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Cloud and aerosol effects on the solar heating rate of the atmosphere

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ABSTRACT

Band-by-band calculations are carried out to investigate the solar heating rate and radiative property of the atmosphere containing various combinations of absorbing gases, aerosols and the cloud. The solar spectrum is divided into nine bands according to the location of the absorbing gases which include water vapor, ozone, oxygen and carbon dioxide. The radiation transfer program takes into consideration the inhomogeneity of the cloudy and aerosol atmospheres, the wavelength dependence of solar radiation and the gaseous absorption within scattering layers. A cumulus in the lower troposphere procreates a heating rate as large as 12° C day^{-1} when the sun is overhead and generates additional heating rates due to ozone in the lower stratosphere. Aerosols are shown to have a pronounced influence on the heating rate of the lower clear atmosphere, but not for coudy conditions. When the sun is close to horizon, the effect of the cloud on the heating rate is shown to be unimportant. We further illustrate that significant under- and overestimation of the atmospheric absorption and reflection, respectively, would be anticipated if the absorption caused by ozone and oxygen is ignored, particularly in cloudy atmospheres.

1. Introduction

The absorption of solar radiation in the earth-atmosphere system represents the initial source of energy which drives the atmosphere in motion. In addition to various gases, the atmosphere is also composed of clouds and aerosol particles. Through these gases and particulates, the absorption and scattering processes take place and the absorbed energy is transferred to differential heat which causes the temperature variation. Although the absorption of solar radiation by water vapor, carbon dioxide, oxygen and ozone has been well understood, effects of clouds and aerosols on the solar heating rate of the atmosphere have not been investigated comprehensively. Clouds cover regularly over about 50% of the planet earth and represent the most important modulators of radiation in the earth-atmosphere system. Thus, it is of vital importance to obtain quantitative information on the influence of the cloud on the solar heating rate under various atmospheric conditions.

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We investigate the solar heating rates and radiative properties of atmospheres containing absorbing gases, aerosols and clouds. The radiation program utilized here follows the previous studies by Liou & Sasamori (1975) and Liou (1976) for aerosol and cloudy atmospheres, respectively. These developments have taken into consideration the inhomogeneity of the plane-parallel atmosphere, the wavelength dependence of solar radiation and the gaseous absorption in scattering layers. We employ a standard atmospheric profile for molecules for Rayleigh scattering calculations. The aerosol concentration profile used is that suggested by McClatchey et al. (1971) for a clear atmosphere whose ground visibility is about 23 km. The refractive indices of aerosols in the solar spectrum and the aerosol particle size distribution are taken from tabulated values provided by Shettle

(1975) for hygroscopic and dust-like particles. As for the cloud, we have selected a typical fair-weather cumulus cloud whose base is placed at 1.7 km with a thickness of 0.45 km. The observed particle size distribution by Batan & Reitan (1957) and the refractive indices of water given by Hale & Querry (1973) are further utilized for single-scattering computations. Water vapor and ozone profiles are taken from the report by McClatchey et al. (1971) for a tropical atmosphere.

Band-by-band radiative transfer calculations are carried out for four model atmospheres consisting of (1) molecules, (2) molecules and aerosol particles, (3) molecules and cloud particles, and (4) molecules plus aerosol and cloud particles. Differential heating rates for the entire solar spectrum in the troposphere and lower stratosphere are presented for a number of solar zenith angles and surface albedos. Reflection, absorption and transmission of solar radiation by the above four model atmospheres are finally evaluated and discussed.

2. Brief description of the radiation program

The basic radiation model used in this study is a modified version of the models of Liou & Sasamori (1975) and Liou (1976). The model utilizes the discrete-ordinate method for monochromatic radiative transfer (Chandrasekhar, 1950; Liou, 1973) with applications to inhomogeneous atmospheres by matching the intensity components at pre-divided homogeneous layers (Liou, 1975). The major improvements of the present model are that the absorption of solar radiation by ozone, oxygen and carbon dioxide is included in the transfer calculations and that the model allows the variations of clouds and aerosols in the atmosphere.

