Cloudiness as a Global Climatic Feedback Mechanism: The Effects on the Radiation Balance and Surface Temperature of Variations in Cloudiness

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ABSTRACT

The effect of variation in cloudiness on the climate is considered in terms of 1) a relation between the radiation balance of the earth-atmosphere system and variations in the amount of cloud cover or effective cloud top height, 2) the effect on the surface temperature of variations in cloudiness, and 3) the dynamic coupling or "feedback" effects relating changes in surface temperature to the formation of clouds. The first two points are studied by numerical integration of a simple radiation flux model, and the third point is discussed qualitatively. Global-average radiation balance calculations show that an increase in the amount of low and middle level cloud cover (with cloud top height and cloud albedo fixed) decreases the surface temperature. But, this result for the global-average case does not hold near polar regions, where the albedo of the cloudy areas can be comparable to (or even smaller than) the albedo of the snow-covered cloudless areas, and where, especially in the winter season, the amount of incoming solar radiation at high latitudes is much less than the global-average value of insolation. The exact latitude at which surface cooling changes to surface warming from a given increase in cloud cover amount depends critically upon the local values of the cloud albedo and the albedo of the cloudless areas that are used in the calculation. However, an increase in effective cloud top height (with cloud cover and cloud albedo fixed) increases the surface temperature at all latitudes.

1. Introduction

The importance of understanding in what ways cloudiness might act as a climatic component has been accentuated recently by concern over the possibility of inadvertent modification to the global climate by man's activities (SMIC Report, 1971). For example, a 2K increase in global surface temperature has been predicted by the radiative-convective model of Manabe and Wetherald (1967) for a doubling in the concentration of atmospheric CO_2. Also a 3.5K decrease in global surface temperature has been suggested by the calculations of Rasool and Schneider (1971) for a factor of 4 increase in atmospheric aerosols. Neither of these studies, however, considered the possibility of any *coupled effects* from simultaneous variations in cloudiness. Could the effects on the global temperature related to increases in these pollutants be offset by a simultaneous change in cloudiness of only a few percent? This probability had been suggested by Möller (1963). Alternatively, could these effects be enhanced by coupled variations in cloudiness? Such a possibility has been discussed by Twomey (1971).

Clouds can affect the radiation balance of a vertical column through the earth-atmosphere system (taken to mean the atmosphere and underlying surface together) by reflecting a large amount of the incoming sunlight back to space, since, on a global average, the albedo of the cloudy part of the earth is about 0.5, compared to the mean albedo of the cloudless fraction of the earth of $\sim$0.14 (London and Sasamori, 1971).

Another well-known consequence of the presence of clouds in the atmosphere is a change in the upward flux of infrared radiation associated with atmospheric temperature. The temperature of the lower atmosphere is observed to decrease vertically at a nearly constant lapse rate. In the troposphere where a substantial amount of clouds exist, the lapse rate is about -6.5K km^{-1}, if *averaged over the whole globe* (Manabe and Wetherald, 1967). Since most clouds are very opaque to the planetary IR radiation, both an increase of the horizontal extent of the cloud coverage and increase of the effective height of the cloud tops reduce the upward flux of the IR radiation escaping from the earth-atmosphere system to space. Thus, there are two competing opposite effects on the global radiation balance from an increase in the *amount* of global cloud cover: 1) an increase in the planetary albedo causing a decrease in available solar energy, and 2) a decrease in the IR radiative loss to space.

The purpose of this paper is to analyze the magnitude of the effect of changes in cloudiness on the radiation

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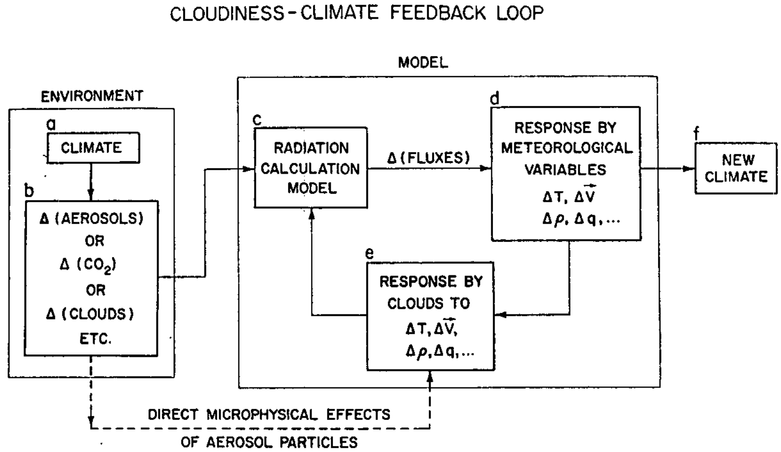

FIG. 1. Flow chart illustrating the possible role of cloudiness as a climatic feedback mechanism. The arrows represent the order in which calculations would be made by a climate model attempting to predict the effect of changes in environment (e.g., $\Delta(CO_2)$ on the climate).

balance by using a simple atmospheric model in which the cloud tops are regarded as a reflecting layer for solar radiation and as a blackbody radiator for the infrared atmospheric radiation. Variations in cloudiness are prescribed independently from the state of the atmosphere in order to extract the rate of change of the net radiative balance as a function of a given perturbation in cloudiness. The magnitude of this can then be compared to the possible effects of carbon dioxide and aerosols. The effect of an increase in the global-average amount of cloudiness on the radiation balance as computed in this study is in general agreement with the results already obtained by Manabe and Wetherald (1967). However, we have extended the calculations to include the effect of variations in cloud top height on the radiation balance, and also we have computed as a function of latitude the local departures from the global-average results.

