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THE EFFECT OF AEROSOLS ON CLIMATE CHANGE

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ABSTRACT

A modified two-stream approximation is presented, which includes the effect of solar zenith angle and is applicable to study the effect of aerosols on both regional and global climate changes. More realistic results are derived. A reasonable critical value of 0.8 for aerosol single scattering albedo to determine whether the aerosols will heat or cool the climate is derived.

INTRODUCTION

In recent years much attention has been devoted to the role of atmospheric aerosols in the thermal state of the earth-atmosphere system. These investigations have been motivated by the possibility that there is a connection between observed climatic trends and a general buildup of atmospheric aerosols in large geographical areas. It is believed that the absorption of solar radiation by an aerosol layer increases the radiative heating of atmosphere, while the backscattering decreases the total amount of energy available to the earth. Chylet *et al*. (1974) assumed the aerosol layer to be a plane-parallel layer, and used a radiative transfer equation to describe the scattering and absorption effects of aerosols. But this equation was integro-differential and could not be solved analytically. They used a two-stream approximation to solve this equation and found the effect of aerosols on climate change to be a function of aerosol layer optical thickness t, aerosol single scattering albedo ω and the earth albedo a. They also stated that "this form of two-stream approximation is applicable only to globally averaged conditions. It does not include the dependence of the heating on the solar zenith angle which is necessary for the study of regional heating effects". In this paper we modify the work of Chylet *et al.*, (1974) in two aspects: (1) we consider the effect of aerosols on regional climate change by including the effect of solar zenith angle), and (2) we consider the effect of aerosols on global climate change by treating the aerosol layer to be a ball crust layer around the earth instead of to be a plane layer, and including the effect of solar zenith angle. Therefore we obtain more realistic results. We also use our modified two-stream approximation to derive a reasonable critical value of aerosol single scattering albedo ω in order to determine whether the aerosols will heat or cool the climate.

THEORY

The radiative transfer equation describing the scattering and absorption effects of aerosols was (Chylet *et al.*, 1974)

$$\mu dI(\mu,t)/dt = I(\mu,t) - 1/2 \int_{-1}^{1} p(\mu,\mu') I(\mu',t)d\mu' \qquad (1)$$

where $I(\mu,t)$ was the specific radiation intensity at optical thickness t, $\mu = \cos\theta$, and θ was the direction with respect to the normal of layer's surface, and $p(\mu,\mu')$ was an appropriate phase function. To determine, R, the albedo of the combined system of earth and an additional aerosol layer, Chylet *et al.*, (1974) used a two-stream approximation by assuming that the radiative intensity in an aerosol layer was isotropic over the upper hemisphere ($\mu > 0$) with the value $I_+(t)$ and over the lower hemisphere ($\mu < 0$) with the corresponding value $I_-(t)$. $I_+(t)$ and $I_-(t)$ are shown in Figure 1. Consequently, the radiative transfer equation could be transformed into a set of two coupled first order differential equations (Chylet *et al.*, 1974)

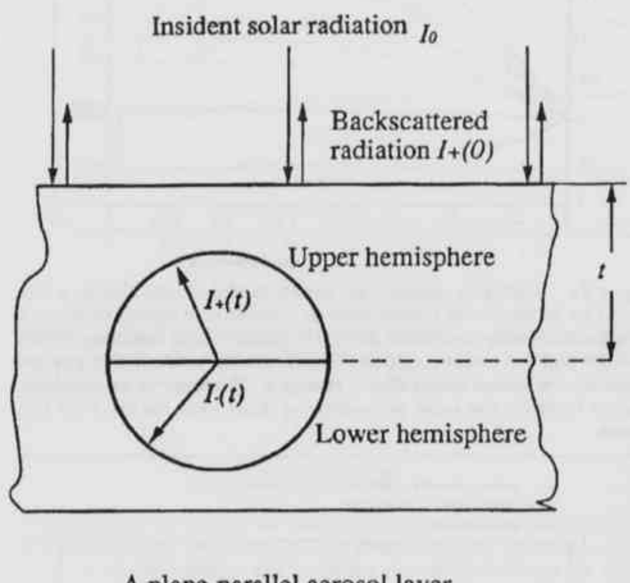

Figure 1. The plane-paralell aerosol layer and the two-stream radiation intensity $I_+(t)$ and $I_-(t)$ as a function of optical thickness t in the aerosol layer. I_0 is the solar radiation intensity which is vertically incident on the top of the aerosol layer, $I_+(0)$ is the intensity of backscattering radiation.

