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An empirical model of optical and radiative characteristics of the tropospheric aerosol over West Siberia in summer

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Abstract

An empirical model of vertical profiles of the aerosol optical characteristics is described. It has been constructed based on the data of multiyear airborne sensing of optical and microphysical characteristics of the tropospheric aerosol over West Siberia at the altitudes 0–5 km. The model proposed allows using season, air mass type, and time of the day as its input parameters. For solar radiative flux calculations, this empirical model has been tested for the mean conditions of the summer season. The available experimental database and the model developed on its basis are shown to be sufficient to perform these calculations taking into account the regional features of West Siberia.

1 Introduction

The atmospheric aerosol is a most variable constituent of the Earth's atmosphere. Radiative models rely on the data on its optical and microphysical properties, with the reliability of the latter determining the adequacy of the models (Haywood and Boucher, 2000; Takemura et al., 2002; Yu et al., 2006; Zhou et al., 2006; Ivlev, 2011).

Given the great diversity of sources, sinks and chemical compositions of the atmospheric aerosol species and strong spatiotemporal variations in their properties, it becomes clear that most reliable data on aerosol optical characteristics and their transformation in the course of geophysical processes could be so far obtained only experimentally. This is especially important in terms of calculations performed for particular regional conditions. In the clear-sky atmosphere, the accuracy of estimating the radiative effects depends on how correctly the aerosol optical depth, scattering phase function, and single scattering albedo are specified (Yu et al., 2006).

This paper describes an approach to simulating the upward and downward solar radiation fluxes using an empirical model of optical characteristics of the tropospheric aerosol over West Siberia for the summer period which we are currently developing.

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2 The approach to developing an empirical model

The measurement instrumentation and the calibration techniques used here were described in detail in earlier publications (Panchenko et al., 1996, 2000), we shall therefore restrict ourselves to a couple of procedural aspects only.

The first aspect consists in the necessity to parameterize the relationship between aerosol characteristics and relative air humidity, which strongly affects transformation of microphysical parameters of aerosol particles. The approach we have developed is based on the principle of separate investigation of variation in the aerosol particle dry matter and aerosol condensation activity under the external factors (Panchenko et al., 2004).

The present-day, real-atmosphere aerosol studies apply a very wide range of methods and facilities to investigate aerosol properties. We can single out two main approaches to estimating its optical characteristics (Panchenko et al., 2004). One of them is based on the data on microphysical and chemical composition of the species and involves a subsequent calculation of the required optical characteristics, and the other relies on the optical measurement data.

An advantage of the "microphysical" approach is that the model data on nearly all required aerosol optical parameters can be obtained in the spectral range of interest. The most serious disadvantage of this approach is that any restriction or distortion of the data on microphysical parameters or shape of the particles gives rise to unpredictable errors in retrieving the optical characteristics; therefore, all parameters estimated with this approach should be carefully validated against the optical measurement data.

The approach to describing the properties of atmospheric aerosol in terms of its optical characteristics in the real atmosphere is free from most shortcomings inherent in the microphysical simulations. The "optical" approach is considered to be disadvantageous in that the data obtained can generally be used only within the range of optical characteristics and spectral interval where the initial observation data were obtained, and its use outside these bounds needs additional investigation and validation.

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The second aspect of our approach is that an empirical model is constructed via combining the microphysical and optical approaches. The presented model of vertical profiles of optical characteristics is based on the data of airborne sensing of optical and microphysical characteristics of the tropospheric aerosol of West Siberia at the altitudes 0–5 km (Panchenko et al., 1998; Panchenko and Terpugova, 2002). It allows using season, air mass type, and time of the day as its input parameters.

The first ("optical") block of the empirical model involves the data of measurements with an airborne nephelometer, recording the aerosol dry matter optical characteristics and the condensation activity parameter. This allows us to retrieve the vertical profile of the "dry" aerosol scattering coefficient at the wavelengths 0.51 μm at the angle 45°, μ_0 (0.51 μm, 45°) (Panchenko and Terpugova, 2002). The second ("microphysical") block represents an empirical model based on the measurements of the vertical profile of the particle size distribution function with a photoelectric counter (Panchenko et al., 1998). In this case, the spectral optical characteristics are calculated in terms of the Mie theory. After that, the μ_0 (0.51 μm, 45°) value, calculated at each specific altitude, is compared to the data obtained with our model, and the corresponding correction coefficient is calculated.

