Optical and Radiative characteristics of the Cloudy and Aerosol Atmosphere

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ABSTRACT

Apparent uncertainties concerning the evaluations of the aerosol and cloud impact on the solar irradiance and the climate characteristics of the Earth's atmosphere in recent studies are determined by the strong time and space variability of the aerosol and cloud optical properties. The reviewed results of the optical parameter derivation were obtained during field radiation experiments carried out by the Institute of Atmospheric Physics, Russian Academy of Sciences and Meteorology Observatory of Moscow State University in different Earth's regions and for different aerosol and cloud types. New results deal with the sensitivity of the solar broad band irradiance to the optical parameters of the aerosols of large biomass burning. The strong sensitivity of the solar radiation fluxes at the atmosphere boundaries and the heating rate inside the atmosphere to the aerosol optical thickness and the single scattering albedo is shown.

Key words: Solar radiation, aerosols, clouds.

1. INTRODUCTION

There are apparent uncertainties concerning the evaluation of the aerosol and cloud impact on the solar irradiance and the atmosphere climate characteristics in recent studies. Firstly, not all optical properties of different aerosol and cloud types are well known due to their strong space, time and spectrum variability. On the other hand, the climate and atmospheric general circulation models (AGCM) employ the approximate radiative transfer codes. Precise methods of the radiation transfer computation are time consuming to be utilized in these codes. Intercomparison of the radiation schemes (Ellingson *et al.*, 1991; Fouquart *et al.*, 1991) have revealed the large discrepancies between their results.

Several radiation field experiments have been made recently to study the optical parameters of the desert, smoke, arctic or urban aerosols and cirrus, cumulus or stratocumulus clouds. The main optical characteristics of the aerosol and cloud layers are the optical thickness (τ), the single scattering albedo (ω) and the asymmetry factor of the scattering phase function (g). The aerosol optical parameters are governed by the particle size distribution and the refractive index of the particle substance. The standard models for the main aerosol types have been proposed by Shettle and Fenn (1976) based on field measurements and calculations based on the Mie scattering theory. Some aerosol models (Blanchet & List, 1983; Sokolik & Golitsyn, 1992) were developed later.

The cloud optical parameters are governed by the cloud liquid water content:

$$LWC = \frac{4\pi}{3}\rho\int r^3 n(r)dr, \qquad (1)$$

and the effective radius of cloud droplets:

$$r_e = \int r^3 n(r)dr / \int r^2 n(r)dr, \qquad (2)$$

where n(r) is the cloud drop size distribution, r (μm) is the cloud droplet radius and ρ (g cm^{-3}) is the density of water. The parameterization for shortwave radiative properties of water clouds for different spectral intervals have been made by Slingo (1989).

It is impossible here to review all papers dealing with the derivation of aerosol and cloud optical parameters and the developing of the radiation codes for models last years. Therefore, only the results obtained by the author and her colleagues at the Institute of Atmospheric Physics, Russian Academy of Sciences in collaboration with the Meteorological Observatory of the Geographycal Department, Moscow State University are described below.

2. REVIEW

The climate model has been developed at the Institute of Atmospheric Physics (Petukhov, 1980) for the investigation of the large-scale and long-term processes in the Earth's climatic system (longer than month). It is thermodynamical and zonally averaged model, where the main equation is the equation of the heat balance and the main variable is the temperature. The main feedback relations are the cloud amount and radiation, temperature and the boundary of the ice cover, the temperature and humidity and precipitation. The cloud amount, moisture content, thickness and height of clouds are included. The dynamical processes over the temperature field in the atmosphere and the underlaying layer are parameterized. This climate model was used to evaluate troposphere aerosol impact on the global albedo and the surface temperature.

