Global climate change: an investigation of atmospheric feedback mechanisms

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ABSTRACT

A global energy balance for the earth-atmosphere system requires knowledge of a number of quantities, two of which are tropospheric lapse rate and cloud altitude. Since both of these quantities could be dependent upon surface temperature, they are potential candidates for feedback mechanisms in surface-temperature sensitivity studies. The present investigation illustrates that the global energy balance is quite insensitive to lapse rate, thus eliminating this as a possible feedback mechanism. Cloud altitude, on the other hand, can be quite important. Employing a three-layer cloud model, and comparing calculated and empirical outgoing flux results, it is suggested that cloud temperatures should be held constant when considering variations in surface temperature, as opposed to the conventional assumption of fixed cloud altitudes. This is consistent with earlier single-cloud modeling (Cess, 1974), and based upon this conclusion it is found that the resulting dependence of cloud altitude upon surface temperature produces a strong positive feedback.

Introduction

It has been suggested that variations in the earth's surface temperature, as well as associated changes in solar absorption by the earth-atmosphere system, are the result of modifications of the earth's albedo due to variability in the amount of particulate matter within the atmosphere. As discussed, for example, by Budyko (1969) and Bryson (1974), the warm period of the 1920's and 1930's is possibly attributed to a corresponding period of minimal volcanic activity, whereas cooling since 1940 might be due to an increase in particulate matter resulting from man's activity, coupled with a more recent increase in volcanic emissions. In addition to increasing the planetary albedo, particulate matter might also enhance the infrared opacity of the atmosphere and thus increase the greenhouse effect. This would tend to raise the mean surface temperature, competing with the albedo modification. The relative role of these two mechanisms is by no means clear, but the study by Rasool & Schneider (1971) suggests that the albedo modification is the dominant mechanism.

Irrespective of the precise details by which particulate matter might influence the global energy balance, an important aspect in understanding global climate changes involves feedback mechanisms which could amplify the influence of variations in factors such as planetary albedo upon the mean surface temperature. Possible feedback mechanisms are discussed in the following section, and the purpose of the present paper is to investigate certain of these mechanisms.

Feedback mechanisms

The infrared opacity of the earth's atmosphere is primarily due to water, both in the form of vapor and clouds, and associated with this are the following possible atmospheric feedback mechanisms: (1) water vapor content, (2) cloud altitude, (3) lapse rate, and (4) cloud amount.

As discussed by Manabe & Wetherald (1967), a decrease in surface temperature reduces the water vapor content of the atmosphere, which results in a decresing greenhouse effect with a further decrease in surface temperature. Thus water vapor content produces a positive feedback mechanism, which according to Manabe & Wetherald (1967) roughly doubles the sensitivity of surface temperature to factors such as planetary albedo, under the assumption of fixed relative humidity.

Conventionally in surface-temperature sensitivity studies, the altitudes of parameterized clouds are held constant, i.e., independent of surface temperature (Manabe & Wetherald, 1967; Rasool & Schneider, 1971; Schneider, 1972). There is no particular rational for this, although a quantitative appraisal of the dependence of cloud altitude upon surface temperature would necessitate a detailed dynamical-hydrological model incorporating the physics of cloud formation. Alternatively, Cess (1974) has illustrated that outgoing flux calculations based upon the assumption of fixed cloud altitude are incompatible with empirical flux results of Budyko (1969), whereas employing a single effective cloud Cess finds that a consistent outgoing flux model is one for which the cloud-top temperature is held constant when the surface temperature is varied. Thus, for example, a decrease in surface temperature would result in a decrease in cloud altitude, and since the cloud contribution to atmospheric opacity diminishes with decreasing altitude, a reduction in surface temperature produces a decrease in greenhouse effect which results in a positive feedback mechanism.

