THE EFFECTS OF CLOUD AND RADIATION

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1. Introduction

Clouds are germane to the modification and redistribution of the total diabatic heating of the atmosphere. We can identify this association in the list of Arakawa (1975) who considered a number of different ways in which clouds might influence atmospheric circulation and climate.

(i) The coupling of dynamic and hydrological processes through the release of latent heat and by evaporation, and by the redistribution of sensible and latent heat and momentum;

(ii) The coupling of radiative and dynamical-hydrological processes in the atmosphere through the reflection, absorption and emission of radiation;

(iii) The coupling of hydrological processes in the atmosphere and in the ground via precipitation, and;

(iv) The influencing of couplings between the atmosphere and the ground through the modification of radiation and the turbulent transfers through the surface.

Diabatic heating is thus a result of the combination of all of these processes and is modified and spatially redistributed by each process. While the individual components of the total heating may be large, the net heating is often just a small residual. Defining these individual components is often greatly complicated by the interactive nature of the various processes that act to produce this heating. Radiative cooling in a cloud top layer, for example, might act to destabilize that layer and thus change the latent heating processes within the cloud top. The need to understand the role clouds play in modifying radiation is thus essential not only because of the modification of the radiative heating field by the cloud itself but also because of the effect of cloud on the other forms of heating.

This lecture concentrates on providing some understanding of the radiative processes that occur in clouds and discusses how they might be modeled. These radiation processes are discussed in terms of the following parameters

(i) The surface radiation budget ....Rs

(ii) The planetary radiation budget ......Rp

(iii) The radiative heating and cooling potential of the cloud layer .....Qr

Thus the effect of clouds on radiation in this lecture is understood to mean the effect of cloud on each of these parameters.

The bibliography included at the back of this paper is not meant to provide an exhaustive literature review of the subject, but it will provide the reader with sufficient diversity to follow up on various different issues introduced here. I have also included additional references which supplement some of the issues discussed.

2. A simple models of Rs

In describing the effect of cloud on the surface radiation budget ,we consider a very simple two layer model of the atmosphere as shown in Fig 1. Each layer is characterized by a specific value of transmittance Tr_L for

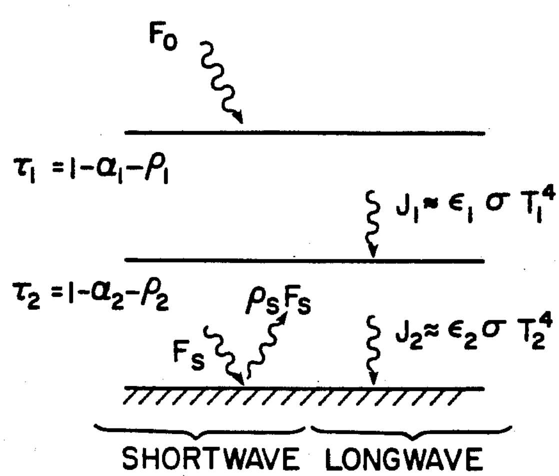


Fig. 1 Schematic of a simple two layer radiative transfer model. Note symbolic of differences from the text: $\tau_1 = Tr_{s,1}$, $F_o = H_o$, $J_1 = \Sigma_1$, $\rho_s = \rho_g$.

longwave radiation and Tr_S for shortwave radiation (the latter being a function of both the reflectance ρ and absorptance A of the Layer) and a source of longwave radiation Σ (which is related to the emittance ε of the layer) It is assumed that the longwave transmittance is a function only of layer emissivity.

We can immediately write the net radiative fluxes into the surface as

$$H_L = Tr_{L,2}\,\Sigma_1 + \Sigma_2 - \varepsilon_g\,\sigma T_g^4 \qquad (a)$$

$$H_S = Tr_{S,2}\,Tr_{S,1} H_o\,(1-\rho_g)\,(1+\rho_1\rho_2) \qquad (b) \qquad (1)$$

where the numerical subscripts refer to the layer, ρ_g is the surface albedo, ε_g is the surface emissivity and Ho the flux of solar energy indicent on layer 1.

These equations are a very simple statement on the radiant energy transfer within the atmosphere and are perhaps the simplest form of a radiative transfer equation (the factor $1+\rho_1\rho_2$ in (1b) represents the single reflection of radiation from the surface to the atmosphere and back with higher order multiple reflections between the atmosphere and surface neglected).

For radiative equilibrium, the surface radiation budget becomes

$$R_S = H_L + H_S = 0 \qquad (2)$$

and on combining (1) and (2) we obtain

$$\varepsilon_g\sigma\,T_g^4 = Tr_{S,1}\,Tr_{S,2}\,H_o(1-\rho_g)\,(1+\rho_1\rho_2) + Tr_{L,2}\,\Sigma_1 + \Sigma_2 \qquad (3)$$

Consider now how this radiative equilibrium budget may change for a given change of cloud in layer 1. This change is assumed to occur in some cloud property which at this stage will remain unspecified. For example, the change may result by an increase or decrease in cloud amount, cloud opacity or cloud liquid water.If we take Δ (p) to refer to the cloud induced difference in some general radiation budget parameter p, then a surface heating or cooling trend may result if

$$\Delta\varepsilon_g \sigma T_g^4 > 0 \quad \text{(cloud induced surface heating trend)}$$
$$\Delta\varepsilon_g \sigma T_g^4 < 0 \quad \text{(cloud induced surface cooling trend)} \tag{4}$$

and it follows

$$\Delta\ \sigma T_g^4 \lessgtr 0 \tag{5}$$

if

$$\Delta\left[Tr_{L,2}\ \Sigma_1 + \Sigma_2\right] \lessgtr -\Delta\left[Tr_{s,2}\ Tr_{s,1}\ H_o(1-\rho_g)(1+\rho_1\rho_2)\right]$$

and by ignoring higher order products involving $\rho_1\rho_2$ (5) becomes

$$Tr_{L,2}\ \sigma T_1^4\ \Delta\varepsilon_1 \ \underset{\text{warming}}{\overset{\text{cooling}}{\lessgtr}}\ H_o(1-\rho_g)\Delta\rho_1 \tag{6}$$

with the following assumptions

(i) $\Delta\left[Tr_{L,2}\ \Sigma_1 + \Sigma_2\right] \sim Tr_{L,2}\ \Delta\Sigma_1 \sim Tr_{L,2}\sigma T_1^4 \Delta\varepsilon_1$

(ii) $\Delta\left[Tr_{S,2}Tr_{S,1}\ H_o(1-\rho_g)(1+\rho_1\rho_2)\right] \sim H_o(1-\rho_g)\Delta\left[Tr_{s1}Tr_{s,2}\right] \sim -H_o(1-\rho_g)\Delta\rho_1$

where $\Sigma_1 \cdot \varepsilon_1\sigma T_1^4$ and we assume that T_1 and $Tr_{L,2}$ are unchanged and that the only change to the total atmospheric shortwave transmittance (i.e. $Tr_{S,1} x Tr_{S,2}$) results from the increased albedo of layer 1. Whether or not cloud increases the total shortwave absorptance of the atmosphere is debatable but we assume that it does not.

Thus (6) suggests that the heating or cooling potential at the surface induced by a change in some cloud parameter is largely determined by the associated change in the emittance ($\Delta\varepsilon_1$) and reflectance ($\Delta\rho_1$) of that layer. It is stressed that a unique relationship (or set of relationships) exists between $\Delta\varepsilon_1$ and $\Delta\rho_1$ or more specifically between cloud emittance and cloud reflectance. The proper incorporation of cloud radiation effects in a GCM must include such a relationship in order to parameterize the effects of cloud (especially thin cloud) on Rs realistically.

Examples of the relationship between $\Delta\varepsilon_1$ and $\Delta\rho_1$ are shown in Fig. 2 for high and low cloud and for a specified solar zenith angle (note we expect a family of such curves for different solar zenith angles). Also superimposed on this diagram are the diagonal lines that represent the particular $\varepsilon_1 - \rho_1$ relationship that produce a null effect of cloud on Rs. The slope of these lines are determined from (6) as

$$C_g = \frac{Tr_{L,2}\ \sigma T_1^4}{H_o\ (1-\rho_g)} \qquad (7)$$

for the two values of ρ_g specified on the diagram, and are thus influenced by the height of the cloud (through T_1 and $\Delta\varepsilon_1$), the surface albedo ρ_g and latitude (through Ho). It is clear that from the $\varepsilon_1 - \rho_1$ relationship(s) that we expect from terrestrial clouds, it is unlikely that the compensating effects between changes in long- and short-wave radiation will completely balance out.

