NUMERICAL MODEL OF THE SOLAR RADIATION FIELD IN THE AEROSOL ATMOSPHERE

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This paper represents some results of numerical calculations of the visible solar radiation field in the "Earth – aerosol atmosphere" system made using Monte Carlo method. The results obtained are intended for making comparative calculations and estimating the influence of individual parameters of the aerosol atmosphere on the radiation fields.

1. INTRODUCTION

Some results of numerical experiments on investigating radiation characteristics of the "Earth – aerosol atmosphere" system by the Monte Carlo method are described in this paper. This is not a detailed description of radiation fields. The calculational results are intended for illustrating the potentials of statistical simulation and a software package developed on the base of it in the Computing Center, Siberian Branch of the Russian Academy of Sciences (Novosibirsk). The solution of several typical problems of atmospheric optics was used as an example.

The aforementioned software package has been developed for solving a wide variety of atmospheric–optical problems from the simplest ones treated here to the complicated stochastic problems of radiative energetics of cloudy atmosphere, satellite–based optical sounding of the atmosphere and ocean, location, and vision. At the same time, the material presented here can be useful for both comparative calculations and quantitative estimate of the effect of individual optical parameters of atmospheric aerosol on radiation fields.

A great number of papers concerns with numerical and analytical solution of the problem on solar radiation transfer in a plane–parallel stratified atmosphere. A sufficiently detailed bibliography is given in, e.g., Refs. 1–3. The specific features of many of these papers are idealization of real optical parameters of the atmosphere and almost uncontrollable errors. The Monte Carlo method is free from these disadvantages. Therefore it is useful in making detailed calculations which could be a basis for estimating the accuracy of approximate methods and determining some regularities in spectral and angular structures of radiation field.

The visible radiation field ($0.4 \le \lambda \le 0.8$ μm) which is of great importance in atmospheric optics is considered in this paper. It should be noted here that to compare efficiencies of numerical and analytical methods in atmospheric–optical problems, a great number of calculations of fluxes and intensities of visible solar radiation for simplified models of the atmosphere have been made on suggestion of International Association of Meteorology and Atmospheric Physics (IAMAP). The results were published in Refs. 3 and 4. Similar work in which I also participated, was performed on initiative on All–Union Interdepartmental Seminar on Radiant Heat Exchange.[5] It should be noted that the methods of calculations are not described here. The algorithms of statistical simulations in the aforementioned problems of atmospheric optics are described at length in Refs. 6 and 7.

The detailed description of calculational results is very lengthy, so only some interesting, from my viewpoint, dependences are considered here. The results represented are a part of numerical radiative model of the Earth's atmosphere which was developed at Computing Center (Novosibirsk).

2. OPTICAL MODELS OF THE ATMOSPHERE

The initial and most important step in constructing a numerical model of atmospheric optical radiation field is a choice of optical model which is the most adequate to real conditions. A great variety of nature and spatiotemporal variations of aerosols have stimulated the development of a great amount of optical models of aerosol atmosphere. The models[8–12] are in most common use now therefore they were used in the present calculations.

It is also of interest now to develop models of aerosol atmosphere as evidenced by recent fundamental papers devoted to this problem.[13–17] Finer examination of the microphysical and chemical properties of aerosols as well as geographical and seasonal variations of atmospheric aerosols have been done in this papers.

The calculations under consideration were made at six wavelengths of the visible solar spectrum range: 0.4, 0.5, 0.55, 0.6, 0.7, and 0.8 μm. The vertical profiles of the coefficients of molecular scattering $\sigma_m(\lambda, z)$ and absorption $k_m(\lambda, z)$ borrowed from Ref. 8 were used here. The data of Refs. 10–12 were used as profiles of coefficients of aerosol scattering $\sigma_a(\lambda, z)$ and absorption $k_a(\lambda, z)$. In Ref. 10 an optical model of aerosol atmosphere closed with respect to all of the initial parameters is given. A three–layer model with the following vertical division: $0 \le z \le 5$ km, $5 < z \le 17$ km, $17 < z \le 50$ km is proposed for taking account of vertical variation of the aerosol scattering phase function $g_a(\lambda, z, \mu)$ (μ is the cosine of the scattering angle). The calculations based on the model[12] are given for the wavelengths $\lambda = 0.4$ and 0.55 μm to estimate sensitivity of the radiation field to variations of aerosol parameters.