In real atmosphere, however, the global distribution of cloudiness is itself a consequence of the general circulation which, in turn, is driven by virtue of the latitudinal radiative imbalance between the incoming solar and the outgoing atmospheric IR radiation. Thus, in a study of the sensitivity of the global climate to changes in cloudiness, it is first necessary to determine the initial magnitude of such changes in cloudiness on the radiation balance. After this, however, it must be established how other coupled *simultaneously* interacting processes, or "feedback mechanisms," might act to dampen ("negative" feedback) or amplify ("positive" feedback) that initial effect on the climate. Perhaps, at this point it might be useful to illustrate the role of cloudiness as a climatic feedback mechanism with a simplified flow chart (Fig. 1). What is "flowing" on the chart is the order in which changes in the environment are encountered and the immediate effects of these changes, first on the radiation fluxes, and then on the meteorological variables calculated by our model. Box a on the left side of Fig. 1 is labeled climate, by which is meant all the meteorological variables that describe the climate at present. Then, a perturbation (Box b) to the original environment (Box a) is introduced, and a new environment is defined (Boxes a and b), from which a new climate (Box f) will eventually result. To attempt to predict whether this pertubation in environment would result in a new climate, we construct a climate model, inside of which three subsections have been identified (Boxes c, d and e). In Box c, the effect of the change in the environment [e.g., Δ (CLOUDS)] on the radiation fluxes is calculated by a *radiation model*. Then, a response by other meteorological variables [such as temperature (T), winds ($\mathbf{V}$), density (ρ), specific humidity (q), etc.] to the change in radiation fluxes caused by the change in environment is calculated in Box d, which represents the *dynamical part* of the climate model. In many past works that have attempted to relate changes in environment to the climate, the flow of the calculations has ended here at Box d and the "new climate" has been inferred directly from Box d. For example, neither the calculations of Rasool and Schneider (1971) for the effect of Δ(AEROSOLS) on ΔT nor the radiative-convective model of Manabe and Wetherald (1967), for the calculation of the effect of increases in CO_2 on surface temperature, included the possible feedback effects of clouds on the radiation calculation. The latter possibility is represented in Fig. 1 by Box e, and illustrates schematically that the possible *feedback effect of cloudiness* on the climate must be included in the flow of the calculations. But, in some recent studies of the general circulation (e.g., Kasahara and Washington, 1971; Arakawa *et al.*, 1968), the global cloud distribution has been computed as a function of the hydrological cycle associated with the atmospheric general circulation. However, the computation of cloudiness in these models has been based upon parameterization techniques in which the amount and distribution of clouds were related only to the *large-scale* fields of relative humidity, vertical motion, and atmospheric stability.

Finally, the possibility of direct *microphysical modification* to the clouds from changes in aerosols is shown symbolically on Fig. 1 by the dashed line between Boxes b and e. Since, in reality, clouds are made up of many water droplets or ice crystals, which have condensed out onto aerosol particle nuclei, the possibility of this direct microphysical modification must also be included in the feedback loop (Twomey, 1971). However, direct microphysical effects are far too complex and occur on scales far too small to be explicitly included in large-scale climate models, and such effects can, at present, be treated only by simplified parameterizations, if they are treated at all.

The flow of the calculations discussed in this paper is primarily from Box a to Box b and through Box c. In Section 2 we introduce briefly the radiation model adopted in the present study. In the model the global mean state of the atmosphere (i.e., climate) is represented by a single temperature distribution in the vertical direction, in which the total solar absorption by the atmosphere and the surface below is equated to the infrared radiative loss from the top of the atmosphere. The model will be utilized to calculate the rate of change of the net radiative balance resulting from a unit change in the amount of cloud cover for a single layer of clouds of various heights. If we assume further that the adjustment to the temperature profile of the model atmosphere resulting from a perturbation in cloudiness takes place while the vertical temperature lapse rate assumed in the unperturbed atmosphere is conserved, we are then able to estimate the possible variation of the most important climatic variable: the temperature of the atmosphere at the earth's surface. Then, the rate of change in surface temperature can be formulated as a function of variations in cloudiness (Section 3). However, it is important to recognize that the assumption that the mean lapse rate is conserved should be classified as a dynamical consideration and, as such, it properly belongs in the domain of Box d and not in the radiation calculations of Box c.

In Section 4, the analysis will be extended to a more realistic case in which the distribution of incoming solar radiation and the distribution of vertical temperature and the amount of water vapor depend on both latitude and season. The assumption that a variation in cloudiness does not alter the slope of the atmospheric temperature profile is also retained for the zonal-average computations of Section 4, but the vertical atmospheric temperature distribution used for the zonal cases are the individual zonal climatological averages, not the global-average lapse rate of $-6.5\mathrm{K\ km^{-1}}$. However, in an individual zonal-average case, no statement about the effect on *temperature* from a change in the radiation balance can be justified with the static radiation model used in this paper. This is because any change in the zonal-average radiation balance arising from a variation in zonal-average cloudiness could be compensated for non-uniquely by either a change in the temperature of the particular zonal belt, or alternatively, by a change in some other heat balance component—such as the transport of sensible or latent heat from a neighboring zone. The radiation balance is, of course, only one component of the overall heat balance of the earth-atmosphere system. It is the heat balance that determines the climate [see, for example, Sellers (1965) for a general discussion of the relationship between the climate and the heat balance].

In what ways changes to the radiation balance resulting from variations in cloudiness might also affect the fluid dynamic motions of the system, which in turn could have additional feedback effects on the climate, is not treated here quantitatively [see the general discussions of climate feedback mechanisms in Schneider and Kellogg (1972) and Chap. 6 of SMIC (1971)]. We plan in future works, however, to study quantitatively the possible coupling between the amount of cloudiness and the atmospheric temperature, considering the subsequent radiative imbalances and the resulting changes to the circulation and to the hydrological cycle by using a closed system of the atmosphere simulated by a numerical model of atmospheric circulation [a preliminary result is reported by Schneider and Washington (1972)].

