A Diffuse Thin Cloud Atmospheric Structure as a Feedback Mechanism in Global Climatic Modeling

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ABSTRACT

This paper describes the first step in the development of another global climatic model in which the structure of the atmosphere and consequently its optical properties are dynamically coupled to the surface temperature. Rather than considering clouds as discrete entities, we structure the atmosphere as a diffuse thin cloud by utilizing the fundamental thermodynamics of the cooling of moist air of fixed surface relative humidity maintaining vertical mechanical equilibrium. Vertical convective thermal mixing is parameterized as is the amount of condensate that is "rained" out. The remaining condensate is distributed as spherical droplets by an assumed distribution function.

The modified two-stream approximation employing a Gaussian quadrature is used to solve the radiative transfer equation. The reflectivity and transmissivity of the model atmosphere and a given amount of aerosol are then calculated. These quantities, together with a parameterization of surface reflectivity to surface temperature, serve to determine the total albedo to solar radiation. The infrared flux is calculated employing the emissivity technique of Rodgers. The radiative dynamical coupling to surface temperature is such that the solar energy absorbed descreases and the emitted infrared increases with an increase in surface temperature, each with about the same magnitude of 0.0026 cal cm^{-2} min^{-1} $(°K)^{-1}$. Thus both provide stabilizing negative feedback.

In applying the diffuse cloud model atmosphere to climate assessment we have at this stage considered only global annual average surface temperature, calculating that temperature which gives radiation balance. The sensitivity of the "climate" to variations in aerosol optical density, atmospheric carbon dioxide, and the solar constant is calculated and the results are comparable to those obtained by others using very different models. In general, our model exhibits slightly greater stability.

1. Introduction

There has been in recent years a growing concern over possible inadvertent climate alteration by man's activity (SMIC, 1971; Matthews *et al.*, 1971). As a result, there has been considerable effort devoted to developing predictive global climatic models (Budyko, 1969, 1972; Sellers, 1969, 1973), or to otherwise assessing the climatic effect of atmospheric pollutants (see, e.g., Manabe, 1971; Lamb, 1970; Rasool and Schneider, 1971; Bryson, 1972; Mitchell, 1970). This effort has been useful in providing tentative predictions and has certainly stimulated more interest and even controversy. However, the climatic models have relied heavily on simplified empirical parameterizations and, in general, none of the assessments have been very inclusive of many of the earth-atmosphere dynamic feedback mechanisms. For instance, one of the most important factors potentially affecting the radiation balance of the earth-atmosphere system is clouds because of their high reflectivity in the visible spectrum and absorption-emission in the infrared. The important cloud parameters are their extent or the fractional cover, their height or temperature, and their optical thickness. Yet none of the models incorporate these parameters in any dynamical sense. Instead clouds are considered constant in extent (usually 50% covered), of fixed albedo and height according to type, and are assumed black to infrared except for the high cirrus (Manabe and Strickler, 1964). Schneider (1972) has provided a thorough discussion of cloudiness as a feedback mechanism in climatic modeling.

It is the purpose of this paper to present a simple annual global average climatic model which attempts to place quantitatively the existence of condensed water in the atmosphere in a theoretical and dynamical way related to surface temperature with a minimum number of parameters. Although in this first paper we will limit our calculations to only annual global averages, it should be clear that one may readily extend the model atmosphere to other average surface temperatures, such as zonal, or zonal and latitudinal and even seasonal, and thereby achieve an extended range of dynamical coupling. We will in this present model introduce variations in aerosol, carbon dioxide and energy input to allow a comparative assessment with other models.

We define the earth-atmosphere system as a thin spherical shell that includes all matter involved in the radiative and thermal properties on a climatic time scale. The energy balance equation is then written in terms of annual global averages, denoted $\langle\ \rangle_{g,y}$, as

$$\left\langle\frac{\partial Q}{\partial t}\right\rangle_{g,y}=\frac{S_0}{4}(1-\langle\alpha\rangle_{g,y,\cos\psi})-\langle E\rangle_{g,y}, \tag{1.1}$$

where Q is heat, t time, S_0 the solar constant, α the albedo whose annual global average is also weighted by $\cos\psi$, ψ being the zenith angle of the sun, and E the infrared emission. From the model atmosphere, which is developed in the next section, the thermodynamic quantities contributing to $\langle\alpha\rangle_{g,y,\cos\psi}$ and $\langle E\rangle_{g,y}$ may be calculated as functions of $\langle T(0)\rangle_{g,y}$, the annual global average surface temperature. Then with appropriate radiative transfer calculations the steady-state solution to Eq. (1.1), i.e., that $\langle T(0)\rangle_{g,y}$ which cause. $\langle\partial Q/\partial T\rangle_{g,y}$ to vanish, is sought.

2. Model atmosphere

The real atmosphere is either clear or cloudys Instead of considering it divided between these two states we will structure a diffuse thin cloud model atmosphere in the vertical dimension. Although this model may be viewed as a time and space average of the real atmosphere, we do not structure it in that manner. We wish to exploit the fundamental thermodynamics of cloud formation, namely the cooling of moist air to temperatures below the saturation point. We also wish to relate the principle optical properties of the atmosphere to the single variable, surface temperature $T(0)$.