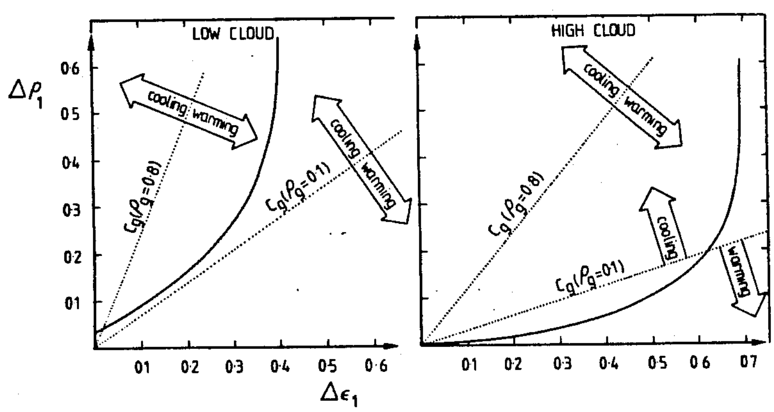


Fig. 2

Examples of the relationship between $\Delta\rho_1$ and $\Delta\varepsilon_1$ for 100mb thick low and high cloud centered respectively at 950 and 350mb. The slopes of the diagonal lines, defined in (7), were determined using R_o=340W.m^{-2}, ρ_g =0.1 and 0.8 and T_1=291°K (low) and 230°K (high). Note $\Delta\varepsilon_1$ (low) < $\Delta\varepsilon_1$ (high) due to moisture differences of respective layers.

We can also infer the following from (6)

(1) The heating tendency at the surface is defined by a combination of two factors that vary with altitude, namely a decrease of cloud temperature (T_1) and an increase in cloud (or layer) emittance ($\Delta\varepsilon_1$) with increasing altitude while the cooling trend is governed by an increase in layer reflectance (i.e. of $\Delta\rho_1$). The fact that high clouds produce a heating tendency results from the relatively large change in the emittance of a relatively dry layer. Introduction of a cloud in the upper layers of the atmosphere can amount to a substantial increase in the emittance of that

layer. This effect of this increase on Rs is partially offset by the descreased temperature of the layer (at most a factor of 4). The imbalance between $\Delta\rho_1$ and $\Delta\varepsilon_1$ is largest for thin clouds and so high thin cloud has the largest surface heating potential.

(ii) The cooling trend, goverened by the loss of solar energy at the surface, is influenced by 3 factors namely Ho, $\Delta\rho_1$, and the value of ρ_g each of which depends on latitude. Typical of polar regions are large values of ρ_g and small values of H_o and even relatively small increases in the long wave radiation at the surface (through $\Delta\varepsilon_1$) provides a potential to induce a heating tendency. In fact, the effect of cloud on Rs can change sign under conditions typical of polar regions depending on the magnitude of ρ_g. This is obvious by reference to Fig.3 which shows the difference in the overcast to clear sky radiative equilibrium surface temperature (as derived from a radiaitve equilibrium model with convective adjustment) as a function of surface albedo for values of H_o typical of latitudes indicated.

We can interpret these curve in terms of a critical surface albedo which from (6) is

$$\rho_{gc} = 1 - \frac{Tr_{L,2}\sigma T_1^4 \Delta\varepsilon_1}{Tr_{S,2} H_o \Delta\rho_1} \qquad (8)$$

such that

$\rho_g < \rho_{gc}$ (cloud-surface albedo induced cooling)

$\rho_g > \rho_{gc}$ (cloud-surface albedo induced heating)

<u>Thus the effect of cloud on Rs and the feedback between Rs and surface albedo (such as described in relation to the CO_2 - climate problem) is inseparable from and requires the careful consideration of cloud effects.</u>

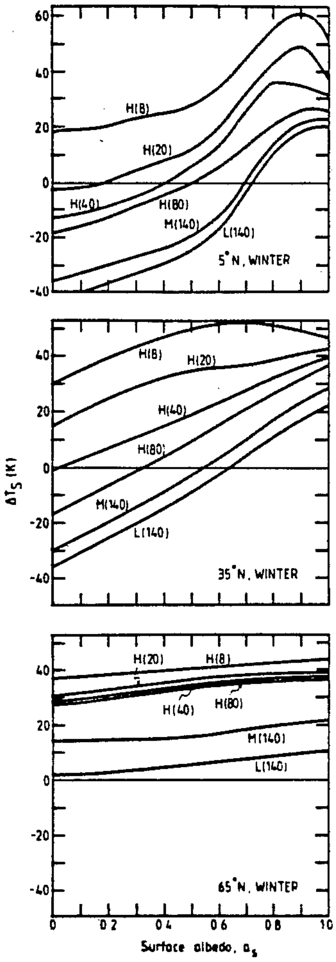


Fig. 3 Equilibrium surface temperature difference ΔT_s (= ΔT_g in text) between clear and overcast sky for 3 latitudes shown as a function of surface albedo a_s (= ρ_g in text) (Stephens and Webster, 1980).

TABLE 1

Values of the cloud climate factors ΔR_p and C_p obtained from a number of sources including the present theoretical study. The numbers shown in relation to the latter are for optical thicknesses (τ) typical of terrestrial clouds.

	ΔR_p ($W.m^{-2}$)	C_p	Source
(a) Other Studies			
	-35	-1	Ellis (1978)
	-55	-0.7	Schneider (1972)
	1	0	Cess (1976)
	-65	-2	Ohring & Clapp (1980)
	-	<0	Hartman & Short (1980)
(b) Present Study			High Cloud
	+14	+0.4	τ = 0.3
	+5	+0.1	τ = 2
	-27	-0.33	τ = 13
			Low Cloud
	-60	-3.8	τ = 13
	-90	-3.9	τ = 39

3. A simple model of Rp

The planetary budget Rp can be similarly derived from the simple two layer model introduced above. Using simple radiative transfer notions, the reflected flux at the top of the atmosphere is

$$H_S \sim H_o(\rho_1 + Tr_{S,1}\rho_2 + (Tr_S)^2\ \rho_g \qquad (9)$$

and the longwave flux emmitted to space is

$$H_L = Tr_L \sigma T_g^4 \varepsilon_g + Tr_{L,1}\varepsilon_2 \sigma T_2^4 + \varepsilon_1 \sigma T_1^4 \qquad (10)$$

where Tr_S and Tr_L are the short- and long-wave transmisions through the entire atmosphere. All other symbols are as described above. If we define the planetary radiative budget as

$$R_p = H_o - (H_S + H_L) \qquad (11)$$

then by incorporating (9) and (10) we arrive at

$$R_p \sim H_o(1-\rho_1-Tr_{S,1}\rho_2 - (Tr_s)^2\rho_g - T_L\sigma T_g^4\varepsilon_g^4 - Tr_{L,1}\varepsilon_2\sigma T_2^4 - \varepsilon_1\sigma T_1^4 \quad (12)$$

Invoking the same arguments as above, a cloud induced change in Rp (which we denote as Δ Rp) can be approximated as

$$\Delta R_p = -H_o\Delta\rho_1(1+\rho_g\Delta\rho_1-\rho_2) + \Delta\varepsilon_1(\sigma T_1^4-\sigma T_g^4) \quad (13)$$

Following Hartman and Short (1980), we define a "cloud factor" as

$$\Delta R_p = C_p\Delta H_L, \qquad C_p = \frac{-\Delta H_S}{\Delta H_L} + 1 \quad (14)$$

This factor attempts to assign some quantitative measure of the relative effects of the short- and long-wave components of Rp that result due to cloud change. For example, Cp<0 when Δ Rp<0 and the loss of radiation by an increase of reflection is larger than the associated decrease in longwave radiation emitted to space. For 0<Cp<1, (i.e. Δ Rp>0) the reverse applies and clouds act to enhance the so-called greenhouse effect.

A number of different estimates of Cp can be deduced from various studies and some are summarized in Table 1. From the estimates listed on the table, the general concensus seems to suggest that Cp<0 (i.e. clouds tend to introduce an overall cooling trend to the earth-atmosphere climate system). However, from our simple analysis, we can formally derive Cp as

$$C_p = \frac{-H_o\Delta\rho_1(1+\rho_g\Delta\rho_1-\rho_2)}{\Delta\varepsilon_1(\sigma T_1^4-\sigma T_g^4)} +1 \quad (15)$$

and it is evident that an increase of cloud at specific levels (especially high cloud) results in positive values of Δ Rp.

The following question is often posed. Can we deduce Δ Rs from some estimate of Δ Rp which may be obtained for example, from satellite measurements? The expressions derived above for Δ Rp and Δ Rs are a little misleading as the cloud temperature relevant to Rs. actually refers to the temperature at or near cloud base whereas the cloud top temperature is infered in the Δ Rp expression. Thus unless these temperatures are correlated in some way, or other information is used to infer cloud base temperature from cloud top temperature, then Rs and Rp remain uncorrelated.