The solar spectrum is divided into nine spectral intervals according to the positions of the absorption bands. The central wavelengths of these bands are 0.3, 0.5, 0.7, 0.94, 1.1, 1.38, 1.87, 2.7 and 3.2 μm. The first two bands are associated with ozone absorption, while the third band is primarily due to oxygen absorption. The rest of these bands are dominated by water vapor absorption. We have used the solar constant and solar spectrum recently derived by Thekaekara (1974) to compute the percentage of solar flux within each band. The absorption coefficients for the ozone in the Chappius band and the u.v. Hartley and Huggins bands are taken from Vigroux (1953) and Inn & Tanaka (1953), as presented in the *Handbook of Geophysics* (1960). Incorporation of the ozone absorption into the transfer program for cloud-free and cloudy atmospheres is straightforward, since its absorption coefficients are continuous. The empirical formula for the absorption of molecular oxygen derived by Yamamoto (1962) is adopted i- this study. As for water vapor and carbon dioxide, empirical absorptivities determined by Howard et al. (1956) are utilized. There are a number of weak carbon dioxide bands in the solar spectrum, namely, 1.4, 1.6, 2.0 and 2.7 μm bands. However, we find that the absorption contribution due to carbon dioxide in the solar spectrum can be completely ignored in the troposphere, where water vapor absorption is dominant, and the lower stratosphere, where the ozone is of prime importance.

Incorporation of water vapor, oxygen and carbon dioxide absorption in scattering atmospheres is accomplished by the exponential fitting of band transmissivities as described by Liou & Sasamori (1975). Effect of the pressure dependence on absorption in an inhomogeneous atmosphere has been taken into consideration. For each band, except 0.3 and 0.5 μm, a set of equivalent absorption coefficients k_m and weights w_m are derived. It follows that these bands are divided into a number of sub-bands from which monochromatic transfer calculations involving the gaseous absorption and scattering and absorption of aerosols and cloud particles can be carried out. Mie scattering calculations for the central wave-

Table 1. *Refractive indices of aerosols and water*

Central wavelength	Aerosols		Water	
	Real	Imaginary	Real	Imaginary
0.3	1.53	0.008	1.349	1.60×10^{-8}
0.5	1.53	0.005	1.335	1.00×10^{-9}
0.7	1.53	0.007	1.331	3.35×10^{-8}
0.94	1.52	0.016	1.327	2.93×10^{-6}
1.1	1.52	0.017	1.326	6.39×10^{-6}
1.38	1.51	0.020	1.321	1.38×10^{-4}
1.87	1.45	0.014	1.309	6.07×10^{-4}
2.7	1.40	0.055	1.188	0.019
3.2	1.43	0.008	1.478	0.0924

lengths of the bands are made for the aerosols and cloud particles. It is assumed that the size distribution of aerosols is height-independent, and may be described by the power law distribution. The real and imaginary refractive indices of aerosols covering the solar spectrum employed in this study are listed in Table 1. Basically, these values are for hygroscopic and dust-like particles (Shettle, 1975). Also presented in the table are the refractive indices for water (Hale & Querry, 1973) associated with scattering calculations involving cloud droplets.

The size distribution used for cumulus is the observed spectrum obtained by Battan & Reitan (1957) with a total number density of 300 cm^{-3}. The phase functions and scattering and extinction cross-sections are evaluated for each central wavelength of the band. In view of the slow varying refractive indices of water and aerosols in the solar spectrum, these parameters for a given wavelength are assumed to be constant over the corresponding band.

The atmospheric water vapor and ozone concentration profiles used in this investigation are based on the climatological data for a moist tropical atmosphere compiled by McClatchey et al. (1971). For carbon dioxide and oxygen, constant mixing ratios of 0.033% and 20.95% are assumed in the calculations. The total precipitable water in the moist tropical atmosphere is about 5.2 gm cm^{-2}. Note that the water vapor, temperature as well as molecular density profiles have been presented in a previous paper by Liou (1976).