The average global aerosol optical thickness $\bar{\tau}_{a,\lambda_0}$ for the Earth's atmosphere is assumed to be equal 0.125 (Toon & Pollack, 1976). It may be expected that $\bar{\tau}_{a,\lambda_0}$ will increase in the future due to the industrial contamination of the atmosphere. It is difficult to evaluate the mean global value of the aerosol single scattering albedo averaged over solar spectrum in present. The derivation of this parameter from radiation measurements performed recently in different regions gave comparable values (approximately $\bar{\omega}_a \approx 0.95$) according to Tarasova *et al.* (1992b). The increase of the soot loading of the atmosphere due to the anthropogenic aerosols can decrease the aerosol single scattering albedo to approximately 0.5 (Shettle & Fenn, 1976). The changes of the mean global albedo and surface temperature shown in Table III due to the different aerosol loading of the troposphere comparing with the case without aerosol were calculated utilizing the climate model.

The impact of the troposphere aerosol on the atmosphere-surface albedo and solar absorption has been taken into account only. The much smaller aerosol influence on the thermal radiation was neglected. The change of the mean global albedo and surface temperature strongly depends on the aerosol optical properties. Therefore, it is necessary to know aerosol optical properties more precisely in different regions and for different seasons to evaluate the possible impact of the troposphere aerosol on Earth's climate.

The following methods were proposed to derive the aerosol and cloud optical properties from the integral and broad band direct beam and diffuse solar irradiance measurements at the Earth's surface. The main parameter strongly influencing the transfer of the solar radiation in the Earth's atmosphere is the optical thickness of the aerosol or cloud layer:

$$\tau = \int_{z_1}^{z_2} \varepsilon(z)dz\,, \tag{3}$$

where $\varepsilon(z)$ is the extinction coefficient at the height z. The aerosol optical depth strongly depends on the solar wavelength due to the small sizes of aerosol particles and can be written as follows:

$$\tau_{a,\lambda} = \tau_{a,\lambda_0}\left(\frac{\lambda_0}{\lambda}\right)^{\alpha}, \tag{4}$$

where τ_{a,λ_0} is the aerosol optical thickness at the wavelength 0.5 μm and α is the wavelength exponent varying from 0.5 to 2 depending on the aerosol particle size. Therefore, it is necessary to know two parameters τ_{a,λ_0} and α to determine solar extinction by aerosols in the Earth's atmosphere.

TABLE I
Simulated change of the mean global albedo (Δ A) and mean global surface temperature (Δ T_s, °K) due to aerosol loading of the Earth's atmosphere with different optical properties.

$\bar{\tau}_{a,\lambda 0}$	$\bar{\omega}_a$	Δ A	Δ T_s
0.125	0.95	0.018	−3.4
0.28	0.95	0.038	−6.8
0.28	0.8	0.013	−2.6
0.28	0.5	−0.018	2.7

It is easy to derive aerosol optical thickness from the measurements of the direct beam solar radiation at the definite wavelength without water vapor and ozone absorption using the Buger's law:

$$I = I_0 \exp(-\tau_{a,\lambda}/\mu_0)\exp(-\tau_{m,\lambda}/\mu_0)\,, \qquad (5)$$

where I_o is the solar direct beam irradiance at the top of the atmosphere, μ_o is the cosine of the solar zenith angle, $\tau_{a,\lambda}$ and $\tau_{m,\lambda}$ are the optical depths of the aerosol and molecular extinction at the wavelength λ. The coefficient α can be determined from the spectral solar irradiance measurements at different wavelengths according equation (4).

The measurements of the broad band direct beam solar radiation have been carried out by the Institute of Atmospheric Physics (IAP) and by the Meteorological Observatory of the Geographycal Department, Moscow State University(MO MSU) in different regions of the Earth. The methods were developed to obtain τ_{a,λ_0} from the integral (IR: from 0.3 to 4 μm) and visible (VIS: from 0.38 to 0.71 μm) solar irradiance using detailed calculations of molecular, aerosol, ozone and water vapor extinction (Tarasova & Yarkho, 1991; Abakumova *et al.*, 1992). The error of derivation due to unknown parameters α and the water vapor amount in the atmosphere column was evaluated.

