temperature, and the rate of supply of water at the ocean surface. The initial condition for the ocean model is a resting ocean with a horizontally uniform distribution of temperature and salinity. At the end of the numerical integration of the ocean model over the equivalent of 60 years with fixed boundary conditions, an equilibrium is attained in the upper layers with only minor changes taking place below the main thermocline. An extensive ice pack with a thickness of 1–4 m forms in the northern part of the ocean.

In the final part of the calculation interaction between the atmosphere and the ocean is allowed. Since different types of fluid motion occur in the atmospheric and ocean models, the atmospheric model requires approximately 40 times more computation to integrate over a given time period as the ocean model. According to the stage I results of the numerical integration of the atmospheric model, the thermal relaxation time of the atmosphere is of the order of 1 year. On the other hand, an estimate of the ratio of heating to heat capacity of the ocean indicates that the thermal relaxation time of the ocean is of the order of centuries. Obviously, it was not feasible to make a synoptic calculation of the interacting models over a long enough period for the oceanic part of the model to reach adjustment. In order to optimize the amount of the computation, the coupling between the atmospheric part and the oceanic part of the model is adjusted such that the evolution of the former during 1 atmospheric year is coupled with that of the latter during 100 oceanic years. For example, the atmosphere on 0th, 0.5th and 1st atmospheric year interacts with the ocean on 0th, 50th and 100th oceanic year of the time integration, respectively. The average temperature of the upper 50 m of the ocean is taken to be representative of the surface mixed layer. The temperature of this surface layer is used as the lower boundary condition of the atmospheric model. The rates of supply of heat, momentum and water to the ocean surface, which are computed in the atmospheric model, serve as the upper boundary condition for the ocean model. The running time-mean operator is applied to vertical fluxes to avoid an overresponse of the ocean model to features caused by individual synoptic disturbances in the atmospheric model.

The latitude-height distributions of temperature of the joint model shown in Fig. 2 are zonal averages at the end of the numerical integration of the final part of the computation. Since the zonally averaged temperatures over the land surface are much lower than those over the sea at higher latitudes, the isotherms do not exactly coincide at the air-sea interface in Fig. 2. This distribution is computed as an average state for the last two-sevenths of the period of the final stage of the time integration. This integration of the joint model is performed over 1 year of atmospheric time, which is equivalent to 100 years for the ocean model. It required about 1200 hours of computing on a UNIVAC 1108.

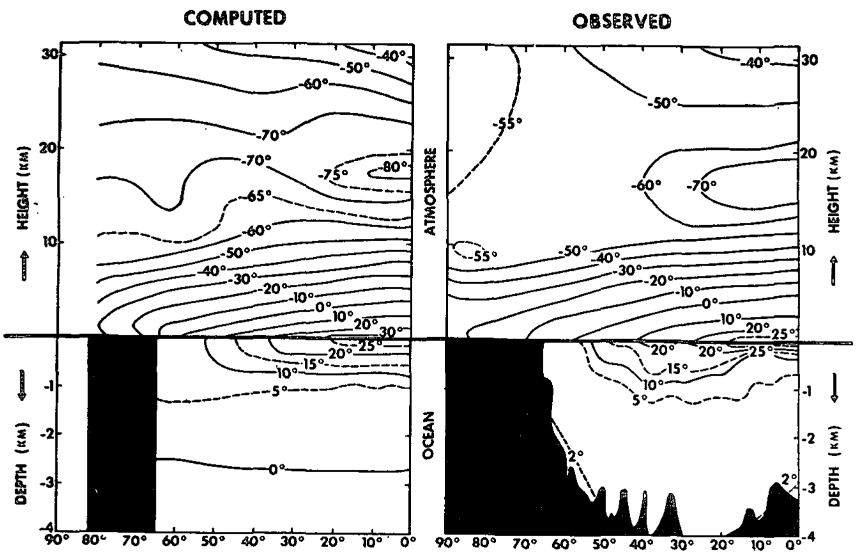

FIG. 2. Zonal mean temperature of the joint ocean-atmosphere system, left-hand side. This distribution, which is the average of two hemispheres, represents the time mean over two-sevenths of the period of the final stage of the time integration. The right-hand side shows the observed distribution in the Northern Hemisphere. The atmospheric part represents the zonally averaged, annual mean temperature. The oceanic part is based on a cross section for the western North Atlantic from Sverdrup *et al.* (1942).

