A Simple Method to Compute the Change in Earth–Atmosphere Radiative Balance Due to a Stratospheric Aerosol Layer

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ABSTRACT

A simple three-layer model of the earth–atmosphere system, including the ground, troposphere and stratosphere, with their interactions, is developed. The model permits the radiative characteristics of both the troposphere and stratosphere to be separately adjusted to describe any atmospheric state. The accuracy is tested against the more precise computations of Herman *et al.* (1976), which take into account the aerosol profile and the angular variation of reflectance at the top of the troposphere. Analytical expressions are obtained for the albedo variation due to a thin stratospheric aerosol layer. The physical procedures are outlined, as well as the influence of the main parameters: aerosol optical thickness, single scattering albedo and asymmetry factor, and sublayer albedo.

The method is applied to compute the variation of the zonal albedo and the planetary radiation balance due to a stratospheric aerosol layer of background H_2SO_4 droplets and of volcanic ash. The resulting ground temperature perturbation is evaluated, using a Budyko type climate model.

1. Introduction

The role of stratospheric aerosols in modifying the earth's radiation balance, and therefore the global climate, is expected to be important due to their long term of residence and extended transport in the stratosphere.

This influence is twofold; a modification of solar reflection and an increase of infrared opacity. This last effect leads to a warming, the degree of which depends on the aerosol infrared absorption, but it has been shown (Luther, 1976; Harshvardhan and Cess, 1976) to be smaller than the effect on solar reflection. The first model calculations, including a crude treatment of the radiative problem, have shown that the aerosol loading generally enhances the planetary reflection, and therefore causes a cooling of the earth's surface and the troposphere. However, they have pointed to a possible inverse effect, warming instead of cooling, when there is absorbing aerosol above clouds, or a highly reflecting ground surface (Charlson and Pilat, 1969; Atwater, 1970; Rasool and Schneider, 1971; Schneider, 1971). Further, more detailed studies have confirmed this result (Yamamoto and Tanaka, 1972; Chylek and Coakley, 1974; Reck, 1974; Wang and Domoto, 1974; Herman and Browning, 1975; Coakley and Grams, 1976; Pollack *et al.*, 1976). Herman *et al.* (1976) have demonstrated the sensitivity to solar elevation angle using a complete model of the earth–atmosphere system, including tropospheric aerosols; the radiative transfer computations use the Gauss–Seidel iteration technique (Herman and Browning, 1965) which is assumed to give results to within better than 1% accuracy, or so-called "exact data." More recently, Harshvardhan (1979) has examined the albedo sensitivity of the earth to a stratospheric aerosol layer, both on the global and zonal scales. He uses a simple analytical formulation, based on primary scattering, with a first-order expansion in τ for the thin stratospheric scattering layer, and assumes the albedo of the sublayer (ground and troposphere) is known. Satellite-derived planetary albedos are used in the numerical computations; the aerosols are assumed to be 75% H_2SO_4 non-absorbing droplets.

Concerning the methodologies, the previous studies can be divided into two main groups. Some methods work out a complete treatment of the radiative transfer problem in the real atmosphere; they lead to credible results, but are computer time-consuming. The other methods use an approximate treatment of the radiative transfer in the thin stratospheric aerosol layer above an unperturbed atmosphere with supposedly known albedo.

In this paper we have worked out an approximate treatment of the whole problem, including the ground, troposphere and stratosphere (Section 2). The purpose was to have at our disposal, for climatological studies, a fast code which will allow us to vary the ground and tropospheric characteristics. The method used both for the troposphere and stratosphere is the two-stream approximation (Irvine, 1968) with a small modification, which will be explained later.

As well as the two-stream approximation, which

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is well-known, the method introduces further simplifications to the physical problem by neglecting the vertical nonhomogeneity of the atmosphere and the angular variation of the reflectances. The possible influence of these approximations has been carefully checked in comparison with the accurate results published by Herman *et al.* (1976) in a realistic case. Results of this comparison are shown in Section 3. Until now, tests have been limited to a cloudless atmosphere, without gas absorption, although the method can include these parameters.

We have restricted the applications presented here to the case of a variable stratospheric loading above a known unperturbed atmosphere. Our purpose was to analyze the climatic impact of stratospheric aerosols of two types, background H_2SO_4 particles and volcanic ash, as they are described in the models proposed by the IAMAP radiation commission (McClatchey *et al.*, 1980); it constitutes a very partial answer to the recommendations of the World Climate Research Programme (1980). The influence of the aerosol layer on the zonal albedo and on the planetary radiation balance, including the infrared opacity increase, is considered in Section 4. Finally the aerosol perturbation, both of the albedo and of the emissivity, is introduced into a Budyko type climate model; the resulting surface temperature variation is computed (Section 4).

2. Modeling and general formulation

As our purpose is to study the influence of stratospheric aerosol, which is strongly dependent on the reflectances of the ground and troposphere, we have devised a three-layer atmosphere model comprising a reflecting surface, troposphere and stratosphere and including interactions between the three components.

In order to define our notations and briefly describe the method, let us first consider a scattering atmosphere above a Lambertian ground (or sublayer). The planar albedo is defined by

$$A(\mu_0) = \frac{1}{\mu_0 f} \int_0^{2\pi} \int_0^{+1} \mu I^+(0; \mu, \phi) d\mu d\phi, \qquad (1)$$

where Arccos μ_0 is the solar zenith angle, Arccos μ is the observation zenith angle, ϕ is the azimuth angle between the observation and the sun vertical planes, $I^+(0; \mu, \phi)$ is the reflected radiance in the direction (μ, ϕ), and f is the solar irradiance.

The spherical or planetary albedo is derived from

$$A_p = 2 \int_0^1 \mu_0 A(\mu_0) d\mu_0 , \qquad (2)$$

where we have also introduced a local albedo $A(\bar{\mu}_0)$, where $\bar{\mu}_0$ is the average daily value of μ_0 for the considered latitude and date.

The planar albedo and the spherical albedo are expressed by considering the successive interactions between the atmosphere and the ground (or sublayer) (Tanre *et al.*, 1979) as

$$A(\mu_0) = S(\mu_0) + [A_G T(\mu_0) \bar{T}/(1 - A_G \bar{S})], \qquad (3)$$

$$A_P = \bar{S} + [A_G \bar{T}^2/(1 - A_G \bar{S})], \qquad (4)$$

where $S(\mu_0)$ is the intrinsic atmospheric contribution to the flux reflectance, $\bar{S}$ is the atmosphere spherical albedo (with no ground reflection), $T(\mu) = e^{-\tau/\mu} + t(\mu)$ is the total (direct + diffuse) atmospheric flux transmittance, τ is the atmospheric optical thickness, $\bar{T}$ is the average (over μ) flux transmittance, and A_G is the ground (or sublayer) albedo.