The outgoing flux further depends upon lapse rate. The tropospheric lapse rate is sufficient for outgoing flux calculations, since stratospheric opacity plays a minor role (Cess, 1974). This lapse rate is evidently dependent upon both water vapor content and large scale dynamics. From previous discussion the dependence upon water vapor content could clearly result in the lapse rate being a function of surface temperature, while the possible dependence upon large scale dynamics might introduce a further dependence upon surface temperature. The lapse rate is thus a potential candidate for a feedback mechanism.

Cloud amount constitutes an additional feedback mechanism, but in the absence of a detailed cloud formation model it is impossible to be quantitative concerning the dependence of cloud amount upon surface temperature. This difficulty is compounded by the possibility that particulate matter could influence cloud amount (Hobbs et al. 1974). The effect of varying cloud cover upon surface temperature has been studied by Manabe & Wetherald (1967) and Schneider (1972), but these investigations do not answer the question as to the influence of cloud amount as a feedback mechanism. This is potentially a very important effect, but as pointed out by Schneider (1972), sufficiently realistic models capable of investigating this mechanism are not currently available.

In the present paper we investigate two of the four proposed feedback mechanisms, cloud altitude and lapse rate, by employing simple radiation flux calculations. As previously discussed, the role of water vapor content has been considered by Manabe and Wetherald (1967), while the effect of cloud amount is as yet unresolved. With respect to cloud altitude, the present paper will extend the earlier approach of Cess (1974), by which calculated radiation fluxes are compared with empirical results, thereby permitting a rough appraisal of possible models for the dependence of cloud altitude upon surface temperature. Before proceeding with this, however, it is first necessary to investigate the role of lapse rate as a feedback mechanism, since this mechanism would be inherent in any empirical flux formulation.

Lapse rate

Detailed knowledge of the dependence of tropospheric lapse rate, $\Gamma = -dT/dz$, upon surface temperature is not readily available, although Stone (1973) does discuss the interaction of large scale dynamics upon lapse rate. Alternatively, a qualitative appraisal of the influence of Γ upon the global energy balance may be achieved by simply holding the surface temperature, T_s, constant and arbitrarily varying Γ. For this purpose we will employ a single effective cloud, which is black in the infrared with a cloud-top temperature $T_c = 244°$K (Cess, 1974). The outgoing infrared flux from the earth-atmosphere system, F, may be expressed as

$$F = c_1 - c_2 A_c \tag{1}$$

where A_c denotes the fraction of cloud cover, while c_1 and c_2 depend upon T_s and Γ, while c_2 is additionally a function of T_c. With regard to feedback mechanisms, we are also concerned with determining $\partial F/\partial T_s$, or

$$\frac{\partial F}{\partial T_s} = \frac{\partial c_1}{\partial T_s} - \frac{\partial c_2}{\partial T_s} A_c \tag{2}$$

The procedure for evaluating c_1 and c_2 is described by Cess (1974), and this accounts for water vapor content as a feedback mechanism by employing a fixed distribution of relative humidity as suggested by Manabe & Wetherald (1967). Cloud altitude as a feedback mechanism is also taken into account by holding the cloud-top temperature, rather than cloud-top altitude, constant when determining $\partial c_2/\partial T_s$ (Cess, 1974). This is discussed in more detail in the following section.

Results for c_1 and c_2 given in the units of cal cm^{-2} min^{-1}, together with their derivatives are listed in Table 1 for several values of lapse rate. Clearly both F and $\partial F/\partial T_s$ are quite insensitive to Γ, and there is a logical physical reason for this. If the lapse rate is increased, then the outgoing flux comes from a colder atmosphere, such that F should be reduced. Countering this effect, however, is the fact that a colder atmosphere has a lower water vapor content with a correspondingly smaller infrared opacity, so that the outgoing flux is enhanced since it comes from lower and thus warmer regions of the atmosphere. These two competing mechanisms essentially cancel one another. This argument pertains solely to the troposphere, but as previously stated, stratospheric opacity plays a minor role in outgoing flux calculations (Cess, 1974).