Consider a volume of moist air as a two-component, potentially two-phase system. Let n_1 be the number of moles of well-mixed dry air with a fixed carbon dioxide mixing ratio assumed to exist only in the vapor phase α. Let $n_{2\alpha}$ and $n_{2\beta}$ be the moles of water vapor and condensed water, respectively. We neglect interfacial surface effects. At $z=0$, $n_{2\beta}=0$, and $n_{2\alpha}$ relative to n_1 in the volume is determined solely by $T(0)$, the surface pressure $P(0)$, and the extent of departure from equilibrium with surface water, namely the relative humidity γ. Thus

$$n_{2\alpha}(0)=\frac{P_{2\alpha}(0)}{P(0)-P_{2\alpha}(0)}n_1, \tag{2.1}$$

with

$$P_{2\alpha}(0)=\gamma P_0\exp\left\{\frac{\Delta H_1[T(0)-T_0]}{RT_0T(0)}\right\}, \tag{2.2}$$

where P_0 is the equilibrium vapor pressure of water at reference temperature T_0, ΔH_1 is the enthalpy of evaporation, and R the gas constant.

To structure the atmosphere the system is expanded quasi-adiabatically and considered to be in mechanical equilibrium in the gravitational field so that

$$dP(z)=-\rho(z)gdz, \tag{2.3}$$

where

$$\rho(z)=\frac{n_1M_1+n_2(z)M_2}{n_1+n_2(z)}\rho'(z)=\bar{M}(z)\rho'(z), \tag{2.4}$$

with M the molecular weights, $\bar{M}(z)$ the mean molecular weight, $n_2(z)=n_{2\alpha}(z)+n_{2\beta}(z)$, and g the acceleration of gravity. Assuming the vapor phase behaves as an ideal gas and that the volume $V_\beta\ll V_\alpha$,

$$\rho'(z)=\frac{[n_1+n_2(z)]P(z)}{n_\alpha(z)RT(z)}, \tag{2.5}$$

with

$$n_\alpha(z)=n_1+n_{2\alpha}(z).$$

For a truly adiabatic process the entropy S of the system is constant. However, we parameterize a quantity of entropy exchange with the environment such that

$$TdS=AdT(z)\neq 0, \tag{2.6}$$

adjusting A to make $-dT(z)/dz=6.5\text{K km}^{-1}$, the standard atmosphere lapse rate, for temperatures above the dew point. The parameter A may be regarded as relating to the vertical convective heat flow.

Assuming the system to be in heterogeneous equilibrium as a function of z, P or T, the Gibbs equation, together with Eqs. (2.3)–(2.6) and the differential form of Eq. (2.2), suffice to determine $dT(z)/dz$, $dn_{2\beta}(z)/dT(z)$, and the densities $\rho_1(z)$, $\rho_{2\alpha}(z)$ and $\rho_{2\beta}(z)$. We have

$$\frac{dT(z)}{dz}=-\frac{\left[\frac{n_{2\alpha}(z)\Delta H_i}{n_\alpha RT(z)}+1\right]\left[\frac{n_{2\beta}(z)}{n_\alpha(z)}+1\right]\bar{M}(z)g}{c_p+\frac{n_{2\beta}(z)}{n_\alpha(z)}c_\beta-\frac{A}{n_\alpha(z)}+\frac{n_{2\alpha}(z)}{n_1R}\left[\frac{\Delta H_i}{T(z)}\right]^2}, \tag{2.7}$$

$$\frac{dn_{2\beta}(z)}{dT(z)}=\frac{n_\alpha(z)\left[c_p-\frac{\Delta H_i}{T(z)}\right]+n_{2\beta}(z)c_\beta-A}{\Delta H_i+\frac{n_1}{n_{2\alpha}(z)}RT(z)}, \tag{2.8}$$

where ΔH_i for $i=2$ is the molar enthalpy of sublimation, c_p is the molar heat capacity of the vapor phase at constant pressure, and c_β the molar heat capacity of the condensed phase (assumed the same for liquid and ice). At temperatures larger than the dew point, Eq. (2.7) reduces to

$$\frac{dT(z)}{dz}=-\frac{\bar{M}g}{c_p-\frac{A}{n_\alpha}}. \tag{2.9}$$

We now introduce a third parameter, f_r, which is the fraction of condensate removed from the system, or "rained" out. From the remaining fraction, $1-f_r$, $\rho_{2\beta}(z)$ may be calculated. It is assumed that the condensed phase exists as spherical droplets whose number density size distribution is given by the function (Deirmendjian, 1964)

$$N(z,r)=\beta(z)r^6\exp(-1.5r), \quad (2.10)$$

with

$$\beta(z)=\frac{3\rho_{2\beta}(z)}{4\pi\int_0^\infty r^9\exp(-1.5r)dr}, \quad (2.11)$$

where r is the drop radius (μm).

Since neither (2.7) nor (2.8) can be integrated analytically, numerical methods are used. We have found it convenient to employ analytic integrals over a finite-difference interval, treating the slowly varying variables as constant and then updating these variables and reiterating. Calculations are made up to about the 100-mb pressure level and the residual quantities are treated as if distributed at that temperature. We turn our attention now to radiative transport.

3. Radiative transport calculations

In this section we outline the radiative calculations employing the variables, $T(z)$, $P(z)$, $\rho_1(z)$ [for carbon dioxide], $\rho_{2\alpha}(z)$, $\rho_{2\beta}(z)$ and $N(z,r)$, obtained from the diffuse cloud model atmosphere derived in the previous section. In addition, we introduce aerosols into the atmosphere and consider ozone as a fixed emitter in the region of the tropopause. We also introduce a parameterization of the surface reflectivity as a function of $T(0)$.

