Increase of Global Albedo Due to Air Pollution

GIICHI YAMAMOTO AND MASAYUKI TANAKA

Geophysical Institute, Tohoku University, Sendai, Japan

(Manuscript received 26 January 1972, in revised form 19 July 1972)

ABSTRACT

The effect of an increase of particles in the atmosphere on the global albedo and accordingly on the thermal regime of the earth is studied by solving the equation of radiative transfer in model turbid atmospheres.

Realistic model atmospheres with respect to size and vertical distributions of aerosol as well as reflectivity of the earth surface are assumed, and reflectivity at the top of the atmosphere, transmissivity at the earth surface, and absorptivity of turbid atmospheres are calculated as a function of atmospheric turbidity and the complex refractive index of the aerosol. It is shown that the thermal effect of increasing atmospheric turbidity is greatly affected by the imaginary part of the refractive index. Thus, if it takes a small value as is believed so at present, the earth-atmosphere system cools off with increase of turbidity, while if its value is large ($n_i > 0.05$, n_i being the imaginary part of the complex refractive index), heating of the earth-atmosphere system is expected due to increasing turbidity.

1. Introduction

Recent developments in man's activities, particularly in his industrial activity, has caused considerable changes to his environment. An increase of particles in the atmosphere is one of the noticeable changes. This causes a change of the radiation field in the atmosphere, which necessarily leads to a change of the heat budget of the earth-atmosphere system. Recently, Rasool and Schneider (1971) made an estimate of this effect by using the two-stream approximation in solving the equation of radiative transfer in a turbid atmosphere. A more sophisticated investigation of this problem is attempted in this study.

2. Model turbid atmospheres

In order to solve the equation of radiative transfer in a turbid atmosphere, the amount of particles, their vertical and size distributions, as well as their average complex refractive index should be specified.

The vertical distribution of particles has been compiled by Elterman (1964), and is shown in Fig. 1 in the form of the relative concentration, together with that of air molecules. There might be some question as to the seemingly unusual increase of particles in the stratosphere in the distribution shown in Fig. 1, and whether such a distribution would cause any local effects in the radiative field in the stratosphere. However, since our concern in this study is directed to the investigation of albedo to space due to the whole air column, we adopted this profile without further considerations. In this distribution the Rayleigh scattering due to air molecules predominates in the stratosphere, while the role of Mie scattering due to particles becomes important in the troposphere.

The size distribution of particles generally follows the Junge distribution (1955). In the present study the "haze C" model proposed by Deirmendjian (1964), which simulates the Junge distribution fairly well, is adopted in computing the phase function due to particles. This is given by

$$n(r)=\begin{cases} C\times 10^4, & \text{for } 0.03\mu\leqslant r\leqslant 0.1\mu \\ Cr^{-4}, & \text{for } 0.1\mu\leqslant r\leqslant 10\mu \end{cases} \tag{1}$$

where C is a constant and $n(r)dr$ is the number of particles of radii between r and $r+dr$, included in the whole air column.

The optical thickness due to particles, τ_M, is given by

$$\tau_M(\lambda)=\int_0^\infty \pi r^2 Q(\alpha,m)n(r)dr, \tag{2}$$

where $\alpha=2\pi r/\lambda$, and $Q(\alpha,m)$ is the efficiency factor of the Mie scattering. Because of the optical ineffectiveness of small particles (say, $r<0.1\,\mu$), and of the scarcity of large particles (say, $r>10\,\mu$), $n(r)$ in Eq. (2) can be replaced by cr^{-4} over the whole range of the integral. Then we have

$$\tau_M(\lambda)=\beta\lambda^{-1}, \tag{3}$$

where

$$\beta=2\pi^2 C\int_0^\infty Q(\alpha,m)\alpha^{-2}d\alpha. \tag{4}$$

Unauthenticated | Downloaded 02/27/24 09:40 AM UTC

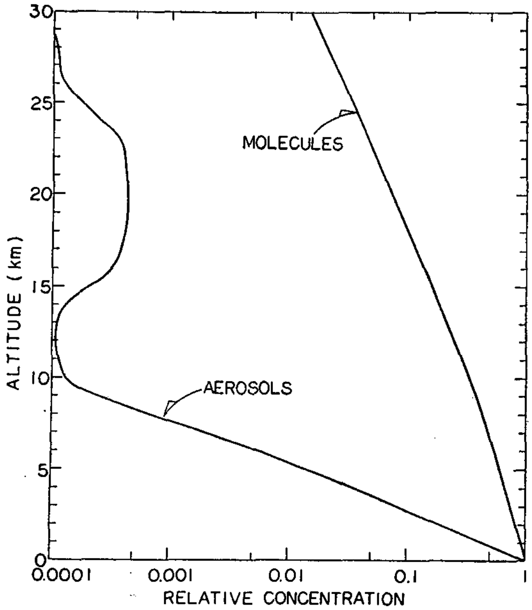

FIG. 1. Vertical distribution of the relative concentration of particles and air molecules.

In this study β is taken, instead of the aerosol amount, as a parameter representing the turbidity of the atmosphere. This quantity β has a meaning similar to that of the turbidity coefficient of Ångström, β_A, which is defined by $\tau_M(\lambda)=\beta_A\lambda^{-1.3}$.