2. The model

For the evaluation of $F_\nu\uparrow$, the infrared flux [for the wavenumber interval $(\nu, \nu+\Delta\nu)$] emitted to space by the earth-atmosphere system, Eq. (1) is solved by numerical integration:

$$F_\nu\uparrow=\left\{B_\nu(T_s)\tau_{\nu s}+\int_{\tau_{\nu s}}^{1}B_\nu(T_z)d\tau_{\nu z}\right\}(1-A_c)+\left[B_\nu(T_c)\tau_{\nu c}+\int_{\tau_{\nu c}}^{1}B_\nu(T_z)d\tau_{\nu z}\right]A_c, \quad (1)$$

where T_z is the temperature of the atmosphere at any height z, $B_\nu(T_z)$ is the blackbody radiation in a frequency interval $(\nu, \nu+\Delta\nu)$ for temperature T_z, the subscript c refers to the z level of the cloud tops, the subscript s refers to the earth's surface, $\tau_{\nu z}$ is the transmission of the atmosphere to diffuse radiation of frequency $(\nu, \nu+\Delta\nu)$ between height z and infinity $(\tau_{\nu\infty}=1)$, and A_c is the fraction of sky covered by a layer of clouds. The first term in the braces is the amount of surface radiation that escapes through the atmosphere directly to space, and the second term the atmospheric emission to space. In the brackets the first term is the amount of cloud top radiation that escapes to

TABLE 1. Parameters used for the global-average model atmosphere.

Parameter	Value
Surface temperature (T_s)	288K
Surface pressure (p_s)	1013 mb
Tropospheric lapse rate ($\partial T_z/\partial z$)	−6.5K km^{-1}
Stratospheric temperature	218K
Relative humidity, surface	75%
Vertical distribution of water vapor mixing ratio (w_z)	$w_s(p_z/p_s)^4$
CO_2 amount	300 ppm

space, and the second term the atmospheric emission of IR radiation to space that originates in the fraction of the atmosphere lying above the clouds.

To perform the calculation, a "model atmosphere" is adopted, which reflects the present-day global average conditions (Table 1).
Relative humidity is kept constant rather than absolute humidity, as suggested in Manabe and Wetherald (1967).

The total infrared flux

$$F_{IR}=\int_{10}^{2400\ \mathrm{cm}^{-1}} F_\nu\uparrow d\nu,$$

leaving the earth-atmosphere system from the model described in Table 1 was computed (with 50% cloud cover and effective cloud top height of 5.5 km) to be 0.3450 cal cm^{-2} min^{-1}. The computational details are described further in Rasool and Schneider (1971) and Prabhakara and Rasool (1963). This value for the total outgoing IR radiation flux will be in balance with the incoming solar radiation for a planetary albedo α_p of 31% (based on a solar constant of 2.00 cal cm^{-2} min^{-1}). This value of α_p is close to the value observed by satellites of 30% (Vonder Haar and Suomi, 1971) and to the value of 33% computed theoretically by London and Sasamori (1971). A planetary albedo of 31% can be reconciled by assuming a value of 50% for the albedo of the cloudy fraction of the earth and a value of 12% for the albedo of the cloudless part of the earth. Although a single layer of clouds with an albedo of 50% is used here to evaluate the effect of changes in the average amount of cloudiness on the radiation balance, the real situation is, of course, more complicated. For the real earth, clouds exist in many overlapping layers, each of differing height and albedo [some recent observations of these parameters are reported by Drummond and Hickey (1971)]. Furthermore, the effective albedo of a particular cloud layer depends upon the albedos of the earth's surface and the various cloud layers below, and can be computed only by accounting for the multiple reflections between the cloud layer and underlying reflectors [e.g., as was done in an elementary way for the case of an absorbing and reflecting aerosol layer by Schneider (1971)]. In this paper the single cloud layer with effective albedo of 50% represents a statistical ensemble or global average of many cloud layers of various albedos. In addition, the effect of variations in cloud amount on the local radiation balance will depend upon the local average solar zenith angle. This dependence is particularly important for large zenith angles typical of polar regions. Nonetheless, a single cloud layer is still quite useful in studying the effect of changes in the average amount of cloudiness on the radiation balance. Although the total outgoing infrared flux calculated by the radiation model is near the observed value, two other comparisons must be made to test the suitability of the radiation model for the calculation of infrared flux to space under the condition of *changing* cloud cover and tropospheric temperature. First, the sensitivity of the outgoing flux to variations in cloud amount should be evaluated, i.e., $F_{IR}(A_c)$. The curve labeled "effective cloud top height = 5.5 km" on Fig. 2 gives the value of the IR flux to space computed by the model as a function of the amount of cloud cover A_c, where the atmospheric variables in this case are for the "model atmosphere" as given in Table 1. As can be seen on Fig. 2, F_{IR} decreases with A_c. This decrease occurs since the increase in cloud cover substitutes as a radiator of blackbody radiation the colder cloud tops for the warmer underlying surface. The figure shows a linear decrease in infrared flux with an increase in cloud amount, and for the 5.5-km cloud height curve the rate of decrease is