The starting point is the integro-differential radiative transfer equation specialized for plane parallel geometry (Chandrasakar, 1950) from which we wish to derive expressions to calculate the total albedo, $\langle\alpha\rangle_{g,y,\cos\psi}$ and $\langle E\rangle_{g,y}$ appearing in Eq. (1.1) both as functions of $\langle T(0)\rangle_{g,y}$. In the visible spectrum we omit the Planck source term and consider that radiation is only nonconversatively, anisotropically scattered, whereupon the modified two-stream Gaussian quadrature approximate solution may be readily obtained (Sagan and Pollock, 1967). Employing the single boundary condition, $t(\tau)=1$, $\tau=0$, where τ is the optical depth, the transmissivity $t(\tau)$ and reflectivity $r(\tau)$ are

$$t(\tau_e)=\frac{4u}{(u+1)^2\exp(\tau_e)-(u-1)^2\exp(-\tau_e)}, \quad (3.1)$$

$$r(\tau_e)=\frac{(u^2-1)[\exp(\tau_e)-\exp(-\tau_e)]}{(u+1)^2\exp(\tau_e)-(u-1)^2\exp(-\tau_e)}, \quad (3.2)$$

$$u^2=\frac{1-\omega_0+2b\omega_0}{1-\omega_0}, \quad (3.3)$$

$$\tau_e=\sqrt{3}u(1-\omega_0)\tau(z), \quad (3.4)$$

where ω_0 is the single-scatter albedo and b the asymmetry factor.

The optical depth of the diffuse cloud to visible radiation may now be evaluated, in that for spherical droplets

$$\tau(z)=\int_0^z \kappa_\nu\rho_{2\beta}(z)dz=\pi\int_0^z\int_0^\infty r^2Q(\alpha)N(z,\alpha)d\alpha dz, \quad (3.5)$$

where

$$\alpha=\frac{2\pi r}{\lambda}, \quad (3.6)$$

and where $Q(\alpha)$ is the scattering-absorption efficiency; $Q(\alpha)$ has been determined by Mie scattering theory, and tabulated by Twomey and Howell (1965) assuming an index of refraction of 1.33 (we assume for simplicity that ice and water droplets are identical). $N(z,r)$ is given by Eq. (2.10).

The transmissivity and reflectivity of a given density of aerosol is evaluated separately as though it occurred in a clear atmosphere or a separate layer. It is independent of the surface temperature $T(0)$.

The total albedo is finally evaluated by considering multiple reflection and transmission of the aerosol layer and diffuse cloud layer above a given earth surface reflectivity. We assume an arithmetic average of the two situations, aerosol above the diffuse cloud and aerosol below the cloud, as being representative of their being mixed. With r_c, r_a and r_s designating the reflectivities of cloud, aerosol and surface, respectively, and t_c and t_a the corresponding transmissivities, the total albedo α is

$$\alpha=\tfrac{1}{2}(R_{ca}+R_{ac}), \quad (3.7)$$

with

$$R_{ca}=r_c+\frac{t_c^2[r_st_a^2+r_a(1-r_ar_s)]}{(1-r_ar_s)(1-r_cr_a)-r_cr_st_a^2}, \quad (3.8)$$

and R_{ac} a similar expression obtained by interchanging the subscripts a and c. The albedo so evaluated is a function of $T(0)$ and we assume it to be the $\langle\alpha\rangle_{g,y,\cos\psi}$ appearing in (1.1).

To calculate the infrared emission we assume that the scattering source term in the integro-differential transfer equation may be neglected in comparison to the absorption-emission term. This is justified in that Yamamoto *et al.* (1970) and Hunt (1973) have shown that scattering in clouds comprises only 3–5% of the total interactions. Furthermore, the amount of condensed water in the diffuse cloud is only about 1% of the total. This allows us to treat the condensed

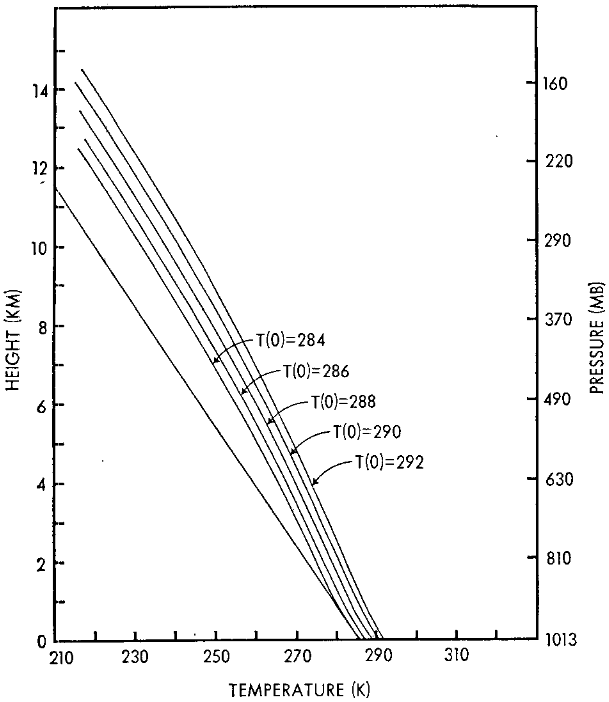

FIG. 1. The temperature-height profile for the model atmosphere at the set of surface temperatures indicated. A slope of -6.5K km^{-1} is also shown.

water as vapor with only small error. We then employ the emissivity formulation of Rodgers (1967) to calculate the radiation fluxes in the presence of water and carbon dioxide. Overlap of bands of these two absorbers is handled as by Rodgers. Ozone is treated separately and is assumed to have a fixed emissivity similar to that calculated by Staley and Jurica (1972) and to be all at the 100-mb temperature.