4. Some properties of ε_1 ρ_1 and Qr.

The need to describe cloud albedo and cloud emittance is central to the understanding of the effects of cloud on Rp and Rs and Qr. Some general characteristics of these parameters which should be incorporated into their parameterization will now be discussed briefly.

In general, the cloud albedo increases with increasing cloud "opacity" and increases with decreasing solar elevation. We can characterize this opacity in terms of the cloud optical thickness τ which is defined such that the fraction $\exp(-\tau)$ of incident radiation that is directly transmitted through the cloud. It is believed that for terrestrial clouds , $5 < \tau < 100$, although cirrus may occur below this range (say $1 < \tau < 5$). Given certain assumptions, this parameter can be related to the vertically integrated cloud liquid water W.

Fig.4 presents some examples of the variation of cloud albedo with optical thickness (or liquid water path) for different values of solar zenith angle Θ_0 As we will see below, the parameters W and Θ_0 are necessary but not fully sufficient to define cloud albedo uniquely. Similarly, cloud emissivity can be shown to vary with optical thickness (or W) as shown in Fig. 5. If we are to speculate on the role of W as some sort of feedback mechanism in model simulations of changing climate (e.g., implied in Charlock 1981) then for those models with predictive cloud liquid water, it is imperative that the behaviour of both ρ and ε with W be incorporated in the model correctly (i.e. $\delta\rho/\delta w$ and $\delta\varepsilon/\delta w$ must be properly defined).

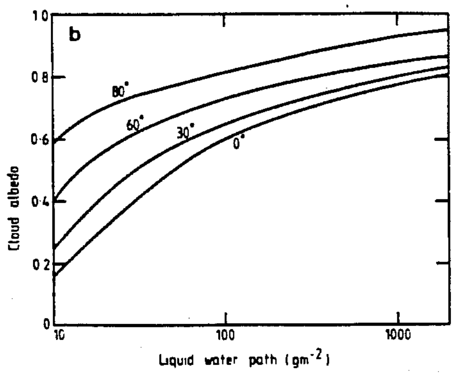


Fig. 4 Cloud albedo as a function of liquid water path W for various zenith angles (W is directly related to optical thickness) (Stephens and Webster, 1980).

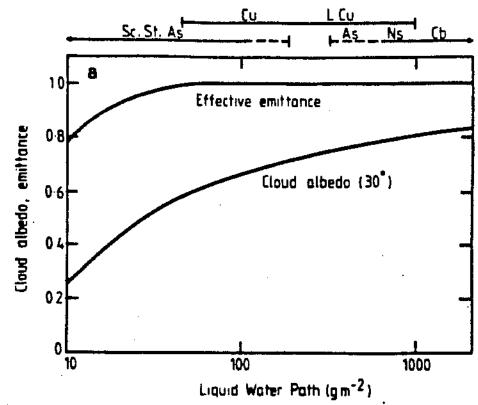


Fig. 5 Cloud emittance as a function of W composed to an albedo-W relation for $\theta_o = 30°$. Bars on upper abscissa denote approximate W ranges for various cloud species (Stephens and Webster, 1980).

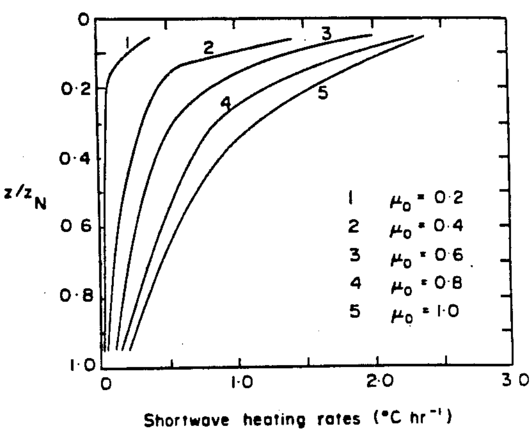


Fig. 6 Solar heating rates in a model Sc cloud for different solar zenith angles. The surface albedo is 0.3 and cloud is 0.5 km thick with a base of 1 km. Vertical scale is a normalized depth varying from 0 (cloud top) to 1 (cloud base (Stephens, 1978).

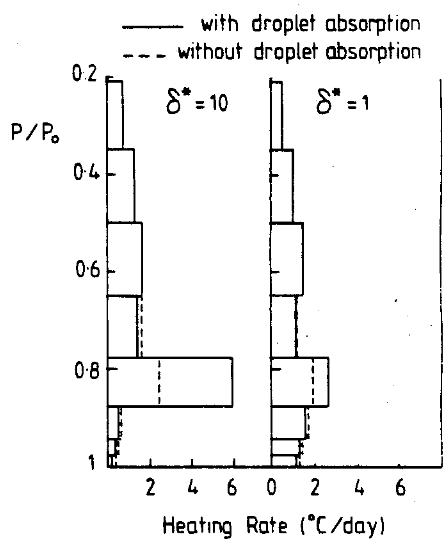


Fig. 7 Solar heating profile for $\theta_o = 0$ and two different cloud optical depths $\delta^* = 1$ and 10 ($\delta^* = \tau_o$ in text) with (solid) and without (dashed) droplet absorption (modified from Fouquardt and Bonnel, 1980).

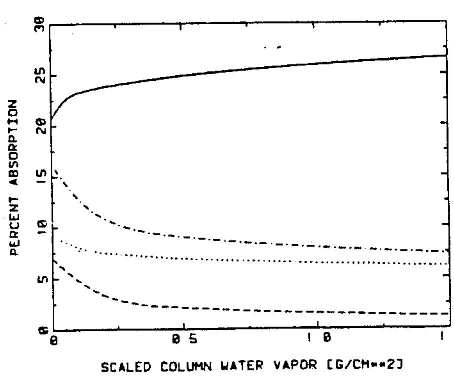


Fig. 8 Integrated absorption, as a percentage of the extraterrestrial insolation, versus scaled water vapor amount above the top of a typical 1 km thick stratus with cloud top altitude of 2 km and overhead sun: Cloud plus above-cloud absorption (solid); total cloud absorption (dot-dash); cloud droplet absorption (dots); cloud vapor absorption (from Davies et al., 1984).

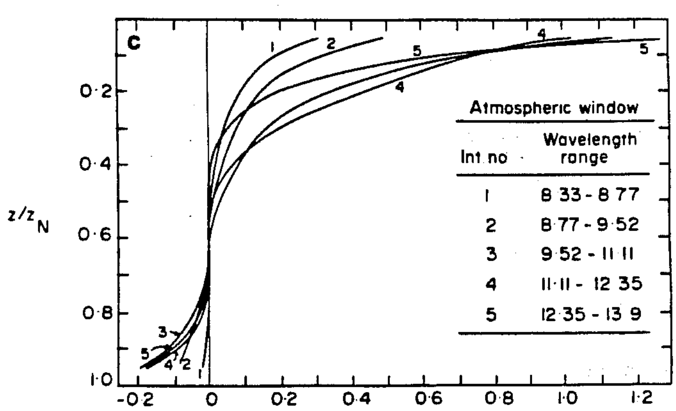


Fig. 9 The contribution to the total IR cooling by each interval of the vibration band (a) rotation band (b) and the atmospheric window region (c) (from Stephens, 1978).

So far we have said little about the actual radiative budget of the cloud layer. The radiative processes within the cloud itself produce strong radiative cooling and heating tendencies of the cloud layer as well as producing modifications to the heating and cooling of the atmosphere surrounding the cloud. Fig. 6 shows the solar heating within a model Sc cloud for different values of μ_o (=cos Θ_o). The maximum heating (ie the maximum absorption of solar energy) occurs in the upper layer of the cloud especially for small μ_o and is a decreasing function of μ_o The absorption of solar radiation within the cloud also has an impact on the atmosphere surrounding the cloud. As an example, Fig 7 shows the solar Qr profile of an atmosphere containing a cloud of optical thickness τ =10 and τ =1. The solid line profiles are calculated Qr's for both cloud droplet and water vapour absorption while the dashed line applies to the case when droplet absorption is artificially suppressed. The major effect of cloud droplet absorption is to enhance the heating of the cloud layer but even when this absorption is suppressed, Qr is enhanced by increasing the path length through the cloud. Note also how Qr is reduced below the cloud as the available energy is absorbed in the cloud and above and little of this energy penetrates below the cloud. It is clear from this diagram that the actual absorption by the droplets cannot be neglected. This point is again emphasized in Fig 8 in which the cloud layer shortwave absorption is shown as a function of the moisture above the cloud. The diagram shows liquid water absorption, vapour absorption in the cloud, the total cloud absorption and the total atmospheric absorption (cloud+above cloud absorption). We can infer from this diagram that the solar Qr in cloud will vary with altitude for relatively dry overlying atmospheres.