Monthly mean values τ_{a,λ_0} have been obtained from long-term measurements of VIS radiation carried out during the period from 1971 to 1989 at MO MSU (Abakumova *et al.*, 1992). The mean aerosol optical depth for all this period is given in Table II. Other values τ_{a,λ_0} shown in Table II for α=1 were obtained from the integral radiation measurements made by IAP and MO MSU during radiation field experiment in April and May 1986 in Moscow region (Abakumova *et al.*, 1989), during large biomass burning in Moscow region in August 1972 (Feigelson *et al.*, 1986), during the ship expedition to Tropical Atlantic Ocean from February to October 1987 (Abakumova *et al.*, 1989) and from June to August 1979 (Krasnokutskaia *et al.*, 1985) and the ship expedition to Atlantic Ocean, Azores region from September to December 1991 (Tarasova *et al.*, 1992a).

The solar absorption by aerosols and cloudiness is mostly determined by the particle single scattering albedo defined as follows:

$$\omega_\lambda = \frac{\sigma_\lambda}{\varepsilon_\lambda}\,, \qquad (6)$$

where σ_λ and ε_λ are the scattering and the extinction coefficients for the aerosol or cloud particles at the wavelength λ. This parameter for the clear sky aerosol atmosphere can be obtained from the simultaneous measurements of solar direct beam and diffuse broad band irradiance (Tarasova *et al.*, 1992b). The method is valid for the hazy atmosphere ($\tau_{a,\lambda_0} > 0.3$) and for the following variations of some other atmosphere and surface parameter: averaged over solar spectrum surface albedo varying from 0.05 to 0.2, solar zenith angle from 0.5 to 0.8, wavelength exponent α from 0.5 to 2, water vapor amount from 0.5 to 4 g/cm^2.

The averaged over solar spectrum aerosol single scattering albedo ω_a given in Table II was retrieved from the long-term radiation measurements performed at MO MSU during summer from 1955 to 1991, during the dust aerosol radiation field experiment in Dushanbe region (Central Asia, Former Soviet Union) in September 1988 and during large biomass burning in Moscow region in August 1972.

The main parameter influencing the transfer of solar radiation through the cloud layer is the cloud optical thickness. It can be calculated in the climate models and AGCM using the following relation:

$$\tau_c = \frac{3}{2\rho}\frac{LWC\,dz}{r_e}\,, \qquad (7)$$

TABLE II

Aerosol optical thickness $\tau_{a,\lambda 0}$, aerosol single scattering albedo ω_a of the atmosphere column and the optical thickness τ_c of extended clouds derived from ground based and ship board broad band solar irradiance measurements carried out in different regions by Meteorology Observatory of Moscow State University (MO MSU) and by Institute of Atmospheric Physics (IAP).

$\tau_{a,\lambda 0}$	Time	Region	Type	Measurements
0.24	1971-1989	Moscow	Urban	MO MSU
0.21...0.25	1986, April, May	Moscow region	Rural	MO MSU
0.5...1.1	1972, August	Moscow	Forest smoke	MO MSU
0.1...0.3	1979, June-August	Tropical Atlantic Ocean	Ocean	IAP
0.49...0.43	1987, February April	Tropical Atlantic Ocean	Saharian dust	IAP
0.09...0.27	1991, Sept.-December	Atlantic Ocean, Azores region	Ocean	IAP
ω_a				
0.9...0.95	1955-1991, summer	Moscow	Urban	MO MSU
0.85...0.99	1972, August	Moscow	Forest smoke	MO MSU
0.9...0.95	1988, September	Dushanbe, Central Asia	Dust storm	IAP
τ_c				
40	1980-90, summer	Moscow	Stratocumulus	MO MSU
9...47	1992, October	Moscow region	Stratocumulus	MO MSU

TABLE III

Parameters a, b, c, d used in formulas (8)-(10) for three intervals of cloud optical thickness τ and for three ranges of solar spectrum: integral (IR), visible (VIS) and ultraviolet (UVR).