For our three-layer model, we applied the above formulation in a first step to the troposphere, where A_G holds for the ground albedo, and in a second step to the stratosphere, where A_G has to be replaced by the albedo due to the troposphere–ground system $A_t(\mu_0)$ for the planar albedo calculations, and A_t for the spherical albedo. In the following, subscripts t and s will refer to the tropospheric and stratospheric quantities, respectively.

It must be noted that, even if the ground can be reasonably assumed to be Lambertian on a large scale, this is not true for the ground–troposphere system. In this case, Eqs. (3) and (4) are no longer exact. The exact analysis leads to more complex equations involving the sublayer bidirectional reflectance $\rho_t(\mu_0, \mu, \phi)$. Average values $\langle \rho_t(\mu_0) \rangle$ over μ can be introduced, but they are defined with different weighting functions for each mechanism of ground–atmosphere interaction (Deschamps *et al.*, 1980), so a simple combination of the terms as in Eq. (3) cannot be achieved. An approximation consists of replacing all the $\langle \rho_t(\mu_0) \rangle$ by the sublayer albedo $A_t(\mu_0)$, which leads to equations similar to Eqs. (3) and (4) (Harshvardhan, 1979). We will also use this approximation, which our tests (see Section 3) have proved to be valid.

The scattering function $S(\mu)$ and the diffuse transmission function $t(\mu)$ are expressed using the modified two-stream approximation (Irvine, 1968).

In the non-conservative case, the two-stream formulation is much more difficult to handle than in the conservative case, and we have tried the more approximate expressions

$$\left.\begin{aligned} S(\mu_0) &= \omega S_{\omega=1}(\mu_0) \\ t(\mu_0) &= \omega t_{\omega=1}(\mu_0) \end{aligned}\right\}, \qquad (5)$$

which should hold when the single scattering albedo ω is near to 1, or when primary scattering is predominant, i.e., when τ is small. Table 1 shows the comparison of the diffuse transmission computed by the two-stream method with exact values, including a separate test of Eq. (5) for four optical thicknesses and two zenith angles.

TABLE 1. Comparison of the diffuse transmission $t(\mu_0)$ computed exactly (1); by using the approximation $t(\mu_0) = \omega\, t_{\omega=1}(\mu_0)$ where $t_{\omega=1}(\mu_0)$ is the exact computation (2); and by the two-stream method including the previous approximation (3).*

	$\omega = 1.0$	$\omega = 0.9$	$\omega = 0.8$	$\omega = 0.7$	$\omega = 0.6$
(1)	0.0089	0.0080	0.0071	0.0062	0.0053
$\tau = 0.010$ (2)	0.0089	0.0080	0.0071	0.0062	0.0053
$\mu_0 = 0.999$ (3)	0.0082	0.0074	0.0065	0.0057	0.0049
(1)	0.0452	0.0405	0.0358	0.0312	0.0266
$\tau = 0.052$ (2)	0.0452	0.0407	0.0362	0.0316	0.0271
$\mu_0 = 0.999$ (3)	0.0415	0.0373	0.0332	0.0290	0.0249
(1)	0.1078	0.0960	0.0845	0.0732	0.0622
$\tau = 0.130$ (2)	0.1078	0.0970	0.0862	0.0755	0.0647
$\mu_0 = 0.999$ (3)	0.0993	0.0893	0.0794	0.0965	0.0596
(1)	0.3452	0.2980	0.2548	0.2150	0.1780
$\tau = 0.520$ (2)	0.3452	0.3107	0.2762	0.2416	0.2071
$\mu_0 = 0.999$ (3)	0.3206	0.2885	0.2565	0.2244	0.1923
(1)	0.0150	0.0135	0.0120	0.0105	0.0090
$\tau = 0.010$ (2)	0.0150	0.0135	0.0120	0.0105	0.0090
$\mu_0 = 0.523$ (3)	0.0155	0.0140	0.0124	0.0109	0.0093
(1)	0.0744	0.0665	0.0586	0.0509	0.0433
$\tau = 0.052$ (2)	0.0744	0.0670	0.0595	0.0521	0.0446
$\mu_0 = 0.523$ (3)	0.0771	0.0694	0.0617	0.0540	0.0463
(1)	0.1704	0.1506	0.1315	0.1131	0.0953
$\tau = 0.130$ (2)	0.1704	0.1534	0.1363	0.1193	0.1023
$\mu_0 = 0.523$ (3)	0.1774	0.1597	0.1419	0.1242	0.1064
(1)	0.4537	0.3810	0.3176	0.2617	0.2120
$\tau = 0.520$ (2)	0.4537	0.4083	0.3630	0.3176	0.2722
$\mu_0 = 0.523$ (3)	0.4788	0.4309	0.3830	0.3351	0.2873

* Size distribution $n(r) = 0 < r < 0.02\ \mu m$
$n(r) = 10^{-4}C \quad 0.02\ \mu m \leq r \leq 0.1\ \mu m$
$n(r) = Cr^{-4} \quad 0.1\ \mu m \leq r \leq 10\ \mu m$
$n(r) = 0 \quad r > 10\ \mu m$
Refractive index = 1.50.

The accuracy of the two-stream method, coupled with the approximation [Eq. (5)], remains better than 10% for $\omega \geq 0.6$, $\tau \leq 0.5$ and normal incidence. When μ_0 decreases, the conditions on ω or τ become a little more restrictive. But, as the aerosol single scattering albedo is generally >0.9 in the solar spectrum, the method should apply with good accuracy for most atmospheric problems.

For small optical thickness, as in the stratosphere, a further approximation is introduced by expanding all functions to the first order in τ. This leads to the following simple expressions, for the albedo perturbation $\Delta A = A_s - A_t$, due to a thin stratospheric aerosol layer of optical thickness τ_s and single scattering albedo ω_s:

$$\Delta A(\mu_0)/\tau_S = \omega_S b_S[1 - A_t(\mu_0)]\left[\frac{1}{\mu_0} - 2A_t(\mu_0)\right] - (1 - \omega_S)\left(\frac{1}{\mu_0} + 2\right)A_t(\mu_0), \quad (6a)$$

$$\Delta A_P/\tau_S = 2\omega_S b_S(1 - A_t)^2 - 4(1 - \omega_S)A_t\,, \quad (6b)$$

with

$$b_S = \tfrac{1}{2}(1 - g_s), \quad (7)$$

and where g_s is the phase function asymmetry factor and A_t is the sublayer spherical albedo.