Since the effect of snow and ice cover is to greatly increase the earth surface reflectivity, forming the principle positive feedback mechanism of some of the global climatic models, we now introduce our own parameterization of r_s as a function of $\langle T(0)\rangle_{g,y}$. We divide the earth surface into a south and north polar ice areas, both having surface reflectivity of $r_{\rm ice}$, and all the region in between with a surface reflectivity r_0. We parameterize the mean latitudinal extent of ice cover as

$$\langle\theta\rangle_S = 65-0.1[\langle T(0)\rangle_{g,y}-288], \tag{3.9}$$

$$\langle\theta\rangle_N = 65-2.0[\langle T(0)\rangle_{g,y}-288], \tag{3.10}$$

where 65 is the present latitudinal extent and 288 the present value for $\langle T(0)\rangle_{g,y}$. The slope is determined from ice-extent, ocean-level relations suggested by Hollin (1962) and the temperature, ocean-level data of Mörner (1973) and Fairbridge (1972) for the South Pole and from Budyko (1972) for the North Pole. From (3.9) and (3.10) we have

$$\langle\theta\rangle = \tfrac{1}{2}(\langle\theta\rangle_S+\langle\theta\rangle_N). \tag{3.11}$$

It follows that

$$\langle r_s\rangle = [1-\sin(\langle\theta\rangle)]r_{\rm ice}+[\sin(\langle\theta\rangle)]r_0, \tag{3.12}$$

and this value is used in (3.8) to arrive at the total albedo.

4. Results

The values for the various quantities and parameters used in the calculations are given in Table 1. We

TABLE 1. Values of the thermodynamic and radiative constants and the various parameters employed in the calculations.

Symbol	Quantity	Value	Dimensions
S_0	solar constant	1.95	cal cm^{-2} min^{-1}
ϵ	surface emissivity	0.95	
σ	Stefan-Boltzmann constant	8.126×10^{-11}	cal cm^{-2} min^{-1} (°K)$^{-4}$
γ	relative humidity at surface	0.75	
P_0	reference vapor pressure of water	145.8	cal m^{-3}
T_0	reference temperature for P_0	273.16	°K
c_p	heat capacity of air at constant pressure	6.96	cal mol^{-1} (°K)$^{-1}$
c_β	heat capacity of condensed water	18.0	cal mol^{-1} (°K)$^{-1}$
ΔH_1	enthalpy of evaporation of water	10,751	cal mol^{-1}
ΔH_2	enthalpy of sublimation of ice	12,186	cal mol^{-1}
R	gas constant	1.99	cal mol^{-1} (°K)$^{-1}$
A	entropy transfer parameter	3.7	cal mol^{-1} (°K)$^{-1}$
g	acceleration of gravity	980	cm sec^{-2}
M_1	molecular weight of dry air	29	gm mol^{-1}
M_2	molecular weight of water	18	gm mol^{-1}
f_r	"rain" out parameter	0.9998766	
ω_0	single-scattering albedo		
	water drops	0.999	
	aerosol	0.95	
b	asymmetry factor		
	waterdrops	0.078	
	aerosol	0.18	
$r_{\rm ice}$	surface ice reflectivity	0.60	
r_0	surface reflectivity (non-ice)	0.07	
$P(0)$	pressure at surface $z=0$	1013	mb
$\tau_{\rm aerosol}$	optical depth of present day aerosol	0.1	
ϵ_{O_3}	ozone emissivity	0.072	
	CO_2 mixing ratio	0.00049	gm gm^{-1}

present first those results which illustrate the structure of the diffuse cloud model atmosphere in terms of the temperature and water content and how these variables are affected by $T(0)$. In Fig. 1, the temperature is plotted as a function of height for five surface temperatures at 2K intervals centered about 288 all at the same surface relative humidity of 0.75. Also shown is the standard atmosphere lapse rate line. The heat released on condensation above dew point heights is evident by a change in slope which returns toward the $-6.5\mathrm{K\ km^{-1}}$ lapse rate at the higher altitudes. The lines diverge due to the greater absolute humidity at the surface with increasing temperature and the subsequent release of larger amounts of latent heat.

In Fig. 2 is plotted the density of condensed water as a function of height, again at the five surface temperatures 284–292K and constant relative humidity. A value of 0.9998766 for the "rain" parameter f_r has been used. The diffuse cloud has an abrupt lower bound but its density asymptotically approaches zero with height. The slight change in curvature of each of the curves between 2000 and 4000 m is due to a shift from ΔH_1 to ΔH_2, i.e., enthalpy of evaporation to sublimation. Fig. 3 depicts the mixing ratio of total water to air as a function of height at the set of surface temperatures. The strong dependency on surface temperature should be noted. Fig. 4 compares the empirical mixing ratio function employed by