It in turn follows that, with respect to a global energy balance, lapse rate does not constitute a significant feedback mechanism. This does not, of course, imply that a change in lapse rate could not indirectly influence the global energy balance, for example by altering large scale dynamics (Stone, 1973) and thus cloud amount.

Table 1. *Values of the outgoing flux parameters for $T_s = 288°K$ and $T_c = 244°K$*

Γ(°K/km)	c_1	c_2	$\partial c_1/\partial T_s$	$\partial c_2/\partial T_s$
6.0	0.370	0.104	0.0027	0.0025
6.5	0.367	0.102	0.0028	0.0025
7.0	0.366	0.102	0.0026	0.0024

Cloud altitude

Cess (1974) has recently illustrated that either a single effective cloud or Manabe and Wetherald's (1967) three-layer cloud model give results for the outgoing infrared flux which are in excellent agreement with empirical findings of Budyko (1969). Furthermore, employing a single effective cloud Cess has shown that $\partial c_2/\partial T_s$, which denotes the cloud correction to $\partial F/\partial T_s$, also agrees with Budyko's result providing the cloud-top temperature is held constant as surface temperature is varied, in contrast to the conventional assumption in which the cloud-top altitude is held constant (Manabe & Wetherald, 1967; Rasool & Schneider, 1971; Schneider, 1972).

The purpose of the present section is twofold. First, we will show that the model of fixed cloud-top temperature, when applied to the more realistic three-cloud model of Manabe & Wetherald (1967), again yields excellent agreement with Budyko. Secondly, it will be shown that the corresponding model of a fixed cloud-bottom temperature is qualitatively consistent with empirical results for net surface radiation. The parameters describing the three-cloud model of Manabe & Wetherald are listed in Table 2, where $A_{c,i}$ denotes the cloud cover fraction for each cloud layer, while the cloud-top and cloud-bottom temperatures have been evaluated from the corresponding altitudes given by Manabe & Wetherald with $T_s = 288°K$ and $\Gamma = 6.5°K/km$. All three cloud layers are assumed to be black in the infrared. Although this might not be the case for the highest cloud layer, the results apply to nonblack clouds by simply replacing cloud amount by the product of cloud amount times cloud emissivity.

For multiple clouds layers, the outgoing flux may be expressed as (Cess, 1974)

$$F = c_1 - \sum_i c_{2,i} \bar{A}_{c,i} \tag{3}$$

where $c_{2,i}$ is determined by treating each cloud layer individually, while $\bar{A}_{c,i}$ represents the cloud cover fraction which is not overlapped by an upper cloud (or clouds), calculated from

Table 2. *Parameters for the three-cloud model of Manabe & Wetherald*

i	Cloud-top temperature (°K)	Cloud-bottom temperature (°K)	$A_{c,i}$	$\bar{A}_{c,i}$
1	223	223	0.228	0.228
2	261	261	0.090	0.069
3	270	277	0.313	0.220

the $A_{c,i}$ values in accordance with Manabe & Strickler (1964). Values for $c_{2,i}$ have been determined previously (Cess, 1974), and employing the same calculation procedure we have varied T_s so as to evaluate $\partial c_{2,i}/\partial T_s$, with these results given in Table 3.

With reference to equations (2) and (3), it follows that

$$\frac{\partial c_2}{\partial T_s} A_c = \frac{\partial c_2}{\partial T_s} \sum_i \bar{A}_{c,i} = \sum \frac{\partial c_{2,i}}{\partial T_s} \bar{A}_{c,i} \tag{4}$$

and the value for $\partial c_2/\partial T_s$, together with previously determined results for c_1, c_2 and $\partial c_1/\partial T_s$ (Cess, 1974), are listed in Table 4 and compared with the empirical results of Budyko (1969). Again the units for c_1 and c_2 are cal cm^{-2} min^{-1}.