$$\frac{\partial F_{IR}}{\partial A_c}\approx -0.107. \qquad (2)$$

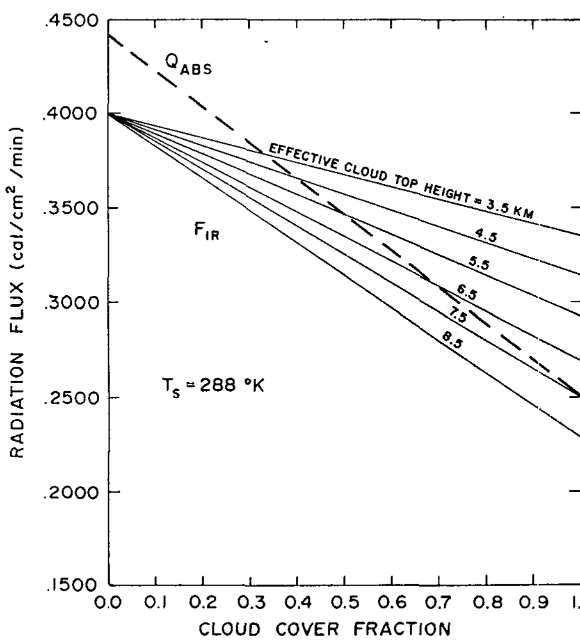

FIG. 2. The infrared flux to space, F_{IR}, emitted from the earth-atmosphere system (described in Table 1) and the absorbed solar energy, Q_{ABS}, as a function of amount of cloud cover and for several values of effective cloud top height.

Eq. (2) can be compared with the empirical formulation of Budyko (1969):

$$I_B = A + BT_s - (A_1 + B_1 T_s) A_c, \tag{3}$$

where I_B is the infrared flux to space, T_s is the surface temperature (°C), and where $A=0.319$, $B=0.00319$, $A_1=0.0684$ and $B_1=0.00228$ are empirically determined coefficients, given here in units of cal cm^{-2} min^{-1}. Applying the applicable parameters of the model atmosphere to (3) and differentiating with respect to A_c, we get

$$\frac{\partial I_B}{\partial A_c} = -0.1026, \tag{4}$$

which is in good agreement with Eq. (2), derived from the model results of Fig. 2 for the case of the model atmosphere with effective cloud top height at 5.5 km. [The agreement between (2) and (4) would be exact for an effective cloud top height a few tenths of a kilometer less than 5.5 km.] Note that in Budyko's empirical formulation, I_B decreases linearly with A_c. This can also be seen for all computed cases of F_{IR} on Fig. 2, and is implied by Eq. (1).

Another test of the accuracy of the radiation model for the purposes of this paper is the sensitivity of the calculated outgoing flux to changes in surface temperature. By differentiating (3), with A_c fixed at 0.5, we have

$$\frac{\partial I_B}{\partial T_s} = +0.0031. \tag{5}$$

By varying T_s in the radiation model [see Fig. 2a in Rasool and Schneider, (1971)] for the present-day case of 5.5 km effective cloud top height and $A_c=0.5$, we find

$$\frac{\partial F_{IR}}{\partial T_s} = +0.0033, \tag{6}$$

which is again in good agreement with the empirical results of Budyko given by (5). Thus, the results for the global-average case obtained with the numerical radiation model for the effect of variations in A_c on F_{IR} will be quite similar to those that could be obtained by use of a single empirical algebraic expression [Eq. (3)]. However, this empirical formulation [Eq. (3)] does not include the effect of changes in cloud top height on the radiation balance. Therefore, the radiation model described above has been adopted to compute the effect on the outgoing IR flux of changes in cloud height as well as in cloud cover, for both global-average and zonal-average cases.

3. Results for the global-average case

This section presents the quantitative relationship between the effective cloud top height and cloud cover fraction that is required to satisfy the condition of global radiation balance between the unreflected part of the incoming solar energy and the outgoing infrared flux to space emitted by the earth-atmosphere system.

The effect on the outgoing IR flux from an increase in the amount of cloud cover is a decrease in F_{IR} (for the fixed tropospheric temperature profile given in Table 1). This was mentioned in the previous discussion of Fig. 2. Furthermore, Fig. 2 shows that the decrease in F_{IR} arising from an increase in A_c is more pronounced for a higher effective cloud top height than for a lower effective cloud top height. This, of course, occurs because the higher cloud tops are colder than the lower ones, and thus emit proportionately less blackbody radiation than the lower clouds or the earth's surface would in their absence.

In order to achieve global radiative equilibrium, the outgoing flux F_{IR} must be equal to the solar energy absorbed in the earth-atmosphere system, Q_{ABS}. The latter depends upon the solar constant Q_{sc}, the albedo of the cloudy fraction of the earth, α_c, and the albedo of the cloudless fraction of the globe, α_s, as follows:

$$Q_{ABS} = \frac{Q_{sc}}{4}(1-\alpha_p), \tag{7}$$

where the total planetary albedo α_p is

$$\alpha_p = \alpha_c A_c + \alpha_s (1 - A_c). \tag{8}$$

As described earlier, α_c is taken to be 0.50 and α_s is assumed in these calculations to be 0.12 for the global average case. For these values of Q_{sc}, α_c and α_s, Q_{ABS} decreases linearly with A_c from 0.4400 cal cm^{-2} min^{-1} with no cloud cover ($\alpha_p=\alpha_s$) to 0.2500 cal cm^{-2} min^{-1} at 100% cloudiness ($\alpha_p=\alpha_c$).