In the ground layer ($0 \le z \le 2$ km) a model of urban aerosol, in troposphere ($2 < z \le 10$ km) a model of tropospheric aerosol, in stratosphere ($10 < z \le 30$ km) for pure atmosphere a model of stratospheric aerosol and for turbid one a model of volcanic aerosol, and in the upper stratosphere ($z > 30$ km) a model of meteoritic dust were used (see Ref. 12, Table VIII). The scattering phase functions for the corresponding models were calculated by the formulas of Mie theory for lognormal and modified gamma distributions with parameters represented in Table VI of Ref. 12. In calculations it was taken into account that in each of the aforementioned models the aerosol represents a composition of aerosols of different nature in the ratios given in Table V of Ref. 12. It is necessary here to specify the relations using which along with the data from Ref. 12 the optical parameters of the atmosphere were calculated.

Let aerosol consist of n fractions of different particles. The following designations are used: $M_i(z)$ is the particle number density (a number of particles per unit volume) of the ith sort at an altitude z ($i = \overline{1, n}$); $\beta_i(\lambda)$ and $\kappa_i(\lambda)$ are microscopic cross sections of radiation scattering and absorption, respectively, with polydispersed particles of the ith sort at the wavelength λ; $g_i(\lambda, \vartheta)$ are the normalized scattering phase functions of polydispersed particles of the ith sort; V_i is the portion of volume of ith particles in the total volume of particles in the mixture; and, $\overline{V}_i$ is the mean volume of ith particles. Then the volume scattering and absorption coefficients and the scattering phase functions of the aerosol mixture are found from the formulas

$$\sigma_a(\lambda, z) = \sum_{i=1}^{n} M_i(z)\, \beta_i(\lambda)\,; \tag{1}$$

$$k_a(\lambda, z) = \sum_{i=1}^{n} M_i(z)\, \kappa_i(\lambda)\,; \tag{2}$$

$$g_a(\lambda, z, \vartheta) = \sum_{i=1}^{n} M_i(z)\, \beta_i(\lambda)\, g_i(\lambda, \vartheta) \,/\, \sigma_a(\lambda, z)\,. \tag{3}$$

The procedure of calculating σ_a, κ_a, and g_a at a wavelength λ is as follows. The values $\beta_i(\lambda)$, $\kappa_i(\lambda)$, and $g_i(\lambda, \vartheta)$ entering into Eqs. (1)–(3) are calculated with respect to the data from Tables VI and VII of Ref. 12 using the formulas of Mie theory (see, e.g., Refs. 9, 18, and 19). Using the known values of $\sigma_a(0.55, z)$ at $\lambda = 0.55$ (see Table VIII) and the data from Table V of Ref. 12 the values M_i entering into Eqs. (1)–(3) are calculated

$$M_i = M_1\, M_{0,\,i}, \tag{4}$$

where $M_{0,\,i} = V_i\, \overline{V}_1 / (V_1\, \overline{V}_i)$;

and, $$M_1(z) = \frac{[\sigma_a(0.55, z) + k_a(0.55, z)]}{\sum\limits_{i=1}^{n} M_{0,\,i}\, [\beta_i(0.55, z) + k_i(0.55, z)]}\,.$$

Substitution of Eq. (4) and β_i, κ_i, and g_i into Eqs. (1), (2), and (3) yields σ_a, κ_a, and g_a. To assess the effect of strong variations in aerosol content in the atmosphere (due to volcanic eruptions) on the optical field structure, the additional calculations for five profiles of the aerosol extinction coefficient taken from Fig. 72 of Ref. 13 were made at $\lambda = 0.55$ µm. In this case, the stratospheric aerosol scattering phase functions were used. To estimate the effect of reflecting properties of the Earth's surface on the radiation field, the model of spectral albedo distribution based on the data from Fig. 29 of Ref. 11 for four types of surface: fresh show, vegetation, aluminium oxide, and water was used.

3. FLUXES AND INTENSITIES OF VISIBLE RADIATION IN AN AEROSOL ATMOSPHERE

Figures 1 and 2 depict sensitivity of vertical profiles of upwelling $F^+(z)$ and downwelling $F^-(z)$ radiation fluxes to variations in aerosol composition of the atmosphere. The results are given in units corresponding to unit power of the source ($\pi S_\lambda = 1$).