The atmosphere is divided into twelve layers, each of which is assumed to be homogeneous with respect to the single-scattering parameters defined below. Except for the three bottom layers where the division is according to the location of the cloud, each layer has a thickness of 3 km. For each layer and spectral interval, the optical depth, the single-scattering albedo and the phase function are defined, respectively, by

$$\Delta\tau_\lambda^l = \Delta\tau^R + \Delta\tau_s^A + \Delta\tau_a^A + \Delta\tau_s^C + \Delta\tau_a^C + k_m \Delta x^l,$$

$$\tilde{\omega}_{0,\lambda}^l = \frac{\Delta\tau^R + \Delta\tau_s^A + \Delta\tau_s^C}{(\Delta\tau^R + \Delta\tau_s^A + \Delta\tau_s^C) + (\Delta\tau_a^A + \Delta\tau_a^C + k_m \Delta x^l)},$$

$$P_\lambda^l(\theta) = \frac{\Delta\tau^R P^R(\theta) + \Delta\tau_s^A P^A(\theta) + \Delta\tau_s^C P^C(\theta)}{\Delta\tau^R + \Delta\tau_s^A + \Delta\tau_s^C},$$

where the superscript l denotes the layer, R, A and C the Rayleigh, aerosol and cloud terms, respectively, the subscript λ denotes the wavelength of the spectral interval, s and a the scattering and the absorption contributions, respectively, and m the number of sub-spectral intervals, k_m represents the equivalent absorption coefficient derived from the exponential fit to the empirical absorptivity, and Δx^l is the pressure corrected path length for gases. For layers with no clouds, $\Delta\tau_a^C = \Delta\tau_s^C = 0$.

Having the single-scattering parameters calculated for the model cloudy and aerosol atmospheres, transfer computations are then followed. The solar radiation program begins with solving the monochromatic transfer equation for plane-parallel homogeneous layers which are predivided according to the cloud location. Upon introducing the appropriate radiation boundary and continuity equations into the analytic solution of the transfer equation, a system of linear equations can be derived and solved by numerical means. Intensity distributions can then be evaluated within the inhomogeneous atmosphere. It follows that upward and downward fluxes can subsequently be obtained. Spectral flux divergences and spectral solar heating rates may now be computed for each layer, as are the total absorption, reflection and transmission values. These may be accomplished by proper summation of the radiation parameters over the sub-spectral intervals according to their percentage of solar flux.

Let index i (=1, 2, . . ., 9) denote the number of the spectral band, $f_{\Delta\nu_i}$ the amount of solar flux within the ith band, and S_0 the solar constant, then the solar heating rate, reflection, total absorption and transmission for the entire solar spectrum are defined, respectively, by

$$\frac{\partial T}{\partial t} = \sum_{i=1}^{9} \left(\frac{\partial T}{\partial t}\right)_i,$$

$$\gamma = \sum_{i=1}^{9} \gamma_i f_{\Delta\lambda_i}/S_0,$$

$$\alpha = \sum_{i=1}^{9} \alpha_i f_{\Delta\lambda_i}/S_0,$$

$$t = \sum_{i=1}^{9} t_i f_{\Delta\lambda_i}/S_0.$$

Note here that α_i is obtained from the net flux divergence for the entire atmosphere and t_i is the total transmission including direct and diffuse components.

3. Results and discussion

Four model atmospheres have been chosen for radiation calculations. The first model consists of only molecules. Thus, Rayleigh scattering and gaseous absorption take place in the atmosphere. In addition to molecules, we introduce an additional aerosol concentration profile typical of that in a clear atmosphere (McClatchey et al., 1971). This constitutes the second model. The third model contains molecules and a cumulus in the lower atmosphere. The final model includes molecules, aerosols and the low cumulus. These four model atmospheres would allow us to investigate the relative contributions of various gases and particulates to the solar heating rate and the radiative property of the atmosphere.