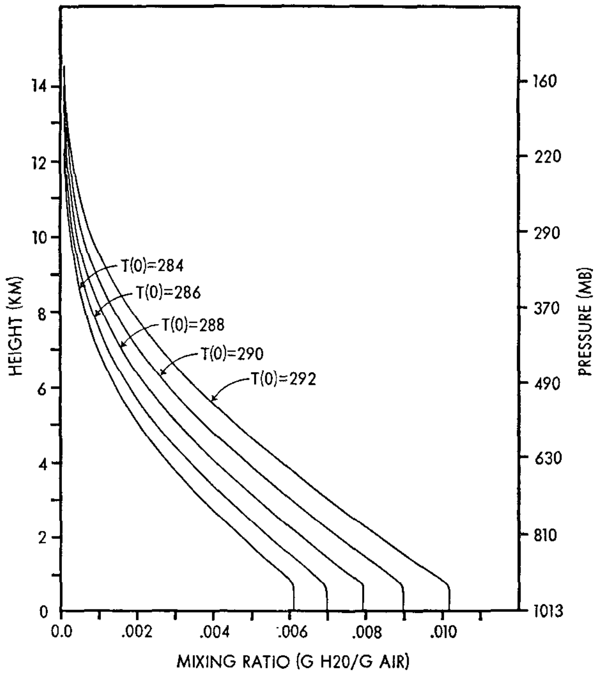

Fig. 3. The total water mixing ratio as a function of height for the model atmosphere at the set of surface temperatures indicated.

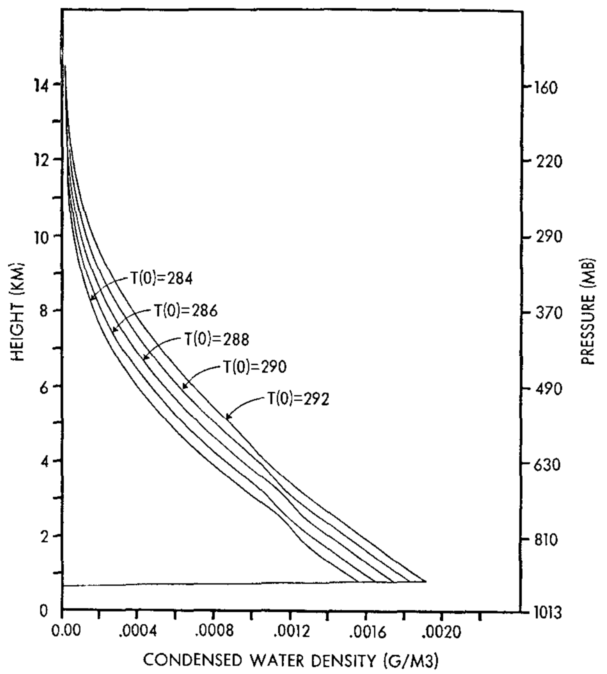

Fig. 2. The condensed water density as a function of height for the model atmosphere at the set of surface temperatures indicated.

Rasool and Schneider (1971), the data taken from Wiscombe (1974) adjusted to the same surface mixing ratio, and our mixing ratio for $T(0)=288\mathrm{K}$. Our diffuse cloud model gives larger mixing ratios than either of the others but resembles the data of the arctic atmosphere derived by Wiscombe rather well, albeit that a linear adjustment has been made. These data represent one of the few sets of observations in an atmosphere in which there is a relatively homogeneous cloud cover. Comparison of our diffuse cloud model with observations on actual clouds (Simpson *et al.*, 1965; Malkus, 1954) shows only qualitative similarity since real well-formed clouds have a much greater condensed water density by about two orders of magnitude.

The structure of the diffuse cloud is, of course, determined in part by the three parameters, γ, A and f_r. The effect of increasing the relative humidity is, as expected, to greatly decrease the absorbed solar energy but with only a very small increase in the infrared emission. Our choice to fix this at 0.75 is to some extent arbitrary but appears to be a reasonable value for the global average. Similarly, the parameter A has been set rather arbitrarily. Rather than giving a lapse rate of $6.5\mathrm{K\ km^{-1}}$ for unsaturated air, one could have adjusted A to give 6.5 for the entire vertical column. Our choice, resulting in the average lapse rate being $5.3\mathrm{K\ km^{-1}}$ up to 100 mb, has the advantage only of simplicity. The "rained" out pa-

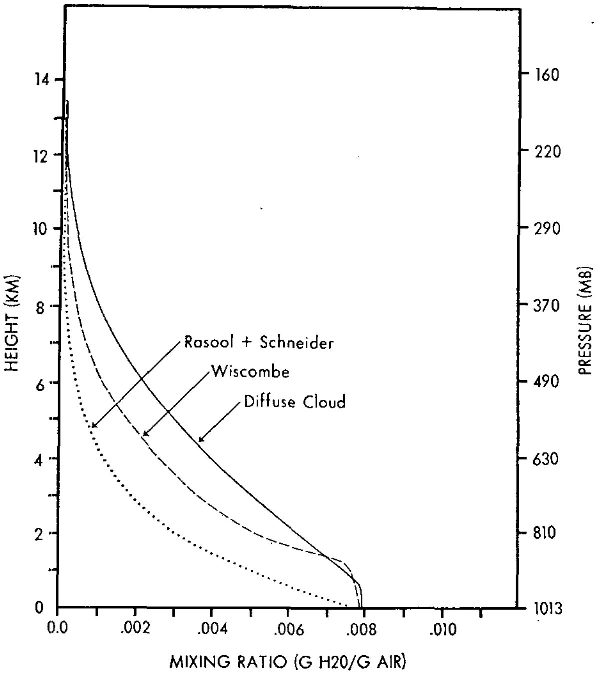

FIG. 4. The total water mixing ratio as a function of height for the model atmosphere at $T(0)=288$K compared with adjusted data from Wiscombe (1974) and the empirical formulation used by Rasool and Schneider (1971).