Examples of the longwave Qr profiles for the same model cloud are shown in Fig 9 for the four different spectral regions indicated. Qr is dominated mainly by the radiant energy exchange that occurs in the so called "atmospheric window" region of the IR gaseous absorption spectrum. This is in direct contrast to clear sky Qr values which are dominated more by the radiant energy exchanges that occur by absorption in the H_2O and CO_2 absorption bands

The contributions of the different spectral regions to Qr change with cloud altitude...low clouds are dominated primarily by cooling to space though the window region, high cloud are additionally affected by the absorption of upwelling longwave radiation from below not only in the window but increasingly in other spectral regions as the cloud elevation increases. Fig 10 reflects this change by illustrating the net longwave radiative budget of the entire cloud layer (in Wm^{-2}) as a function of the midlayer altitude.

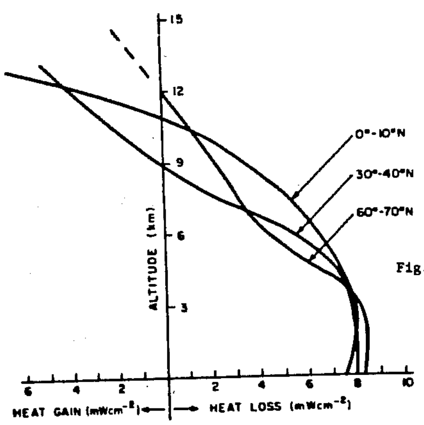


Fig. 10 Net radiative loss from a thin "black" cloud inserted at various heights in the mean atmospheres of the latitude zones (after Paltridge, 1974).

5. A simple model of ε , ρ and cloud heating Qr

The obvious question to ask is what properties of the cloud define cloud albedo and cloud emittance and how are these properties ultimately related to the radiative heating field within the cloud? As an attempt to answer these questions, we will adopt the philosophy described above and consider a simple model of the flow of monochromatic radiative flux H through some level z of a "cloud" (or any general absorbing and scattering layer for the matter). A schematic of such a model is shown in Fig 11. The flow of radiation can be considered in terms of the upwelling irradiance $H^{+}(z)$ and the downwelling irradiance $H^{-}(z)$. In addition, we also consider the following parameters:

(i) "a" as the amount of unit, normally (vertically) incident irradiance absorbed by a horizontal layer of unit thickness as it crosses a that layer

(ii) "b" as the amount of unit irradiance backscattered as it crosses a horizontal layer of unit thickness and

(iii) "D" as a measure of the "diffuseness" of the radiation such that aD is the amount of irradiance lost by absorption from the flow of unit irradiance with a distribution D as it traverses a layer of unit thickness

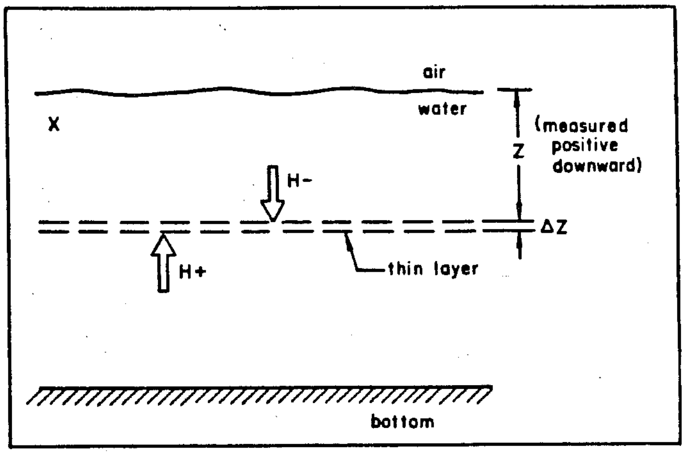


Fig. 11

Setting for the two-flow description of radiative transfer. As implied in the diagram, the setting (and solutions to follow) apply equally to the hydrosphere and atmosphere.

With these definitions in mind and taking the z dependence as applied, we can then define the change of downward irradiance after traversing a layer Δz as

$$\Delta H^{+} = -(aD+b)\Delta z\ H^{+} + bH^{-}\Delta z + (H_{\varepsilon}^{+})\Delta z \qquad (16)$$

where the first term on the right of (16) is the loss by absorption and scattering of H^{+} while the second represents the increase to H^{+} by the backscattering of the downwelling irradiance H^{-}. The term in parenthesis represents the source of H^{+} within the layer and this is described in more detail below. Similarly for H^{-}

$$\Delta H^{-} = (aD+b)H^{-}\Delta z - bH^{+}\Delta z + (H_{\varepsilon}^{-})\Delta z \qquad (17)$$

As $\Delta z \to 0$, we have the set of equations

$$\pm \frac{dH^{\pm}}{dz} = -(aD+b)H^{\pm}(z) + bH^{\pm}(z) + (H_{\varepsilon}^{\pm}) \qquad (18)$$

These equations constitute a two flow model of radiation in a horizontally homogeneous and stratified "cloud". In fact the arguments presented here were those first employed by Schuster in 1905.

5.1 The general solution of the two-flow equations

The general solution to (18) (excluding the source term) is

$$H^-(z) = m_+g_-e^{kz} + m_-g_+e^{-kz} \qquad (19)$$
$$H^+(z) = m_+g_+e^{kz} + m_-g_-e^{-kz}$$

where

$$k = \left[aD(aD+2b)\right]^{1/2}$$

and

$$g_\pm = 1\pm\frac{aD}{k}$$

and the coefficients $m\pm$ are determined from appropriate boundary conditions. The albedo of a cloud layer of thickness d follows as

$$\rho = \frac{H^+(o)}{H^-(o)} = (1-\gamma^2)\ \frac{e^{\tau_k}-e^{-\tau_k}}{(1+\gamma)^2e^{\tau_k}-(1-\gamma)^2e^{-\tau_k}} \qquad (20)$$

and the transmittance of the cloud as

$$T_r = \frac{H^-(\tau_k)}{H^-(o)} = \frac{4\gamma}{(1+\gamma^2)e^{\tau_k}-(1-\gamma^2)e^{-\tau_k}} \qquad (21)$$

where

$$\gamma = aD/k \qquad (22)$$

We also introduce the concept of the "diffuse optical depth" of the cloud as

$$\tau_k = kd \qquad (23)$$

such that the factor $\exp(-\tau_K)$ repesents the transmittance of diffuse radiation through the cloud (this interpretation is evident from (25)) - then for thick cloud (i.e τ_K >>1), the albedo of the layer becomes

$$\rho_\infty = \frac{k-aD}{k+aD} \qquad (24)$$

That is, the albedo is ivariant to cloud thickness. It also follows for thick clouds that

$$H^-(z) = H^-(0)e^{-kz}, \; H^+(z) = \rho_\infty H^-(z) \qquad (25)$$

The implications of these findings are discussed more fully below.

5.2 A particular solution to include the source of radiation as a result of the primary scatter of the directly transmitted sunlight

In the above derivation, the albedo was found to be independent of solar zenith angle μ_o, which is contrary to the behaviour described in relation to Fig 4. We can introduce this functional dependence by considering the shortwave radiation field as divided into a component of diffuse light and a component associated with collimated direct light. Thus the contribution to the general diffuse light field at z by the single scatter of a collimated light source H_o incident on the top of the cloud along the angle whose cosine is μ_o and directly transmitted to z is

$$H_\varepsilon^\pm = H_o \begin{bmatrix} b_o \\ f_o \end{bmatrix} e^{-\alpha z/\mu_o} \qquad (26)$$

where α is the volume extinction coefficient, b_o and f_o describe the forward and backscatter of the collimated flux H_o (the subscript "o" reminds us that these scattering parameters refer to the collimated beam and are therefore functions of μ_o).

The general solution of (18) with (26) follows as

$$H^\pm(z) = m_+ g_\pm e^{kz} + m_- g_\pm e^{kz} - \frac{H_o}{2} C(\mu_o, \pm) e^{-\alpha z/\mu_o} \qquad (27)$$

where

$$C(\mu_o, \pm) = \frac{f_o(b_o)b + b_o(f_o)\left[a+b\pm\alpha/\mu_o\right]}{(k+\frac{\alpha}{\mu_o})^2}$$

and all other parameters are as described above and the factor in brackets () applies for "-". The dependence of the irradiance on μ_o is now apparent through the presence of the last term in (27).