Thus, if an increase in the *amount* of cloud cover occurs, there are two competing factors at work: a decrease in Q_{ABS} and a decrease in F_{IR}. Fig. 2 shows that the *rate* of decrease of solar absorption (i.e., an increase in α_p) with increased cloud amount is faster than *rate* of decrease of infrared flux to space [i.e., $|(\partial Q_{ABS})/(\partial A_c)| > |(\partial F_{IR})/(\partial A_c)|$]. Therefore, these calculations show that for realistic values of the globally averaged parameters, Q_{sc}, α_c and α_s, the net effect on the radiation balance of an increase in the amount of cloud cover (with cloud top height fixed) is a radiative imbalance that causes F_{IR} to exceed Q_{ABS}. When $F_{IR} > Q_{ABS}$ the radiative equilibrium temperature of the earth will eventually be reduced in order to restore the balance: $F_{IR}=Q_{ABS}$. Assuming that radiative and convective processes would be able to conserve the currently observed tropospheric lapse rate of -6.5K km^{-1}, *then the effect of a sustained increase in the average amount of cloud cover of the earth would be a decrease in the global-average surface temperature, provided that the cloud top height and cloud albedo remain unchanged.* This result is in agreement with the global-average results depicted in Fig. 20 of Manabe and Wetherald (1967) for the cases of the relatively bright (i.e., $\alpha_c \gg \alpha_s$) low and middle level clouds. For high, thin cirrus clouds, in which the cloud albedo parameter is not several times larger than α_s, and in which the IR emissivity of the cirrus clouds can be considerably less than unity [Hunt (1972) suggests that this emissivity is always less than 0.7], Manabe and Wetherald show that no *general* statement can be made as to the effect of an increase in cirrus cloud cover on the surface temperature in the absence of better knowledge of the numerical values of these parameters. Better statistical information on the visible albedo of cirrus clouds, their transmittance, the albedo of the atmosphere and surface below them, their height, and their relative "blackness" to IR radiation would be required to determine even the algebraic sign of any possible change in surface temperature occurring as a result of an increase only in the amount of cirrus cover. The theoretical study of Liou (1972) or the observations of Cox (1971), Platt (1971), or Paltridge and Sargent (1971) are some examples of work being done in this regard.

Another possible "change" in cloudiness is a variation in the average or effective height of the cloud tops. Fig. 2 shows that the net effect on the radiation balance of an increase in the effective height of the cloud tops (with the amount of cloud cover or the cloud albedo remaining unchanged), is a radiative imbalance that causes Q_{ABS} to exceed F_{IR}. By applying the arguments (about the conservation of tropospheric lapse rate) of the previous few paragraphs, Fig. 2 then shows that *the effect of a sustained global increase in the effective cloud top height would be an increase in the global surface temperature, provided that the amount of cloud cover and cloud albedo remain unchanged.*

For radiative equilibrium ($Q_{ABS}=F_{IR}$) the curve Q_{ABS} must intersect the lines F_{IR} on Fig. 2. These intersections are possible radiative equilibrium states of cloud cover amount and cloud top height. For the present-day global-average value of 50% cloud cover, Q_{ABS} intersects F_{IR} on Fig. 2 for an effective cloud top height of about 5.5 km.

The main point of the preceding discussion is that a "change in cloudiness" does not necessarily imply that a change in the radiation balance must accompany the change in cloudiness. If both the amount and height of the clouds change in such a way that Q_{ABS} still equals F_{IR}, then no change in the radiation balance need occur. This can be illustrated by cross-plotting from Fig. 2 the intersections of Q_{ABS} with various F_{IR} lines. The result is the curve labeled "$T_s=288$K" on Fig. 3. Repeating this procedure for two other surface temperatures, $T_s=290$K and $T_s=286$K, yields the other two curves seen on Fig. 3 (the other parameters of Table 1 are also applied for these cases). Each curve on Fig. 3 represents the locus of possible cloud top heights and cloud cover amounts for which radiative equilibrium is maintained and there is no change in the net radiation balance (nor, for fixed lapse rate, in surface temperature). However, the arrows on the figure show that *an increase in the*

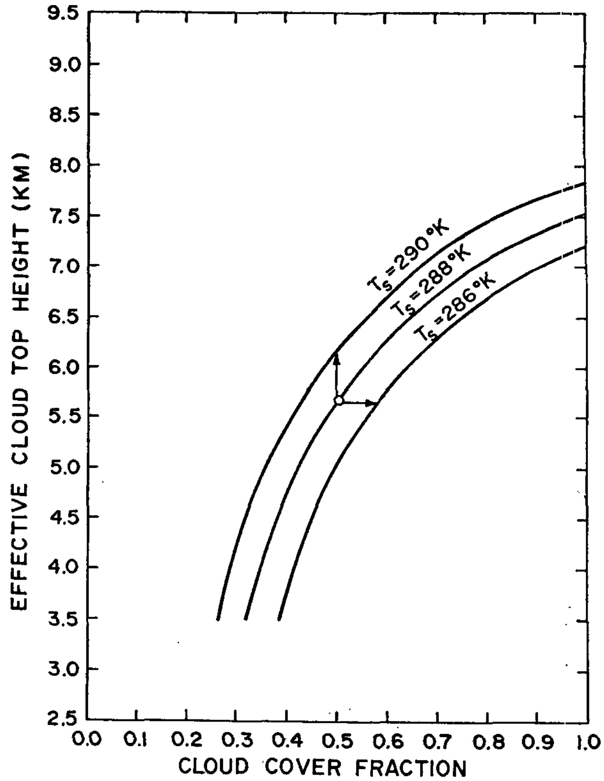

FIG. 3. A cross-plot of the intersections on Fig. 2 of the Q_{ABS} curve with the various F_{IR} curves for three values of surface temperature. Each curve represents a locus of the possible equilibrium values of cloud top heights and cloud cover amounts that are consistent with a constant value of surface temperature.

effective cloud top height by itself from the present-day value of 5.5 km to about 6.1 km could raise the surface temperature by 2K, or that an increase in cloud cover fraction per se from 50% to about 58% should decrease T_s by 2K (assuming a fixed lapse rate).

Therefore, to determine the effect of a change in cloudiness on the radiation balance, it first must be determined whether "change" means one of cloud top height or cloud cover, since Fig. 3 shows that a change in the net radia.ion balance need not occur if both the cloud top height and the amount of cloud cover vary so as to follow one of the curves[2] on Fig. 3.