It has been shown by Jaenicke *et al.* (1971) that the size distribution frequently differs, especially over sea, from that of a power law. This fact suggests the difficulty of representing the size distribution, and accordingly the atmospheric turbidity, by a single parameter such as β or β_A. Presumably two parameters or more would be necessary to represent the size distribution or turbidity correctly. However, since many parameters are involved in this study, a single parameter represensation of turbidity was deemed necessary. The error caused by approximating the observed size distribution curve, which is slightly convex in a log-log diagram over the optically effective range, by an equivalent straight line corresponding to r^{-4}, is not crucial to this study whose main object is to clarify the importance of the imaginary part of the refractive index of particles on the global albedo.

With respect to the real part of the average complex refractive index of particles in the atmosphere, the value of 1.50, which is proposed by Eiden (1966), seems to be reasonable. It was therefore adopted in this study.

As to the imaginary part, little research has been carried out so far. Eiden (1966) estimated it from measurements of the ellipticity of scattered light, to be between 0.01 and 0.1, and Fischer (1970) recently made estimates which ranged between 0.005 and 0.07. In order to cover this range, the following values were assumed for the complex refractive index, m, of particles:

1) $m=1.50-0.0i$ (no absorption by particles)
2) $m=1.50-0.005i$
3) $m=1.50-0.01i$
4) $m=1.50-0.02i$
5) $m=1.50-0.05i$
6) $m=1.50-0.1i$

Radiative transfer in turbid atmospheres in the spectral range from 0.3 to 2.3 μ can then be studied without taking into account absorption by the Chappius band of O_3 and the H_2O, CO_2, and O_2 bands in the near-infrared region. Because of the relatively thin optical thicknesses of the turbid atmospheres, the surface reflectivity A_S plays an important role in the present investigation. The following values were assumed:

1) $A_S=0$, representing a black surface
2) $A_S=0.05$, representing an average ocean surface
3) $A_S=0.15$, representing an average land surface

3. Equation of radiative transfer and a method of solving it

In order to study the problem of radiative transfer rigorously, it is necessary to take into account the polarization of light. However, it was shown by Tanaka (1971b) that the effect of polarization can safely be neglected in evaluating intensity or flux in the radiation field. As we are currently interested in evaluating the flux of radiation normal to the horizontal surface, we shall neglect the effect of polarization. Moreover, we shall consider the diffuse radiation field in a horizontally homogeneous, plane-parrallel atmosphere illuminated by the sun at μ_0 ($=\cos\theta_0$, θ_0 being the zenith angle). The flux of the solar radiation at the top of the atmosphere is assumed to be πF. Then the equation of radiative transfer is given by

$$\mu\frac{dI(\tau,\mu)}{d\tau}=I(\tau,\mu)-\frac{1}{2}\int_{-1}^{+1}p^{(0)}(\tau;\mu,\mu')I(\tau,\mu')d\mu' -\frac{F}{4}e^{-\tau/\mu_0}p^{(0)}(\tau;\mu,-\mu_0), \quad (5)$$

where I and $p^{(0)}$ are, respectively, the intensity of the diffuse radiation field and the phase function independent of the azimuthal angle, and τ is given by

$$\tau(z)=\int_\infty^z[(\sigma_M{}^{(s)}+\sigma_M{}^{(a)})\rho_M(z)+\sigma_R\rho(z)]dz,$$

where $\sigma_M{}^{(s)}$ and $\sigma_M{}^{(a)}$ are, respectively, the Mie mass scattering and mass absorption coefficients of particles, σ_R is the Rayleigh mass scattering coefficient of air, ρ_M the concentration of particles, ρ the density of air, and other terms have their usual meaning. The

Unauthenticated | Downloaded 02/27/24 09:40 AM UTC

boundary conditions are given by

$$\left.\begin{aligned} I(0,-\mu)&=0\\ I(\tau_s,+\mu)&=I_g(+\mu)\end{aligned}\right\}, \tag{6}$$

where I_g is the reflected light from the earth surface ($\tau=\tau_S$).

The phase function $p^{(0)}(\tau;\mu,\mu')$ consists of the sum of that due to the Rayleigh scattering by air molecules and that due to the Mie scattering by particles, i.e.,

$$p^{(0)}(\tau;\mu,\mu')=\frac{\sigma_M{}^{(s)}\rho_M(z)}{(\sigma_M{}^{(s)}+\sigma_M{}^{(a)})\rho_M(z)+\sigma_R\rho(z)}p_M{}^{(0)}(\mu,\mu')$$
$$+\frac{\sigma_R\rho(z)}{(\sigma_M{}^{(s)}+\sigma_M{}^{(a)})\rho_M(z)+\sigma_R\rho(z)}p_R{}^{(0)}(\mu,\mu'),$$

where $p_M{}^{(0)}(\mu,\mu')$ and $p_R{}^{(0)}(\mu,\mu')$ are the normalized phase functions due to particles and air molecules, respectively. The resultant phase function $p^{(0)}(\tau;\mu,\mu')$ satisfies the normalization integral,

$$\frac{1}{2}\int_{-1}^{+1}p^{(0)}(\tau;\mu,\mu')d\mu=\omega(\tau), \tag{7}$$

where $\omega(\tau)$ is the albedo for single scattering of the medium at the level τ, which is given by

$$\omega(\tau)=\frac{\sigma_M{}^{(s)}\rho_M+\sigma_R\rho}{(\sigma_M{}^{(s)}+\sigma_M{}^{(a)})\rho_M+\sigma_R\rho}.$$