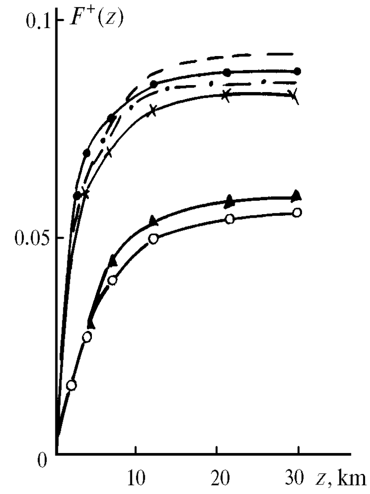

FIG. 1. Vertical profiles of upwelling fluxes of radiation at $\lambda = 0.55$ µm, $\vartheta_0 = 0$, and $A = 0$. Profiles $\Sigma(z)$ from Fig. 72 of Ref. 13: —o— background aerosol (1); —Δ— normal aerosol content (8); —×— and —·—·— aerosol content after volcanic eruption (2 and 3); and —•— and – – – – aerosol content after strong volcanic eruptions (5 and 6). Figures in brackets are the numbers of profiles $\Sigma(z)$ corresponding to those from Ref. 13.

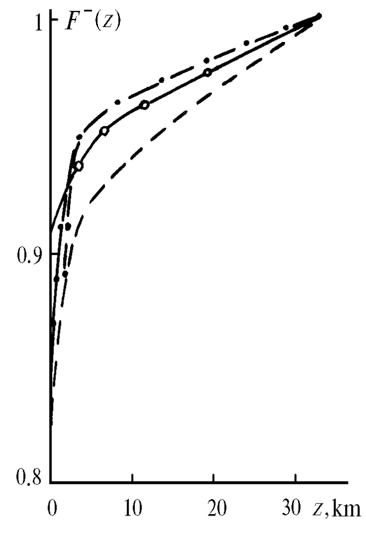

FIG. 2. Vertical profiles of downwelling fluxes. Designations are identical to those from Fig. 1.

The data from Fig. 72 of Ref. 13 are taken as vertical profiles of the aerosol extinction coefficients. The vertical profiles of fluxes are given for models corresponding to average annual conditions, normal content of aerosol, and a turbid atmosphere due to moderate and strong volcanic eruptions. The results show that the fluxes of upwelling radiation are more sensitive to variations in aerosol content in the atmosphere than those of downwelling radiation.

The increase of the aerosol extinction coefficient by one or two orders of magnitude at 15–20 km altitudes with constant profile of survival probability $q(z)$ results in the increase of upwelling fluxes by several tens per cent. The change of downwelling radiation fluxes comprises not more than ten per cent. The fluxes $F^+(z)$ increase sharply as a function of altitude in the ground layer and troposphere. At altitudes higher than 10 km this increase slows down.

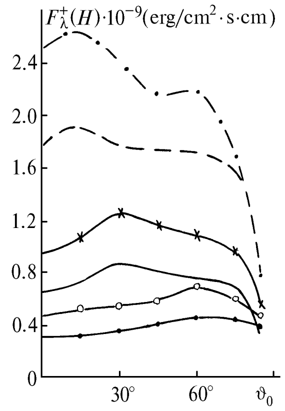

FIG. 3. A plot of radiation fluxes outgoing from the atmosphere vs the angle ϑ_0 for $A = 0$. An optical model is borrowed from Ref. 10.
Designations: —·— $-\lambda = 0.55$ μm; — — — — $\lambda = 0.5$ μm; —×— — $\lambda = 0.55$ μm; ━ — $\lambda = 0.6$ μm; —○— — $\lambda = 0.7$ μm; *and* —•— $-\lambda = 0.8$ μm.

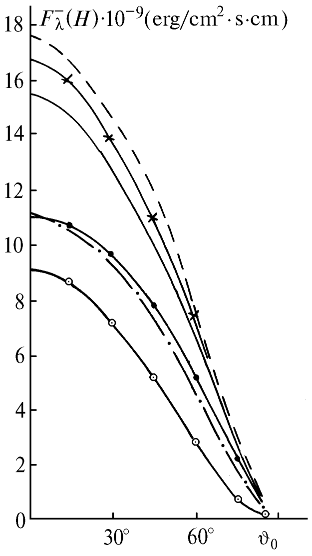

FIG. 4. A plot of a radiation flux coming to the Earth's surface vs the angle ϑ_0. Designations are identical to those from Fig. 3.