3. Test of the method

Many authors have tested the accuracy of the various two-stream methods (Radiation Commission of IAMAP, 1977), which is generally around a few percent and our further simplification [Eq. (5)] has been tested above to remain within the same general order of accuracy. However, these tests generally concern the method itself and not the simplifying assumptions it imposes on the atmosphere. An important assumption is that each atmospheric layer is treated as a whole, thus eliminating the possible influence of the aerosol mixing ratio profile. Moreover, as mentioned previously, an approximation of the sublayer albedo (neglecting the directional effect) is introduced in the modeling of the troposphere–stratosphere interactions.

TABLE 2. Comparison of the reflected fluxes $\mu_0 A_t\ (\mu_0)$ for unit incident solar flux at the top of the troposphere between HBR results (1); and our results (2). The computations are given for the three ground albedos and the three solar zenith angles.

	15°		45°		65°	
A_G	(1)	(2)	(1)	(2)	(1)	(2)
0.1	0.155	0.156	0.132	0.132	0.105	0.103
0.3	0.315	0.312	0.245	0.243	0.167	0.165
0.9	0.860	0.861	0.635	0.632	0.380	0.381

In order to test this method, we have chosen to apply it to the model used by Herman *et al.* (1976, hereafter HBR), who give results at $\lambda = 0.5\ \mu$m obtained by a complete and accurate resolution of the transfer problem.

a. Atmospheric model

TROPOSPHERE:

Molecular optical thickness $\tau^{\text{Rayl}} = 0.1$
Aerosol characteristics:

- optical thickness $\tau^{\text{aer}} = 0.145$
- size distribution $n(r) = Cr^{-3.5}$
- Elterman's profile
- refractive index $m = 1.54 - 0.00i$.

STRATOSPHERE:

Aerosol uniform layer between 15 and 25 km
Size distribution (log normal)

$$n(r) = \frac{C}{(2\pi)^{1/2}\sigma r}\exp\{-[\ln(r/\bar{r})]^2/2\sigma^2\}$$

$$\sigma = 1.3, \quad \bar{r} = 0.3\ \mu\text{m}$$

Real part of refractive index $m = 1.54$.

The imaginary part of the refractive index of the stratospheric particles is kept variable, in order to vary the single scattering albedo ω_s between 1 and 0.6. The stratospheric aerosol mass loading is varied between zero and 3.2 μg m^{-3}, which corresponds to an optical thickness τ_S between zero and 0.1216. Three ground albedos, $A_G = 0.1$, 0.3, 0.9 and three solar zenith angles, Arccosμ_0 = 15, 45 and 65° are considered.

b. Results for the tropospheric albedo

Table 2 shows the comparison of our results with HBR results for the three ground albedos and the three solar zenith angles.

The agreement is always better than 1%, which validates our neglecting of the vertical profile and the use of the two-stream method in the case of a cloudless atmosphere.

c. Results for the stratospheric aerosol influence

The albedo variations ΔA have been computed using Eq. (6a) and also with the primary scattering approximation. As will be discussed in Section 4, compensation between positive and negative terms may lead to $\Delta A = 0$ for a particular equilibrium value of the single scattering albedo ω_S^{eq}. Table 3 compares the values obtained by both methods with HBR exact values. Table 4 gives a similar comparison for the slopes $\Delta A(\mu_0)/\tau_S$ for five values of ω_S between 0.6 and 1. All the comparisons are done for the three ground albedos and the three solar zenith angles.

For 15 and 45° solar zenith distance, the two-stream results are better than the primary scattering results, the equilibrium single scattering albedos are nearly exact, and the slopes agree within better than 10% with the exact values. For 65° solar zenith distance the converse becomes true, with a slightly better agreement of the primary scattering and exact values. In any case, the difference between the two methods is very small. The two-stream method has a slight advantage, considering the larger influence of small solar incidences on the radiation balance. However, the method may fail for near-grazing incidences, where the earth's curvature should also be taken into account. However, the accuracy of the method seems to increase with the ground albedo; this may have some compensating effect at high latitude, where the sun is low, but where the ice- or snow-covered ground is highly reflective (Ellis and Vonder Haar, 1976).

It is important to emphasize that HBR includes the directional variation of the troposphere albedo, and the good agreement with their results proves the validity of the assumption made on this point in our formulation. Finally, Fig. 1 reproduces HBR Figs. 2, 3, 4 in comparison with our results plotted in the same form.

4. Stratospheric aerosol influence on planetary albedo and radiation balance

a. Stratospheric aerosol models

In the following sections we will focus on the perturbation introduced by an increase of the aerosol

TABLE 3. Comparison of the equilibrium single scattering albedo ω_S^{eq} obtained by both methods [two-stream (2); and primary scattering (3); approximations] with HBR exact values (1). The results are given for the three ground albedos and three solar zenith angles. Dashes indicate cases where equilibrium does not exist (always warming).

	15°			45°			65°		
A_G	(1)	(2)	(3)	(1)	(2)	(3)	(1)	(2)	(3)
0.1	0.88	0.88	0.96	0.85	0.87	0.90	0.82	0.87	0.82
0.3	0.98	0.97	—	0.95	0.95	0.98	0.90	0.94	0.92
0.9	—	—	—	—	—	—	—	0.998	—

TABLE 4. As in Table 3, except for the slopes $\Delta A(\mu_0)/\tau_S$. The results are given for the three ground albedos and for five values of ω_s.

	$A_G = 0.1$			$A_G = 0.3$			$A_G = 0.9$		
ω_s	(1)	(2)	(3)	(1)	(2)	(3)	(1)	(2)	(3)
					$\theta_0 = 15°$				
0.6	−0.140	−0.155	−0.185	−0.334	−0.374	−0.405	−0.987	−1.088	−1.096
0.7	−0.093	−0.099	−0.135	−0.268	−0.273	−0.309	−0.770	−0.818	−0.828
0.8	−0.045	−0.043	−0.084	−0.185	−0.172	−0.213	−0.553	−0.549	−0.560
0.9	*	+0.013	−0.033	−0.083	−0.071	−0.117	−0.316	−0.279	−0.292
1.0	+0.064	+0.069	+0.018	*	+0.031	−0.021	−0.051	−0.009	−0.024
					$\theta_0 = 45°$				
0.6	−0.172	−0.196	−0.212	−0.381	−0.436	−0.455	−1.128	−1.223	−1.230
0.7	−0.106	−0.122	−0.142	−0.296	−0.313	−0.336	−0.902	−0.919	−0.926
0.8	−0.035	−0.049	−0.071	−0.184	−0.190	−0.216	−0.621	−0.614	−0.623
0.9	+0.042	+0.025	0.000	−0.057	−0.068	−0.096	−0.338	−0.309	−0.319
1.0	+0.080	+0.098	+0.071	+0.066	+0.055	+0.023	*	−0.005	−0.015
					$\theta_0 \doteq 65°$				
0.6	−0.259	−0.327	−0.287	−0.519	−0.615	−0.588	−1.354	−1.571	−1.569
0.7	−0.142	−0.204	−0.157	−0.360	−0.433	−0.401	−1.039	−1.176	−1.175
0.8	*	−0.081	−0.028	−0.189	−0.251	−0.215	−0.767	−0.782	−0.780
0.9	+0.125	+0.042	+0.102	0.000	−0.070	−0.029	−0.471	−0.388	−0.386
1.0	+0.271	+0.165	+0.231	+0.189	+0.112	+0.157	*	+0.006	+0.009

* Difficult to estimate from the Figures in Herman *et al.* (1976).

stratospheric loading, using mainly Eq. (6), where we will assume the ground-troposphere local albedo $A_t(\bar{\mu}_0)$ as known from experimental observations.