In order to evaluate the diffuse reflected light at the top of the atmosphere and the diffuse transmitted light at the bottom of the atmosphere from Eqs. (5) and (6), the method of the principle of invariance developed by Chandrasekhar (1950) is adopted. Because of the vertical inhomogeneity of the atmosphere, the following functions are necessary: the reflection function $S^{(0)}(0,\tau_S;\mu,-\mu')$ and the transmission function $T^{(0)}(0,\tau_S;\mu,-\mu')$ for the case of no reflecting surface at the bottom of the atmosphere (the standard problem); and $S^{*(0)}(0,\tau_S;-\mu,\mu')$ and $T^{*(0)}(0,\tau_S;\mu,\mu')$, the reflection and transmission functions for the case of the standard problem in which the bottom of the atmosphere is assumed to be illuminated. Here the superscript (0) means that the functions are independent of the azimuthal angle. Then from definitions, we have

$$I(0,+\mu)=\frac{1}{4\mu}S^{(0)}(0,\tau_s;\mu,-\mu_0)F$$
$$+\frac{1}{2\mu}\int_0^1 T^{*(0)}(0,\tau_s;\mu,\mu')I_g(\mu')d\mu'+e^{-\tau_s/\mu}I_g(\mu), \tag{8}$$

$$I(\tau_s,-\mu)=\frac{1}{4\mu}T^{(0)}(0,\tau_s;-\mu,-\mu_0)F$$
$$+\frac{1}{2\mu}\int_0^1 S^{*(0)}(0,\tau_s;-\mu,\mu')I_g(\mu')d\mu'. \tag{9}$$

The reflected light $I_g(\mu)$ at the earth's surface, which appears in Eqs. (6), (8) and (9), is assumed to be independent of μ (Lambert's law of reflection). In this case, according to Chandrasekhar (1950), it is given by

$$I_g=\frac{A_sF[\mu_0e^{-\tau_s/\mu_0}+t(0,\tau_s;\mu_0)]}{1-\bar{s}(0,\tau_s)}, \tag{10}$$

where

$$\left.\begin{aligned}\bar{s}(0,\tau_s)&=\int_0^1\int_0^1 S^{(0)}(0,\tau_s;\mu,-\mu')d\mu d\mu'\\ t(0,\tau_s;\mu_0)&=\frac{1}{2}\int_0^1 T^{(0)}(0,\tau_s;-\mu,-\mu_0)d\mu\end{aligned}\right\}, \tag{11}$$

and A_S is the reflectivity of the earth surface (reflected flux divided by the incident flux).

The next problem is to evaluate $S^{(0)}$, $T^{(0)}$, $S^{*(0)}$ and $T^{*(0)}$. The simultaneous integral equations for them are easily obtained from Eqs. (5) and (6) and the equations of the principle of invariance appropriate to the inhomogeneous atmosphere, by eliminating the field intensity I. There are many methods of solving the problem numerically. Here we shall adopt the matrix method which was originally developed by Twomey *et al.* (1966) for the case of a homogeneously stratified medium and was later generalized by Tanaka (1971a) to be applicable to an inhomogeneously stratified medium and also to the case of polarized light. The details of this method are not given here (see Tanaka 1971a), but essentially it consists of dividing the atmosphere into thin layers and computing the reflection and transmission functions for the whole atmospheric layer by successively building up thicker layers from thinner layers. The starting point is a layer so thin that it is essentially a single scattering layer; the reflection and transmission functions for it are given from the phase function.

If $I(0,+\mu)$ and $I(\tau_S,-\mu)$ are evaluated, the corresponding fluxes are given by

$$\left.\begin{aligned}\bar{R}(\mu_0)&=2\pi\int_0^1 I(0,+\mu)\mu d\mu\\ T(\mu_0)&=2\pi\int_0^1 I(\tau_s,-\mu)\mu d\mu\end{aligned}\right\}. \tag{12}$$

It is evident that these quantities are dependent upon μ_0, although μ_0 is not explicitly involved in $I(0,+\mu)$ and $I(\tau_S,-\mu)$. The flux of the incident solar radiation

Unauthenticated | Downloaded 02/27/24 09:40 AM UTC

referred to the horizontal top surface of the atmosphere is given by $\pi F\mu_0$. Therefore, the diffuse reflectivity $\mathcal{R}(\mu_0)$ and transmissivity $\mathcal{T}(\mu_0)$ referred to $\pi F\mu_0$ are given by

$$\left.\begin{aligned}\mathcal{R}(\mu_0)&=R(\mu_0)/\pi F\mu_0\\ \mathcal{T}(\mu_0)&=T(\mu_0)/\pi F\mu_0\end{aligned}\right\}. \tag{13}$$

An average reflectivity $\bar{\mathcal{R}}$ and transmissivity $\bar{\mathcal{T}}$ over the sunlit hemisphere are then given by

$$\left.\begin{aligned}\bar{\mathcal{R}}&=2\int_0^1 R(\mu_0)d\mu_0/\pi F\\ \bar{\mathcal{T}}&=2\int_0^1 T(\mu_0)d\mu_0/\pi F\end{aligned}\right\}. \tag{14}$$