Fig. 1 shows the solar heating rate for the four model atmospheres when the sun is overhead (cosine of the solar zenith angle $\mu_0 = 1$). Two surface albedos of 0.1 and 0.4 representing, perhaps, the ocean surface and bright sand (or old dirty snow), respectively, are utilized to study the surface reflection on the heating rate. In a clear atmosphere, the heating rate arises mainly from the absorption due to water vapor. For the surface albedo of 0.1, a maximum heating rate of about 3.4 °C day^{-1} is seen at about 2 km. It decreases rapidly with height associated with the decrease of the water vapor concentration. Then the heating rate increases drastically owing to the absorption of solar ultraviolet radiation by ozone. It is the main reason for the increase of the temperature in the stratosphere. When an aerosol distribution whose ground visibility is about 23 km is introduced, the heating rate increases by about 0.5° C day^{-1} near the ground. This increase is caused by the additional absorption by aerosols in the solar spectrum. It is clear that the heating rates caused by aerosols are primarily in the lower atmosphere. Effects of the clouds on the solar heating are of vital significance. The peak heating rate generated within the cloud is caused by the convergence of the net flux at the cloud top and the almost negligible net flux at the cloud bottom. The atmosphere above the cloud experiences the additional heating rate produced by the reflection of the cloud. On the contrary, the heating rate below the cloud reduces to a small value. Introducing additional aerosols to the cloudy atmosphere appears to produce insignificant effect on the heating rate profile. Note that the visible optical depth of the cumulus whose thickness is 0.45 km is about 25, which causes a heating rate as large as of 12° C day^{-1} within the cloud. The absorption of water vapor as well as cloud droplets is the prime responsibility for such a large heating rate. Since the radiative transfer program takes into account simultaneous processes involving absorption of water vapor and scattering and absorption of cloud droplets, it is not possible to separate individual contributions to the heating rate within the cloud. As for the surface albedo of 0.4, increases of the heating rates in the lower atmosphere are evident

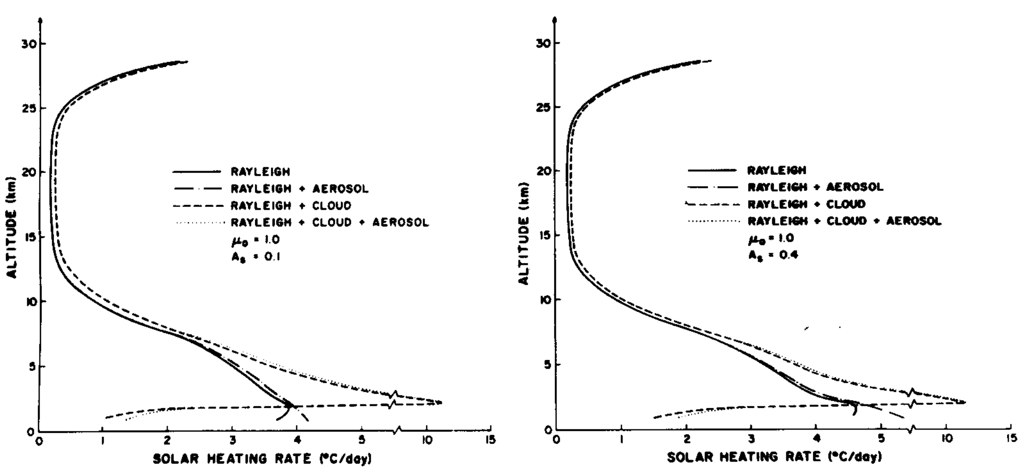


Fig. 1. The solar heating rate for the four model atmospheres when the cosine of the solar zenith angle $\mu_0 = 1$. The left- and right-hand sides of the graph are for surface albedos of 0.1 and 0.4, respectively.

for the clear and aerosol atmospheres. Near the ground level, an increase of about 0.5° C day^{-1} is to be anticipated. We also see a slight increase of heating rates in the lower stratosphere. However, effects of the ground reflection appear unimportant in cloudy conditions owing to the dominant influence of the cloud on radiation processes. It is interesting, though, to notice an increase of the heating rate below the cloud when a higher albedo is encountered.