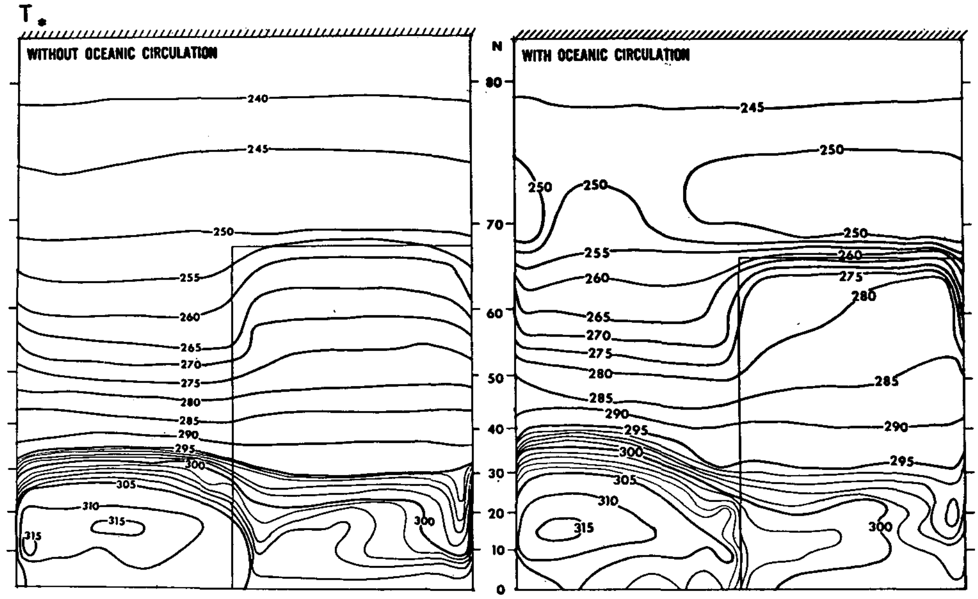

FIG. 3. The surface temperature pattern (°K), the distributions of two hemispheres being averaged. The left-hand side of the figure represents an average over the final 100-day period of the time integration of the atmospheric model (the first stage integration) while the right-hand side is the final 100 atmospheric days (or 28 oceanic years) of the time integration of the joint ocean-atmosphere model (the final stage of the integration).

Even after this very time-consuming calculation, a climatic equilibrium is not completely attained. The atmospheric and ocean surface temperatures are nearly steady, but the mean temperature of the ocean continues to increase at an average rate of 1.3C per century toward the end of the integration. The result is an average net flow of heat into the ocean equal to 0.01 ly min^{-1}, which is about 0.5% of the solar constant. This net warming is associated with the cold water mass near the bottom shown in Fig. 2 and is a relic of the second stage of the calculation, in which fixed surface boundary conditions are maintained at the ocean surface. In short, this result indicates that the thermal relaxation time of the model ocean is longer than 100 years. The free interaction of the atmospheric and ocean models stops the formation of deep water at temperatures $\lesssim$3C, and eliminates the ice pack at polar latitudes. Also, it reduces the snow cover over the continent significantly.

The most interesting result[1] of the calculation is the quantitative demonstration of the effect of ocean currents on the distribution of temperature, relative humidity, and precipitation patterns. This is done by comparing the state, which is obtained from the time integration of the atmospheric model (the first stage) with the state, which is obtained from that of the joint ocean-atmosphere model (the final stage). For example, Fig. 3 shows the surface temperature patterns averaged over the final periods of the first and final stages of the calculation. In the final stage, upwelling near the equator forms a weak temperature minimum there, instead of a maximum. The ocean currents also cause isotherms over the northern part of the ocean to have a much more realistic NE–SW trend. A tight gradient appears along the poleward wall of the ocean in the final stage of the calculation. The rates of precipitation corresponding to the two temperature patterns are shown in Fig. 4. The comparison indicates a drastic reduction of rainfall over the tropical ocean due to the effect of equatorial upwelling. The overall ratio of precipitation falling over land compared to that falling over the sea is very much altered by this effect. Also, a significant increase in the rate of precipitation is evident along the east coast of the continent in the subtropics owing to the effect of the poleward advection of warm water by the so called "subtropical gyre." Another modification of interest caused by the ocean currents is the general poleward shift of the rainbelt in middle latitudes. It is hoped that the experience gained in this preliminary study will be useful in planning and carrying out more extensive climatic calculations in the near future.

[1] In the final stages of the analysis it was discovered that the stress values used as an upper boundary condition over the ocean were multiplied by an extraneous factor, the cosine of latitude. A new run corresponding to three decades of ocean time was made to test the effect of this error. The subarctic ocean gyre became much stronger, while the strength of the subtropical gyre and the thermohaline circulation remained the same. Warming by 1-2C in the upper ocean takes place at both high and equatorial latitudes at the expense of middle latitudes. The level of sensible heat transfer between the ocean surface and deeper layers has roughly the same distribution as in the main run and the main features of Figs. 2–4 are unchanged. More details are given in Manable (1969) and Bryan (1969).

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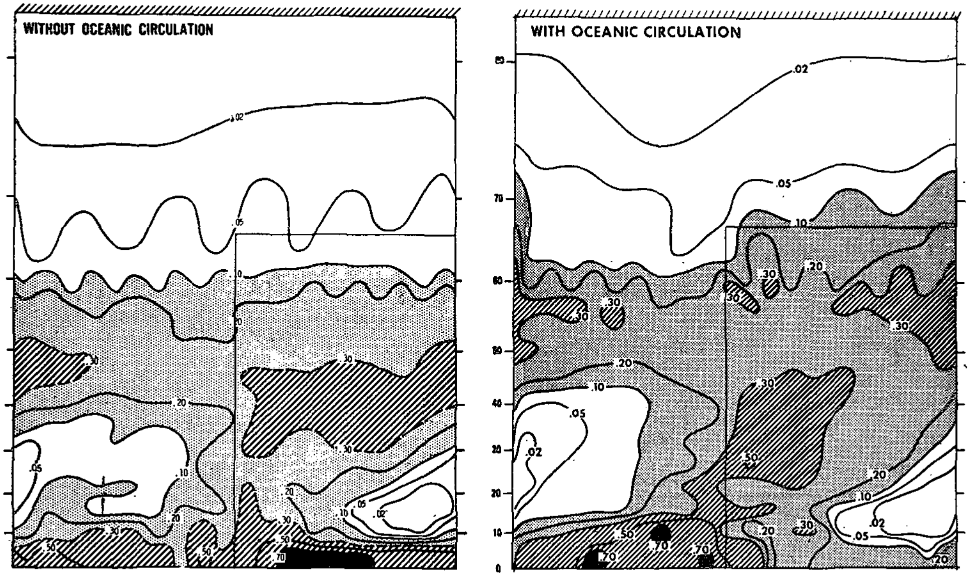

FIG. 4. The distributions of the rate of precipitation (cm day^{-1}) corresponding to the temperature patterns of Fig. 3.

this study and is responsible for creating an ideal environment.

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