What is important in the present study is the excellent agreement involving $\partial c_2/\partial T_s$, since it is this quantity which is indicative of cloud altitude as a feedback mechanism. Alternatively, employing the conventional assumption of fixed cloud-top altitude, one finds $\partial c_2/\partial T_s = -0.0002$ which differs considerably from Budyko. These findings are consistent with the single-cloud modeling of Cess (1974), and they support further the cloud model for which cloud temperature rather than cloud altitude is specified in surface-temperature sensitivity studies. It must be emphasized, however, that this conclusion is model dependent, in that it is based upon comparing calculated and empirical outgoing flux results, with subsequent indirect inference regarding cloud altitude as a feedback mechanism. Certainly Budyko's empirical flux formulation by itself does not constitute a direct observation of cloud-altitude feedback.

Table 3. *Values of $\partial c_{2,i}/\partial T_s$ for fixed cloud-top temperature*

i	$\partial c_{2,i}/\partial T_s$ (cal cm^{-2} min^{-1} °K^{-1})
1	0.0028
2	0.0023
3	0.0021

Table 4. *Comparison of calculated quantities with empirical results of Budyko for $T_s = 288°K$*

	c_1	c_2	$\partial c_1/\partial T_s$	$\partial c_2/\partial T_s$
Calculated	0.367	0.097	0.0028	0.0024
Budyko	0.366	0.102	0.0032	0.0023

Circumstantial evidence additionally supporting the fixed cloud-temperature model follows from considerations of net surface radiation. For this purpose we initially consider each cloud layer separately. With T_c now denoting the temperature of the cloud bottom, including a black cloud within the radiation flux formulation, equation (12), of Cess (1974), and denoting the net upward radiation flux at the surface by I, it readily follows that

$$I/\sigma T_s^4 = \psi(\zeta_s) - [0.75(T_c/T_s)^4 (1/3 + \exp(-\sqrt{\zeta_s^2 - \zeta_c^2})) + \psi(\zeta_c) - 1] A_{c,i}^* \tag{5}$$

where σ is the Stefan–Boltzmann constant, $A_{c,i}^*$ represents the cloud cover fraction which is not overlapped by a lower cloud (or clouds), and

$$\psi(\zeta) = 1 - 0.75 \int_0^\zeta (\xi/\xi_s)^n \exp(-\sqrt{\zeta_s^2 - \xi^2}) \frac{\xi d\xi}{\sqrt{\zeta_s^2 - \xi^2}} \tag{6}$$

while ζ is a dimensionless optical coordinate defined as

$$\zeta = 0.144\sqrt{P_w \tilde{P} H_w}$$

with $P_w \tilde{P} H_w$ in atm^2 cm, where P_w denotes the partial pressure of water vapor, $\tilde{P}$ the effective broadening pressure, and H_w the water vapor scale height, while ζ_c and ζ_s refer to the cloud bottom and surface, respectively. Expressions for these quantities, together with the definition for n, are given elsewhere (Cess, 1974).

Equation (5) may be rephrased as

$$I = I_0(1 - b_i A_{c,i}^*) \tag{7}$$

Table 5. *Cloud parameters for net surface radiation*

i	$A^*_{c,i}$	b_i	$(\partial b_i/\partial T_s)_{T_C}$	$(\partial b_i/\partial T_s)_{z_C}$
1	0.143	0.352	−0.0045	0.0019
2	0.062	0.664	−0.0084	0.0019
3	0.313	0.838	−0.0109	0.0017

where I_0 is the net surface radiation for clear skies. The quantities I_0 and b_i have been evaluated for $T_s = 288°K$, $\Gamma = 6.5°K/km$ and the three cloud layers listed in Table 2. For the clear-sky contribution we find that

$$I_0 = 0.144 \text{ cal cm}^{-2} \text{ min}^{-1}$$

$$\partial I_0/\partial T_s = 0.0016 \text{ cal cm}^{-2} \text{ min}^{-1} \text{ °K}^{-1}$$

while calculated results for the cloud constants are listed in Table 5.