Finally, assume for the moment that we could determine exactly by how much the global cloudiness might, for some reason (such as an increase in aerosols), be changed. We could then compute the effect of this change in cloudiness on the radiation balance. However, as emphasized in the Introduction, to determine the

[2] Budyko (1972, private communication) believes that this might indeed be true for the earth. In discussing a possible relationship between the cloud height and the total cloud amount, Budyko comments that "the data available (always insufficient) indicate that a positive correlation seems to exist between these values. In this case the variations of absorbed radiation and those of outgoing longwave emission would fully compensate for each other. It could decrease the effect of cloudiness on the temperature at the earth's surface, and as a result this effect for average global conditions might fall within the calculation accuracy."

eventual effect of this change in cloudiness on the global climate it would ultimately be necessary to consider the impact of other possible feedback mechanisms that simultaneously might work either to reduce or amplify the initial change in cloudiness.

4. Results for the zonally-averaged cases

In the Introduction it was stated that, based on the use of a static radiation model alone, no general statement could be made that relates variations in zonal-average cloudiness to a consequent change in the zonal-average temperature. This statement is true because a variation in the zonal-average radiation balance could be compensated for by either a change in zonal-average temperature and/or by a change in some other component of the heat budget, such as the sensible heat transport to or from neighboring zonal belts. Furthermore, even to determine the effect of variations in zonal-average cloudiness on the zonal-average radiation balance, it is necessary to specify all of the following parameters as a function of latitude: Q_{ABS}, α_s, α_c, T_z, and the vertical distribution of radiation absorbing gases—particularly water vapor. But the values of these parameters also depend upon the season. The significance of this fact for climate studies is discussed below, and has also been pointed out in an empirical study by Budyko (SMIC Report, 1971, pp. 122–122).

Here, we will compute the effect of changes in cloudiness on the local zonal-average *radiation balance* (not surface temperature) by using reasonable estimates of the above parameters for each zone, and for the summer and winter season. The results given below are not intended to be taken as quantitatively realistic, but rather are meant to show significant qualitative differences between the zonal-average and the global-average cases. It will be shown below that while the area-weighted averages of the zonal results are in qualitative agreement wlth the global-average results described earlier, the zonal-average results for the polar regions are opposite to those of the global-average, especially in the winter.

The zonal-average calculation is made for five zones in the Northern Hemisphere, where the zonal atmospheric temperatures and the zonal-average dew point temperatures (used to determine w_z) needed for the computations by the radiation model [Eq. (1)] are obtained from September 1969 NAVAIR data.

In order to calculate the net flux difference δ_i, the change in absorbed solar flux, $\partial Q_{ABS}/\partial A_c$, and the change in IR flux to space, $\partial F_{IR}/\partial A_c$, must be evaluated, since we define

$$\delta_i \equiv \left(\frac{\partial Q_{ABS}}{\partial A_c}\right)_i - \left(\frac{\partial F_{IR}}{\partial A_c}\right)_i, \tag{9}$$

where the subscript i refers to the particular zonal belt. For the ith zone,

$$(Q_{ABS})_i = Q_{s_i}[1-(A_c\alpha_{c_i}+(1-A_c)\alpha_{s_i})], \tag{10}$$

where Q_{s_i} is the zonal-average value of incoming solar radiation for the ith zone, α_{c_i} is the zonal-average albedo of the cloudy part of the zone, and α_{s_i} is the zonal-average albedo of the cloudless fraction of the zone. Eq. (10) is, of course, the zonal form of Eq. (7). Differentiating (10) yields

$$\left(\frac{\partial Q_{ABS}}{\partial A_c}\right)_i = -Q_{si}(\alpha_{ci}-\alpha_{si}). \tag{11}$$

The IR flux change $(\partial F_{IR}/\partial A_c)_i$ is computed by the radiation model using climatological atmospheric temperature and dew point data (NAVAIR) appropriate to the ith zone, and by assuming that the effective cloud top height remains fixed at 5.5 km. The fraction of the total hemispheric surface area contained in the ith zonal belt is $\bar{A}_i$. The values of the parameters and the results of the computation are given in Table 2 for January and in Table 3 for July. The most important entry in the tables is δ_i, which gives the *net* flux difference (in a 1-cm² vertical column in the ith zone) for an increase in cloud cover from $A_c=0.0$ to $A_c=1.0$. (When δ_i is negative this

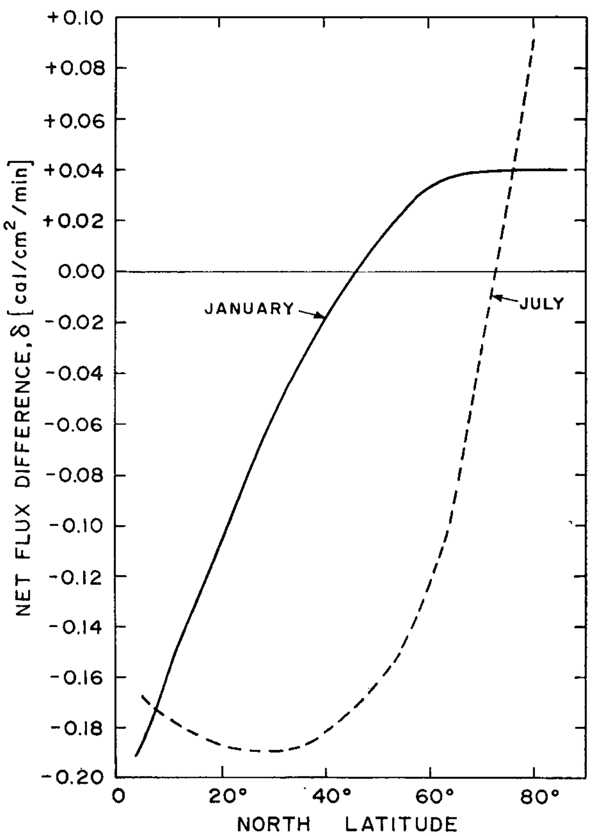

FIG. 4. The net flux difference $\delta[=(\partial Q_{ABS}/\partial A_c)-(\partial F_{IR}/\partial A_c)]$ computed as a function of north latitude. Note that although the hemispheric average value of δ is negative, δ is positive in high latitudes.