τ	spectrum	a	b	c	d	formula
5...20	IR	0.961	−0.145	−0.053	0.004	(8)
	VIS	0.97	−0.117	−0.050	0.005	(8)
	UVR	14.965	14.347			(9)
20...50	IR	5.138	−1.073	−0.856	−0.01	(10)
	VIS	4.412	−0.735	−0.791	0.0002	(10)
	UVR	4.569	−0.447	−0.738	−0.028	(10)
50...100	IR	9.477	−1.447	−1.043	0.024	(10)
	VIS	10.128	−1.903	−0.984	−0.007	(10)
	UVR	12.145	−1.074	−0.983	−0.023	(10)

where r_e (μm) is the effective cloud drop radius defined by relation (2), ρ (g cm^{-3}) is the density of water, LWC is liquid water content (1) and z is the height. It is needed to know the optical thickness for different cloud types to validate the cloudiness fields obtained by models.

It has been proposed to derive the cloud optical thickness from the comparison of the measured and calculated values of the ratio (C_Q) of global solar radiation under the cloudy and clear sky conditions (Tarasova & Chubarova,1994). This technique has been developed for the homogeneous middle and low extended cloudiness over the ocean or snow free surfaces consisted of liquid droplets. Simple formulas were obtained as a result of the least square fitting to the detailed radiation calculations of C_Q:

$$C_Q = (a + b\theta)\exp\{(c + d\cos\theta)\tau_c\}, \quad (8)$$

$$C_Q = \exp\{-\tau / (a(1 + \tau_c / b)^{0.5})\}, \quad (9)$$

$$C_Q = (a + b\theta)\tau^{c + d\cos\theta}. \quad (10)$$

The constant parameters a, b, c, d are given in Table III for three ranges of the solar spectrum: ultraviolet from 0.29 to 0.38 μm (UVR), visible from 0.38 to 0.71 μm (VIS) and integral from 0.3 to 4 μm (IR).

The formulas (8)-(10) can be utilized for the aerosol optical thickness varying from 0.15 to 0.6, aerosol single scattering albedo from 0.9 to 0.95, water vapor content in the atmosphere column from 0.69 to 3.4 g/cm^2, height of the cloud bottom from 1 to 3 km, ozone amount from 0.25 to 0.45 atm.cm., surface albedo from 0.03 to 0.23, the effective radius of cloud droplets from 4 to 12 μm. The error of the cloud optical thickness derivation due to unknown parameters mentioned above was evaluated as 30% when $\tau_c<5$, 15% when $5<\tau_c<100$ and 20% when $\tau_c>100$.

The values of the optical thickness of the extended stratocumulus clouds were retrieved from the long-term measurements of the integral hemispherical solar irradiance carried out at the MO MSU from 1980 to 1990 (Chubarova *et al.*, 1994). The average value over this time interval is given in Table II. The hour mean values of τ_c obtained during cloud radiation experiment in Moscow region during October 1992 (Tarasova & Chubarova 1994) are shown in Table II also.

3. CALCULATION TECHNIQUE

The radiation scheme for solar flux computations (Tarasova, 1992) is based on the vertically inhomogeneous Delta-Eddington code (Joseph *et al.*, 1976) which is the two-stream approach to the radiation transfer integro-differential equation. This equation can be written in the azimuthally averaged form as follows:

$$\mu\frac{dI}{d\tau} + I = \frac{\omega}{2}\int_{-1}^{1}\gamma(\mu,\mu')\,I(\tau,\mu')\,d\mu', \quad (11)$$

where I is azimuthally averaged intensity, $\gamma(\mu, \mu')$ is the scattering phase function, $\mu=\cos\theta$, θ is the scattering angle, τ is the optical thickness, ω is the single scattering albedo. The two-stream Eddington approach leads to a pair of the differential equations:

$$\frac{1}{2}\frac{dX}{d\tau} = -(1-\omega)\,Y + \frac{\omega}{2}S_0e^{-\frac{\tau_0-\tau}{\mu_0}} \quad (12)$$

$$\frac{2}{3}\frac{dY}{d\tau} = -(1-g\omega)\,X - g\,\omega S_0\mu_0\,e^{-\frac{\tau_0-\tau}{\mu_0}} \quad (13)$$

where $X = F^{\uparrow} - F^{\downarrow}$, $Y = F^{\uparrow} + F^{\downarrow}$ and