The IAMAP Radiation Commission (McClatchey *et al.*, 1980) has recently proposed standard atmospheric models for use in radiation calculations. They are based on our best present knowledge and represent a fairly realistic description, even if they cannot be expected to describe all the possible conditions of the real atmosphere. Their main purpose is to furnish a small number of well-defined atmospheres useful for intercomparisons.

In this paper, we are using the two stratospheric aerosol models of the Standard Radiation Atmosphere (SRA). The background stratospheric aerosol (BSA), comprising 75% H_2SO_4 solution, and the volcanic aerosol (VA), are characterized by their complex refractive index (McClatchey *et al.*, 1980, Table 7) and their size distribution given by a modified gamma function

$$n(r) = Ar^{\alpha}\exp(-br^{\gamma}); \qquad (8)$$

the coefficients α, b, γ are listed in Table 5; A is a normalization constant related to the total number of particles, or to the optical thickness. The main parameters of the stratospheric aerosol layer (used in the above formulation) are the optical thickness τ_s, the single scattering albedo ω_s and the backscattering coefficient $b_s = (1 - g_s)/2$, related to the asymmetry factor g_s. They are computed for a series of wavelengths from the Mie theory. As we are interested in the aerosol influence over the whole solar spectrum, we will define the following effective values:

$$\tau_s^{\text{eff}} = \int_0^{\infty} \tau_{s\lambda} f_\lambda d\lambda \left(\int_0^{\infty} f_\lambda d\lambda\right)^{-1}, \qquad (9)$$

$$\omega_s^{\text{eff}} = \frac{1}{\tau_s^{\text{eff}}} \int_0^{\infty} \omega_{s\lambda}\tau_\lambda f_\lambda d\lambda \left(\int_0^{\infty} f_\lambda d\lambda\right)^{-1}, \qquad (10)$$

$$b_s^{\text{eff}} = \frac{1}{\omega_s^{\text{eff}}\tau_s^{\text{eff}}} \int_0^{\infty} b_{s\lambda}\omega_{s\lambda}\tau_{s\lambda} f_\lambda d\lambda \left(\int_0^{\infty} f_\lambda d\lambda\right)^{-1}, \qquad (11)$$

where the weighting function f_λ is the incoming solar monochromatic irradiance at the top of the atmosphere. Table 5 gives the values of ω_s^{eff} and b_s^{eff} for both SRA models. Similarly, we will define an effective infrared optical thickness

$$\tau_{s\text{IR}}^{\text{eff}} = \frac{1}{\sigma T_s^4} \int_0^{\infty} \tau_{s\lambda} B_\lambda(T_s) d\lambda = C_s \tau_s^{\text{eff}}, \qquad (12)$$

weighted by the blackbody radiance B_λ at the stratospheric temperature taken as $T_s = 215$ K. The ratio $C_s = \tau_{s\text{IR}}^{\text{eff}}/\tau_s^{\text{eff}}$ is also given in Table 5.

b. Influence of aerosol on the local albedo

Eq. (6a) has been applied to the monthly average albedo sensitivity for 10° latitude belts. The average daily value of the solar zenith angle ($\bar{\mu}_0$) is taken for the 21st of each month and the mean latitude of the zone. The unperturbed atmosphere albedo $A_t(\bar{\mu}_0)$ is taken to be the mean monthly values quoted in Ellis