3 The empirical model of vertical profiles of aerosol optical characteristics

The above-described generalized model was initially based on the data of airborne measurements performed within 1986–1988. Unfortunately, the airborne instrumentation in that period had no devices to measure the concentration of the light absorbing constituents (soot, or black carbon – BC).

Clearly, in order to perform radiation calculations from the empirical data we have to know the aerosol absorption characteristics. To this aim, regular monthly measurements of the aerosol scattering coefficients of dry particle matter and black carbon mass concentration M_{BC} were initiated in 1999 aboard an AN-30 Optic-E airborne laboratory (Panchenko et al., 2000; Kozlov et al., 2009). The black carbon measurements

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were performed along horizontal flight legs at the altitudes 0, 0.5, 1.0, 1.5, 2.0, 3.0, 4.0, 5.5, and 7.0 km.

3.1 Vertical profiles of aerosol scattering coefficient and black carbon concentration

Before introducing the absorber concentration data into the earlier developed procedure for retrieval of optical characteristics, we compared the seasonal profiles of the aerosol scattering coefficient for two observation periods: 1986–1988 and 1999–2007 (Fig. 1).

It is evident from Fig. 1 that its vertical profiles $\sigma_s(z)$ obtained in the period 1999–2007 agree well with the empirical model developed earlier from the dataset for 1986–1988. This implies that the data on the aerosol absorption properties (mass concentration M_{BC} and relative black carbon content in the submicron aerosol $P = M_{BC}/M_{aer}$, Fig. 2) can also describe the aerosol characteristics in the region under study quite adequately; they represent the main features of the "aerosol weather", and can be used to develop an empirical model.

Numerous comparisons of the calculation results with the near-ground measurement data and the measured spectral dependences of the aerosol optical depth (AOD, $\tau_a(\lambda)$) showed that, within the standard measurement error, the model provides for a retrieval of vertical profiles of the scattering coefficients and the angular dependence of the scattering matrix elements in the wavelength range 0.44–0.87 µm. This is quite consistent with the measurement capacity of the airborne instrumentation (nephelometer and photoelectric counter). Indeed, while adequately representing optical properties of the submicron particles (0.05–0.7 µm) (Panchenko et al., 2004), a nephelometer operating in the visible wavelength range gives no data on the particle size distribution function nor their optical contribution beyond this range, hence it is problematic to correctly reconstruct optical characteristics beyond the visible spectral range. On the other hand, a photoelectric counter measuring the particles in the radius range 0.3–5 µm offers a means to extend the range of estimation of optical characteristics

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towards longer wavelengths. However, measurement of particle concentrations closer to the upper size boundary (taking into account possible distortions in the airborne air intake) becomes increasingly less reliable than in the case where smaller particles are recorded.

At the same time, simulations of the solar radiative fluxes within 0.2 (0.3)–5 µm require data on spectral dependence of the aerosol optical characteristics outside the interval 0.44–0.87 µm. Some aspects of this problem are discussed in Sect. 4 of this paper.

3.2 The model of the particle size distribution function

The aerosol size distribution function at each altitude 0, 0.5, 1.0, 1.5, 2.0, 3.0, 4.0, and 5.0 km was modeled as a sum of two lognormal functions using an empirical microstructure model (Panchenko et al., 1998). Parameters of the fractions (median radius R_i, distribution half-width v_i, and volume concentration of every fraction V_i, $i = 1, 2$) were selected in such a way that the Mie angular scattering coefficient of the dry submicron fraction μ_0 (0.51 µm, 45°) would coincide with the mean seasonal value (Panchenko et al., 1996) at this altitude. (Aerosol particles were assumed to be non-absorbing, with the real part of the refractive index being $n_{aer} = 1.5$.)

Table 1 presents the scattering coefficients (Panchenko et al., 1996) and relative absorber (black carbon) mass concentrations P (Kozlov et al., 2009) in the aerosol dry matter, measured at these altitudes, as well as the distribution parameters R_i, v_i, and V_i adjusted for both particle fractions, $i = 1, 2$.

3.3 Estimation of the complex refractive index

Let us consider black carbon as an absorbing substance.

An estimation of the refractive index raises the question of the form in which black carbon exists in the atmosphere. The estimates of single scattering albedo (Λ) made earlier under different assumptions on the content of black carbon showed that the

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absolute value of albedo and its spectral dependence substantially vary, depending on whether black carbon exists as a separate fraction or is uniformly distributed over the volume of aerosol particles (internal mixture) (Terpugova et al., 2005). Moreover, irrespective of which (internal or external) mixture is hypothesized, the relative content of black carbon may also depend on the particle size.

