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CHAPTER 1

A COUPLED OCEAN–ATMOSPHERE AND THE RESPONSE TO INCREASING ATMOSPHERIC CO_2*

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PLAN OF THE NUMERICAL EXPERIMENTS

The exchanges of heat and carbon between the ocean and atmosphere are crucially important in determining the climatic response to increasing anthropogenic CO_2 in the atmosphere. Subject to an uncertainty caused by a possible growth or decay of the biospheric carbon reservoir, the oceans are estimated to absorb enough CO_2 to reduce the anthropogenic build-up in the atmosphere by as much as 50%. *Sensitivity* studies such as those carried out by Manabe and Wetherald (1975), Manabe and Stouffer (1980), and many others, estimate the equilibrium response to an increase in atmospheric CO_2. The actual timing of the outset of climate change, however, will be controlled by heat exchange with the ocean, which has a heat capacity orders of magnitude larger than the atmosphere. The present study is concerned with the physical processes involved in this transient response.

It is impossible to use historical data to estimate the response time of the coupled ocean–atmosphere system, since the rapid increase of greenhouse gases is an unprecedented event. Transient tracers produced by the bomb tests of the late fifties and early sixties provide one source of information on downward pathways from the ocean surface to deeper layers. The downward movement of these tracers offers an interesting analogue to the penetration of a heat anomaly originating at the surface, with one important difference. The tracers are neutrally buoyant, while a heat anomaly will change the ocean circulation as it is carried downward in the main thermocline. The question then arises as to whether data from the transient tracers (Ostland et al., 1976) can be a quantitatively useful analogue. One of the motivations for this study is to investigate this important question.

At this point, it is crucially important to identify all the significant processes in the transient response of climate to increasing CO_2. For this reason, we have chosen to use models of the ocean and atmosphere that are quite complete physically, but other aspects such as the geometry of continents and oceans are deliberately kept as simple as possible. The actual configuration is shown in Fig. 1. Computations are carried out for a domain spanning 120° of longitude, and extending from equator to pole. Cyclic boundary conditions are assumed in the zonal direction, and mirror symmetry is assumed across the equator. The ocean is confined to a 60° longitude sector which extends from equator to pole. Our geometry is equivalent to both hemispheres having three identical continents

*Some of this material has been published elsewhere. The reader is referred to Bryan et al. (1982) and Spelman and Manabe (1984).

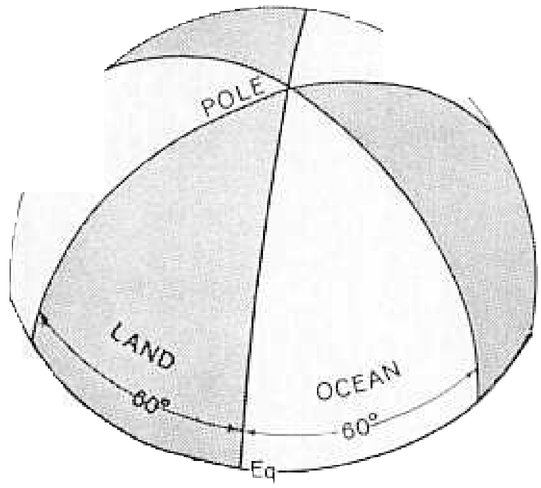

Fig. 1. The simplified geometry of the model. Cyclic symmetry is assumed over a 120° longitude sector. Mirror symmetry is assumed across the equator.

about the size of North America. The three continents are separated by three identical oceans with approximately the width of the North Atlantic.

The atmospheric component of the model is similar to that used in many previous CO_2/climate studies (e.g., Manabe and Stouffer, 1980). Essentially the model is an advanced numerical weather prediction model with the more detailed representation of radiation and boundary layer processes needed for climatic purposes. Clouds are not predicted by the model, but specified from data. This may have an important bearing on the sensitivity of the model (Hansen et al., 1984). We will return to this point in the final section. The atmospheric model is based on a spherical harmonic representation of variables with a resolution of approximately 400 x 400 km in the horizontal plane, and nine vertical levels. The ocean model has the equivalent horizontal resolution but twice the number of grid points in the zonal direction and twelve levels in the vertical. The horizontal resolution in the ocean model is minimal in its ability to resolve ocean currents and should certainly be improved in future studies. Snow over land surfaces, and an ice pack over the sea are included. The sea-ice parameterization was originally specified in Bryan (1969). Snow and ice permit an important albedo feedback which amplifies external influences.

The experimental procedure is illustrated in Fig. 2. Equilibrium climates are calculated for the coupled model corresponding to a normal concentration of atmospheric CO_2, and for the case of four times the normal concentration of atmospheric CO_2. The ordinate in the diagram is the globally averaged sea-surface temperature. The two climate equilibria are illustrated by two horizontal lines. The abscissa represents the time after "switch on"; that is, the time elapsed after the climate corresponding to a normal level of CO_2 is perturbed by impulsively quadrupling the CO_2 level in the model atmosphere. Our study is intended to study the processes that govern the speed and character of the response shown schematically by the dashed line in Fig. 2. Will the delay caused by the ocean's capacity be measured in years, decades, or centuries?