Depicted in Figs. 3–5 are spectral data for fluxes of upwelling $F^+_\lambda(H)$ (at the upper boundary of the atmosphere) and downwelling $F^-_\lambda(0)$ (near the Earth's surface) radiation depending on the angle ϑ_0 of the Sun declination (ϑ_0 is the angle between an external normal to the atmospheric layer and direction to the Sun) for three types of surface: fresh snow, aluminium oxide, and absolutely absorbing surface ($A = 0$). The results are given in erg/cm^2·s·cm. The model from Ref. 10 is used in the calculations. The results testify to strong spectral variability of fluxes. The upwelling fluxes $F^+_\lambda(H)$ for $A = 0$ depend monotonically on the wavelength λ, i.e., as the wavelength increases the fluxes decrease. Wavelength redistribution of upwelling fluxes in the case of reflecting surface is related to spectral variability of surface albedo. The wavelength dependence of downwelling fluxes is more complicated.

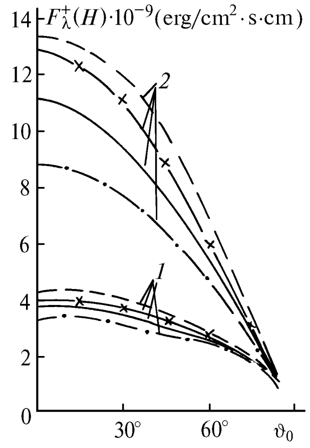

FIG. 5. The same as in Fig. 3 for a two types of surface: 1 – aluminium oxide and 2 – snow.

Figures 6–8 represents the radiant influxes (absorbed solar radiation) $\Pi_k = \pi S_\lambda \cos\vartheta_0 + F^+_\lambda(0) - F^+_\lambda(H) + F^-_\lambda(0)$, radiation budgets $B_\lambda(z) = F^-_\lambda(z) - F^+_\lambda(z)$, and illumination of the Earth's surface E_k *vs* the angle ϑ_0. The calculations were made using the model[12] at $\lambda = 0.4$ and 0.55 μm for four types of surface. The values $B_\lambda(z)$ are given for $z = 0.5$ km. The data show that the aforementioned characteristics strongly depend on the radiation wavelength, angle ϑ_0, and type of surface.

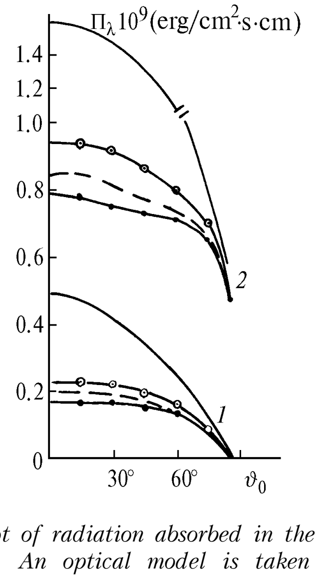

FIG. 6. A plot of radiation absorbed in the atmosphere vs the angle ϑ_0. An optical model is taken from Ref. 12. $\lambda = 0.4$ μm *(1) and* $\lambda = 0.55$ μm *(2);* — — *snow;* —○— — *vegetation;* — — — — *aluminium oxide; and,* —•— — $A = 0$.

A more fine characteristic of radiation field which strongly depend on optical parameters of the atmosphere is the angular structure of brightness of the atmosphere. This is attested by the results represented in Fig. 9. They were obtained at $\lambda = 0.55$ μm for $A = 0$ for the described models of the aerosol extinction coefficients.[13] The figure depicts the radiation intensities $I(z, \mu)$ when $\mu > 0$ for $z = H$, and when $\mu < 0$ for $z = 0$ (μ is the cosine of the latitude angle with respect to the $0z$ axis). Represented in Fig. 10 are the calculational results of brightness of the atmosphere $I(z, \mu, \varphi)$ at $\lambda = 0.4$ and 0.55 μm using the model[12] for two values of the angle ϑ_0, namely, 30 and 60° and two values of azimuth φ, namely, 0 and 180°. The results are given for two types of surface.