Of course, the contribution of the direct solar radiation is not included in $T(\mu_0)$, $\mathcal{T}(\mu_0)$ and $\bar{\mathcal{T}}$. The transmitted flux and transmissivity of the direct solar radiation at $\tau=\tau_S$ are given by $\pi F\mu_0{}^{-\tau_s/\mu_0}$ and $e^{-\tau_s/\mu_0}$, respectively, and the average transmissivity of the direct solar radiation over the sunlit hemisphere is given by

$$2\int_0^1 e^{-\tau_s/\mu_0}\mu_0 d\mu_0=2E_3(\tau_s), \tag{15}$$

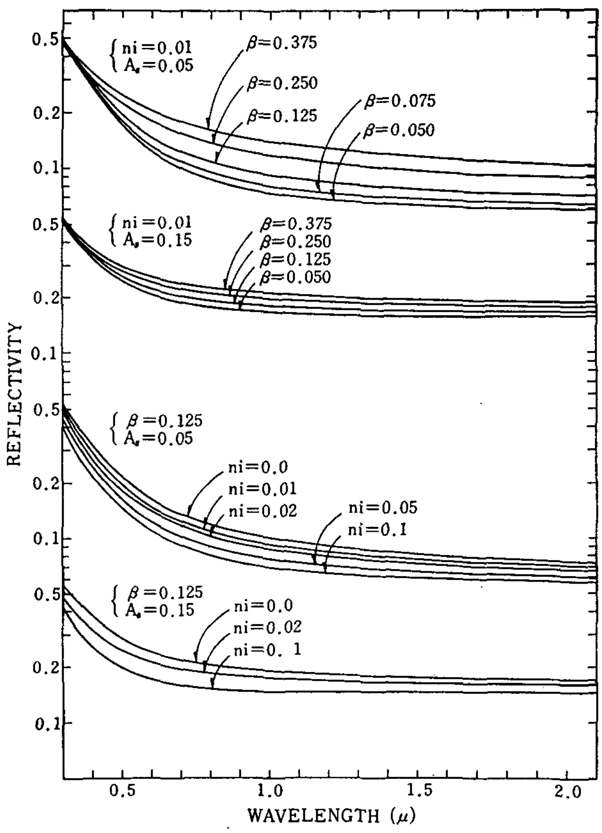

FIG. 2. Average reflectivity over the sunlit hemisphere as a function of wavelength λ for different β, n_i and A_s values.

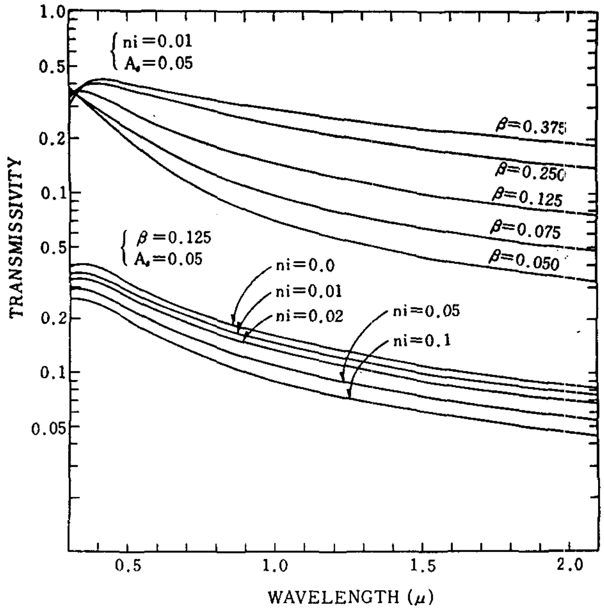

FIG. 3. Average diffuse transmissivity over the sunlit hemisphere as a function of wavelength λ for $A_s=0.05$ and for different β and n_i values.

where $E_3(\tau_S)$ is the exponential integral of the third order.

The quantities $\bar{\mathcal{R}}$ and $\bar{\mathcal{T}}$ and the contribution to the transmissivity due to direct solar radiation shall be used later in considering the global effect.

4. Results of computation

Solutions of the system of equations were obtained for values of λ (μ) = 0.3, 0.45, 0.55, 0.70, 0.90, 1.30, 1.60, 1.90, 2.25; for μ_0 equal to the zeros of the shifted 10th degree Legendre polynomial whose range of normalization is [0,1] instead of the customary range [−1,1]; for β=0.0, 0.05, 0.75, 0.125, 0.25, 0.375; and for the values of A_S and m already given in Section 2. Although values of $\mathcal{R}$ and $\mathcal{T}$ are obtained as functions of λ, μ_0, β, A_S and m, it is not the purpose of the present study to require such detail. Therefore, we shall begin by showing the general feature of $\bar{\mathcal{R}}$ and $\bar{\mathcal{T}}$ (the average values of $\mathcal{R}$ and $\mathcal{T}$ over the sunlit hemisphere). Fig. 2 shows $\bar{\mathcal{R}}$ as a function of λ, taking β, n_i (the imaginary part of m), and A_S as parameters. As can be seen, $\bar{\mathcal{R}}$ decreases with increase of λ, increases with increase of β, decreases with increase of n_i, and increases with increase of A_S. Fig. 3 shows $\bar{\mathcal{T}}$, whose general features with respect to λ, β and n_i are similar to those of $\bar{\mathcal{R}}$. However, in the short wavelength region, the wavelength dependence of $\bar{\mathcal{T}}$ for small values of β and for large values of β reverses. This is due to two combined effects: one that the transmissivity decreases when the optical thickness increases beyond a certain value, and the other that the Rayleigh optical thickness increases with decrease of λ. The effect of A_S on $\bar{\mathcal{T}}$ is qualitatively similar to that on $\bar{\mathcal{R}}$, but