$$(1/2)dI_+(t) / dt = I_+(t) - (1-\beta)\omega I_+(t) - \beta\omega I_-(t) \qquad (2)$$

$$(-1/2)dI_-(t) / dt = I_-(t) - (1-\beta)\omega I_-(t) - \beta\omega I_+(t) \qquad (3)$$

where $\beta\omega$ is the backscattering coefficient. By definition, the albedo R is determined by the relation $I(0) = RI_0$ where I_0 is the solar radiation incident vertically on the top of the aerosol layer. The heating caused by an additional aerosol layer is given by the albedo change a - R, where a is the albedo of the earth. By solving (2) and (3), one can find (Chylet *et al.*, 1974)

$$a - R = [2a(1-\omega) - (1-a)^2\omega\beta] / [(1-\omega) + (1-a)\omega\beta + \alpha/2 \tanh(\alpha t)] \qquad 4)$$

where $\alpha = 2[(1-\omega)/(1-\omega+2\beta\omega)]^{1/2}$ and the solar zenith angle is assumed to be zero. The sign of (4) determines whether an aerosol layer will heat or cool the earth-aerosol system. Since the denominator of (4) is always positive, heating occurs if

$$(1-\omega) / \beta\omega > (1-a)^2/2a \qquad (5)$$

The solid line curves in Fig. 2a and 2b show the albedo change a - R obtained from the unmodified two-stream approximation as a function of t for various values of a, with $\beta\omega = 0.1$, and $\omega = 0.9$ and $\omega = 0.99$ respectively.

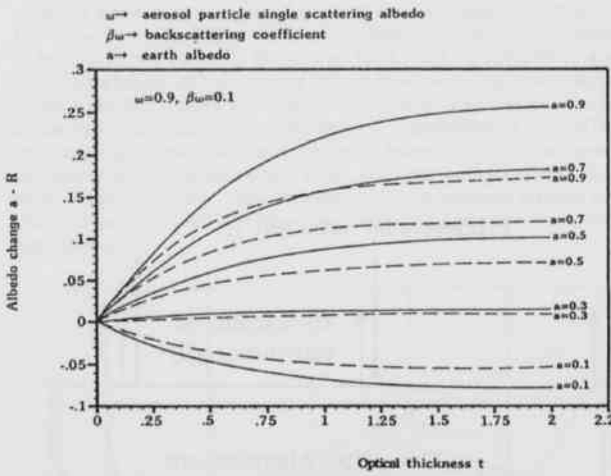

Figure 2a. Solid line curves: two-stream model albedo change a - R caused by an additional aerosol layer as a function of optical thickness t for a backscattering coefficient $\beta\omega = 0.1$, aerosol single scattering albedo $\omega = 0.9$ and for various values of earth albedo a. Dash line curves: modified two-stream model albedo change a - R caused by an additional aerosol layer for the same parameters as those used for the solid line curves.

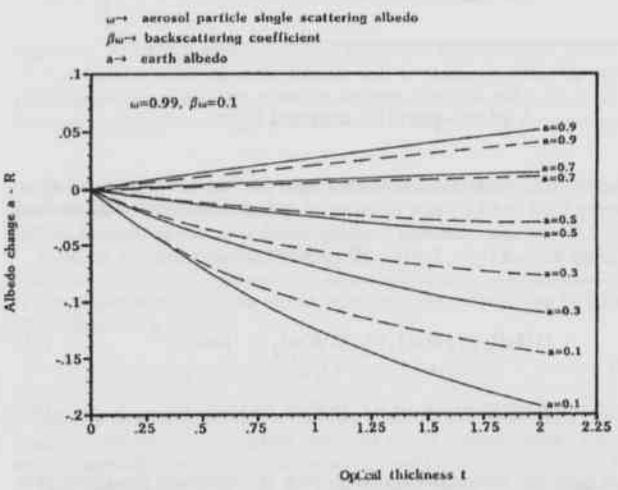

Figure 2b. Solid line curves: two-stream model albedo change a - R caused by an additional aerosol layer as a function of optical thickness t for a backscattering coefficient $\beta\omega = 0.1$, aerosol single scattering albedo $\omega = 0.99$ and for various values of earth albedo a. Dash line curves: modified two-stream model albedo change a - R caused by an additional aerosol layer for the same parameters as those used for the solid line curves.