5.3 A particular solution to include the source of irradiance by thermal emission

Invoking Kirchoff's law, thermal emission can be represented by

$$H_\varepsilon^{\pm} = aB(z) \qquad (28)$$

where B(z) is the Planck black body function and a the absorption coefficient. The general solution of (18) with the source term (28) then follows as

$$H^{\pm}(z) = m_+ g_{\pm} e^{kz} + m_- g_{\pm} e^{-kz} + H_p^{\pm}(z) \qquad (29)$$

where $g_{\pm}$ and k are as defined above, $m_{\pm}$ are derived from boundary conditions appropriate to long-wave radiation and H_p (z) is the particular solution associated with the thermal source term. For an isothermal cloud layer with $B(z) = B_o$, this particular solution assumes the form

$$H_p^{\pm}(z) = B_o/\bar{\mu} \qquad (30)$$

5.4 The three basic optical properties of cloud.

It now appears that we can readily derive the cloud radiative properties that were indentified from the outset as important to our understanding of the effect of cloud on radiation if we use simple two-flow theory and given the parameters D, a, b, b_o, and f_o. As it turns out, these parameters are functions of 3 more basic properties of the cloud (which we call cloud optical properties). These optical properties are

(i) the volume extinction coefficient α (or cloud optical depth τ)

The relevance of this parameter is already apparent from the above discussions. This quantity is a relatively weak function of wavelength (λ). It is also relevant to compare τ with the diffuse optical depth. The inequality $\tau_K < \tau$ always applies as diffusely transmitted radiation includes the increase due to scattering into the forward direction and therefore is attenuated less than direct radiation along any given path.

(ii) the single scatter albedo $\tilde{\omega}_o$.

This parameter is the ratio of the scattering to total extinction by the cloud particles. Thus when $\tilde{\omega}_o = 1$, the cloud is non-absorbing (note the absorption coefficient $a=(1-\tilde{\omega}_o)\alpha$) The single scatter albedo varies strongly with wavelength generally in the range $0.9<\tilde{\omega}_o<1$ for short-wave radiation (mainly in the near IR region with $\lambda > 0.7$ μm) and less than about 0.6 for longwave radiation. That is absorption is much stronger in the long-wave region than for solar radiation. The relevance of this to parameterization will be discussed below.

(iii) the asymetry parameter g.

This parameter is a measure of the degree of anisotropy of scattering. For completely forward scatter, g=1 while g=-1 for total backscatter and g=0 when the scattering is isotropic. For real clouds g=0.7-0.9 and is relatively independant of wavelength.

There are a number of ways that these three parameters can be combined to obtain the various parameters contained within the solutions of the two flow equations(eg Meador and Weaver, 1980). An example is

$$D = 1/\bar{\mu}$$

$$b = \frac{\tilde{\omega}_o}{2}(1-g)$$

$$b_o = \frac{\tilde{\omega}_o}{4}(1-3g\ \bar{\mu}\mu_o)$$

$$f_o = \frac{\tilde{\omega}_o}{4}(1-3g\ \bar{\mu}\mu_o)$$

$$a = (1-\tilde{\omega})\alpha$$

where D is set normally between 1 and 2 (actually it can be shown that D=2 for isotropic radiation fields and D=1 for purely collimated radiation).

5.5 Some simple deductions from two flow theory

As a consequence of the simple solutions described above, it is possible to make a number of informative observations

(a) Solar radiation

If we consider the general solution for the transfer of solar irradiance through a cloud, then it is possible to demonstrate the the albedo for a cloud of thickness d is

$$\rho = \frac{H^{+}(0)}{H^{-}(0)} = \frac{C(\mu_o,-)g_+g_-}{\mu_o\Delta(d)} \left[e^{kd}-e^{-kd}\right]\frac{C(\mu_o,+)}{\mu_o\Delta(d)}\left[\frac{\Delta(d)}{\Delta(d)}e^{-\alpha d/\mu_o} -1\right] \tag{32}$$

and the transmittance is

$$Tr = \frac{H^{+}(d)}{H^{-}(0)} g_+g_-\left[e^{kd} -e^{-kd}\right]e^{-\alpha d/\mu_o} + \frac{C(\mu_o,-)}{\mu_o}\left[\frac{\Delta(0)}{\Delta(d)}e^{-\alpha d/\mu_o}\right] \tag{33}$$

where

$$\Delta(d) = g_+^2\, e^{kd} - g_-^2\, e^{-kd} \tag{34}$$

The absorptance is obtained from the combination of these two quantities. On contemplating the two limits $z\to\infty$ and $z \to 0$, the thick limit is

$$\lim_{kd\to\infty} \rho\to\rho_\infty = \frac{C(\mu_o,-)}{\mu_o}\left[\frac{g-}{g+} - \frac{C(\mu_o,+)}{C(\mu_o,-)}\right] \tag{35}$$

Here we find that the albedo of a semi-infinite cloud is a function only of sun angle (μ_o) and , as remarked above, invariant to cloud depth. It also follows that Tr is a function only of μ_o and not of depth and thus the absorption deep in a thick layer is also invariant to depth. Thus the heating field deep in a thick cloud (recall Qr is related to dA/dz) is therefore zero.This invariance limit of albedo for is apparent in Fig.4 for τ large. The thin layer limit is

$$\lim_{kd\to o} \frac{\rho(kd)}{kd} \to \frac{b_o}{\mu_o}\,\rho(kd)\to\frac{b_o}{\mu_o}\tau_k \tag{36}$$

and as expected, the albedo of a thin cloud is directly proportional to the backscatter coefficient b_o (which is itself a function of μ_o) and varies linearly with τ_K and τ. This linear regime is also evident in Fig 4. We can also deduce that the absorption varies linearly with "cloud" depth and thus Qr is constant through thin cloud. This "linear regime" of cloud absorption is the exact analogue of the "linear regime" encountered in molecular absorption studies while the invariant regime is the analogue of the well known square root law of molecular absorption.

Another implication of the solution described above , and one most important in terms of considering any attempt to parameterize Qr, concerns the extent to which the solar heating extends into the cloud. It is convenient to descibe this in terms of a scale depth of Qr which we formally define as the depth to which the heating Qr is reduced by a factor of 1/e. If we consider a thick cloud such that $\tau_K \to \infty$, it then follows that

$$m_+ = 0$$

and

$$m_- = \frac{H_o}{2} C(\mu_o,-)/g_+$$

and thus

$$H^+(z) = \frac{H_o}{2}\left[\rho_\infty C(\mu_o,-)e^{-kz} - C(\mu_o,+)e^{-\alpha z/\mu_o}\right]$$

$$H^-(z) = \frac{H_o}{2} C(\mu_o,-)\left[e^{kz} - e^{-kz} - e^{-\alpha z/\mu o}\right]$$

The net flux at level z is

$$\bar{H}(z) = H^-(z) - H^+(z)$$

from which we obtain

$$Q_r \to \frac{d\bar{H}}{dz} = \frac{H_o}{2}(1-\rho_\infty)k\, e^{-kz}\, C(\mu_o,-) - \frac{H_o}{2}\frac{\alpha}{\mu_o} e^{-\alpha z/\mu o}\left[C(\mu_o,-) - C(\mu_o,+)\right] \quad (37)$$

Since $k < \alpha$, it can be demonstrated that decay of Qr with depth is governed by the factor $e^{-\tau_K}$ for $\mu_o \to 1$ while Qr is governed by the factor $e^{-\tau}$ when $\mu_o \to 0$. This change in scale of solar heating within cloud as μ_o decreases is obvious from Fig.6. For $\mu_o \to 1$, the scale depth becomes

$$z_k = 1/k \sim \bar{\mu}/\left(\left[(1-\tilde{\omega})(1-\tilde{\omega}_o g)\right]^{1/2} \alpha\right) \tag{38}$$

Fig.12 shows typical values of z_k for different model clouds with given liquid water contents. Generally, z_k is on the order of 1 km and thus in order to resolve cloud heating adequately in a model, the vertical resolution must be of at least $O(z_k)$ unless parameterized in some fashion.