Unauthenticated | Downloaded 02/27/24 09:40 AM UTC

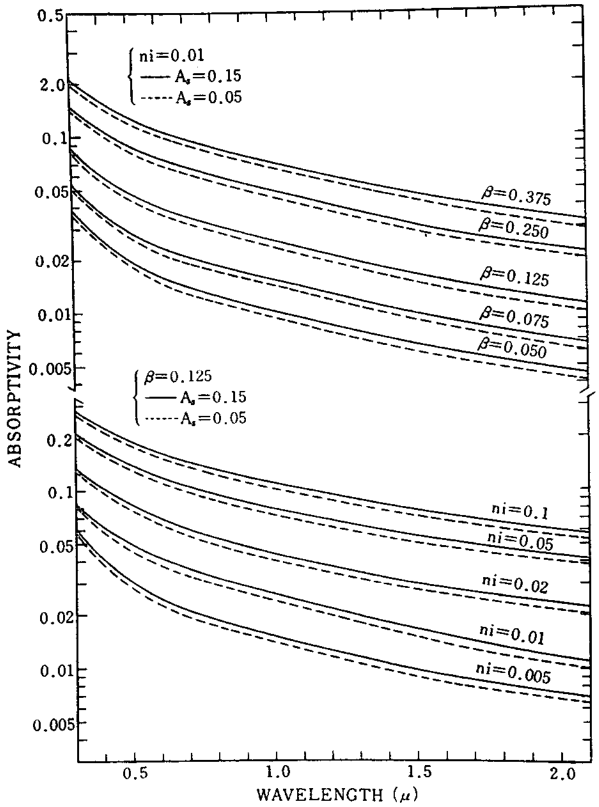

FIG. 4. Average absorptivity over the sunlit hemisphere as a function of wavelength λ for different β, n_i and A_s values.

quantitatively so small that the values for $A_S=0.15$ are not shown. Fig. 4 shows the average absorptivity $\overline{\mathfrak{A}}$ for the whole air column and sunlit hemisphere, which can easily be obtained from the values of $\overline{\mathfrak{R}}$ and $\overline{\mathcal{T}}$ and the solar flux at the top of the atmosphere. The dependence of $\overline{\mathfrak{A}}$ on λ, β, n_i and A_S is qualitatively similar to that of $\overline{\mathfrak{R}}$.

Next, we evaluate the average value of $\overline{\mathfrak{R}}$ over wavelength. This quantity is nothing but the global albedo A_0, which is given by

$$A_0=\int_{0\cdot 3\mu}^{2.3\mu}\pi F(\lambda)\overline{\mathfrak{R}}(\lambda)d\lambda \bigg/ \int_{0\cdot 3\mu}^{2.3\mu}\pi F(\lambda)d\lambda. \tag{16}$$

It should be noticed that the effect of absorption by atmospheric absorption bands is not taken into account in A_0. Fig. 5 shows A_0 as a function of β and n_i for two cases of $A_S=0.05$ and 0.15. An interesting feature is that when n_i is small, A_0 increases with increase of β, while for very large values of n_i, A_0 decreases with increase of β.

In the above calculations we have assumed cloudless atmospheres. Actually, however, about half of the earth surface is covered by clouds. Because of the large optical thickness of normal clouds, which is about two orders greater than that of a cloudless atmosphere, the effect of particles will be reduced in a cloudy atmosphere. If we designate the global albedo of the real atmosphere by A, then we have

$$A=nA_c+(1-n)A_0, \tag{17}$$

where n is the global average cloud amount, and A_c its average albedo. Based on the estimation by Robinson (1966) and Budyko (1969), we shall adopt $n=0.5$ and $A_c=0.5$. If we assume that the overall areal ratio of ocean to land, 0.71:0.29, is applicable to each latitudinal belt, appropriate values of A_0 for given β and n_i values can easily be obtained from the curves of A_0 in Fig. 5. Accordingly, we can estimate the values of A, which are also shown in Fig. 5. It is seen that the effect on A due to changes in β and n_i has been significantly reduced due to the existence of clouds. Still it can be seen that A decreases with increase of β for a very large value of n_i ($n_i=0.1$).

Corresponding to the behavior of A_0, we shall next examine the average transmissivity over the sunlit hemisphere and wavelength (Fig. 6). This figure shows that the average transmissivity of diffuse radiation increases with increase of β, but it decreases with increase of n_i. The average transmissivity of the direct solar radiation versus β is also shown in the figure as a single curve irrespective of the values of n_i. If aerosol amount had been used instead of β as abscissa, separate curves would have been necessary for each n_i value.

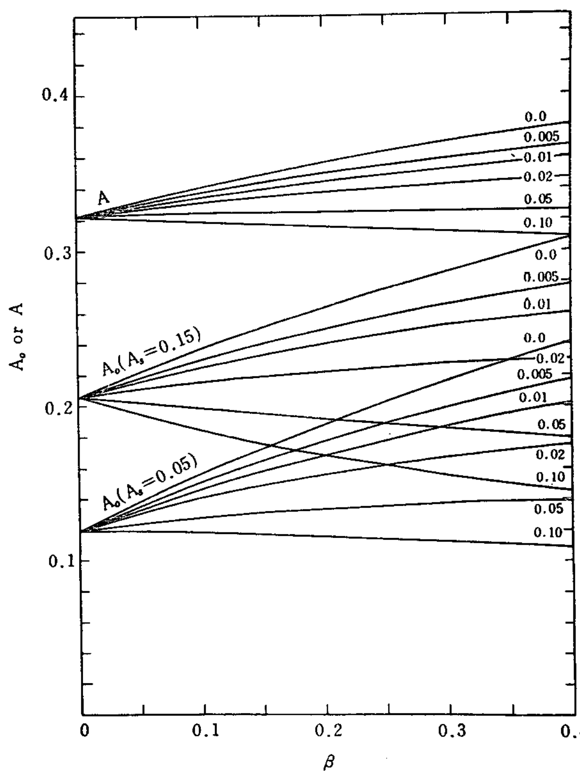

FIG. 5. Albedo of the earth as a function of β for different n_i values. A_0 indicates albedo for cloudless atmospheres, and A albedo which takes into account existence of clouds.