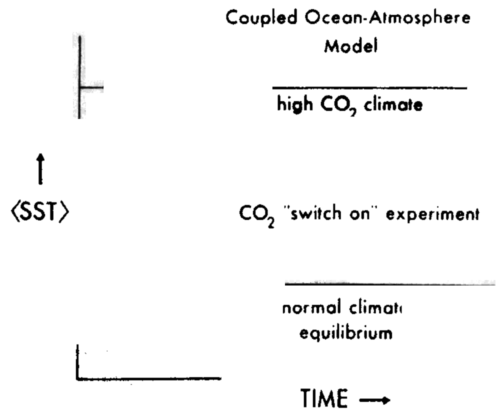

Fig. 2. A schematic diagram of the numerical experiments. The ordinate is the globally averaged sea-surface temperature. The horizontal lines represent climatic equilibria for normal and high atmospheric CO_2. The dashed line represents the "switch-on" experiment.

THE EQUILIBRIUM CALCULATIONS

For fixed lower boundary conditions the calculation of climate equilibrium for an atmospheric model without seasonal variations is relatively simple. Since the response of the atmosphere is only a few months, climatic equilibrium can be reached by simply extending a numerical integration with respect to time over a period of six months to a year. On the other hand, the ocean has a heat capacity orders of magnitude greater than the atmosphere and its thermal relaxation time is correspondingly greater. To integrate a coupled model to climatic equilibrium by straightforward time-stepping would require numerical integration over a period of at least one thousand years. This is clearly not a feasible procedure. To circumvent this difficulty, another method for reaching climatic equilibrium has been devised (Manabe and Bryan, 1969; Bryan et al., 1975; Bryan, 1984). Essentially the method takes into account the fact that different elements of the climate system have very different time scales. By decoupling "slow" and "fast" time-scale processes, each component can be adjusted according to its own natural time scale. Since systems with slow time scales can be integrated with much longer time steps than systems with fast time scales, decoupling allows a much more efficient approach to equilibrium. Once equilibrium is achieved, response experiments are carried out with a fully synchronous version of the coupled model to avoid any possible distortion.

To test the degree of equilibrium achieved, the coupled model was integrated over fifty years in a control run without any change in the atmospheric CO_2 concentration. The net exchange of heat between the ocean and atmosphere was carefully monitored. It was found that the average exchange of heat was less than $0.4\ \mathrm{W\,m^{-2}}$. At this rate of heating, 300 years would be required to change the upper kilometer of the water column by one degree. This is a very small drift compared to the signal imposed by our "switch-on" experiment.

RESULTS OF THE "SWITCH-ON" EXPERIMENT

The result of the "switch-on" experiment is shown in Fig. 3 for the sea surface temperature and the air temperature over both land and sea. The normalized response, R, is:

$$R = \frac{T - T_0}{T_\infty - T_0} \tag{1}$$

where T is the temperature as a function of time after "switch on", T_0 is the initial temperature, and T_∞ is the final equilibrium temperature. The denominator in eqn. (1) is the sensitivity to a change in CO_2, and the numerator is the time-dependent response. The advantage of this normalization is that it allows a direct interpretation of the transient experiment in terms of previous sensitivity studies which evaluate equilibrium response. An important finding of the CO_2 sensitivity studies (Manabe and Wetherald, 1975; Manabe and Stouffer, 1980; Hansen et al., 1984) is a tendency for polar amplification in CO_2-induced warming. As we examine the zonally averaged response in Fig. 3b it can be seen that shortly after "switch on" R is greater at low latitudes. As time goes on, the latitudinal gradient of R gradually diminishes, almost disappearing at 25 years. This means that the transient response at 25 years has nearly the same latitudinal structure as the