$F^{\uparrow} = 2\pi\int_0^1 I(\tau,\mu)\mu d\mu$, $F^{\downarrow} = 2\pi\int_{-1}^0 I(\tau,\mu)\mu d\mu$ are the upward and downward hemispherical fluxes, g is the asymmetry factor of the scattering phase function $g = \int_{-1}^{1}\gamma(\mu,\mu')\mu' d\mu'$. The Delta-Eddington approximation (Joseph *et al.*, 1976) is equivalent to the Eddington approximation with transformed parameters

$$\tau' = (1-\omega g^2)\tau,\ \omega' = (1-g^2)\omega/(1-\omega g^2),\ g' = g/(1+g).$$

The system of the differential equations (12), (13) can be solved analytically for a homogeneous layer and numerically for a set of adjacent homogeneous layers in the atmosphere using the condition of the flux continuity at the boundaries of the layers. The scattering and absorbing properties are constant within each layer.

This technique can be used successfully for calculating the particle scattering and absorption in the atmosphere, but it is difficult to utilize it for the computation of the gaseous absorption. The water vapor rotational/vibrational absorption spectrum consists of closely spaced line structures. The spectral line parameters for 159 000 lines have been compiled by Rothman (1981) and the first compilation has been done by McClatchey *et al.* (1973). Therefore, the line-by-line computations are very time-consuming. The measured broad band transmission function can be used to calculate the gaseous absorption.

There are two main techniques to include the water vapor absorption in the near infrared region of solar spectrum from 0.7 to 4 μm to the multiple scattering calculations. In the first case called the k-distribution or exponential-sum fitting (ESFT) method the broad band transmission function $T_{\Delta\lambda}(u)$ is expended into a series with respect to N exponents (Wiskombe & Evans, 1977):

$$T_{\Delta\lambda}(u) = \sum_{k=1}^{N} f(x_k) \exp(-x_k u), \tag{14}$$

where u is the water vapor amount at the radiation path (water vapor path), x_k is considered as a water vapor broad band absorption coefficients and can be used in equation (11) or (12), (13). So, these equations have to be solved N times taking into account particle and water vapor extiction coefficients. Weights $f(x_k)$ are used to obtain the resulted radiation fluxes.

In the second case called the photon free path method (Feigelson & Krasnokutskaya, 1978; Bakan & Quenzel, 1978) the solar radiation fluxes are computed as follows:

$$F_{\Delta\lambda} = \int_0^\infty J_{\Delta\lambda}(l)\, T_{\Delta\lambda}(w \cdot l)\, dl \tag{15}$$

where $J_{\Delta\lambda}(l)$ is the photon path distribution function which can be computed by the equation (11) written in nonstationary form, w is the water vapor amount, photon path length is $l = \varepsilon\, l'/\tau$, where ε is extinction coefficient, τ is optical thickness, l' is the geometrical photon path length.

The simple approximations concerning these methods can be used for the models to avoid the time consuming computations. The accuracy of these parameterizations is evaluated when compared to available reference calculations and observations only. In the mean photon path (MPP) approach (Veltishev *et al.*, 1990) the transmission function at the mean photon path $T_{\Delta\lambda}(w\bar{l})$ in scattering atmosphere is used. The mean photon path in scattering atmosphere is determined from the available calculations (Feigelson & Krasnokutskaya, 1978; Bakan & Quenzel, 1978). Therefore, equation (15) transforms to the stationary form (11) and can be solved in two-stream approach.