Unauthenticated | Downloaded 02/27/24 09:40 AM UTC

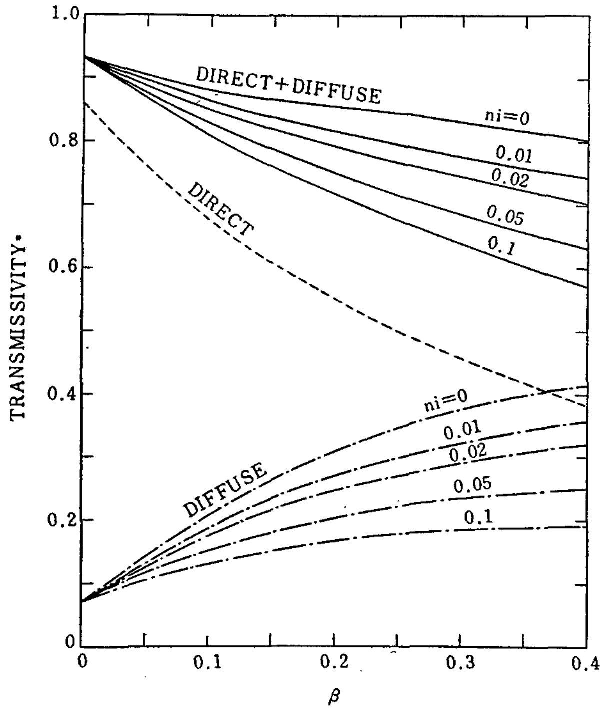

FIG. 6. Transmissivity averaged over the sunlit hemisphere and wavelength as a function of β. Curves denoted DIRECT, DIFFUSE and DIRECT + DIFFUSE show transmissivity for direct, diffuse, and total components, respectively.

Curves indicated by (DIRECT+DIFFUSE) mean the total transmissivity on the earth surface. Because of the predominant contribution of direct solar radiation, it decreases with increase of β and n_i. The global average value of this quantity taking account of clouds has not been estimated.

Fig. 7 presents the heating rate (°C day^{-1}) due to absorption of particles averaged over the globe and whole air column. Evidently, heating rate increases with increase of β, n_i and A_S.

Based on the above estimation of A we shall next discuss the effect of increase of particles on the thermal regime of the earth's atmosphere by use of a simple global model.

An equivalent blackbody temperature T_e (°K) of the earth-atmosphere system, which corresponds to the net incident solar energy, is given by

$$\pi R^2 S(1-A)=4\pi R^2 \sigma T_e^4, \tag{18}$$

where R is the radius of the earth, S the solar constant, and σ the Stefan-Boltzmann constant. Taking the value of $T_e=254.1$K for the molecular atmosphere ($\beta=0$) as a reference, the value of ΔT_e is shown in Fig. 8 as a function of β, with n_i as a parameter. It can be seen from the figure that for small values of n_i, ΔT_e decreases with increase of β, whereas for $n_i=0.05$ no appreciable change of ΔT_e vs β is seen; for values of n_i larger than 0.05, ΔT_e increases with increase of β.

A similar estimation of the global average temperature near the earth surface, T_S, can be made as follows: If we let the global average outgoing longwave flux from the top of the atmosphere be I, then

$$\pi R^2 S(1-A)=4\pi R^2 I. \tag{19}$$

According to Budyko (1969), I is expressed empirically as a function of T_S as

$$I=a+bT_S-(a_1+b_1T_S)n, \tag{20}$$

where n is the global cloud amount, and a, b, a_1 and b_1 are numerical constants. If I is expressed in kcal cm^{-2} month^{-1}, $a=14.0$, $b=0.14$, $a_1=3.0$ and $b_1=0.10$ are the values evaluated by Budyko. Again taking the value of $T_S=292.0$K as a reference for the molecular atmosphere ($\beta=0$), the form of the functional variation of ΔT_S with β and n_i, shown in Fig. 9, is seen to be similar to that of ΔT_e given in Fig. 8. It is noticed, however, that the overall changes of ΔT_S are larger than those of ΔT_e.

