Climate Change and Cloud Feedback: The Possible Radiative Effects of Latitudinal Redistribution[1]

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ABSTRACT

The sensitivity of outgoing longwave flux to changes in cloud cover ($\partial F/\partial A_c$) as defined by Cess (1976) must be evaluated carefully to avoid discrepancies arising from the interchange of averaging conventions. In a recent zonal atmospheric model experiment the global value of $\partial F/\partial A_c$ was different in sign than in other calculations. This difference in behavior was traced to a latitudinal redistribution of cloud amount and height that occurred in the doubled CO_2 experiment. However, when $\partial F/\partial A_c$ was evaluated at individual latitudes and then weighted globally, the value of this parameter was consistent with those found by Cess (1976) and Budyko (1974).

1. Introduction

Budyko (1974) first suggested that changes in the longwave and shortwave radiation fluxes due to cloud-cover change tend to balance or counteract each other. Schneider (1972) also discussed this cloud feedback problem, although, because of the assumptions made, his model did not lead to compensating effects. Cess (1976) has shown that in terms of present day empirical data, changes in global cloud amount appear to produce compensating feedback processes that result in no net modification of the global heat balance. Cess and Ramanathan (1978) have clarified the calculation of the relevant sensitivity parameter, $\partial F/\partial A_c$ (where F is the outgoing longwave radiation and A_c is the cloud-cover fraction), and they have suggested that this term cannot be evaluated from model data unless the calculations include the variations in cloud amount with height as well as in total cloud-cover amount.

The results presented here show that in addition to cloud height and amount, it is important to analyze the latitudinal redistribution of cloudiness that might accompany a climate change such as that which would result from doubled atmospheric CO_2. The cloud distribution with latitude may change in such a way that calculation of a global sensitivity number can give unrepresentative results.

[1] This paper was conceived in Leningrad, USSR, while this author was taking part in the US/USSR Bilateral Exchange for Protection of the Environment.

2. Preliminary sensitivity

The basic model used to test this problem is a version of the zonal atmospheric model developed at the Lawrence Livermore National Laboratory (MacCracken, 1968; Ellsaesser *et al.*, 1976; Potter *et al.*, 1978). During the preliminary stages of model validation and sensitivity studies we have calculated the model response to various atmospheric CO_2 concentrations (Potter, 1979). Some of the results of past experiments suggest that the global cloud cover may respond to a perturbed climate (doubled CO_2) as a result of changing dynamics and hydrology. These variations appear to be unrelated to the seasonal response variations discussed by Cess (1976) based on seasonal changes in cloudiness. The overall sensitivity of the annual average version of the model presently is being investigated and the following results are preliminary.

Solar constant sensitivity studies by others (Wetherald and Manabe, 1975; Ramanathan, 1977) with global atmospheric models have used Northern Hemisphere average surface temperatures in the calculation of various sensitivity parameters. Similar calculations have been performed with our zonal model. For a solar constant increase of 2%, for the Northern Hemisphere alone, we found a surface temperature increase of 2.32 K. Based on these results, the model appears to have responded in much the same way as that of Wetherald and Manabe (1975). Proceeding in a manner similar to the regressions computed by Budyko (1974) and Cess

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TABLE 1. Value of C, where $A = Cr^2$.

Latitude	Height (mb) 200	400	600	850
90°N	0.043	0.200	0.300	0.500
80°N	0.043	0.300	0.400	0.700
70°N	0.043	0.300	0.496	0.920
60°N	0.137	0.300	0.603	1.020
50°N	0.150	0.300	0.689	1.060
40°N	0.150	0.300	0.683	1.030
30°N	0.150	0.246	0.616	0.884
20°N	0.150	0.192	0.540	0.800
10°N	0.150	0.346	0.446	0.800
0°	0.150	0.443	0.420	0.800
10°S	0.150	0.346	0.446	0.800
20°S	0.150	0.192	0.540	0.800
30°S	0.150	0.246	0.616	0.884
40°S	0.150	0.300	0.683	1.030
50°S	0.150	0.300	0.689	1.060
60°S	0.137	0.300	0.603	1.020
70°S	0.043	0.300	0.496	0.920
80°S	0.043	0.300	0.400	0.700
90°S	0.043	0.200	0.300	0.500

(1976), the control run value of $\partial F/\partial A_c$ was found to be -110 W m^{-2}.