TABLE 2. Zonal-average parameters and results for January.

Zonal belt i	0–10N 1	10–30N 2	30–50N 3	50–70N 4	70–90N 5
$\bar{A}_i$	0.1735	0.3263	0.2658	0.1737	0.0604
α_{s_i}	0.08	0.095	0.11	0.19	0.535
α_{c_i}	0.5	0.5	0.5	0.5	0.5
Q_{s_i}	0.6166	0.4621	0.2649	0.0637	0.0000
$(\partial Q_{ABS}/\partial A_c)_i$	−0.2590	−0.1868	−0.0954	−0.0197	0.0000
$(\partial F_{IR}/\partial A_c)_i$	−0.0740	−0.0810	−0.0750	−0.0540	−0.0390
δ_i	−0.1850	−0.1058	−0.0204	+0.0343	+0.0390
$\delta_i\bar{A}_i$	−0.0321	−0.0345	−0.0054	+0.0060	+0.0024
$[(\partial F_{IR}/\partial z)_{4.5}^{5.5}]_i$	−0.0087	−0.0093	−0.0085	−0.0074	−0.0067

increase in cloud amount causes a decrease in the net radiational energy available to the ith zone, and when δ_i is positive there is an energy flux surplus in the ith zone.) This result is plotted in Fig. 4. Note that the effect on the net zonal radiation balance, δ_i, of an increase in cloud amount for constant cloud top height, depends strongly upon the latitude, and for polar latitudes the effect is reversed (δ_i changes from negative to positive values). A glance at Table 2 shows that the reversal in sign occurs north of $\sim 45°$ latitude because the $(\partial F_{IR}/\partial A_c)_i$ term in Eq. (9) becomes larger in magnitude than the $(\partial Q_{ABS}/\partial A_c)_i$ term. This is for two reasons: 1) Q_{s_i} is relatively small in high latitudes in winter, and 2) the difference $\alpha_{c_i}-\alpha_{s_i}$ is small or even *negative* near the poles where the albedo of the clouds can be less than the albedo of the underlying snow and ice. For the July case, Fig. 4 shows that δ_i becomes positive north of $\sim 70°$ latitude. Thus, in the summer case the reversal in the sign of δ_i occurs much closer to the poles than it does for the winter case. This is because, unlike January, the July value of Q_{s_i} is very large in polar regions and the reversal in the sign of δ_i is a result only of the assumption that $\alpha_{s_i} \gtrsim \alpha_{c_i}$ in the polar region.

The numerical values of α_{s_i} used here are the yearly average values of Sellers (1965), and do not reflect any seasonal changes in the surface albedo that might be expected in higher latitudes. If included, this seasonal effect would probably tend to make δ_i somewhat more positive in high latitudes in winter, and more negative near the pole in summer. In fact, the exact latitude of the cross-over point from which δ_i changes from negative to positive values depends critically on the relative values of the parameters α_{s_i}, α_{c_i}, Q_{s_i}. Furthermore, the albedos α_{s_i} and α_{c_i} depend upon the average solar zenith angle, particularly in polar regions. Bowling (1972)[3] has communicated that she has drawn similar conclusions from studies now in progress.

To compute the hemispheric average $\bar{\delta}$, the zonal values are area weighted as follows:

$$\bar{\delta}=\sum_{i=1}^{5} \delta_i \bar{A}_i.$$

[3] Private communication.

For January $\bar{\delta}=-0.0636$ and for July, $\bar{\delta}=-0.1544$. The average of these values yields the yearly-average, global-average value, $\bar{\delta}=-0.109$ cal cm^{-2} min^{-1}. This value of $\bar{\delta}$ is the difference between global-average absorbed solar flux and outgoing IR flux when A_c is varied from 0.0 to 1.0 for fixed T_z. From Fig. 2, the yearly-average and global average value of the net flux difference,

$$\bar{\delta}_1=\frac{\partial Q_{ABS}}{\partial A_c}-\frac{\partial F_{IR}}{\partial A_c},$$

is $\bar{\delta}_1=(0.2500-0.4400)-(0.2930-0.4000)=-0.0830$ cal cm^{-2} min^{-1}. This value is near to the area-weighted average, $\bar{\delta}=-0.109$ cal cm^{-2} min^{-1}, obtained above from averaging the zonal cases of this section. Exact agreement between these numbers cannot be expected since the zonal average computations of this section are based upon zonal and seasonal average estimates of the parameters α_{c_i} and A_{c_i} that are even less well known than the yearly- and global-averaged estimates of these parameters used earlier in the radiation computation for the global averages. Also, as pointed out above, the same values for α_{s_i} are used in the January and July cases. The uncertainty inherent in the values of all these parameters suggests that the shape of the curves in Fig. 4, especially for latitudes poleward of about 45N, should be considered only as qualitative. Nevertheless, these parameters are close enough to the actual values to permit us still to draw some interesting general conclusions from Fig. 4.

The important inference that can be taken from the zonal-average computations of this section is that while they agree with the global-average case in predicting that the effect on the *global* climate of a uniform increase in the amount of global cloud cover will be a decrease in global temperature (since $\bar{\delta}<0$), they also show that just the opposite effect should be felt locally in high latitudes, particularly in the winter months. [Cox (1971) also found that the effect of variations in cloudiness on the radiation balance could change algebraic sign with latitude. His conclusions were based in part on measurements of the infrared emissivity of cirrus clouds.] Thus, a uniform global increase in cloud cover amount could decrease the temperature in low and

TABLE 3. Zonal-average parameters and results for July.