As Rasool and Schneider (1971) have made a similar estimation of T_S as a function of the optical thickness of the particles (shown in Fig. 2b of their paper), it will be interesting to compare their results with ours. They used the albedo for single scattering, ω_0, as a parameter instead of n_i. Assuming the Junge distribution and $n_r=1.50$, their values $\omega_0=0.99$ and 0.90 correspond to $n_i=0.001$ and 0.013, respectively, at $\lambda=0.55\ \mu$. The optical thickness, τ_{vis}, which they used as a measure

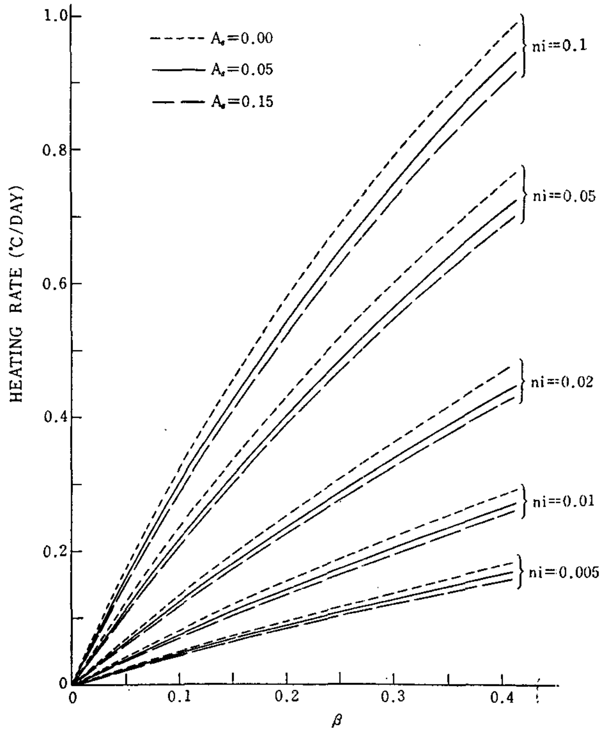

FIG. 7. Average heating rate of the earth's atmosphere as a function of β for different n_i values.

Unauthenticated | Downloaded 02/27/24 09:40 AM UTC

of turbidity, is related to β by $\beta=0.55\ \tau_{vis}$ from Eq. (3). Under these transformations, comparison is made for the change of τ_{vis} from 0.1 to 0.3 or of β from 0.055 to 0.165. The result is $\Delta T_S=-3.4$K by Rasool and Schneider, and -3.9K by Yamamoto and Tanaka for $\omega_0=0.99$ or $n_i=0.001$, and $\Delta T_S=-2.3$K by Rasool and Schneider, and -2.4K by Yamamoto and Tanaka for $\omega_0=0.90$ or $n_i=0.013$. These results should be considered good agreement. A comment to their result is that the representation of their Figs. 2a and 2b is somewhat misleading. Since they took present conditions as a reference, their result leads to two different flux or T_S values for the aerosol-free atmosphere. This seems to be contradictory, because the flux or temperature of the aerosol-free atmosphere will not depend on ω_0 or n_i of the aerosols.

5. Discussion

The above results indicate that in order to understand the probable climatic effect of an increase of particles in the atmosphere, it is necessary to determine the average value of n_i more accurately than hitherto estimated. The fact that large volcanic eruptions in the past have caused cooling of the average surface temperature allows us to infer that the value of $n_i>0.05$ is improbable and that $n_i\approx 0.01-0.02$ seems to be most likely.

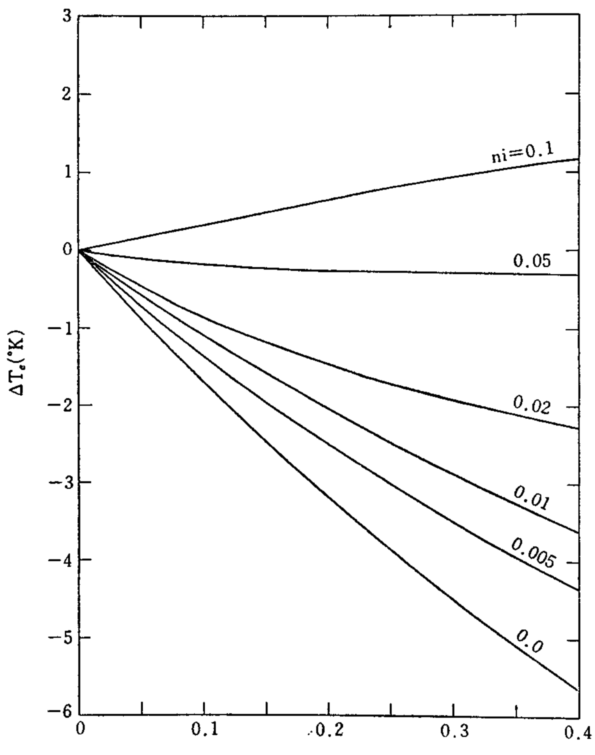

FIG. 8. Effective blackbody temperature, ΔTe, as a function of β for different n_i values.

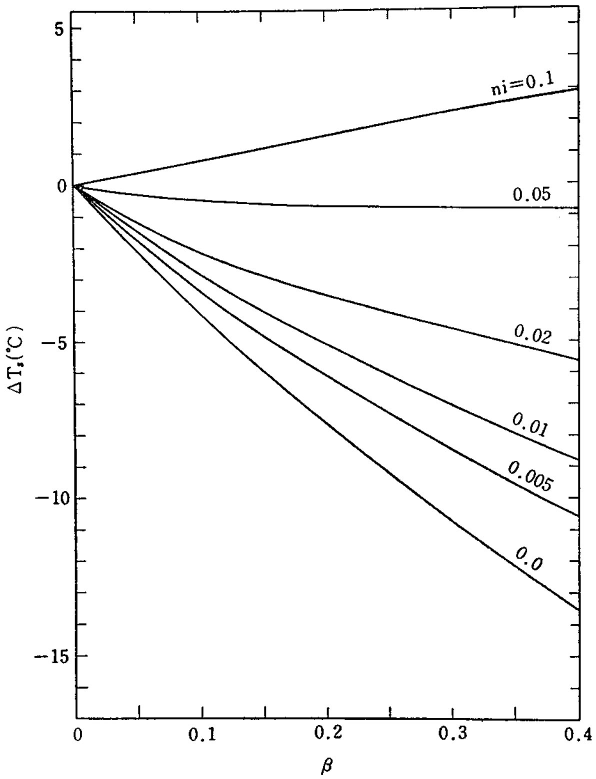

FIG. 9. Surface temperature, ΔTs, as a function of β for different n_i values.