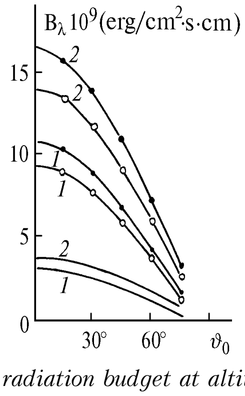

FIG. 7. A plot of radiation budget at altitude z = 0.5 km vs the angle ϑ_0. Designations are identical to those from Fig. 6.

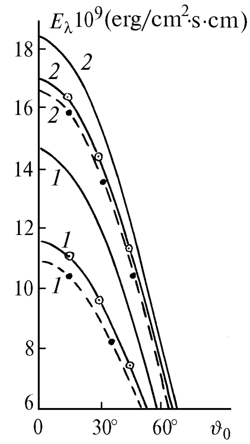

FIG. 8. A plot of illumination of the Earth's surface vs the angle ϑ. Designations are identical to those from Fig. 6.

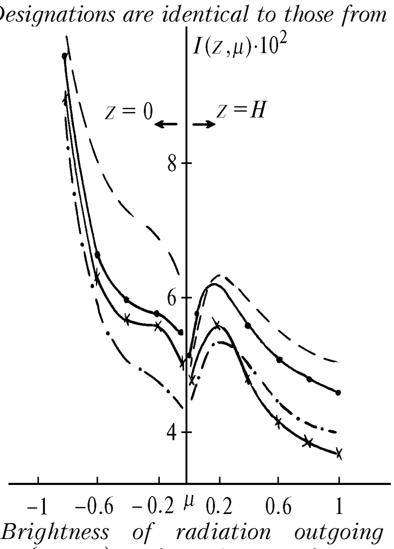

FIG. 9. Brightness of radiation outgoing from the atmosphere ($z = H$) and coming to the Earth's surface ($z = 0$) vs the value of μ. Designations are identical to those from Fig. 1.

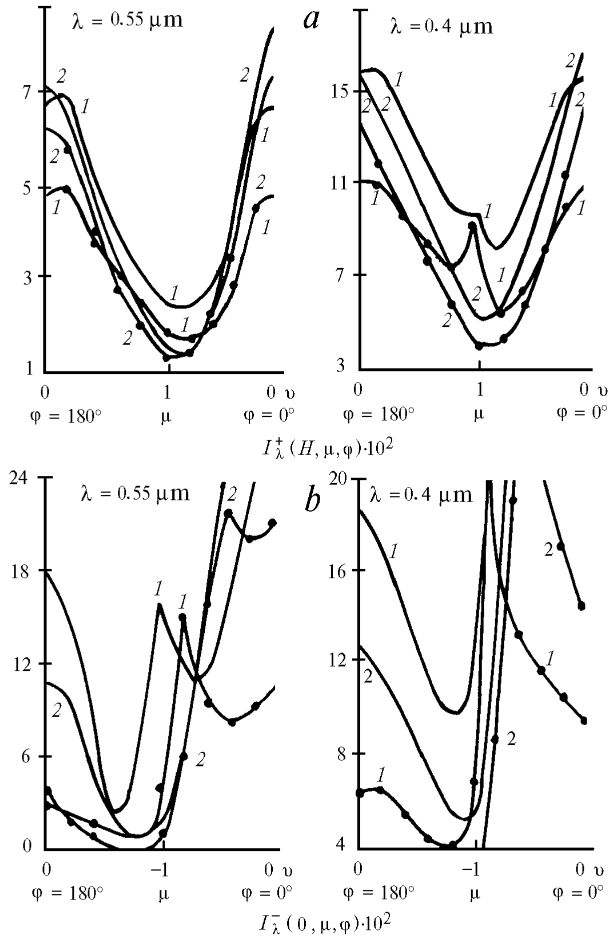

FIG. 10. Brightness of radiation outgoing from the atmosphere (a) and coming to the Earth's surface (b) vs μ and φ. The model is borrowed from Ref. 12. Curve 1 – ϑ = 30°; 2 – ϑ = 60°; a solid line – snow, and a dotted line – A = 0.

The calculational results described in this paper give an idea of variability of basic characteristics of the visible radiation field as a function of wavelength, optical properties of the atmosphere, and conditions of illumination and measurements.

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