The data on particular experiments dealing with the black carbon particle size distribution are available in the literature (Hitzenberger and Tohno, 2001; Kozlov et al., 2002]). Their results indicate that particles with the radius ~0.1–0.12 µm contain the largest amount of black carbon, and its relative content decreases as the particle size is increased. Based on the data of the cited works, we performed calculations using a hypothesis that 90 % of black carbon is in the submicron fraction and 10 % of black carbon is in the coarse fraction.

3.3.1 Dry aerosol

Black carbon in aerosol particles was assumed to be uniformly distributed (no size dependence) within each fraction, and the relative black carbon content was taken as a function of the median radius of the fraction. In addition, in order to calculate the relative black carbon mass concentration P from the airborne measurements use was made of the mass concentration of the submicron aerosol fraction M_{aer} obtained from the data of nephelometric measurements of the aerosol scattering coefficient. Therefore, in the refractive index calculations, the P values were taken with respect to the content of submicron aerosol particles.

In the first stage, the volume of black carbon was selected for its relative mass concentration at each altitude to coincide with the experimental mean seasonal values (Terpugova et al., 2005). The volume concentration of black carbon in every i-th fraction ($V_{\text{BC},i}$) was calculated using the formula

$$V_{\text{BC},i} = P \times C_i \times V_1 \times \rho_{\text{aer}} / \rho_{\text{BC}}, \quad i = 1, 2, \tag{1}$$

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where C_i is the black carbon fraction in the i-th aerosol fraction ($C_1 = 0.9$, $C_2 = 0.1$, in accordance with the hypothesis accepted), V_1 is the volume concentration of the submicron fraction, ρ_{BC} and ρ_{aer} are the densities of black carbon and non-absorbing aerosol, respectively (see notations in Table 1). The refractive index of the mixture in each fraction ($n_i - \chi_i \times i$, $i = 1, 2$) was determined according to the internal mixture rule (as a sum of the refractive indices of black carbon and non-absorbing aerosol, taken with the weights corresponding to the relative volume contents of black carbon at each altitude):

$$n_i = \left(n_{aer} \times V_{aer,i} + n_{BC} \times V_{BC,i}\right) / \left(V_{aer,i} + V_{BC,i}\right), \quad (2)$$

$$\chi_i = \chi_{BC} \times V_{BC,i} / \left(V_{aer,i} + V_{BC,i}\right), \quad (3)$$

where $V_{aer,i}$ is the volume concentration of the non-absorbing aerosol in the i-th fraction.

The calculations using Eqs. (1)–(3) were performed assuming the densities of black carbon and non-absorbing aerosol to be $\rho_{BC} = 1.85\,g\,cm^{-3}$ and $\rho_{aer} = 1.5\,g\,cm^{-3}$, respectively, and the complex refractive index of black carbon was set to be $1.8–0.74\,i$ (Gelencser, 2004).

3.3.2 Wet aerosol

In the next stage, we re-calculated the refractive index of dry aerosol in accordance with the mean seasonal relative humidities RH at different altitudes that were obtained during airborne sensing (Panchenko et al., 1996).

The dependences of aerosol parameters (radius, volume concentration, and scattering coefficient) on relative humidity were approximated by the Hanel formula (Hanel, 1976). In particular, for the angular scattering coefficient it is given by

$$\mu_{wet} = \mu_0 \times (1 - RH(z))^{-\gamma(z)}, \quad (4)$$

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where μ_0 is the scattering coefficient for zero relative humidity, $RH(z)$ is the mean seasonal relative humidity at altitude z, $\gamma(z)$ is the mean seasonal parameter of the condensation activity at the respective altitude.

The γ values at different altitudes, except for those at the ground level ($z = 0$), are some of the input parameters of the empirical model; they were obtained in the period of airborne measurements (Panchenko et al., 1996). The mean seasonal data of multiyear measurements in Tomsk were taken as the near-ground γ value (Panchenko et al., 2005). In the summer period, the condensation activity parameter was $\gamma = 0.3$ at $z = 0$, and $\gamma = 0.5$ at all the other altitudes.