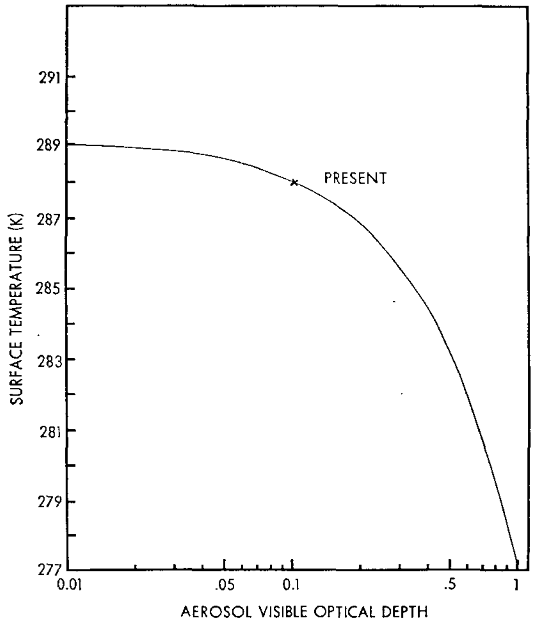

FIG. 6. Calculated values of the radiative steady-state surface temperature as a function of the optical depth of aerosol.

rameter very sensitively affects the atmospheric albedo as may be suspected from its value being stated to the seventh place, but affects the infrared emission only insensitively. Because of these facts we have used it strictly as an adjustable parameter to force the calculated albedo to correspond to satellite-derived data, followed by only minor adjustment in ozone emissivity to compensate in the infrared for radiation balance. Of course, it would be desirable to have all three of these atmospheric parameters in some way dynamically related to surface temperature in a rational way. However, at this stage we have acceded to simplicity and held them constant.

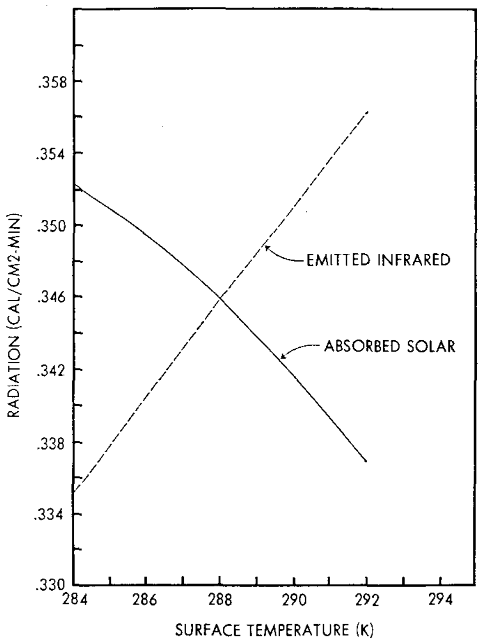

FIG. 5. Calculated values of the absorbed solar and emitted infrared flux given by the model atmosphere as a function of surface temperature. The point of intersection is the point of radiation balance or steady state.

We now illustrate the dynamical relations of the solar energy absorbed and infrared energy emitted as a function of surface temperature. These relations are plotted in Fig. 5. Over this relatively small temperature range both curves are nearly linear. The point of intersection is the steady state which for these calculations was set by parametric adjustment at about $T(0)=288$K.

The stability around this steady state to variations in temperature is easily ascertained by taking the variation of (1.1):

$$\delta\left\langle\frac{\partial Q}{\partial t}\right\rangle_{g,y}=\left[-\frac{S_0}{4}\frac{\partial\langle\alpha\rangle_{g,y,\cos\psi}}{\partial T}-\frac{\partial\langle E\rangle_{g,y}}{\partial T}\right]\delta T. \quad (4.1)$$

The algebraic sign of the quantity in brackets evaluated at the steady state determines its stability. If it is negative it is stable and if positive, unstable. From Fig. 5 it is obvious that both terms in the bracket contribute about equally to stability with magnitudes of about 0.0026 cal cm^{-2} min^{-1} (°K)$^{-1}$.

At this point it is clear that to assess the effect on the climate of any change in conditions as determined by our model, one would introduce the alteration, such as a change in aerosol amount, holding the other parameters at the original steady-state values, and then find that temperature giving a new steady state.

In Fig. 6 we present the results of altering atmospheric aerosol from the assumed present-day value of about 0.1 optical depth units. With the parameters chosen for ω_0 and b for a typical aerosol, any amount of aerosol results in more backscatter than in solar absorption, thus leading to a net cooling of the earth-atmosphere system. A doubling produces about a 1K decrease in mean annual global surface temperature, whereas a fourfold increase produces somewhat more than a 3K decrease.

As may be seen in Fig. 7, a doubling of CO_2 increases the mean annual global surface temperature according to our dynamical model by about 0.7K, but a sixfold increase only increases the temperature 1.7K. The nonlinearity is due to saturation of the 15-μm band. Fig. 8 illustrates the result of varying the solar constant by $\pm 3\%$ giving a near linear slope of only about 0.67K percent^{-1} or 37K cal^{-1} cm^2 sec which is an indication of the remarkable stability of the earth-atmosphere system to changes in solar input.