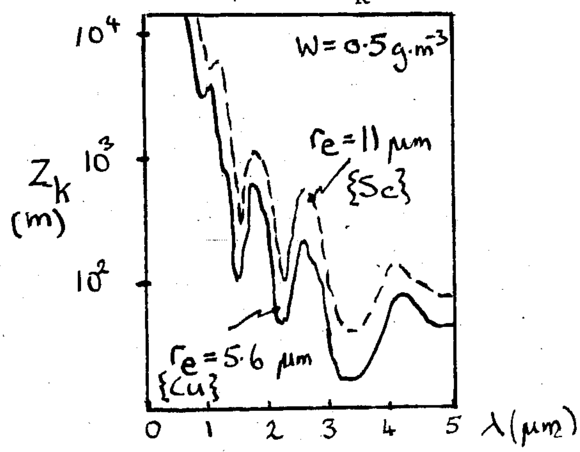


Fig. 12

The scale depth (defined in text) for shortwave heating for clouds of a given droplet size distribution typical of cloud types listed. Z_k is inversely proportional to cloud liquid water for the given distribution and is shown as a function of λ. (r_e defined in text)

(b) Infrared radiation

Long-wave radiative properties of clouds are usually considered in terms of an emissivity as we have done in the simple models of Rs and Rp. As shown in Fig. 9, the long-wave radiative characteristics of cloud are dominated by the radiative interactions confined to the "atmospheric window" region between about 8 μm to 13 μm. For the sake of illustration, and without any loss of generality, we can assume the atmosphere to be transparent in this spectral region and approximate

$$H^-(0) = 0 \qquad \text{cloud top}$$

$$H^+(d) = H_g \qquad \text{cloud base}$$

for the fluxes incident on the cloud boundaries. Here, H_g is the flux emitted from the surface below the cloud. For an isothermal cloud (29) can be written as

$$\begin{aligned} H^+(z) &= \text{Tr}\, H_g + \text{Tr}(\gamma-1)B_o + B_o(1-\gamma e^{-kz}) \\ H^-(z) &= \rho H_g + \rho(\gamma-1)B_o + B_o(1-e^{-kz}) \end{aligned} \tag{39}$$

where

$$\gamma = g-/g+$$

and ρ and Tr are the general diffuse reflectance and transmittance functions as defined above. If we use the following definition for emissivity

$$\varepsilon = \frac{H^-(d)}{B_o} \qquad (40)$$

then from (39)

$$\varepsilon = \rho\, H_g/B_o + \rho(\gamma-1)R + (1-e^{-kz}) \qquad (41)$$

and by further assuming that $\rho \sim o$

$$\varepsilon = (1-e^{-kz}) \qquad (42)$$

This form of emissivity has been extensively used in analysis of observational data and as a means of parameterizing cloud IR radiative transfer (as we will discuss in the next lecture). This emissivity is usually expressed as a function of cloud liquid water path W ($= \int w dz$, where w is the liquid water content) rather than of z. That is

$$\varepsilon = 1-e^{-KW} \qquad (43)$$

and several estimates of K have been derived from both theory and observation (a summary of some of these is shown in Table 2). In this formula, K assumes

Table 2

K	Source	Type of measurement
Boundary layer cloud		
0.13–0.16	Stephens (1978)	Theoretical
0.13	Platt (1976)	Vertical narrowband (10–12 μm) radiance
0.11–0.15	Schmetz *et al.* (1981)	Vertical narrowband (11 μm) radiance
0.13	Bonnel *et al.* (1980)	Vertical narrowband (8–14 μm) radiance
0.08	Stephens *et al.* (1978)	Broad band hemispheric irradiance
Cirrus cloud		
0.08	Paltridge and Platt (1981)	Vertical narrowband (10–12 μm) radiance
0.056	ibid	Broad band hemispheric irradiance
0.076–0.096	Griffith *et al.* (1980)	Broad band hemispheric irradiance

the gestalt of an absorption coefficient (by virtue of the assumption that $\rho = 0$) .By comparing (42) to (43)and given that $g \to 1$, then k also assumes the property of a diffuse absorption coefficient

$$k = \frac{1}{\mu}\left[(1-\tilde{\omega}_o)(1-\tilde{\omega}_o g)\right]^{1/2} \alpha \sim \frac{a}{\mu}(1-\tilde{\omega}_o) = a/\bar{\mu}$$

It is relevant to point out that even for small values of ρ , when the ratio H_g/B_o can be large (such as in the case of high cold cloud over a warm underlying surface) and the first term on the right hand side of (39) large. Thus ε is no longer a simple exponential function ; in fact ε can greatly exceed unity. This is exemplified in Fig. 13 in which the ratio ρ is expressed as a function of $H_g - B_o$. For relatively thick, high cloud (e.g. $\tau > 2$), reflection contributes substantially to the down-welling cloud base flux producing "superblack" clouds.

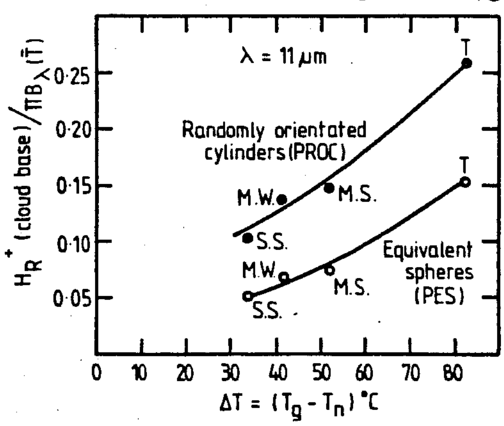


Fig. 13 The contribution to the emittance (as defined by (40)) by the reflection of longwave radiation from cloud base (i.e., first term on the right of (41)) as a function of ground to cloud temperature (i.e., of H_g to B_o). (Stephens, 1980).

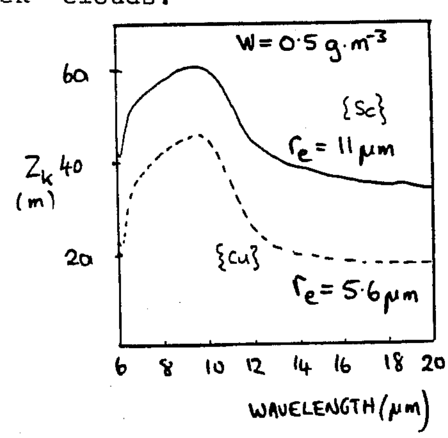


Fig. 14 Same as for Fig. 12 but for infrared radiation.

As a final illustration of the simple theory outlined above, we consider a cloud as a semi-infinite layer and invoking (39) we can duduce the following relationship

$$Q_r \to -kB_o e^{-kz}$$

Once again the scale depth of longwave heating in the cloud is determined by the factor 1/k. This parameter (z_k) is shown in Fig. 14 as a function of wavelength for different cloud models. In the infrared, z_k is smaller than the case for solar radiation by virtue of the enhanced particle absorption.

Thus to resolve the longwave heating within cloud and for the given values of liquid water taken to represent the clouds of Fig. 14, the vertical resolution of the model needs to be $0(\sim 50m)$.

6. Some uncertainties in the characterization of the basic cloud optical properties.

The preceding discussions demonstrate that the effect of cloud on the radiative budgets at the surface, top of the atmosphere and of the cloud layer itself can ultimately be defined (within the constraints of the simple model) in terms of three optical properties ω_0, α and g (or cloud optical depth). The question that we must address with the view of parameterizing cloud-radiation interaction is what cloud parameters define these optical properties and can they be taken as invariant from cloud type to cloud type. If we take 0 to represent one of the three optical properties, then the generic functional relationship applies

$$0 = E(\sigma, r, n(r), \lambda, m, \xi, q) \tag{45}$$

where
σ = particle shape characteristics
r = " size "
n = " number density
λ = wavelength of radiation
m = composition of particle
ξ = orientation of particle
q = environmental conditions (such as specific humidity)

and the function E(...) is taken to represent an appropriate solution to Maxwell's equations for a given particle shape. All but the last parameter characterize the microphysics of the cloud. How important are these individual parameters in defining $\tilde{\omega}_0$, α and g and can they be described in terms of parameters which more readily represents a bulk property of the cloud (such as cloud liquid water)? Unfortunately, the answer to these questions is not clear cut but is somewhat conditional as the following illustrations show. These examples apply only to water clouds for which we can assume the droplets to be spherical. For a given λ the variables σ, ξ and m are then irrelevant.