3. Change in cloud

The cloud-cover prescription is based on a function of the relative humidity data of Crutcher (1971) and of the cloud data by London (1957) such that

$$A = \mathrm{Min}[0.7, C(\theta, p)r^2], \tag{1}$$

where A is the cloud fraction at the appropriate latitude (θ) and pressure level (p), r is the relative humidity and C is a coefficient which varies spatially (Table 1). In the case of overlapping cloud cover, it is assumed that there is a random correlation between the amount of cloudiness at different layers. The total cloud fraction at a given latitude is then

$$A_c = 1 - \prod_{i=1}^{4} (1 - A_i), \tag{2}$$

where A_i is the cloud fraction at the ith level.

Cloud absorption of incident solar radiation is pressure dependent [0.01 at 200 mb, 0.04 at 400 and 600 mb and 0.10 at 850 mb (modified from Houghton, 1954)] and upward scattering is dependent on solar zenith angle for the three lowest cloud layers (Cess, 1976). The top cloud layer (cirrus clouds) has a fixed albedo (0.21).

Fig. 1 shows the change in cloud cover due to doubled atmospheric CO_2 at each of the heights where clouds are computed, and the change in total cloud cover. The change in total cloud cover also includes a contribution from clouds that occur over the mountain fraction at each latitude.

The two most significant features of the change in cloud distribution are the increases in cloud cover at 850 mb at 60°N and 40°S. These changes are apparently related to peculiarities of the model and to the ocean/land distributions for both hemispheres. The large cloud increases occurred in latitudes that tended to be stable in the lower layers, in a fashion similar to that described by Wetherald and Manabe (1980). Increased CO_2 warmed the atmosphere which lead to increased water vapor through enhanced evaporation. This, in turn, warmed temperatures further and reduced the Bowen ratio. The increased water vapor in the lower layers produced higher relative humidities and increased cloudiness through Eq. (1). Because the upper layers (600 mb and above) at these latitudes were convectively stable, the increase in water vapor did not penetrate to the higher levels.

Fig. 2 demonstrates the changes in stability (temperature lapse) for the two cases. The curves for the temperature difference between 1000 and 400 mb ($\Delta T_{1000-400}$ mb) in the control run show that in the high latitudes of the Northern Hemisphere, the vertical gradient between 1000 and 400 mb is smaller (i.e., the troposphere is more stable) than for the midlatitudes of both hemispheres. The midlatitudes are also less stable than the tropics or subtropics. The perturbed case produced more stable conditions (1000–400 mb) at most latitudes but less stable conditions at 60°N. However, stability increased at

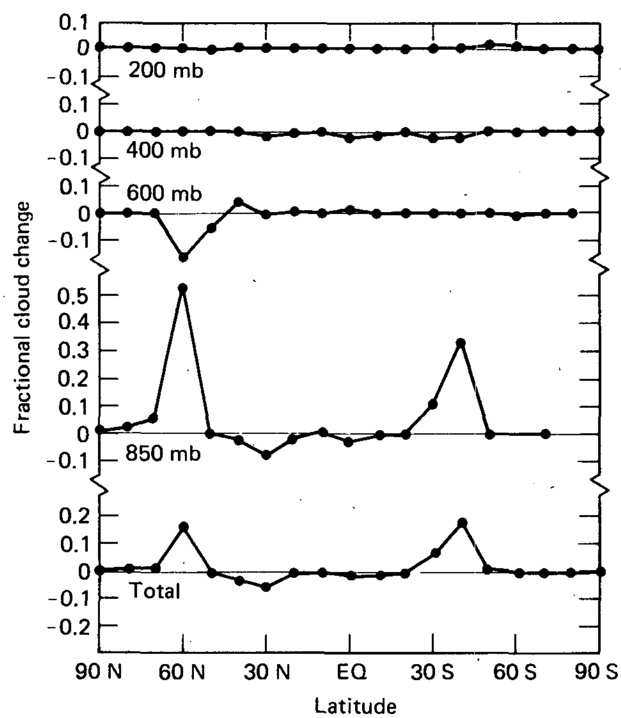

FIG. 1. Change in cloud cover due to doubled atmospheric CO_2 at each of the heights where clouds are computed. The lowest line is the change in total cloud cover.

60°N from 850–400 mb (Fig. 2) indicating that convection was more active below and less active above 850 mb, trapping and increasing water vapor and therefore clouds at 850 mb.

The cloud increase at 40°S (850 mb) is not as easily explained. We suspect that the reason is much the same as for 60°N except that the Antarctic land mass in the Southern Hemisphere has the effect of pushing the transition zone between the midlatitudes and subtropics equatorward. Although the displacement of this zone appears rather exaggerated, comparison of the three sets of curves in Fig. 2 shows that the temperature differences in the Northern and Southern Hemispheres become more symmetric higher in the atmosphere ($\Delta T_{600-400}$ mb). Therefore, the asymmetric response in cloud cover due to increased CO_2 is likely due to the asymmetric surface boundary conditions (i.e., differences in fractional ocean/land/mountain distribution between the Northern and Southern Hemispheres). A similar difficulty in interpretation exists in the $4 \times CO_2$ case in Potter (1980) where these transitional latitudes also caused anomalous temperature responses.