Our work is as follows. For a regional situation, the aerosol layer can be treated as a plane-parallel layer shown in Fig. 3. But the solar radiation incident angle changes from $-\pi/2$ to $\pi/2$ during the day time. Therefore the effect of solar zenith angle must be included for a more accurate calculation. We assume that the intensity of solar radiation incident on the top of the aerosol layer at angle θ to be $I(\theta) = I_0 \cos\theta$ because the intensity attenuated by the atmosphere path is proportional to cosθ. The albedo R becomes a function of θ, R(θ). The aerosol layer can be treated as the sum of many small pieces of aerosol layer shown in Fig. 3 by dash lines. For a given point in the aerosol layer, the optical thickness is $t' = t/\cos\theta$ shown in Fig. 3. There is no significant error if the

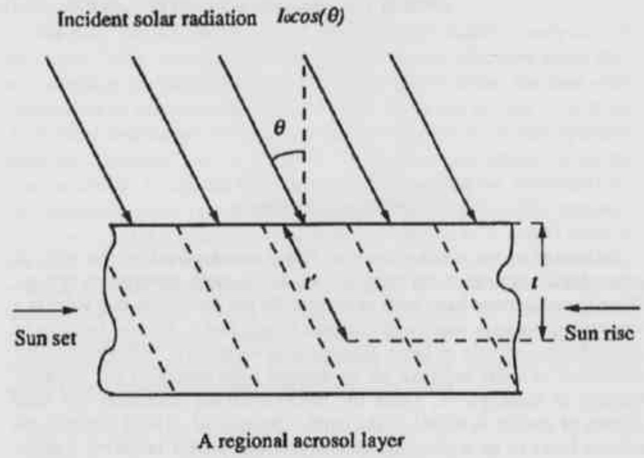

Figure 3. Solar radiation is incident at different angles on the top of a regional aerosol layer during a day.

aerosol layer pieces are small enough. Using the two-stream approximation for every small piece of aerosol layer we obtain a modified equation

$$a - R(\theta) = [2a(1-\omega) - (1-a)^2\beta\omega] / [(1-\omega)+(1-a)\beta\omega + \alpha/2 \tanh(\alpha t/\cos\theta)] \quad (6)$$

where we take the direction of incident radiation to be the normal direction thereby (6) is independent of azimuthal angle, and assume that a is not a function of θ. The average albedo change a - R for one day is

$$a - R = \int_{-\pi/2}^{\pi/2} [a - R(\theta)] \cos\theta \, d\theta \Big/ \int_{-\pi/2}^{\pi/2} \cos\theta \, d\theta \qquad (7)$$

The integration range from $-\pi/2$ to $\pi/2$ is corresponding to sun rise and sun set. It is difficult to solve (7) analytically, but we can solve it numerically. The numerical results of (7) which include the effect of solar zenith angle are shown by dash line curves in Fig. 2a and 2b for the same parameters as those used for the solid line curves in Fig. 2a and 2b. We can see that there are apparent differences between the solid line curves and dash line curves in Fig. 2. The dash line curves obtained from our modified equation are not so widely spread as the solid line curves are, which are obtained from the unmodified two-stream approximation. This difference means that the effect of aerosols on climate change obtained by us is relative weak. This result of ours is logical because our modification includes the effect of solar zenith angle and the solar radiation intensity is weak in the morning and evening times, consequently the heating or cooling of aerosols are weak. From (7) we can find that the critical condition given by (5) for determining whether the aerosols will heat or cool the climate does not change.

When considering the effect of aerosols on the global climate change, the aerosol layer should no longer be considered as a plane layer, but should be a ball crust layer around the earth surface shown in Fig. 4. At

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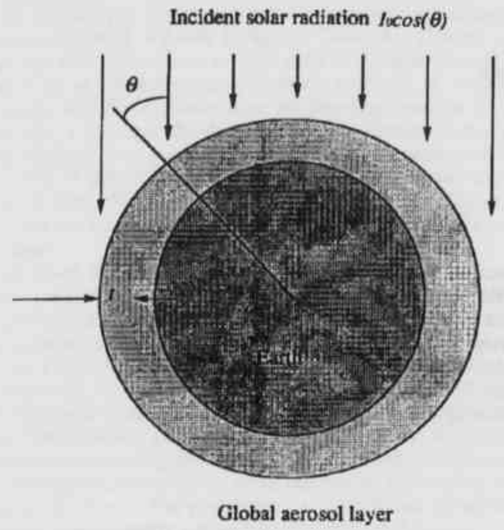

Figure 4. Solar radiation is indident on the top of a global aerosol layer. At any time the incident angle depends on the incident positions.