Fig. 2 illustrates the solar heating rates for the four model atmospheres under the same conditions as those in Fig. 1, except for a cosine of the solar zenith angle of 0.6. The solar heating rates generally decrease because the available solar flux normal to the top of the atmosphere is reduced by a factor of μ_0. For the clear and hazy atmospheres, we see that the heating rate differences between the two cases are approximately a factor of 0.6. This reveals that effects of the variation of the sun's zenith angle from 0° to about 30° on the absorption and scattering processes are not very significant. Note here that the surface reflection effect is also clearly shown in this diagram. In a cloudy atmosphere, however, we see a larger reduction of the heating rate above, within and below the cloud when $\mu_0 = 0.6$. The additional decrease is probably due to the increased multiple scattering processes within the cloud layer when the sun's zenith angle is low. Lacis & Hansen (1974) also computed the solar heating rate in cloudy atmospheres without considering the absorption contributions from aerosol and cloud particles owing to their parameterization treatment of water vapor absorption within scattering atmospheres. Utilizing a mid-latitude winter atmosphere whose surface albedo is 0.07, they showed a heating rate of about 2.3° C day^{-1} for the cloud layer with a visible optical depth of 8 located at 3 km when $\mu_0 = 0.5$. Our calculations employing $A_s = 0.1$, $\mu_0 = 0.6$, a tropical atmosphere and a visible cloud optical depth of 25 give a heating rate of about 5.5° C day^{-1}. Beside the different atmospheric water vapor profile and the cloud optical depth employed in the calculations, the distinct difference seems to be due, in part, to the neglection of cloud droplet absorption. The importance of cloud droplet absorption has also been noted recently by Welch & Zdunkowski (1976). Influence of the low zenith angle can be realized in Fig. 3 where a $\mu_0 = 0.2$ is employed in the calculation. For cloudy atmospheres, it is evident that scattering processes dominate and the heating rate within the cloud reduces drastically. The surface reflection effect on the heating rate appears insignificant when the sun is close to the horizon.

Results of the transmission, reflection and absorption of the four model atmospheres are depicted in Fig. 4. For clear and hazy atmospheres denoted by the solid and dashed lines, respectively, the dependence of the transmission on the cosine of the solar zenith angle is clearly illustrated. For a clear atmosphere whose surface albedo is 0.1, the atmospheric transmission is about 61% when the

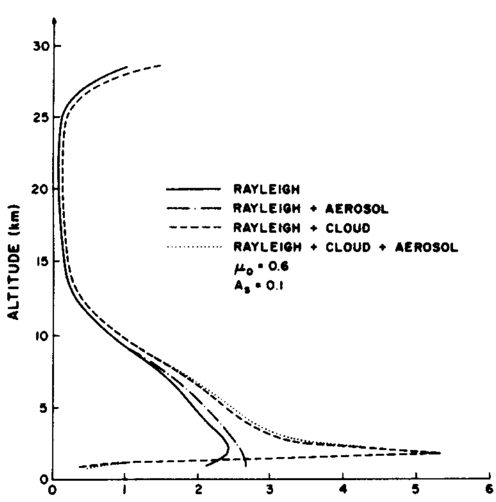
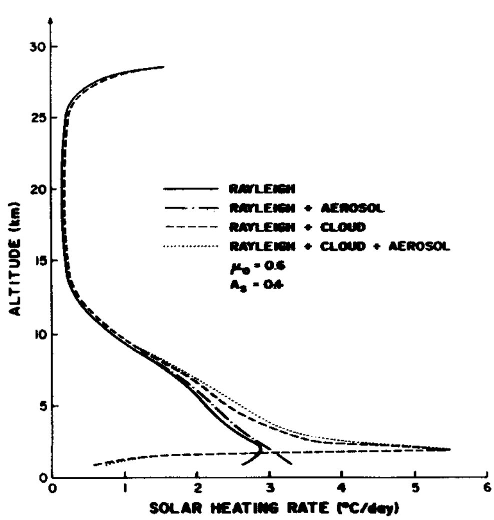


Fig. 2. Same as Fig. 1; except for $\mu_0 = 0.6$.