It will then be of interest to compare cooling due to increase of particles with heating due to increase of CO_2. According to Manabe and Wetherald (1967), a doubling of the CO_2 concentration will increase the average surface temperature by 2C. An increase of atmospheric CO_2 from 320 ppm (present concentration) to 378 ppm [estimated concentration for the year 2000 AD (Machta, 1971)] will warm the surface layer by about 0.5C. In the case of particles it is difficult to estimate the present global average value of β with accuracy comparable to the case of CO_2. However, referring to the estimation of β by Yamamoto *et al.* (1968) from IGY data primarily from land stations, and taking account of clearer air conditions over the ocean, the present average value of β is estimated to be about 0.05–0.075. Then a doubling of β will decrease the average surface temperature by $1.3-1.8$C if $n_i=0.01$, and by $1.0-1.2$C if $n_i=0.02$. While these values are smaller than that due to the CO_2 effect, it is difficult to determine which of the two effects, CO_2 and particles, is predominant since the rate of secular increase of the global average value of β is entirely unknown. However, it is noted that the local variability of β is evidently large compared to that of CO_2, and that, as far as Japan is concerned, the recent increase of β is so rapid that the average β value will double in less than 20 years (Yamamoto *et al.*, 1971). Global monitoring of particle concentration will be essential to the study of man's

Unauthenticated | Downloaded 02/27/24 09:40 AM UTC

impact on climate. With respect to the method of monitoring, it is felt that while the measurement of direct solar radiation is a simple and time-proved method of evaluating atmospheric turbidity, observations should be made over the solar radiation spectrum in order that the particle size distribution can be determined, along with a more accurate value of β.

Acknowledgments. This work was supported by the Ministry of Education of Japan, under Grant 91410.

REFERENCES

Budyko, M. I., 1969: The effect of solar radiation variations on the climate of the earth. *Tellus*, **21**, 611–619.

Chandrasekhar, S., 1950: *Radiative Transfer*. Oxford, Clarendon Press, 393 pp.

Deirmendjian, D., 1964: Scattering and polarization properties of water clouds and hazes in the visible and infrared. *Appl. Opt*., **3**, 187–196.

Eiden, R., 1966: The elliptical polarization of light scattered by a volumn of atmospheric air. *Appl. Opt*., **5**, 569–575.

Elterman, L., 1964: Rayleigh and extinction coefficients to 50 km for the region .27 μ to .55 μ. *Appl. Opt*., **3**, 1139–1147.

Fischer, K., 1970: Measurements of absorption of visible radiation by aerosol particles. *Atmos. Phys*., **43**, 244–254.

Jaenicke, R., C. Junge and H. J. Kanter, 1971: Messungen der Aerosolgrössenverteilung über dem Atlantik. *"Meteor" Forschungsergeb*., *Reihe B*, No. **7**, 1–54.

Junge, C., 1955: The size distribution and aging of natural aerosols as determined from electrical and optical data on the atmosphere. *J. Meteor*., **12**, 13–25.

Machta, L., 1971: The role of the oceans and the biosphere in the carbon dioxide cycle. Paper presented at Nobel Symposium, 16–20 August, Gothenburg, Sweden; also *Inadvertent Climate Modification SMIC*, M.I.T. Press, 233–240.

Manabe, S., and R. T. Wetherald, 1967: Thermal equilibrium of the atmosphere with a given distribution of relative humidity. *J. Atmos. Sci*., **24**, 241–259.

Rasool, S. I., and S. H. Schneider, 1971: Atmospheric carbon dioxide and aerosols—effects of large increases on global climate. *Science*, **173**, 138–141.

Robinson, G. D., 1966: *Solar Radiation*. Amsterdam, Elsevier, 347 pp.

Tanaka, M., 1971a: Radiative transfer in turbid atmospheres, I. Matrix analysis for the problem of diffuse reflection and transmission. *J. Meteor. Soc. Japan*, **49**, 296–312.

——, 1971b: Radiative transfer in turbid atmospheres, II. Angular distribution of intensity of the solar radiation diffusely reflected and transmitted by turbid atmospheres. *J. Meteor. Soc. Japan*, **49**, 321–332.

Twomey, S., H. Jacobwitz and H. B. Howell, 1966: Matrix method for multiple-scattering problems. *J. Atmos. Sci*., **23**, 289–296.

Yamamoto, G., M. Tanaka and K. Arao, 1968: Hemispherical distribution of turbidity coefficient as estimated from direct solar radiation measurements. *J. Meteor. Soc. Japan*, **46**, 278–300.

——, —— and ——, 1971: Secular variation of atmospheric turbidity over Japan. *J. Meteor. Soc. Japan*, **49**, 859–865.

Unauthenticated | Downloaded 02/27/24 09:40 AM UTC