any given time the solar radiation is incident on the top of the aerosol layer at different angles depending on the incident positions. The spatial average albedo change a - R can be obtained from equation

$$a - R = \int_{-\pi/2}^{\pi/2} [a - R(\theta)] \cos\theta \, d\theta \int_0^{2\pi} d\phi \Big/ \int_{-\pi/2}^{\pi/2} \cos\theta \, d\theta \int_0^{2\pi} d\phi$$
$$= \int_{-\pi/2}^{\pi/2} [a - R(\theta)] \cos\theta \, d\theta \Big/ \int_{-\pi/2}^{\pi/2} \cos\theta \, d\theta \quad (8)$$

where θ is the polar angle and ϕ is the azimuthal angle. (8) is the same as (7) because the situation is symmetry about the solar radiation incident direction and independent of ϕ. The situation for the global aerosol effect becomes the same as that for the regional aerosol effect. For a global aerosol effect we take a spatial average, while for a regional aerosol effect we take a time average. Thus the results shown by the dash line curves in Fig. 2 obtained from (7) for the regional aerosol effect can be taken as the results for the global aerosol effect.

Fig. 5 shows the dependence of earth albedo a on the earth surface structure (Chylet *et al.*, 1974). Using the following data: oceans occupy 70% of the earth surface with albedo $a_o = 0.05$. farmland and urban areas occupy 10% of the earth surface with albedo $a_f = 0.2$, deserts occupy 5%

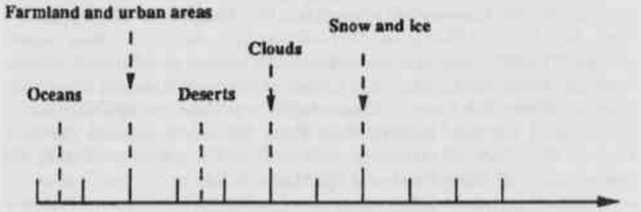

Figure 5. The dependence of earth-atmosphere system albedo on the earth surface structure.

of the earth surface with albedo $a_d = 0.35$. snow and ice occupy 15% of the earth surface with albedo $a_S = 0.7$, and clouds cover 10% of the whole earth surface areas with albedo $a_c = 0.5$, we obtain the average earth albedo

$$a = (a_0*70\%+a_f*10\%+a_d*5\%+a_s*15\%)*90\%+a_c*10\%=0.2 \quad (9)$$

Fig. 6 shows the albedo change a - R obtained from (7) as a function of optical thickness t for different aerosol single scattering albedo ω and ωβ = 0.1. We see from Fig. 6 that the critical value of ω (for a - R = 0) to determine whether the aerosols will heat or cool the climate is about 0.8 which agrees well with the result given elsewhere (Hansen *et al.*, 1979). Chylet et al (1974) did not calculate the critical value of ω in their paper.

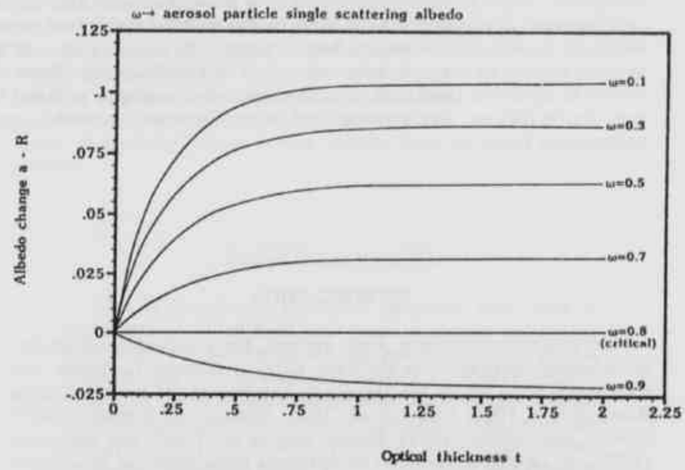

Figure 6. Critical aerosol single scattering albedo ω to determine whether heating or cooling by an aerosol layer for an average earth albedo a = 0.2.

CONCLUSION

In this paper we modify the two-stream approximation (Chylet *et al.*, 1974) by including the effect of solar zenith angle for regional aerosol effect on climate change, and including the effects of both the solar zenith angle and the earth surface curvature for global aerosol effect on climate change. These modifications widen the applicable range of the two-stream approximation and give more accurate results, but the critical condition for determining whether an aerosol layer will heat or cool climate is not changed. Using our modified model and a global average earth albedo of a = 0.2, we obtain a reasonable critical aerosol single scattering albedo value of ω= 0.8 to determine whether aerosol heating or cooling occurs.

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