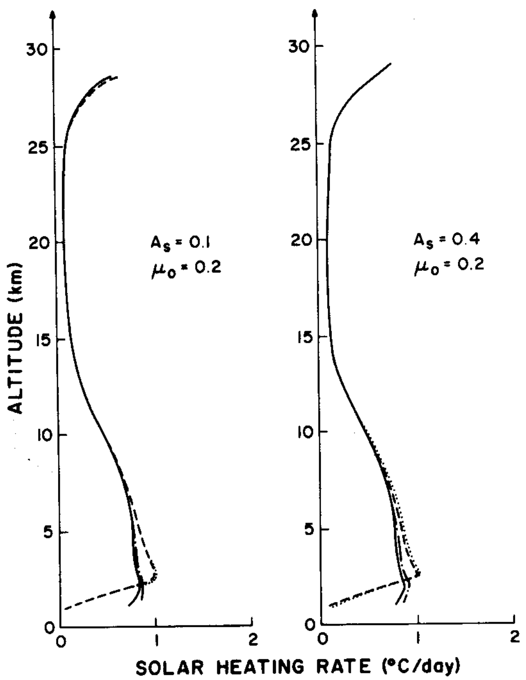


Fig. 3. Same as Fig. 1, except for $\mu_0 = 0.2$.

sun is overhead and decreases to about 50% when $\mu_0 = 0.5$. With the additional load of aerosol particles in the troposphere the transmission reduces by about 3–5% for most of the solar zenith angle. Clouds block most of the solar radiation as evident from our daily experience. Even with a thin cumulus whose thickness if 0.45 km in the atmosphere, the transmission of solar radiation yields a value of no more than about 18% or so. Comparisons with the previous investigation by Liou (1976) whose investigation did not consider the absorption by ozone, oxygen and carbon dioxide we find our present transmission values are about 2% lower than his values. In a cloudy atmosphere effects of aerosols, whose main concentration is below the cloud, on the transmission of solar radiation seem to be quite small with an additional decrease of about 1%. A larger surface albedo of 0.4 increases the transmission of solar radiation for all the four model atmospheres arising from the reflection component of solar radiation from the surface. As for the reflection values when $A_s = 0.1$, the clear atmosphere reflects about 12% of the incoming solar radiation when the sun is overhead. The reflection values increase when the sun moves to the horizon position owing to the longer scattering path length. Adding aerosols in the troposphere indicates an increase of reflection by about 1–4% with the maximum increase occurred at about $\mu_0 = 0.3$. In the cloudy atmosphere containing a cumulus cloud, reflection values range from 42–47% depending upon the position of the sun. Interestingly, adding aerosols in that cloudy atmosphere reduces the atmospheric reflection for μ_0 ranging from 1 to about 0.1. The surface reflection ($A_s = 0.4$) shows an important effect on the increase of the atmospheric reflection in clear and hazy conditions. It is so significant that the increases are as large as 15%. We also notice that the additional aerosol loading in clear and cloudy atmosphere is to reduce the atmospheric reflection for most of the solar zenith angle.