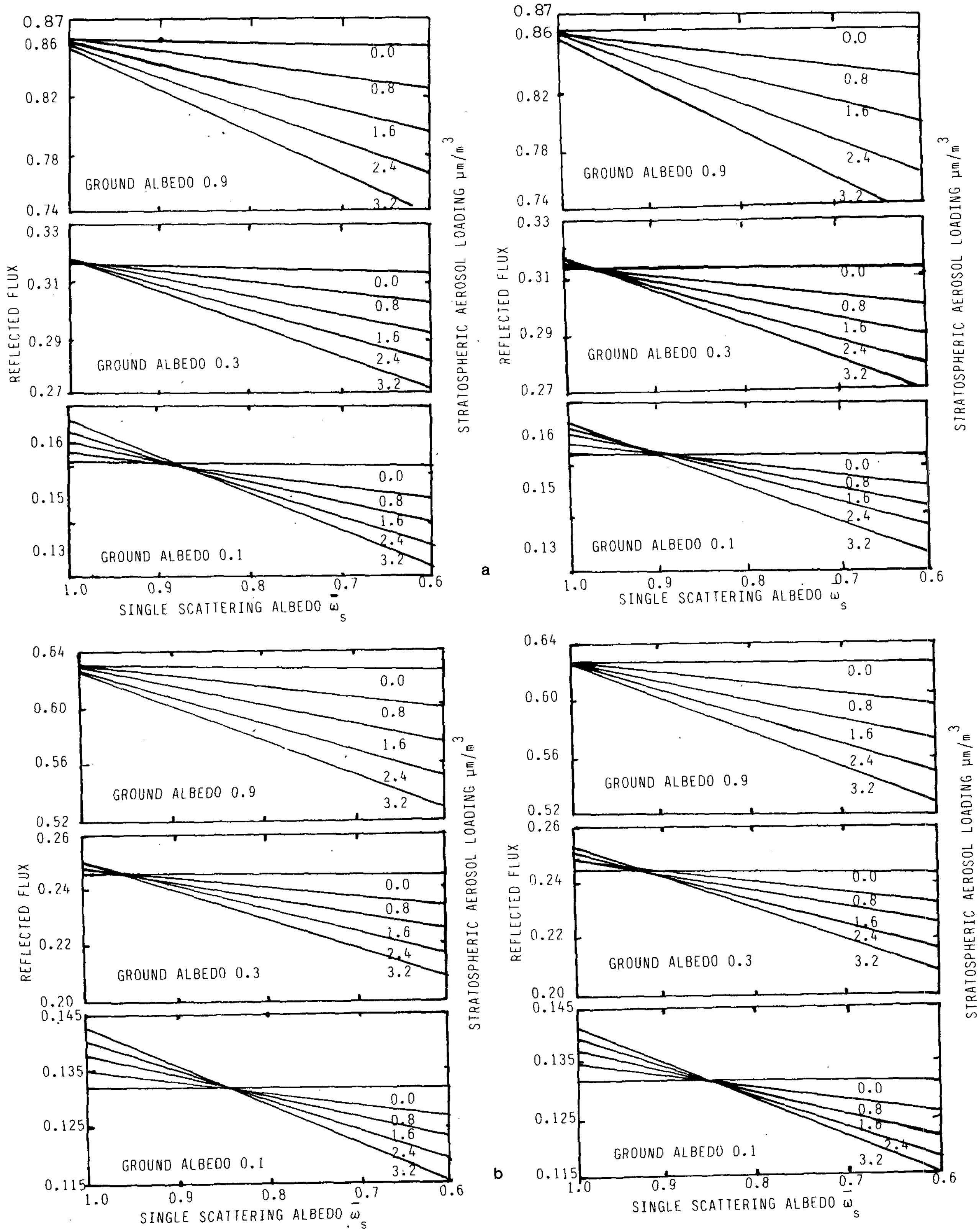

FIG. 1. Comparison of reflected flux as a function of the atmospheric aerosol single scattering albedo, for five values of the stratospheric aerosol loading for unit incident solar flux. Left: HBR results; right: our results. Incident solar angle θ_0 = 15° (a), 45° (b), 65° (c).

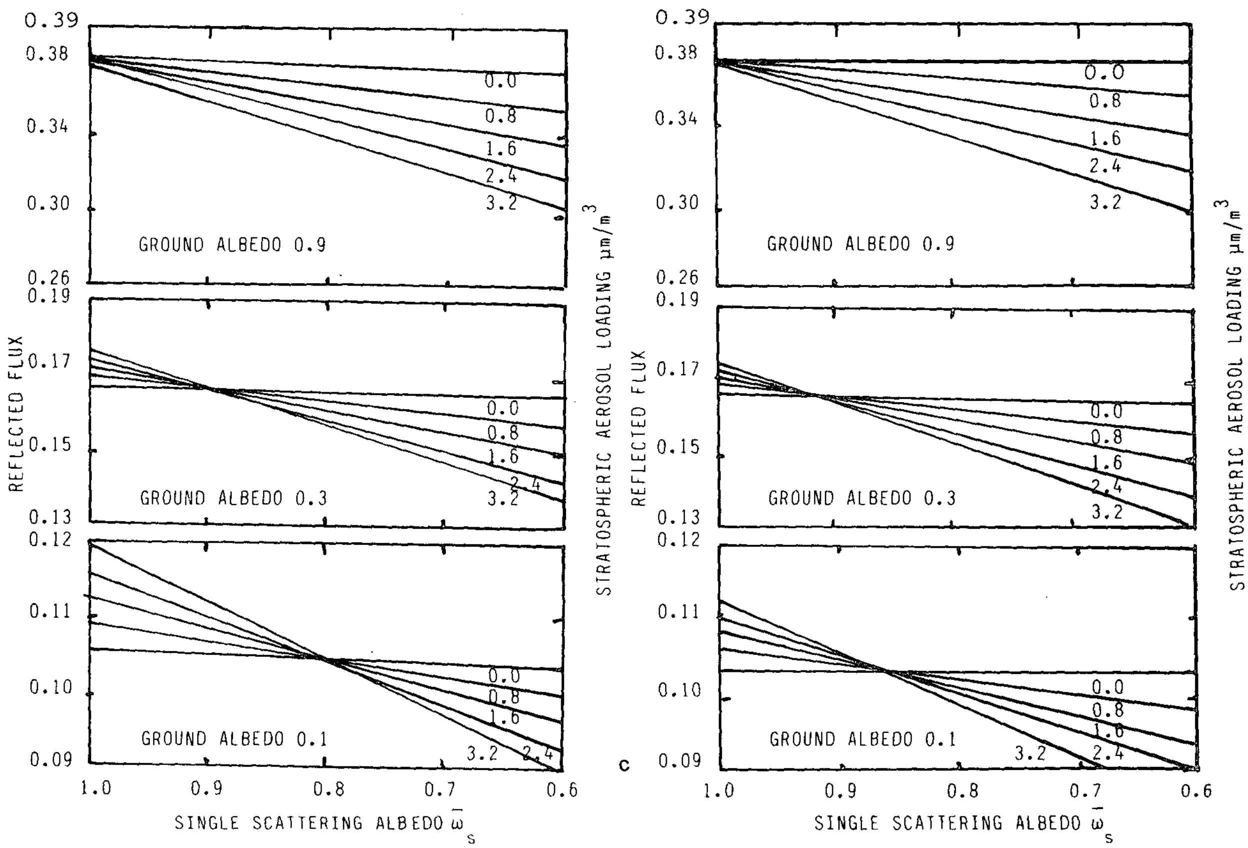

FIG. 1. *(Continued)*

and Vonder Haar (1976). As this albedo is only known by its average value over the solar spectrum, it seems more sensible to use for τ_s, ω_s and b_s the average effective values of the stratospheric parameters as defined by Eqs. (9), (10) and (11) rather than to repeat the computations for a series of wavelengths.

A comparison with Harshvardhan's (1979) results for the same case ($\tau_s = 0.03$, $\omega_s = 1$, $b_s = 0.135$) has shown a very good agreement, although Harshvardhan averaged the albedo itself on the solar spectrum instead of the parameters.

Fig. 2 shows the results obtained for the two SRA aerosol models for both hemispheres; the stratospheric optical thickness has been kept constant at $\tau_s = 0.03$, which is ~10 times the present value. A comparison of the two diagrams of Fig. 2, together and with Harshvardhan's (1979) results, outlines the influence of the aerosol model. Of course, the three contours display the same general pattern, with a large seasonal variation at high latitudes due to the solar incidence variation, a slight seasonal variation at mean latitudes where the winter increase of albedo is partly compensated by the decrease of $\bar{\mu}_0$, and a nearly constant value of ΔA at low latitudes where both A and $\bar{\mu}_0$ do not vary much. The influence of the rather strong absorption assumed in the SRA volcanic model, results in a strong reduction of the albedo increase; with volcanic aerosols, one can even trace in the Northern Hemisphere a 0% isopleth corresponding to a complete compensation and a slight decrease of albedo at high latitudes in summer. The backscattering coefficient has a small influence, ΔA, increasing slightly with b_s.