Fig 15 presents a spectrum of $\tilde{\omega}_o$ for visible and near IR (NIR) radiation as a result of droplet absorption alone (solid curve) and combined droplet plus vapour absorption (dashed curve). In the visible region (containing approximately 50% of the solar energy) $\tilde{\omega}_o$ =1 while typically $\tilde{\omega}_o$ >0.9 for NIR. It is important to stress that the absorption bands of liquid water (indicated by ther dips in the solid curve) closely overlap those of water vapour (enhanced dips in the dashed curve). This feature together

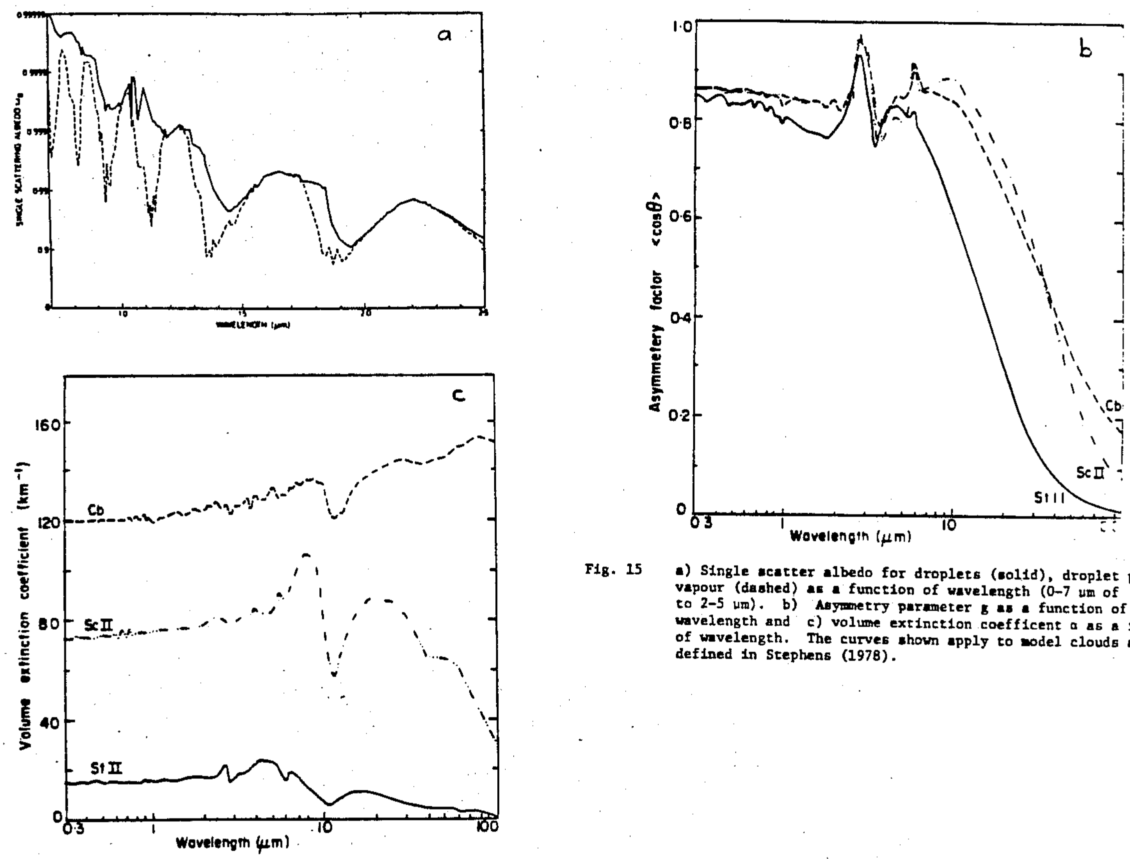


Fig. 15 a) Single scatter albedo for droplets (solid), droplet plus vapour (dashed) as a function of wavelength (0-7 um of to 2-5 um). b) Asymmetry parameter g as a function of wavelength and c) volume extinction coefficent α as a function of wavelength. The curves shown apply to model clouds as defined in Stephens (1978).

with the weakness of absorption compared to scattering (ie $\tilde{\omega}_o$ near 1) complicates the calculation of shortwave absorption. The absorption in the clear sky above cloud affects the amount of energy deposited within the cloud layer as both phases of water compete for the same energy and once removed above the cloud can no longer be absorbed in the cloud or below the cloud.

By contrast, the IR region is dominated by droplet absorption which exceeds that by vapour (at least in the "window region" where the radiant energy exchange is dominant) and simple absorption approximations have proven to be a viable way of approximating the radiative transfer processes at these wavelengths.

Also included in Fig 15 are the spectra of α and g both demonstrating an insensitivity to wavelength.

To illustrate the effect of α, g and $\tilde{\omega}_o$ on the optical properties of the cloud, it is convenient to consider the size distrubtion of cloud droplets as a modified gamma function

$$n(r) = \text{constant } r^{(1-3b)/b} e^{-r/br_e}$$

where n(r) is the volume concentration of droplets, r_e the effective radius of the distribution and b the variance of the distribution. Fig 16 a and b demonstrate the effect of both r_e and b on the cloud optical properties. In summary, these properties are fairly insensitve to b (the width of the distribution) but depend on r_e The absorption of the cloud (as indicated by $1-\tilde{\omega}_o$) increases for increasing effective droplet size. It can be demonstrated that this absorptance goes approximately as $r_e^{3/2}$ and so absorption within a thick cloud will double as r_e varies from 5μm to 8μm. Thus the impact of large droplets (e.g. precipitation) on cloud absorptance may be substantial. Another effect discussed by Twomey concerns the indirect effect of pollution on the cloud in providing greater numbers of condensation nuclei thus for producing colloidally stable clouds - i.e. for a given liquid water, r_e will be smaller and the total number density of droplets substatially larger. For fixed values of cloud liquid water, the optical thickness increases, short-wave absorption decreases (by virtue of the relation shwon in Fig. 16b for r_e) and the albedo increases. Note that α varies approximately as $1/r_e$ when liquid water is fixed. In fact it was demonstrated by Stephens (1978) for a vertically homogeneous cloud with liquid water path W, $\tau \sim 3W/2r_e$.

A more direct effect of pollution on cloud radiative properties is evident if we consider droplets containing impurities, such as small carbon particles. The scattering of such heterogeneous particles is complex and has not been satisfactorily addressed although approximate theories do provide us with some clue to the potential effect of dirty droplets on cloud radiative

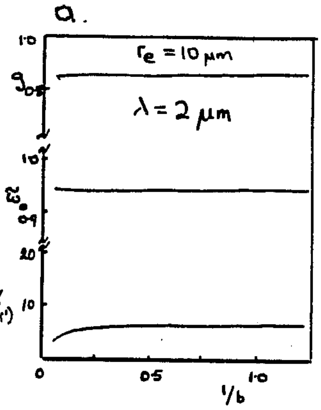


Fig. 16a The cloud optical properties as a function of 1.b for the values of r_e and λ given.

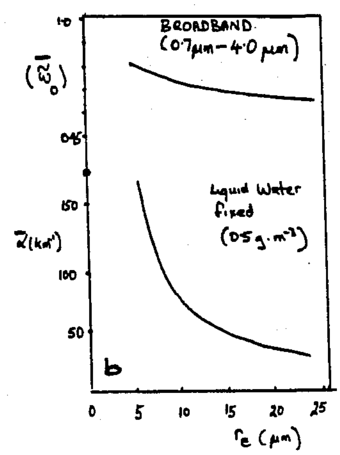


Fig. 16b The broadband (0.7 μm–4.0μm) average of the single scatter albedo and extinction coefficient as a function of r_e. Note cloud absorption is proportional to $1-\bar{\omega}_0$.

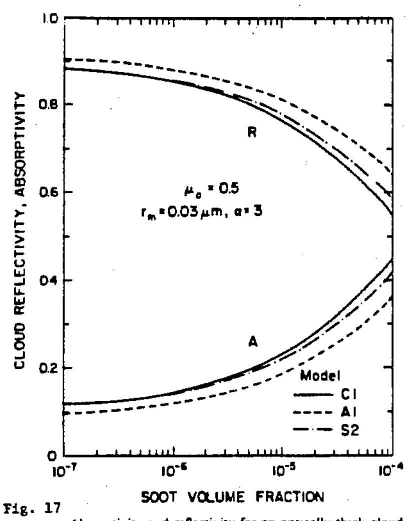


Fig. 17 Absorptivity and reflectivity for an optically thick cloud at $\lambda = 0.5\,\mu m$ as a function of soot volume fraction. The cosine of the solar zenith angle is taken to be 0.5. An amount of carbon between 5×10^{-6} and 1×10^{-5} by volume is required, depending on the cloud drop-size distribution, to obtain reflectivity of 0.8 for thick clouds.

(from Chylek et al., 1984).

properties. Fig. 17 indicates an extreme sensitivity of cloud albedo and absorptance to the volume fraction of soot included in the droplet. This diagram presents results for λ =0.5um where there is no absorption by pure water (or water vapour). Since there is a significant amount of solar energy available in this spectral region, even a small amount of absorption can produce a large effect on Qr.

The relevant optical properties of ice crystal clouds are unfortuneately functions of all parameters listed above. It is fair to say that these optical properties can only be derived for the most ideal of crystal shapes. The application of these ideal theories to real ice clouds is thus uncertain. Also basic knowledge of particle size distributions n(r) and ice water contents, while improving, is still fairly rudimentary. The need for more observations to support or refute the use of ideal approximations to E(...) is obvious. In the interim, we can conveniently approximate this function by Mie theory (i.e. assume the crystals as spherical ice particles). With this assumption, most of the above discussion applies directly to ice clouds. Even the spectral disribution of $\tilde{\omega}_0$ is similar to that of water droplets (except for a few notable spectral regions).