To take into account the temperature and pressure height dependence in the atmosphere the, empirical scaling of a water vapor amount is used:

$$w(z) = w_0(z)\left(\frac{p(z)}{p_0}\right)^n \left(\frac{T_0}{T(z)}\right)^m, \tag{16}$$

where p_0=1013.25 mb, T_0=273.15 K, n and m are constants determined empirically by different authors.

The radiative transfer computations in scattering atmosphere are performed for four spectral intervals: 0.2...0.7, 0.7...1.2, 1.2...2.5, 2.5...4 μm taking into account molecular scattering and aerosol and cloud scattering and absorption. Then, the results are summed with the weight of solar spectral irradiance at the top of the atmosphere (World Climate Research Programme 7, 1986) to get the visible and near infrared solar fluxes.

To perform the calculations of the solar irradiance the water vapor transmission function in the near infrared solar spectrum (Feigelson & Krasnokutskaya, 1978) is employed. It approximates the results of the solar absorptance measurements performed by Golubitsky and Moskalenko (1968):

$$T(u) = 0.0772 \exp(-58.3u) + 0.145 \exp(-1.45u) + 0.778 \exp(-0.019u), \tag{17}$$

where water vapor content u is less than 10 cm and

$$T(u) = 0.69 - 0.004u, \tag{19}$$

where u > 10 cm. Then, the MPP method is utilized. Therefore, the radiative fluxes taking into account the absorption by water vapor in the NIR solar spectrum are:

$$F_{NIR}^{wv}(z) = F_{NIR}(z)\, T(u\bar{l}), \tag{20}$$

where $F_{NIR}(z)$ is the solar NIR flux in scattering atmosphere without water vapor, u (g/cm^2) is the water vapor amount, $\bar{l}$ is the magnification of the water vapor path due to the multiple scattering determined from the calculations of the mean photon path in the aerosol and cloudy atmosphere (Feigelson & Krasnokutskaya, 1978; Bakan & Quenzel, 1978).

4. RADIATION SENSITIVITY STUDIES FOR THE AEROSOL OF LARGE BIOMASS BURNING

Biomass burning is a strong source of the aerosol emission into the troposphere (Kaufman *et al.*, 1992). Smoke aerosol particles absorb and scatter solar radiation changing the radiative balance at the atmosphere boundaries and heating rate inside the Earth's atmosphere during the daytime. Their impact on the thermal radiation is negligible due to the small size of smoke particles. Therefore, the increase of the aerosol loading of the atmosphere has to change the diurnal variation of the radiative balance at the ground and the daytime surface air temperature.

The Smoke, Clouds and Radiation – Brazil (SCAR-B) Project was carried out from 15 August through 20 September 1995 in Central Brazil and Southern Amazonas by National Aeronautics and Space Administration (NASA), USA, in collaboration with University of Vashington, USA, National Institute for Space Research (INPE), Brazil, University of Sao Paulo (USP), Brazil, and Special Agency of Brazil (AEB). The main goal of this project is to estimate the smoke effect on the atmosphere radiation balance and it's climate characteristics. The remote and in situ measurements of the microphysical, optical and chemical properties of the smoke produced by biomass burning have been done. The scientific results of this experiment will be published later. Sensitivity studies performed here can help to evaluate the needed accuracy of aerosol optical property derivation.

As example of the possible smoke influence on the radiation balance the calculations have been made to obtain the solar fluxes at the atmosphere boundaries and the heating rate inside the atmosphere for station Alta Floresta (10S, 56W) on 15 August 1995. Solar radiation scheme described in session 3 was employed. Four models of the smoke optical parameters shown in Table IV are proposed on the base of previous observations taken place during large biomass burning near Moscow, Russia, in August 1972 (Feigelson *et al.*, 1986) and Biomass Burning Airborne and Spaceborne Experiment in the Amazonas (BASE-A) in September 1989 (Kaufman *et al.*,1992).