Zonal belt i	0–10N 1	10–30N 2	30–50N 3	50–70N 4	70–70N 5
$\bar{A}_i$	0.1735	0.3263	0.2658	0.1737	0.0604
α_{s_i}	0.08	0.095	0.11	0.19	0.535
α_{c_i}	0.5	0.5	0.5	0.5	0.5
Q_{s_i}	0.5784	0.6634	0.6807	0.6421	0.6622
$(\partial Q_{ABS}/\partial A_c)_i$	−0.2429	−0.2687	−0.2655	−0.1991	+0.0232
$(\partial F_{IR}/\partial A_c)_i$	−0.0750	−0.0820	−0.0830	−0.0770	−0.0650
δ_i	−0.1679	−0.1867	−0.1825	−0.1221	+0.0882
$\delta_i\bar{A}_i$	−0.0291	−0.0609	−0.0485	−0.0212	+0.0053
$[(\partial F_{IR}/\partial z)_{4.5}^{5.5}]_i$	−0.0088	−0.0096	−0.0096	−0.0086	−0.0082

middle latitudes, while simultaneously increasing the temperature at higher latitudes. This condition would reduce the equator-to-pole temperature gradient, which is a primary driving force of the atmospheric general circulation. A possible feedback effect of cloudiness on the climate is clearly apparent from this example, since the equator-to-pole temperature contrast strongly affects the baroclinic stability of the atmospheric circulation.

The high latitudes account for only a relatively small percentage of the total hemispheric surface area. Therefore, the results for high latitudes cannot significantly change the global-average conclusions that an increase in cloud amount decreases the global temperature. Nevertheless, the polar regions may be more climatically significant than their relative surface area might suggest. It has been suggested by Budyko (1969), Sellers (1969), SMIC (1971) and others that the extent of the polar ice cover is probably highly sensitive to small changes in the heat budget. Furthermore, because of the relatively high albedo of the polar ice compared to open ocean and because of the difference in the evaporability of ice covered sea and open sea, small changes in the extent of the ice fields could have a far-reaching effect on the climate elsewhere (Donn and Ewing, 1968). Therefore, an important conclusion of this study is that *global-average models of the radiation balance are probably not sufficient to determine the average effect on the global climate of a given change in the amount of cloud cover, since the zonal-average results show that polar regions behave oppositely to the global-average, and since small changes in the heat balance of polar regions could have a substantial effect on the extent and movement of the polar ice.* The inherent non-linearities of the system, of which the polar ice feedback effect is an important example, dictate that future climate models include spatial scales considerably finer than global averages.

The last row in Tables 2 and 3 gives the change in IR flux to space computed for an increase in cloud top height from 4.5 to 5.5 km, while A_c is held fixed at a value of 0.5. The result for each zonal-average case is roughly the same as for the global-average case: a decrease in F_{IR} with increasing cloud top height. This is expected for both zonal-average and global-average cases since the atmospheric temperature at 5.5 km is colder than its temperature at 4.5 km, and in the calculations α_{c_i} has been assumed to be unchanged by the increase in cloud top height. [If, however, an increase in cloud top height were also to be accompanied by an increase in cloud thickness, then the albedo of the cloud could be increased (Twomey, 1971). This then would identify a negative feedback effect, since an increase in cloud top height *increases* the radiational energy available to a vertical column through the earth-atmosphere system by decreasing the infrared flux escaping to space, whereas an increase in the cloud albedo would *decrease* the amount of solar energy absorbed in the column.]

5. Concluding remarks

The main points that have been discussed in this paper can be summarized as follows:

1. To determine the effect of variations in cloudiness on the radiation balance of a vertical column through the earth-atmosphere system it is necessary to specify the season and geographic location of the column in question, and to establish whether the "variation in cloudiness" refers to a change in the amount of cloud cover and/or a change in the effective cloud top height or cloud albedo.
2. No general statement can be made that relates variations in cloudiness in a particular zonal belt to a consequent change in zonal-average temperature, since a variation in the zonal-average radiation balance caused by the change in cloudiness could be compensated for non-uniquely by either a change in the zonal radiative temperature and/or a change in some other component of the zonal heat balance.
3. The effect of an increase in the value of the effective cloud top height is an increase in the surface temperature, provided that all other factors remain unchanged.
4. The effect of an increase in the global-average cloud cover amount (but not necessarily of an increase in the amount of thin cirrus clouds by themselves) is a decrease in the global-average surface temperature, provided that all other factors remain unchanged.
5. The global-average result given above (point 4) does not apply in high latitudes, especially in the winter months, but, rather, the opposite effect is computed for polar regions. Since, for example, slight variations in the polar heat balance are believed to have a substantial effect on the fields of polar ice, which in turn could have a significant influence on the climate elsewhere, global-average radiation models are not sufficient to determine even the average effect on the global climate of a given change in the amount of cloud cover.
6. Better statistical data on the absorptivity, reflectivity and transmissivity of clouds to both visible and infrared radiation fluxes should be obtained, as well as improved data on the thickness, height, amount and distribution of clouds. This is especially needed for cirrus clouds.
7. The numerous, nonlinear and coupled interrelationships between the factors that determine the climate suggest that in order to understand the possible role of cloudiness as a global climate feedback mechanism, it is necessary to have a realistic, large-scale, radiative and dynamical model of the land-ocean-atmosphere system, that includes some microphysical aspects of cloud formation. Such a detailed model is clearly not presently at hand. Meanwhile, we can still learn a great deal with existing models, and even simple radiation balance calculations can be useful in studying the first-order effects of variations in cloudiness on the global climate.

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