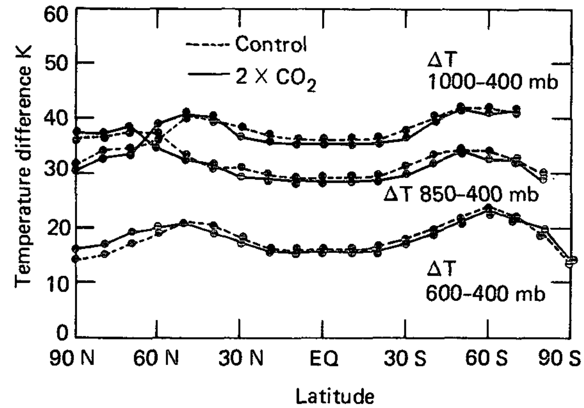

FIG. 2. Temperature difference between 1000 and 400 mb; 850 and 400 mb, and 600 and 400 mb. Dashed line is control and solid line is for the $2 \times CO_2$ experiment.

4. Contribution of various mechanisms to the change in the planetary energy budget

One method of assessing the sensitivity of the longwave loss to cloud amount change ($\partial F/\partial A_c$) from external forcing is to calculate the contribution of individual components to the total change in longwave loss. This can be done by comparing the equilibrium state of the model for the control and doubled atmospheric CO_2 concentrations ($2 \times CO_2$) runs. The change in the net radiation at the top of the atmosphere (δR) is given by

$$\delta R = \delta S - \delta F, \quad (3)$$

where δS is the change in the net solar flux and δF the change in outgoing longwave flux at the top of the atmosphere. These changes may be expressed in terms of the primary individual contributing factors, such that

$$\delta S = \delta_r S + \delta_{A_c} S + \delta_\alpha S, \quad (4)$$

$$\delta F = \delta_{CO_2} F + \delta_T F + \delta_r F + \delta_{A_c} F, \quad (5)$$

where the subscripts r, CO_2, T, A_c and α represent water vapor mixing ratio, CO_2, temperature, cloud amount and surface albedo, respectively.

These individual contributions have been calculated in the following way: 1) both the control and perturbed annual-average climates were integrated until an equilibrium climate was reached; 2) using only the short and longwave radiation routines of the zonal model, individual sets of data fields generated by the perturbed model run (e.g., T, r, A_c, α or CO_2) were substituted into the control data set and used to calculate new radiative conditions (all other data fields being held fixed). The results of each of the five radiation calculations are presented in Table 2.

5. Discussion

Of primary interest in this paper are the effects of changes in cloud amount on the outgoing longwave radiation at the top of the atmosphere. Since A_c increased in the $2 \times CO_2$ experiment, Table 2 indicates that an increase in cloud amount caused the net solar radiation to decrease (reflecting more solar radiation back to space), but the outgoing longwave radiation at the top of the atmosphere to increase. Since clouds act to keep the longwave radiation emitted below the clouds from escaping to space, we might expect an increase in cloud amount to reduce the longwave loss. Because the model's global results show the opposite tendency, a fuller explanation must be sought.

For the $2 \times CO_2$ run, cloudiness generally increased in the mid and high northern latitudes and in the mid-southern latitudes. Since the model computes the albedo of the three lowest cloud layers as a function of zenith angle [higher zenith angle leads to higher albedo based on Cess (1976)], planetary albedo is more sensitive to cloud increases in high latitudes than in low latitudes. In the lower latitudes where cloud cover decreased very slightly, the local planetary albedo also decreased, but insufficiently to compensate globally for the increase in albedo at higher latitudes. This overall reduction in albedo occurred even though the area of the decrease in cloud cover was substantially greater than the area where the increase occurred. Thus, as expected, the effect of a slight increase in global cloud cover was to make $\delta S < 0$.

The lower latitudes, which have a higher radiating temperature, experienced a decrease in cloud cover and therefore an increase in outgoing longwave ra-

TABLE 2. Change in solar, longwave and net radiation [S, F and $(S - F)$ in W m^{-2}] from the control model due to the separate calculations of the perturbation changes in various contributing factors, i.e., T, CO_2, r, A_c and α. Solar (longwave) radiation changes are positive if the earth gains (loses) energy.