In the first case, the aerosol optical properties were retrieved from the ground based integral solar irradiance measurements using the comparison of the measured and calculated values of the direct beam and diffuse radiation. The solar heating rate given in Table IV was calculated. In the second case, the aerosol optical thickness was derived from ground based spectral radiation observations and from aircraft measurements of the aerosol extinction coefficient. The single scattering albedo

TABLE IV

Smoke optical thickness $\tau_{a,\lambda 0}$ at the wavelength 550 nm, wavelength exponent α, single scattering albedo ω_a, effective radius of aerosol particles r_e, water vapor content in the atmosphere column u (g/cm^2) and average solar radiative heating rate inside the atmosphere $\frac{\partial T}{\partial t}$ (K/day) at solar elevation 44 degrees obtained by Feigelson *et al.* (1986) shown as case I and obtained by Kaufman *et al.* (1992) shown as case II.

Case	$\tau_{a,\lambda 0}$	α	ω_a	r_e (μm)	u (g/cm^2)	$\frac{\partial T}{\partial t}$ (K/day)
I	0.5...1.2 0.7...1.5	1 2	0.85...0.95			1...6
II	0.5...1.0	1.5...2.2	0.89...0.91	0.22...0.45	2...4	

was calculated by Mie scattering theory taking into account the amount of the graphitic carbon in aerosol particles. Therefore, four aerosol optical models given in Table V are proposed to evaluate the sensitivity of the solar radiation fluxes to aerosol optical thickness τ_{a,λ_0} of total atmosphere column at the wavelength λ_o= 0.55 μm and to single scattering albedo ω_a of aerosol particles averaged over the solar spectrum and smoke layer. The influence of other aerosol and atmosphere parameters on the solar radiative hemispherical fluxes at the Earth's surface is much weaker (Tarasova *et al.*, 1992b).

TABLE V
Four models of smoke optical parameters proposed for solar radiation sensitivity studies.

Model	τ_{a,λ_0}	ω_a
I	0.3	0.95
II	0.8	0.95
III	1.5	0.95
IV	1.5	0.85

The height distribution of the smoke aerosol was chosen due to the measurements of the vertical profiles of the aerosol extinction coefficient (Kaufman *et al.*, 1992). It is shown in Table VI for models II, III, VI. The height distribution used in model I is in accordance with the background aerosols (McClatchey *et al.*, 1972).

The asymmetry factor of aerosol phase function averaged over the solar spectrum was chosen to be equal g=0.65 (Tarasova *et al.*, 1992) and wavelength exponent α=1. The profile of the Tropical Atmosphere (McClatchey *et al.*, 1972) was employed for the calculations. Surface albedo

TABLE VI
The height distribution of the smoke aerosol optical thickness for the models II, III, IV.

Height (km)	Aerosol optical	thickness τ_{a,λ_0}
	Model II	Models III, IV
1.5...3.0	0.15	0.3
1.0...1.5	0.1	0.2
0...1.0	0.45	0.9

is 0.06 in the visible region of the solar spectrum and 0.22 in the near infrared (Briegleb *et al.*, 1986) valid for the tropical evergreen broad-leaved forest.

The upward and downward solar irradiance at the top of the atmosphere and at the surface calculated for different solar zenith angles are shown in Fig. 1 and Fig. 2, respectively. The difference of

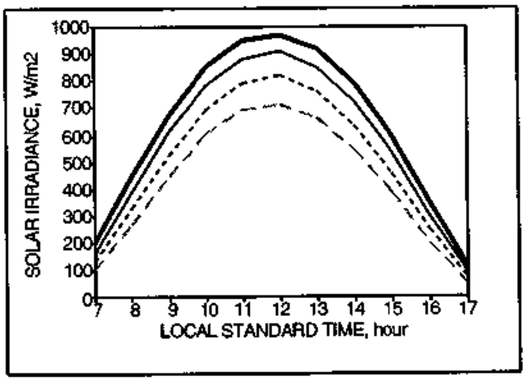

Fig. 1 — The downward solar total irradiance at the Earth's surface calculated using four smoke aerosol optical models from Table V and Table VI. Heavy solid curve is for model I, solid: model II, dotted: model III, dashed: model IV.