Lastly, we discuss the relative contributions of gases, aerosols and cloud particles on the total absorption within the atmosphere. For a clear atmosphere with a surface albedo of 0.1, the atmosphere absorbs about 34% when the sun is overhead and about 56% when the sun is close to the horizon. Input of a light aerosol concentration reveals an increase of absorption of about 3–5%. In the previous study by Liou & Sasamori (1975), the total absorption in a clear atmosphere containing a light aerosol concentration when the sun is overhead is about 27%. This value underestimates the total absorption by about 7% owing to the neglection of the ozone and oxygen absorption in their study. When the cloud is introduced to a clear atmosphere the total atmospheric absorption increases to about 39% when the sun is overhead and to about 55% when the sun moves to the horizon. Comparison with the earlier study by Liou (1976) who obtains the total absorption of about 27% we see the significance of neglecting the absorption contribution due to ozone and oxygen. The Hartley and Huggins bands of ozone (0.3 μm band) consist of about 7% of the total solar flux a portion of which is absorbed in the stratosphere. The absorption of solar radiation by oxygen A band is equally important in view of the large portion of solar flux within the band. That the absorption of oxygen in the 0.7 μm band reduces significantly the reflection of solar radiation in cloudy atmospheres. Introducing a light aerosol concentration to the cloudy atmosphere shows a slight increase of total absorption which is less than the increase produced in the clear atmosphere. Note that the increase of the total atmospheric absorption in cloudy atmospheres when the sun moves toward the limb is not as drastic as compared with that in clear and aerosol atmospheres. It is likely that the scattering processes involving clouds dominate the atmospheric radiation field. The domination of the cloud scattering is also evident when a higher surface albedo of 0.4 is considered. The total absorption does not increase appreciably in cloudy atmospheres. The absorbed solar energy within the atmosphere represents the initial energy source which drives the atmosphere in motion.

Finally, we note that absorption values obtained in the present transfer calculations for a clear atmosphere apparently are larger than those previously reported. In this investigation, however, absorption due to water vapor, carbon dioxide, oxygen and ozone along with multiple scattering effects of molecules are simultaneously taken into consideration. In addition, the transfer program also includes the surface reflection effect. It is likely that multiple scattering and surface reflection play significant roles in the increase of total absorption within a clear atmosphere.

4. Conclusion

The solar heating rates and radiative properties of the troposphere and lower stratosphere are investigated for four model atmospheres containing various combinations of absorbing gases, aerosols and cloud particles. The solar spectrum is divided