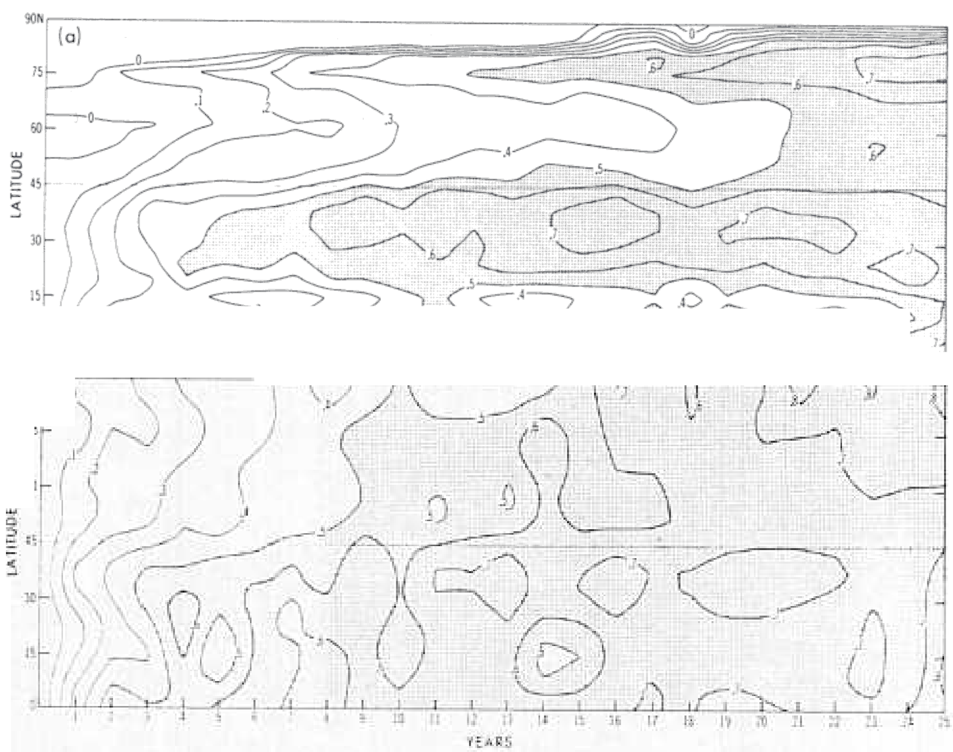

Fig. 3. The zonally averaged normalized response based on one year averages. The ordinate is latitude and the abscissa is time after "switch on". (a) The sea-surface temperature; (b) air temperature over both land and sea.

equilibrium response, including the polar amplification noted earlier. It also suggests that the pattern of transient response to a slow build-up of CO_2 will be similar to the patterns found in sensitivity studies, as long as the time scale of CO_2 increase is longer than 20–30 years.

As mentioned previously, transient tracers provide the only observed analogue for the downward penetration of a large surface heat anomaly caused by a build-up of atmospheric CO_2. Simple one-dimensional models (Oeschger et al., 1975) have been fitted to transient tracer data and used to estimate the delayed response of a coupled ocean–atmosphere model (Cess and Goldenberg, 1981). Our model allows us to actually compare the downward penetration of a tracer and the downward penetration of a heat anomaly. In the control run for normal atmospheric CO_2 a uniform tracer is "switched on" at the upper boundary. The scale depth, d, of penetration may be defined as follows:

$$d = \frac{1}{\mu_s} \int_{-H}^{0} \mu \, dz$$

where μ is the horizontally averaged value of the tracer as a function of depth, and μ_s is the corresponding surface value. H is the total depth of the ocean, independent of position.

In Fig. 4 the penetration depth of the tracer for the control run is shown as a function of time after the tracer is "switched on" at the surface. Tracer input is confined to the top level of the ocean model. Another curve shows the corresponding penetration of a heat anomaly when atmospheric CO_2 is "switched on". Intuition would suggest that the buoyancy associated with a heat anomaly would slow down penetration relative to a passive tracer. Such a feedback has been suggested by Harvey and Schneider (1985), who have incorporated it in a simple, one-dimensional ocean–atmosphere model. In our present calculations the feedback is in the reverse sense. Surface heating causes a greater penetration of a heat anomaly than would be expected for a passive tracer. Vertical transport takes place in three ways: advection, convection and diffusion. If the model

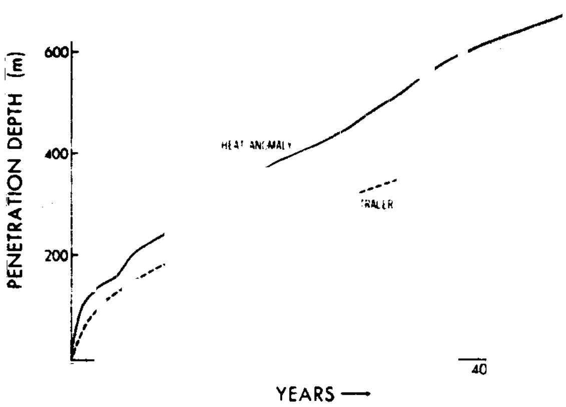

Fig. 4. The penetration depth in meters as a function of time after "switch on". The tracer experiment corresponds to normal atmospheric CO_2.

response was linear, a temperature anomaly would be carried downward by these three types of vertical transport in exactly the same way as a transient tracer introduced at the surface. The nonlinear effect is mainly due to the suppression of convection by the temperature anomaly. In the model, diffusion is independent of changes in the density structure and analysis shows that the vertical advection is also relatively independent of small changes in stratification. Since convection at high latitudes ordinarily transfers heat upward, the suppression of convection traps heat below the surface that has been transported to the subArctic gyre from lower latitudes. Thus, the heat anomaly is carried downward by the same linear processes which carry a tracer downward. Superimposed is another mechanism which involves the suppression of convection which in the undisturbed state cools subsurface waters in the subArctic gyre.

This finding provides an example of the unexpected results which can be obtained in the study of coupled models. Recently Hansen et al. (1984) have pointed out that cloud feedback may greatly increase climate sensitivity to increasing atmospheric CO_2. If this is true, the time scale of climatic response will also be strongly influenced, and the results given here which neglect cloud feedback may be underestimated. This is just one of the many interdisciplinary problems that are awaiting investigation by coupled models of the atmosphere and ocean.

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