Eq. (6a) makes it easy to understand the various physical processes; the first term on the right-hand side corresponds to the radiation scattered by the aerosols; it comprises the atmospheric upward scattering $\omega_s b_s/\mu_0$ (which is a gain, increasing the albedo), the loss by scattering before and after reflection $-\omega_s b_s(1/\mu_0 + 2)A_t(\mu_0)$, and the gain $2\omega_s b_s A_t^2(\mu_0)$ due to the radiation scattered downward by the layer and reflected twice by the sublayer. The second term $(1 - \omega_s)(1/\mu_0 + 2)A_t(\mu_0)$ corresponds to the loss of radiation due to the aerosol absorption. A similar interpretation can be made for the spherical or planetary albedo [Eq. (6b)]. In the case of the planetary albedo, it is easy to find from Eq. (6b) the equilibrium value of the single scattering albedo leading to a perfect compensation ($\Delta A_P = 0$), as

$$\omega_s^{eq} = \frac{2A_t}{b_s(1 - A_t)^2 + 2A_t}, \qquad (13)$$

TABLE 5. Characteristics of the aerosol models (see text).

Aerosol model	α	b	γ	ω_s^{eff}	b_s^{eff}	$C_s = \tau_{\text{IR}}^{\text{eff}}/\tau_s^{\text{eff}}$
Background stratospheres aerosol	1.0	18	1.0	0.998	0.189	0.0383
Volcanic aerosol	1.0	16	0.5	0.994	0.169	0.0217

which varies from 0 for $A_t = 0$ to 1 (conservative scattering) for $A_t = 1$.

For the planar or local albedo, the equilibrium value depends both on A_t and μ_0; it may even not exist, as can be understood by considering that the scattering term [first term in Eq. (6a)] is either positive or negative, according to the sign of $1/\mu_0 - 2A_t(\mu_0)$; for $\mu_0 > \frac{1}{2}A_t(\mu_0)$, which is possible if $A_t(\mu_0) > 0.5$, even non-absorbing aerosols ($\omega_s = 1$) give a decrease of A. As the loss of energy is actually related to the absorption of the scattered light in the sublayer, the equilibrium value reappears at $\omega_s^{\text{eq}} = 1$ for $A_t(\mu_0) = 1$. The general behavior of $\Delta A(\mu_0)$ when varying ω_s, μ_0 and $A_t(\mu_0)$ is illustrated in Fig. 1.

c. Influence of aerosol on the planetary radiation balance

The global planetary balance expresses the equilibrium between the absorbed solar flux and the infrared emitted flux; it can be written as

$$f(1 - A_P) = 4\sigma T_P^4, \tag{14}$$

where A_P is the planetary (or spherical) albedo, T_P is an effective planetary temperature, f is the solar irradiance, and σ is the Stefan–Boltzman constant.

The addition of a stratospheric aerosol layer modifies the planetary albedo by ΔA_P [see Eq. (6b)] and the emitted infrared flux by

$$\Delta F_{\text{IR}} = [\epsilon_s \sigma T_s^4 + (1 - \epsilon_s)\sigma T_P^4] - \sigma T_P^4, \tag{15}$$

where ϵ_s is the emissivity of the aerosol layer assumed to be at a uniform temperature T_s different from T_P.

In the longwave spectrum, the aerosol scattering can be neglected, and expanding the transmission to the first order in τ, a mean effective emissivity is given from the infrared effective optical thickness defined by Eq. (12) as

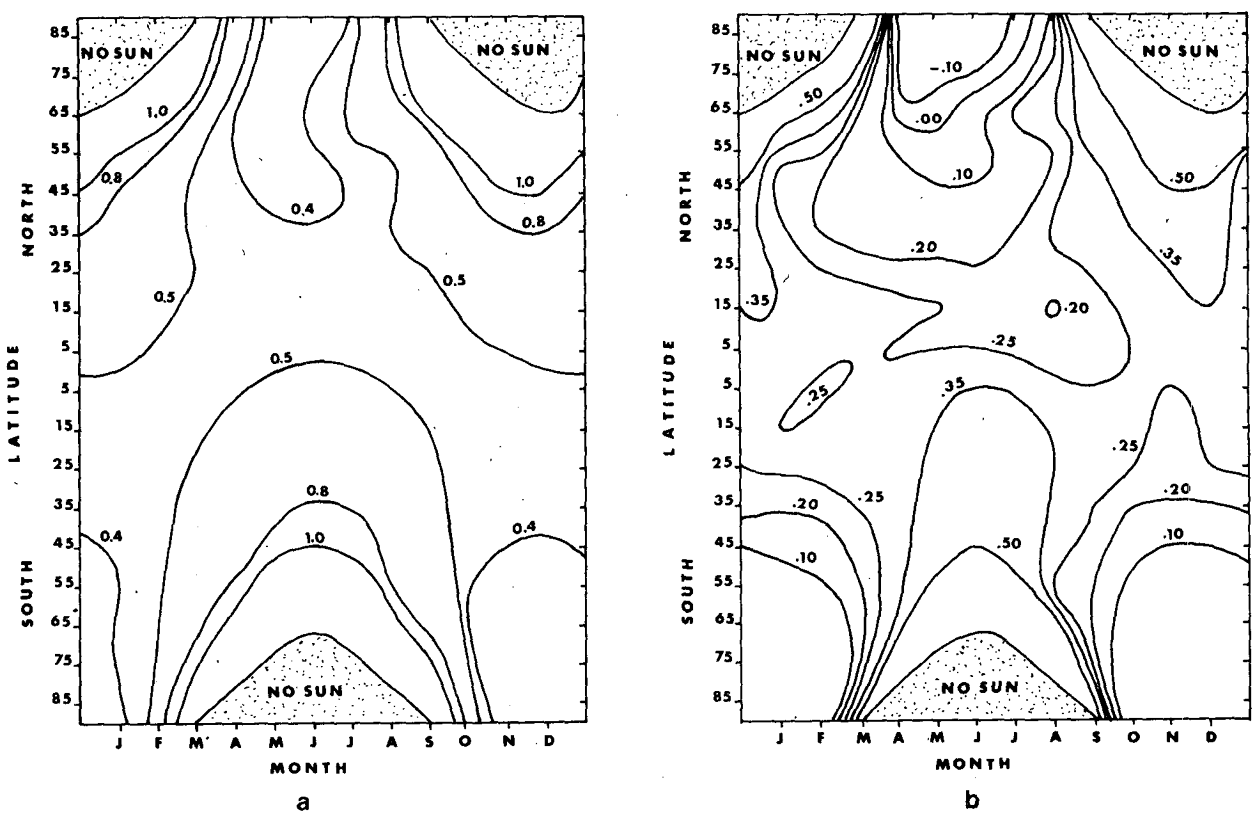

FIG. 2. Change in zonal albedo (percent) with addition of a stratospheric aerosol layer ($\tau_s = 0.03$) for the two SRA models. (a) Background aerosols; (b) Volcanic aerosols.