	$\delta_x y$				
y	T	CO_2	r	A_c	α
S	—	—	0.34	−0.57	0.39
F	3.53	−2.55	−1.74	0.76	—
$(S - F)$	−3.53	2.55	2.08	−1.33	0.39

diation. In a similar manner, the increase in midlatitude cloudiness decreased F. Consequently, locally the effect of changes in cloud amount on F was opposite to the global effect. Because the increase in F in the lower latitudes more than compensated for the decrease in the midlatitudes, there was a net global increase in F although there was a net increase in $A_c(\partial F/\partial A_c > 0)$. By using the control and perturbed cases to calculate the individual latitudinal differences for F and A_c we can see the effect of different averaging approaches. Table 3 shows δF, δA_c and $\partial F/\partial A_c$ using the two climate states (control and $2 \times CO_2$) to calculate the differences. The area weighted average of δF is 0.76 W m^{-2} (the same as in Table 2), the area weighted average of δA_c is 0.01, but the area weighted average of $\partial F/\partial A_c$ is not the same as the value that is calculated by taking the ratio $\overline{\delta F}$ to $\overline{\delta A_c}$. The latitudinal redistribution of cloud amount is the primary reason for this feature of the model response.

The positive values of $\partial F/\partial A_c$ at 90 and 60°N in Table 3 indicate that the cloud-amount increased and the cloud-heights lowered. This had the effect of relocating the clouds at a lower, warmer atmospheric level and thus increasing the longwave flux to space at those particular latitudes. In addition, each latitude appears to have a unique $\partial F/\partial A_c$, although the primary reason for the latitudinal variation in this parameter is the variation in cloud height.

These results indicate that one cannot relate climatic change simply to variations in global mean cloud cover. One must also allow for the latitudinal redistribution of the cloud amount and change in cloud height in considering effects on S and F. When $\partial F/\partial A_c$ was calculated from the control and $2 \times CO_2$ simulations based on individual zonal statistics similar to the regression calculations of Budyko (1974) and Cess (1976), the general longwave sensitivity remained nearly the same, i.e., $\partial F/\partial A_c(\text{control}) \approx \partial F/\partial A_c(2 \times CO_2)$.

6. Conclusions

Our zonal atmospheric model appears to respond to solar constant changes in much the same manner as described by Wetherald and Manabe (1980). The $\partial F/\partial A_c$ term calculated from a global climate change ($2 \times CO_2$) gave a somewhat surprising value that would not normally be considered valid based on the traditional view concerning the effect of clouds on the longwave radiation. However, a closer look provides insight into why the model acted as it did.

These results suggest that with a major climate change, as envisioned with doubled CO_2, clouds may be redistributed latitudinally and by level and the associated longwave flux to space may change sufficiently to alter the perceived model sensitivity even though the *relationship* between clouds, shortwave and longwave radiation remains relatively constant at each latitude. However, when the values of $\partial F/\partial A_c$ were calculated at individual latitudes and then averaged to obtain a global value, the global longwave sensitivity is then consistent with that calculated using the regression approach utilized by Cess (1976) and Budyko (1974).

In general, this analysis helps confirm that the sensitivity parameters as pointed out by Cess (1976) are consistent between models of different dimensionality. It also focuses on the question of the global redistribution of clouds, which may be a significant feedback process that could profoundly affect regional climatic patterns.

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TABLE 3. Values of δF, δA_c and $\partial F/\partial A_c$ by latitude. The control and $2 \times CO_2$ cloud amounts were used to calculate the contribution of clouds to the change in F.

Latitude	δF (W m^{-2})	δA_c	$\partial F/\partial A_c$
90°N	0.009	0.007	1.33
80°N	−0.021	0.010	−2.09
70°N	−0.514	0.012	−42.81
60°N	1.904	0.153	12.44
50°N	1.457	−0.008	−182.18
40°N	0.367	−0.035	−10.49
30°N	4.580	−0.061	−75.07
20°N	0.953	−0.016	−59.56
10°N	0.697	−0.003	−232.43
0°	2.930	−0.027	−108.47
10°S	2.278	−0.018	−126.55
20°S	0.442	−0.005	−88.32
30°S	−0.349	0.060	−5.81
40°S	−5.416	0.180	−30.09
50°S	−0.277	0.002	−138.30
60°S	0.260	−0.001	−260.33
70°S	0.198	−0.001	−197.58
80°S	0.379	−0.003	−126.28
90°S	−0.002	0.001	−2.32
Global average	$\overline{\delta F} = 0.763$	$\overline{\delta A_c} = 0.010$	$\partial F/\partial A_c = -99.05$

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