7. Horizontally inhomogeous cloud systems: is cloud amount an optical property?

So far all our discussions assume or imply that the interaction of radiation with cloud is primarily one dimensional with horizontal variations negligible. This is clearly not reasonable as the horizontal variability of cloud can be as extreme as the variation in the vertical. We are therefore left to ponder the following series of questions;

(i) how can we calculate the albedo, emittance and heating of a region

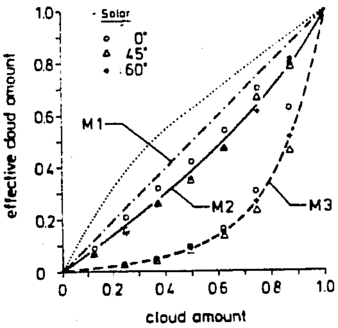
(ii) can the simple relationships that we derive for homogenous clouds be applied to the non-homogeneous atmosphere containing broken cloud (can we employ a domain averaged optical depth for example in these relationships)

(iii) or can we simply weight the properties derived from 1D theory by cloud amount and what is the relevance of cloud amount to the basic problem of radiative transfer?

We are only now developing the theoretical tools that are appropriate to address this series of questions. The following illustrations will be used in an attempt to provide some sort of answer to these queries. Consider the following in the context of (iii) above: the albedo of a broken cloud layer (given that there is no reflection from below the cloud) might be approximated as

$$\rho = A \frac{N_e}{N} \rho_\pi$$

where ρ_π is the albedo of horizontally infinite cloud of the same vertical depth of the broken cloud, A is the fraction of the area of interest covered



Solar effective cloud amount versus actual cloud amount. Different signs denote wavelength-integrated Monte Carlo results for 3 zenith angles. Lines pertain to parameterizations for different cloud models --- M1 (area weighting with cloud amount). —— M2 (cloud grows only in horizontal dimensions) M3 (cloud grows in three dimensions). ······ solar effective cloud amount after Harshvardhan (1982)

Fig. 18 (from Schmetz, 1984).

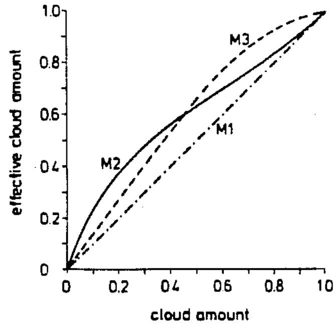


Fig. 19 Infrared effective cloud amount versus actual cloud amount (Schmetz, 1984).

by cloud and the ratio Ne/N is a factor that is introduced to represent the differences that might arise from the horizontal inhomogeneity of our cloudy domain when compared to the common homogeneous cloud assumption. The product Ae=ANe/N is the "effective" cloud amount and it is customary to assume Ne/N=1. As it turns out, this ratio is not unity for a regular array of cube clouds and it seems to vary with the way the clouds are distributed within the area. Figs.18 and 19 illustrate Ae versus A for different cloud distributions as derived from Monte Carlo calculations. M1 represents the usual solid cloud assumption (i.e. Ne/N=1), M2 an array of clouds of fixed vertical extent but with a horizontal extent that varies with A while M3 apply to cubic cloud again with dimensions varying with A. The relationship between Ae and A was derived for albedo (Fig. 18) and emittance (Fig. 19). These diagrams suggest that there is little justification in simply weighting the clear and cloudy radiative properties in a linear manner with cloud amount (i.e. M1 curves). It is also evident that there is not a unique relationship between cloud amount A and the domain averaged albedo and emittance. We are thus led to conclude that cloud amount is not a fundamental radiative property of an inhomogeneous cloudy atmosphere. The parameter does not appear in the governing equations that describe the radiative transfer through such a medium and all treatments of a partially cloudy environment employing A are based on empiricism and faith.

The second issue raised implies the use of some spatial average of the optical properties in the relationships derived from 1-D radiative transfer theory. It is clear that this approach will also be unsuccessful; ε and ρ are non-linear functions of these optical properties. Thus

$$\bar{\rho} = \int f(\tau)d\tau / \int d\tau \neq \rho(\bar{\tau})$$

where f is taken to represent such a relationship and where the integral applies over our domain of interest. The issue raised above in relation to (iii) and the results shown in Figs. 18 and 19 suggest that under conditions of horizontal non-uniformity even the function f is not applicable.

Given these comments, it would seem that much of the discussion of this lecture is inapplicable to conditions of broken cloud. However this is not the case. This has support from the following two considerations:

(i) the three optical properties $\tilde{\omega}_o$, α and g are equally relevant and basic to 3-D radiative transfer and

(ii) it has been demonstrated in recent theoretical work on this subject that solutions can be formulated in such a way that the resemble the 1-D counterparts (Preisendorfer and Stephens, 1984).

As further support, consider a partially cloudy environment as one in which the basic optical properties vary, then a cloud (for instance) exists where perturbations of α (or τ) exist which are superimposed on some background value of α_o Adopting this view Stephens (1985) obtained quite general solutions of the multi-dimensional radiative tranfer equation. Fig 20 is offered as an example of a particular solution for a roll cloud (a 2-D cloud with a Gaussian variation of τ in the horizontal). The domain averaged albedo is shown as a function of the averaged optical depth $\bar{\tau}$ and is compared to the albedo derived for a horizontally uniform cloud with the same averaged optical depth. The diagram shows that both the linear and invariant limits exist for this 2-D cloud and that for small optical depths (τ<5).

$$\rho_{\pi}(\bar{\tau}) \sim \bar{\rho}$$

as a result of the linearity of the relationship between ρ and τ. The invariant limit differs from the homogenous cloud limit and is sensitive to the spatial variations of the cloud optical properties.

In these theories, the spatial variation of cloud appear as "apparent" optical properties in the solutions thus suggesting that it may be possible to incorporate broken cloudiness directly into the definition of $\tilde{\omega}_0$, α and g An example would be

$$\alpha(\text{broken cloud}) = \bar{\alpha} + \ell(\bar{\alpha}, \sigma_\alpha \text{ ....})$$

where $\ell(\bar{\alpha}, \sigma_\alpha,)$ is some function representing the distribution of α within the domain.

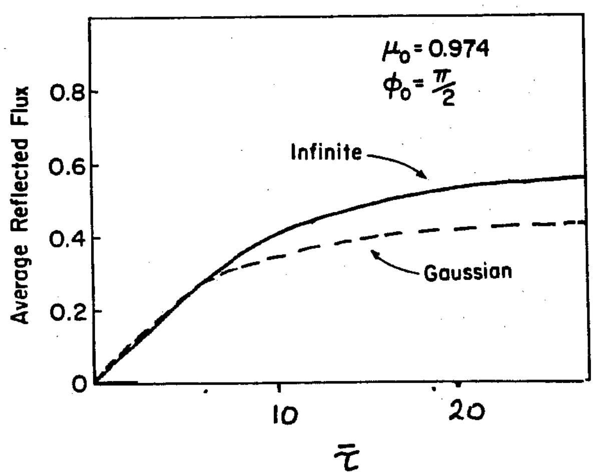


Fig. 20 The domain averaged albedo as a function of average optical thickness for a horizontally homogeneous cloud (infinite) and a cloud with a Gaussian optical depth distribution in the horizontal (modified from Stephens, 1985).

8. Summary

In the lecture we introduced simple models of the surface radiation budget, the planetary radiation budget and cloud radiative heating rates in order to describe the effect of certain cloud properties on these parameters. From these models the change in the albedo and emittance of a layer were identified as providing the major impact of cloud on the radiative budgets at the boundaries of the atmosphere. In further developing a simple "two flow" model of radiative transfer in a cloud layer we could demonstrate that the requisite properties of cloud that describe such a change could be derived in terms of 3 given cloud optical properties - single scatter albedo, cloud optical depth (or equivalently extinction coefficient) and the asymmetry parameter. We also demonstarted simply how the albedo and emittance of cloud varies as a function of a number of different parameters and introduced the notion of scale depth to characterized the depth over which radiative heating and cooling applies within a cloud.

From the discussions presented and the examples shown, it is clear that one of the important aspects of clouds on radiation is that they redistribute radiant energy in both vertical and horizontal. The treatment of the horizontal redistribution of radiation to date involves empirical relationship using a "cloud amount" parameter. It is suggested that this parameter is in fact not fundamental to radiative transfer and that other means of treating an inhomogeneous atmopshere are possible.

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