TABLE 6. Modification of the planetary budget ΔB (W m^{-2}) for the two SRA aerosol models with τ_s = 0.03.

Aerosol model	Scattering $-2\omega_s b_s \tau_s (1-A_P)^2 f$	Solar absorption $4(1-\omega_s)\tau_s A_P f$	Emission $(1-A_P)\tau^{IR} f$	ΔB
Background stratospheres aerosol	−28.5	0.0923	1.56	−26.8
Volcanic aerosol	−26.4	2.28	0.887	−23.2

$$\epsilon_s = 2C_s\tau_s . \quad (16)$$

Eq. (14) with Eq. (15) yields a planetary budget

$$\Delta B = -f\Delta A_P + 4\epsilon_s\sigma(T_P{}^4 - T_S{}^4), \quad (17)$$

which is a variation from the $\Delta B = 0$ equilibrium value; ΔB is a flux per unit surface area and is counted as positive if there is a net gain by the planet.

For the sake of simplicity we will assume $T_s^4 = T_P{}^4/2$, which is approximately verified ($T_s \approx 215$ K, $T_P \approx 255$ K), and using Eq. (16) and (14) into Eq. (17) yields

$$\Delta B = -f\Delta A_P + f(1 - A_P)C_s\tau_s = -f(\Delta A_P + \Delta A'_P), \quad (18)$$

where ΔA_P is given by Eq. (6b); $\Delta A'_P$, due to the infrared greenhouse effect of the aerosols, can be taken into account by a further "equivalent variation of albedo"

$$\Delta A'_P = -(1 - A_P)C_s\tau_s . \quad (19)$$

Table 6 gives the modification of the planetary budget due to a stratospheric aerosol layer of optical thickness (averaged over the solar spectrum) $\tau_s = 0.03$ for both the SRA aerosol models; the planetary albedo $A_P = 0.309$ is taken from Ellis and Vonder Haar (1976). The contributions due to aerosol scattering, aerosol solar absorption and aerosol infrared emission are shown separately. It is clear that the backscattering term is dominant. The infrared contribution is at least one order of magnitude smaller, which *a posteriori* justify our approximate treatment of that contribution. The absorption term, which is of course zero for a conservative layer ($\omega_s = 1$), can become significant for absorbing aerosols, as in the volcanic model.

d. Climatic impact of the stratospheric aerosols

Using a global climate model, the modification of the planetary radiation balance, due to the stratospheric aerosols, can be translated into a mean ground temperature variation. A simple Budyko-type model (Budyko, 1969; Chylek and Coakley, 1975) will be applied with the two SRA aerosol models, and the possible feedback due to the variation of the polar ice cap limit will be emphasized.

In the climate model of Budyko (1969), the energy balance at any latitude ϕ = Arcsin$x$ is expressed by

$$\frac{f}{4}S(x)[1 - A(x)] - F_{IR}(x) = D(x), \quad (20)$$

where $(f/4)S(x)$ is the annual average incoming solar flux at the top of the atmosphere at the latitude x, $A(x)$ is the annual average value of the local albedo $A(\bar{\mu}_0)$, $F_{IR}(x)$ is the annual average outgoing infrared radiation flux, and $D(x)$ is the heat divergence resulting from the horizontal redistribution by the oceanic and atmospheric circulations.

Budyko assumes the empirical expressions

$$F_{IR}(x) = a + BT(x), \quad (21)$$

$$D(x) = \beta[F_{IR}(x) - \bar{F}_{IR}], \quad (22)$$

$$\bar{F}_{IR} = a + B\bar{T}, \quad (23)$$

where $\bar{T}$ is the hemispherical (or planetary) mean surface temperature, $T(x)$ is the mean surface temperature at the latitude x, and $a = 202$ W m^{-2}, $B = 1.46$ W m^{-2} K^{-1}, $\beta = 2.61$ are empirical coefficients.

On the planetary scale, radiative equilibrium is achieved and the integration of Eq. (20) over a hemisphere with Eq. (23) leads to an expression which fixes the planetary average surface temperature $\bar{T}$:

$$\frac{f}{4}(1 - A_P) = \bar{F}_{IR} = a + B\bar{T}, \quad (24)$$

where A_P is the planetary albedo defined previously.

When a stratospheric aerosol layer is added to the unperturbed atmosphere, the planetary albedo is changed from A_P to $A_P + \Delta A_P + \Delta A'_P$, with ΔA_P given by Eq. (6b) and $\Delta A'_P$ by Eq. (19). This leads to an expression for surface temperature variation (averaged over the planet):

$$\Delta\bar{T} = -\frac{f}{4B}(\Delta A_P + \Delta A'_P). \quad (25)$$

The variation of $\Delta\bar{T}$ versus τ_s was first computed for the aerosol model used by Harshvardhan and Cess (1976) and characterized by ($\omega_s = 1$, $b_s = 0.185$, $C_s = 0.0408$). The agreement is within a few percent. In Fig. 3, $\Delta\bar{T}$ is plotted versus τ_s for the two SRA models. According to many authors' usage and in order to make results clearer, the scale is extended to $\tau_s = 0.1$, although such a high value (approximately 40 times the present one) seems unlikely to be realized. All results are for $A_P = 0.309$ taken from Ellis and Vonder Haar (1976). For a given optical thick-

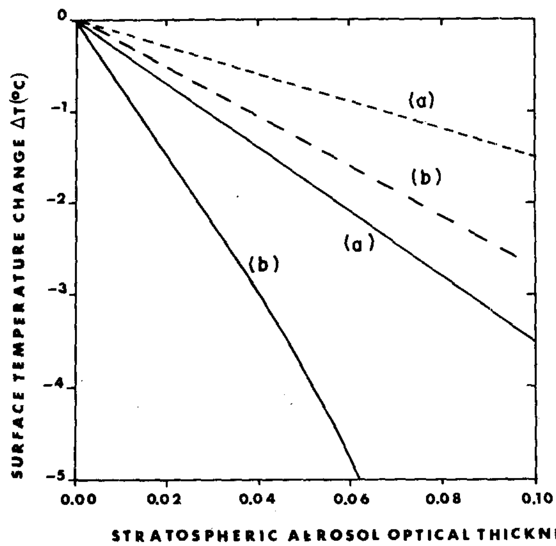

FIG. 3. Variation of the average planetary surface temperature $\Delta\bar{T}$ versus the stratospheric aerosol optical thickness for the SRA aerosol models (infrared included). (a) Without ice feedback, (b) With ice feedback. Solid lines indicate background aerosols; dashed lines, volcanic aerosols.

ness, the cooling is smaller for the volcanic aerosol ($\Delta T = -0.45°C$, for $\tau_s = 0.03$) than for the background aerosol ($\Delta T = -1.1°C$ for $\tau_s = 0.03$); this last model gives the same results as the Harshvardhan and Cess model. This difference between aerosols is mainly due to the compensating influence of solar radiation absorption when ω_s decreases.