Finally in Fig. 9 we note the effect of an increase in global dissipation of energy at the surface, say from the activities of man. Again the stability of the system is evident in the slope, the effect also only amounting to about 37K cal^{-1} min cm^2.

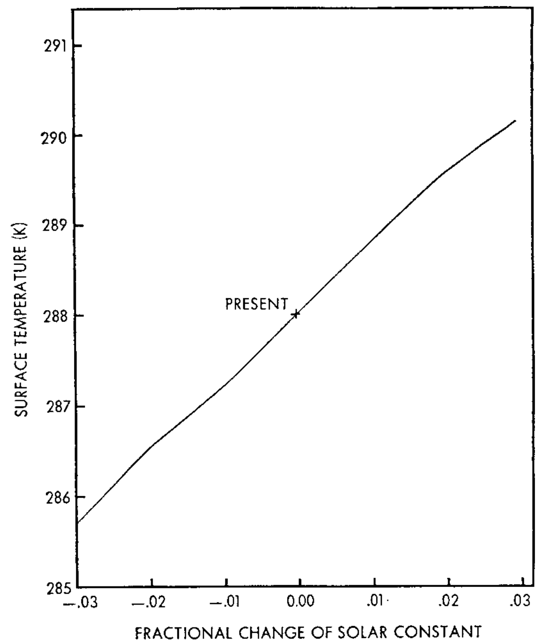

Fig. 8. Calculated values of the radiative steady-state surface temperature as a function of fraction changes in solar constant around the adopted value of 1.95 cal cm^{-2} min^{-1}.

5. Discussion

In the foregoing we have presented the structure of a diffuse cloud model atmosphere dynamically coupled to the single annual global average surface temperature. The "cloud" problem, we reiterate, is a

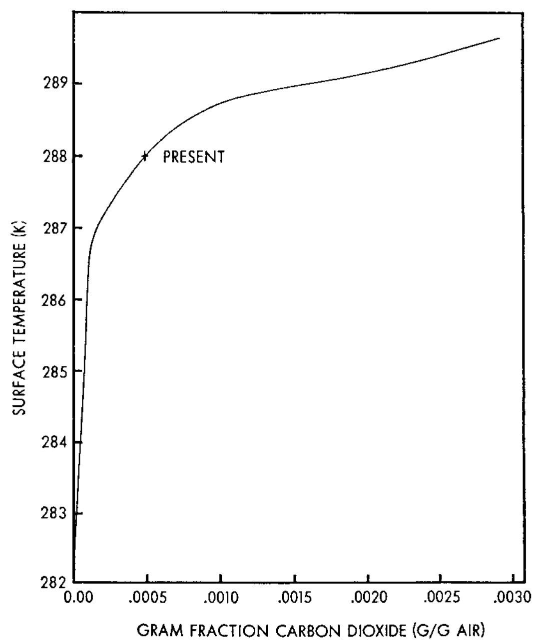

Fig. 7. Calculated values of the radiative steady-state surface temperature as a function of atmospheric CO_2 mixing ratio.

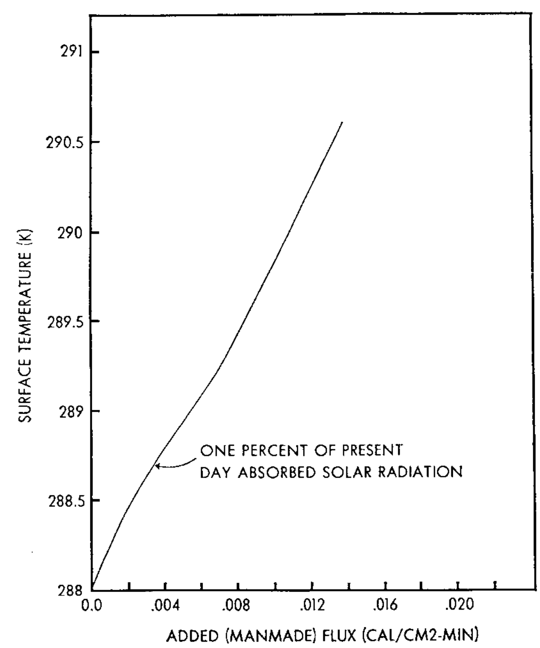

Fig. 9. Calculated values of the radiative steady-state surface temperature as a function of added surface energy (man-made).

central problem in global climatic calculation not only because of its prime role in affecting the radiation balance but also because there has been no good theoretical or empirical way to dynamically couple the effect. Our approach has been to exploit the fundamental thermodynamics of condensation, the cooling of moist air, to uniquely provide an atmospheric structure. No matter how one visualizes the nature of the cooling process, that is, whether it represents the adiabatic expansion of a convective updraft or the polar flow of air driven by the latitudinal imbalance of radiation, the fundamental thermodynamics is there. We have chosen the vertical quasi-adiabatic concept for it is more easily visualized and leads naturally to a vertically structured atmosphere in mechanical equilibrium. Of course, no claim can be made that the atmosphere so structured is a model of the real atmosphere, where clouds form in discrete patches of varying extent, thickness and height, persist for periods of time, and "rain" out or otherwise dissolute. However, even though one may speculate that it may represent a reasonable time-space average, it is the dynamical process that is probably more representative of the real atmosphere and this, not in its detail and idealizations, but only in the fundamental thermodynamical relations. It is these relations, together with three adjustable parameters, that structures the atmosphere as a diffuse thin cloud.