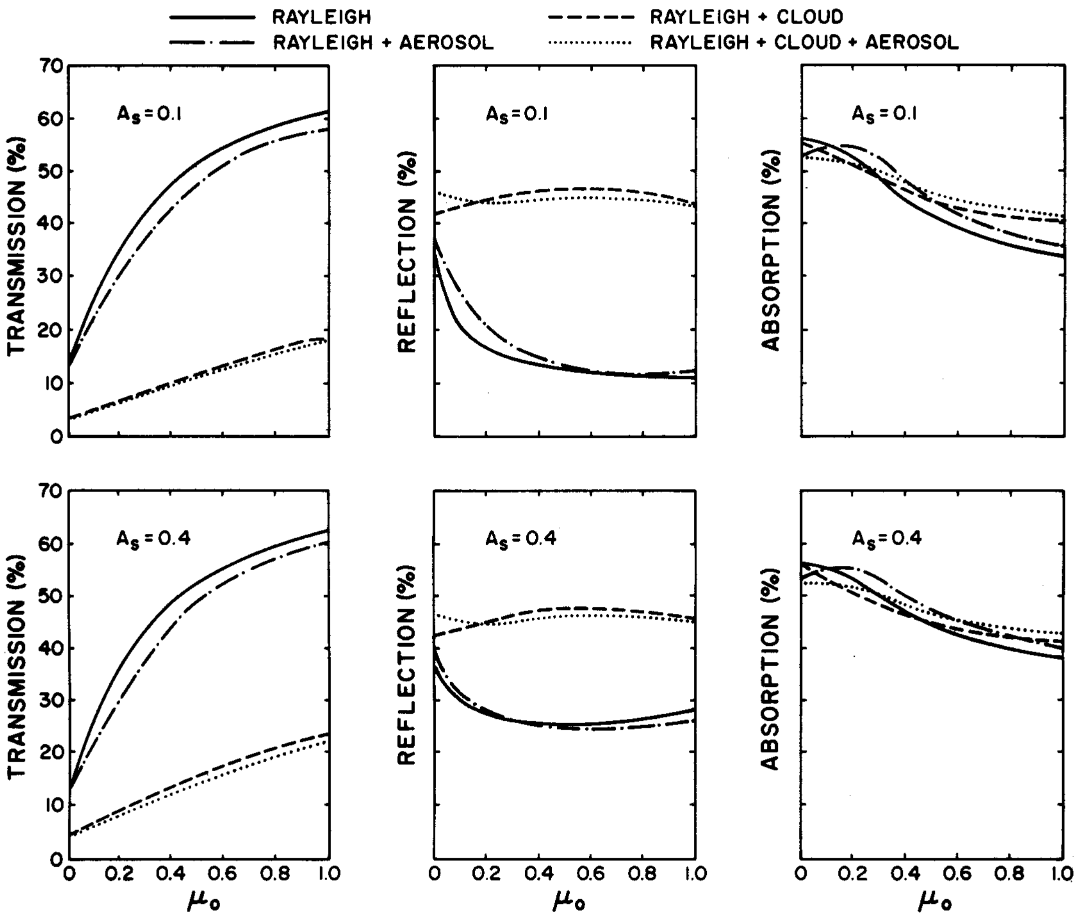


Fig. 4. Transmission, reflection and absorption of the Rayleigh (solid lines), Rayleigh and aerosol (long and short dashed lines), Rayleigh and cloudy (dashed lines), and Rayleigh aerosol and cloudy atmospheres (dotted lines) as functions of μ_0 for two surface albedos of 0.1 and 0.4.

into nine bands according to the location of the absorbing gases which include ozone, oxygen, water vapor and carbon dioxide. Inhomogeneous atmospheres are divided into twelve layers, each of which is considered to be homogeneous with respect to the single scattering properties. Band-by-band transfer calculations are carried out in which the inhomogeneity of the cloudy and hazy atmospheres, the wavelength dependence of solar radiation and the gaseous absorption within scattering layers are taken into account.

The significant results based on the present calculations are summarized below.

(1) A cumulus cloud in the lower troposphere generates a heating rate as large as of 12° C day^{-1} when the sun is overhead. The contributions to the heating are caused by the absorption of water vapor as well as cloud droplets.
(2) Aerosols have a pronounced influence on the heating of a clear atmosphere. In a cloudy atmosphere, however, their effect on the solar heating rate appears small.
(3) The solar heating rates are shown to decrease rapidly when the sun moves to the horizon position. Effects of the cloud on the heating rate when the sun is low is insignificant.
(4) A tropospheric cloud also generates additional heating rates due to ozone in the lower stratosphere.
(5) Surface albedo effects are shown to be important in clear and hazy atmospheres, but not in cloudy conditions.
(6) Absorption due to ozone and oxygen is shown to have pronounced effects on the total atmospheric absorption and reflection. Underestimations of the total atmospheric absorption by about 7 and 13% would be anticipated in clear and cloudy atmospheres, respectively, if the absorption of solar radiation by ozone and oxygen is ignored.

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ВЛИЯНИЕ ОБЛАКОВ И АЭРОЗОЛЯ НА СКОРОСТЬ НАГРЕВАНИЯ АТМОСФЕРЫ СОЛНЦЕМ

Проведены спектральные расчеты для исследования радиационных свойств атмосферы и скорости ее нагревания Солнцем при различных сочетаниях поглощающих газов, аэрозолей и облачности в атмосфере. Солнечный спектр разделен на 9 спектральных участков в соответствии с полосами поглощения атмосферных газов, в состав которых включены водяной пар, озон, кислород и углекислый газ. Программа расчета радиационного переноса учитывает неоднородность облачности и аэрозольной составляющей, зависимость интенсивности солнечного излучения от длины волны, а также газовое поглощение в рассеивающих слоях. Кучевые облака в нижней тропосфере обеспечивают скорость нагревания порядка 12°C/сутки при Солнце в зените и вызывают дополнительное нагревание озоном нижней стратосферы. Показано, что аэрозоль дает заметный вклад в скорость нагревания нижней безоблачной атмосферы, но в случае облачной атмосферы его влияние мало. Когда Солнце близко к горизонту, эффект облаков в скорости нагревания, как показывают расчеты, несущественен. Далее показано, что значительная недо- и переоценка, соответственно, имеют место при неучете поглощения, вызываемого озоном и кислородом, особенно, в облачной атмосфере.