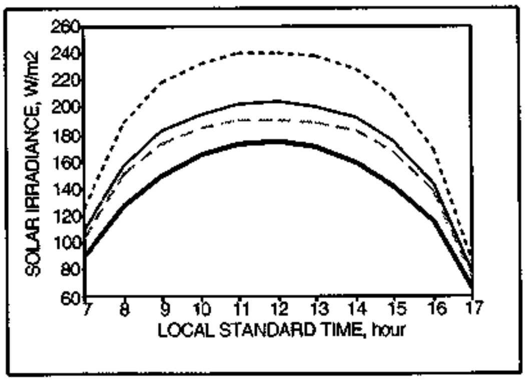

Fig. 2 — The upward solar total irradiance at the atmosphere top calculated using four smoke aerosol optical models from Table V and Table VI. Heavy solid curve is for model I, solid: model II, dotted: model III, dashed: model IV.

the downward solar hemispherical flux in W/m^2 at the ground computed for models II and I, III and I, IV and I is shown in Table VII and can be considered as the radiative forcing of smoke aerosols. The computed heating rate profile due to solar absorption is shown in Fig. 3 for solar zenith angle 75^0 and in Fig. 4 for solar zenith angle 25^0. The

TABLE VII
The difference of the downward solar irradiance in W/m^2 calculated using aerosol optical models I and II, I and III, I and IV from Table V and VI for different local time and cosine of solar zenith angle cos θ.

Local time	cos θ	II-I	III-I	IV-I
7	0.250	36	68	96
12	0.912	64	156	256
17	0.159	21	38	55

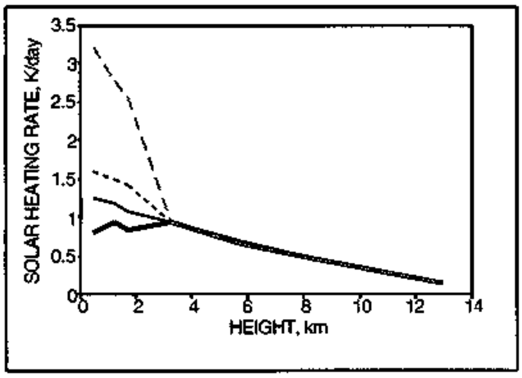

Fig. 3 — The vertical profile of the solar heating rate in the smoke aerosol atmosphere. Heavy solid curve is for model I, solid: model II, dotted: model III, dashed: model IV. Cosine of solar zenith angle is 75°.

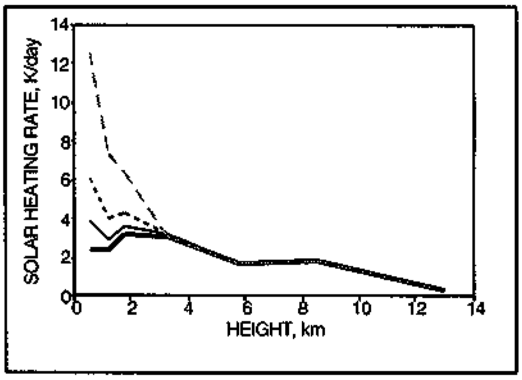

Fig. 4 — As in Fig. 3 but for cosine of solar zenith angle is 25°.

strong sensitivity of the solar irradiance to the smoke aerosol optical thickness and the single scattering albedo is demonstrated.

The observation data obtained during SCAR-B mission give us the opportunity to investigate in details the diurnal variation of the radiation balance due to the smoke loading of the atmosphere. In order to achieve this objective, the aerosol optical parameters such as the aerosol optical thickness and the single scattering albedo have to be retrieved from direct beam and diffuse solar irradiance measurements carried out at the SCAR-B surface network using sun/sky scanning spectral radiometers, shadowbands and hemispherical spectrometers. The calculations of the smoke aerosol impact on the solar irradiance performed here can help to evaluate the needed accuracy of the parameter derivation.

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