At this stage of development, we have coupled the diffuse cloud model only to the annual global average temperature. Many other averages may be envisaged such as zonal average temperatures, with the parameters zonally determined. Overall feedback under these conditions would undoubtedly be quantitatively different. In another vein, no matter what spatial average is used, horizontal homogeneity need not be adhered to. For one could divide the atmosphere into clear areas and cloudy areas and with appropriate parameterization have the feedback coupling alter either the extent of cloud cover at fixed optical depth, or alter the optical depth maintaining fixed fractional cover, or even a combination of these. Alternatively, one may parameterize a fractional area distribution of different but fixed surface relative humidities which would lead to a like distribution of "cloud" optical depths and heights, perhaps more akin to the usual concept of the atmosphere. In each of these cases, the overall dynamical feedback would probably differ. However, in the absence of a clear rationale to the contrary we have at this stage again bowed to simplicity and kept parameterization to a minimum.

A comment on the vertical distribution of relative humidity is in order. In the diffuse cloud model atmosphere the relative humidity increases with height to the dew point. At all heights above the dew point it is 100%. This distribution is not at all like the hemispheric means obtained by Telegadas and London (1954) or Murgatroyd (1960) which suggest a linear decrease with decreasing pressure (Manabe and Wetherald, 1967). However, rather than the average relative humidities (which must be $<100\%$), it is the vertical distribution of the average absolute humidity (including water content of clouds) relative to the equilibrium water vapor pressure of the vertical temperature profile, that would better represent the situation the diffuse cloud model reflects.

Perhaps somewhat more disturbing is the lack of a clearly defined "top" in the diffuse cloud model, even though a sharply demarked "bottom" does exist. The absence of a "top" and the fact that the thin cloud is not completely opaque to infrared considerably alters the emissivity properties from those usually envisaged and parameterized into atmospheric radiative calculations by others. However, would not a long-term average over a given region blur the "tops" of the many different cloud types with their differing heights and would not the cloudless areas further blur the "blackness?" This is, of course, speculative, but the fact that the model diffuse-cloud atmosphere adjusted parametrically to give the correct global albedo also gives very nearly the correct value for the global infrared emission lends credence. The fact that the change in emission with a change in surface temperature [0.0026 cal cm^{-2} sec^{-1} $(°K)^{-1}$] is nearly quantitatively the same as that given by a variety of other models (Sellers, 1969; Budyko, 1969) [including a fixed 50%, fixed height, nondynamic cloud model (Rasool and Schneider, 1971)] also lends credence.

The stability of the climate due to the negative feedback in our diffuse cloud model has been noted. We have seen that the dynamical relations affecting the solar energy absorbed and the infrared energy emitted both contribute about equally to the stability. As noted above, the slope for the infrared emission agrees well with several other models and it would appear that any reasonable means of calculating the infrared emission or parameterizing the emission as a function of surface temperature will give a negative feedback of about this magnitude. On the other hand, the amount of solar energy absorbed has been more difficult to dynamically couple with surface temperature largely because of the absence of suitable mechanisms to deal with cloud cover and reflectivity. The empirical parameterization used by Budyko (1969) and Sellers (1969, 1973) both lead to a positive feedback for the coupling, in that the zonally average albedos change only as a result of change in surface snow or ice. Cloud cover change is lacking even though the absolute humidity is a strong function of surface temperature. Rasool and Schneider have assumed a fixed cloud cover with fixed cloud albedo and a fixed surface reflectivity so that the feedback is zero. Computed on the basis of the annual glodal

TABLE 2. Temperature change (°K) for given environmental variations.

Variation	Diffuse cloud model	Budyko (1969, 1972)	Sellers (1973)	Rasool and Schneider (1971)	Manabe and Wetherald (1967)
Doubling of CO_2 concentration	+0.7	>+4	+0.1	+0.8	+2.3*
Doubling of aerosol optical depth	−1.1		<−5	−1.8	
1% decrease in S_0	−0.7	−5	−5		−1.3
1% increase in S_0	+0.7		+0.9		+1.3

* Fixed relative humidity.

average temperature our diffuse cloud climatic model exhibits considerably more stability under present-bay conditions than others, especially that of Faegre (1972).

Although, admittedly, the results presented herein are only preliminary in the sense that they are based only upon the annual global average temperature, it is of some interest to compare our results on varying certain environmental conditions with some of those obtained by others with quite different models. This comparison is made in Table 2. It should be noted that the diffuse cloud model predicts a temperature change somewhat smaller in magnitude than any of the other models for each of the environmental variations, with one exception. In this one case Sellers (1973) himself attributes the small temperature increase attending a doubling of CO_2 to an oversimplification of his radiative calculations. It would appear that the greater climatic stability shown by our model is again evident in these comparisons.

In concluding this discussion and paper we should like to emphasize that the objective of developing a global climatic model that is, insofar as possible, based upon first principles has not been reached. The present work represents only a step and even this step could probably have been improved by employing alternate means of calculating the radiative transport, such as to include molecular absorption in the solar flux and nonconservative anisotropic scattering in the infrared flux. More important, however, since the error made here is probably small in terms of the dynamical aspects, is the need to extend the dynamical coupling to zonal and/or seasonal averages and perhaps even to the zonal divided in oceanic and continental averaged surface temperatures. Work in this direction is currently underway.

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