In the Budyko climate model, it is possible to infer the stratospheric aerosol influence on the polar ice cap limit and its feedback effect on the surface temperature. Budyko uses a very approximate representation of the unperturbed atmospheric albedo, with a discontinuity at the ice cap edge latitude x_s. His albedo values lead to a planetary albedo of 0.328 for $x_s = 72°$ (present climate value), which is slightly higher than the Ellis and Vonder Haar value $A_P = 0.309$. We have checked that such a change of A_P modifies the $\Delta\bar{T}$ by <0.1°C for $\tau_s = 0.03$, before using the Budyko model to infer the ice edge shift feedback on the surface temperature variations due to the stratospheric aerosol increase. Using Eq. (20) at $x = x_s$ and Eq. (24), one can compute x_s and $\bar{T}$ (or their variations) due to the stratospheric aerosols layer.

Results of $\Delta\bar{T}$ versus τ_s, including the ice feedback, are shown in Fig. 3. For $\tau_s = 0.03$, $\Delta\bar{T}$ increases to −2.15°C (instead of −1.1°C without feedback) for the background aerosol, and to −0.8°C (instead of −0.45°C) for the volcanic aerosol, that is, approximately by a factor of 2. It must be kept in mind that this factor of 2 is probably an upper limit, because the Budyko model is known to show a particularly high sensitivity.

Because the single-scattering albedo is the dominant parameter, we have varied it for both models, keeping b_s and C_s constant. The results for both models are similar, as expected. They are plotted on Fig. 4, which shows the variation of the average surface temperature $\Delta\bar{T}$ versus ω_s for three values of τ_s; for comparison, one curve without ice feedback is given.

The calculated equilibrium value of ω_s corresponds to an exact compensation between the different aero-

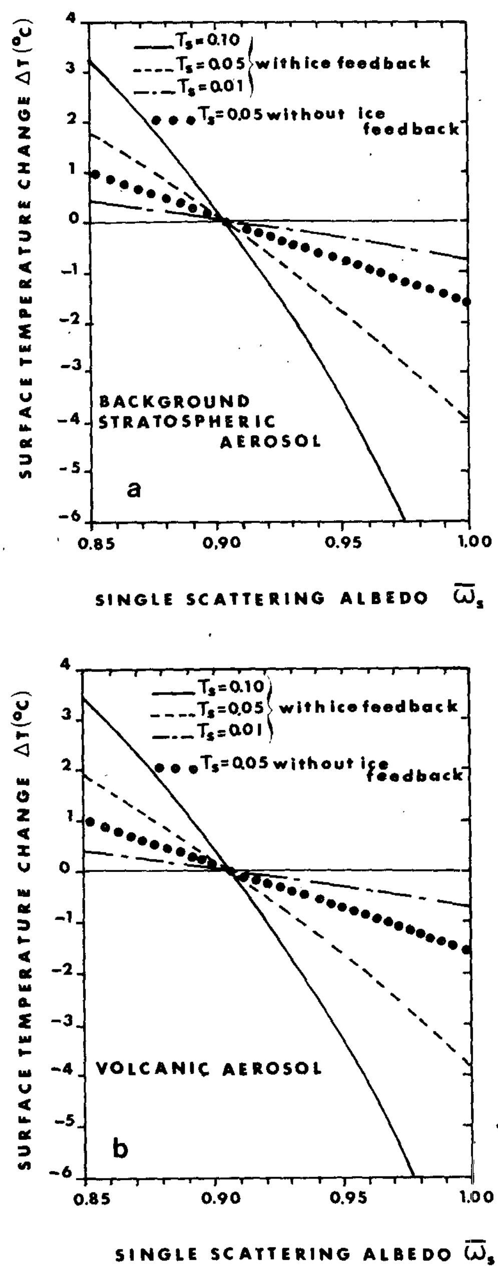

FIG. 4. Variation of the average planetary surface temperature $\Delta\bar{T}$ with the single scattering albedo.

sol effects, including the IR greenhouse effect. This value is found close to 0.905 for both models (low sensitivity to b_s and C_s), placing a boundary between the cooling due to the most frequent aerosols, such as those of both our models, and the possible warming by very strongly absorbing aerosols.

5. Conclusion

A simple formulation has been used to describe the whole atmosphere, including the interactions between the ground, the troposphere and the stratosphere and based on a slightly modified two-stream approximation. Therefore, the radiative characteristics of either the troposphere or stratosphere can be separately adjusted to describe any atmospheric state. The accuracy has been checked on a realistic model by comparison with exact computational results (Herman *et al.*, 1976). The local and planetary albedos are retrieved almost exactly and the variations of albedo due to a stratospheric aerosol layer are reproduced with an accuracy of 10% in the worst cases and generally much better. We believe that we have thus filled a gap between the accurate but computer-time-consuming methods and the fast approximate methods, which generally assume that the stratospheric layer above the ground–troposphere system is characterized by a known and angular-independent albedo. Numerical tests have demonstrated the validity of the method as a whole, including both the mathematical treatment of radiative transfer (two-stream approximation) and more important, the modeling of the physical problem (neglect of vertical nonhomogeneity, angular variation, etc.).

The climatic impact of a stratospheric aerosol layer has been investigated, considering the two aerosol models of the Standard Radiation Atmosphere proposed by the IAMAP Radiation Commission (McClatchey *et al.*, 1980) as standard reference cases. The background stratospheric aerosol model (H_2SO_4 particles) leads to results quite similar to those found previously by Harshvardhan (1979) and Harshvardhan and Cess (1976), whereas the volcanic aerosol model has a much smaller influence, due to the opposite and partially compensating influence of the aerosol scattering and absorption of solar radiation. In this last case, one can even observe a slight decrease of the local albedo at high latitudes in summer, which means a heating effect of the aerosol. However, on the planetary scale, both models lead to a loss of energy of respectively 27 and 23 W m^{-2}, for an optical thickness of 0.03 ($\sim$10 times the percent value), including the small infrared greenhouse effect i.e., cooling occurs. The possibility of heating will appear only for much more absorbing aerosols with a single-scattering albedo < 0.9. Using a simple Budyko-type model, this can be related to surface temperature decreases of 1.1 and 0.45°C, respectively, for a layer of background and volcanic aerosols with an optical thickness of 0.03. Such a cooling, if maintained over a long period of time, could lead to a shift of the ice cap edge towards the equator, which in turn would amplify the temperature decrease.

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