As in the outgoing flux calculation, it is the quantity $\partial b_i/\partial T_s$ which represents the importance of surface temperature upon cloud models, and $\partial b_i/\partial T_s$ has been calculated for both a fixed cloud temperature and a fixed cloud altitude, denoted respectively by $(\partial b_i/\partial T_s)_{T_C}$ and $(\partial b_i/\partial T_s)_{z_C}$. What is significant is that the two cloud models yield different signs for $\partial b_i/\partial T_s$, regardless of cloud layer.

Empirical coefficients characterizing the influence of cloudiness on net surface radiation have been determined by Berland & Berland (as quoted by Kondratyev, 1969). While their results are for specific latitude zones and do not apply for global mean conditions, comparison of the cold to warm half-year results illustrates that $\partial b_i/\partial T_s$ is always negative, irrespective of latitude zone or cloud layer. This is consistent with the $(\partial b_i/\partial T_s)_{T_C}$ results of Table 5, but is contrary to the $(\partial b_i/\partial T_s)_{z_C}$ values. While this sort of comparison is by no means conclusive, it does lend support to the premise that cloud-bottom temperature should be held constant in surface-temperature sensitivity studies.

As a final point it is worth comparing the present calculation for net surface radiation with climatological data. Equation (7) applies to multiple clouds providing b_i is replaced by b, where

$$b \sum_i A^*_{c,i} = \sum_i b_i A^*_{c,i}$$

and from Table 5 $b = 0.68$. It in turn follows that for global mean conditions with $A_c = 0.5$, $I = 0.095$ cal cm^{-2} min^{-1}, which agrees favorably with the value 0.099 cal cm^{-2} min^{-1} from climatological data (Sellers, 1965).

Concluding remarks

It is worthwhile to reemphasize a point made earlier by Cess (1974) concerning cloud altitude as a feedback mechanism. A convenient parameter for illustrating feedback mechanisms is $\partial T_s/\partial T_e$, where T_e is the effective temperature of the planet with $F = \sigma T_e^4$. Since the outgoing infrared flux is equal to absorbed solar radiation, then changes in factors such as planeatary albedo would alter T_e, with the corresponding change in surface temperature being governed by $\partial T_s/\partial T_e$.

For clear skies, equations (1) and (2) and Table 4 yield $\partial T_s/\partial T_e = 2.0$, which clearly illustrates the influence of water vapor content as a positive feedback mechanism. Considering next fixed cloudiness with $A_c = 0.5$, then for the fixed cloud temperature model, $\partial T_s/\partial T_e = 3.2$, and we see that cloud altitude contributes an additional and significant positive feedback. This is not the case, however, for the conventional by apparently less appropriate model of fixed cloud altitude, for which $\partial T_s/\partial T_e = 1.7$.

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ИЗМЕНЕНИЕ ГЛОБАЛЬНОГО КЛИМАТА: ИССЛЕДОВАНИЕ АТМОСФЕРНЫХ МЕХАНИЗМОВ С ОБРАТНОЙ СВЯЗЬЮ

Глобальный баланс энергии для системы земля–атмосфера требует знания ряда параметров, два из которых — вертикальный градиент температуры в тропосфере и высота облаков. Так как оба этих параметра могут зависеть от температуры поверхности, они являются потенциальными кандидатами на роль механизмов с обратной связью в исследованиях чувствительности температуры поверхности. Данное исследование показывает, что глобальный баланс энергии совершенно не чувствителен к вертикальному градиенту температуры и последний, таким образом, не может служить механизмом с обратной связью. С другой стороны, высота облаков может быть очень важной. На основе трехслойной модели облачности и сравнения вычисленных и эмпирических потоков уходящей радиации высказывается предположение, что температура облаков должна считаться постоянной при изменении температуры поверхности в противоположность обычному допущению о фиксированных высотах облаков. Это согласуется с предыдущим моделированием однослойной облачности (Сесс, 1974) и позволяет сделать вывод, что зависимость высоты облаков от температуры поверхности приводит к сильной положительной обратной связи.