The angular scattering coefficient μ_0 (0.51 µm, 45°) was the main characteristic measured in the airborne experiments. For the submicron aerosol fraction, this characteristic is linearly related to the particle volume concentration; therefore, for the wet aerosol the following formula is valid:

$$V_{i,\mathrm{wet}} = V_i \times (1 - \mathrm{RH}(z))^{-\gamma(z)}, \quad i = 1, 2, \tag{5}$$

where V_i and $V_{i,\mathrm{wet}}$ are the volume concentrations of the respective "dry" and "wet" i-th fractions.

The refractive index $n_{i,\mathrm{wet}} - \chi_{i,\mathrm{wet}} \times i$, $i = 1, 2$, was determined under the assumption that during wetting the volume of the fraction increases due to water uptake only:

$$n_{i,\mathrm{wet}} = (n_i V_i + n_\mathrm{w} V_\mathrm{w}) / V_{i,\mathrm{wet}}, \tag{6}$$

$$\chi_{i,\mathrm{wet}} = \chi_i V_i / V_{i,\mathrm{wet}}. \tag{7}$$

Here, $n_\mathrm{w} = 1.33$ is the real part of the refractive index of the liquid water and $V_\mathrm{w} = V_{i,\mathrm{wet}} - V_i$ is the condensed water volume, i.e. an increment in the volume of aerosol particles.

In selecting a method how to account for the effect of the relative air humidity on the aerosol microphysical characteristics, we considered the following two cases:

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1. Only particles in submicron fraction are wetted.

2. Both fractions are wetted.

The calculated values of the complex refractive index for the two fractions at different altitudes are presented in Table 2.

The Mie calculations of the optical characteristics and their comparison with the data of in situ measurements of the spectral and angular characteristics of the scattered radiation showed that the first case (wetting for the wide lognormal distribution of the submicron fraction only) shows a better agreement with the measured characteristics. It is for this reason that we used it to calculate the optical characteristics.

3.4 Vertical profiles of aerosol optical characteristics

This section presents calculations of the spectral dependence of the aerosol optical depth $\tau_a(\lambda)$ and vertical profiles of the single scattering albedo and the asymmetry factor of the scattering phase function in the altitude range 0–5 km (Fig. 3).

The normalized spectral behavior of AOD $\tau_a(\lambda)/\tau_a$ ($\lambda = 0.55$ µm) in the wavelength range 0.37–1.02 µm and the Ångström exponent α (0.44–0.87 µm) are in a good agreement with the data of multiyear sun-photometer measurements on the territory of Siberia (Sakerin et al., 2007) (Fig. 3a). Moreover, the α (0.44–0.87 µm) value is close to the Ångström exponent in the OPAC model of continental aerosol (continental average, RH = 70 %, Hess et al., 1998) and is equal to ~1.4.

The aerosol single scattering albedo $\Lambda(\lambda, z)$ varies quite substantially with altitude: irrespective of the wavelength, the maximum values $\Lambda(\lambda, z) \sim 0.92$–$0.93$ are observed in the mixing layer 0.5–1.5 km; outside this layer they decrease towards the Earth's surface and with increase in altitude (Fig. 3b). At the same time, the altitude variations in the asymmetry parameter of the scattering phase function AF(λ, z) are much smaller, not exceeding ~0.02 in the entire wavelength range 0.44-0.87 µm (Fig. 3c). Note for comparison that the data of multiyear airborne measurements over Oklahoma in the summer period indicate that single scattering albedo and asymmetry factor in the

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visible wavelength range vary but slightly as the altitude changes within ~0.5–4.5 km (Andrews et al., 2011).

4 Radiation calculations

The broadband solar radiation fluxes in the molecular-aerosol, plane-parallel atmosphere were calculated using the Monte Carlo algorithms developed earlier (Zhuravleva et al., 2009). Molecular absorption was accounted for by dividing the wavelength interval 0.2–5.0 μm into 31 bands, with the transmission function in each of them approximated by a finite exponential series (k-distribution method). Molecular absorption coefficients were calculated taking into account the real profiles of pressure, temperature, and water vapor concentration from the HITRAN2008 database and MT_CKD v.2.4 continuum model (http://rtweb.aer.com/continuum_frame.html). The comparisons showed that the numerical simulation results are in a satisfactory agreement with the results of line-by-line calculations and the data of field measurements (Zhuravleva and Firsov, 2004; Tvorogov et al., 2008; Chesnokova et al., 2009).

For the computations presented below, the transmission function was calculated using a regional model of temperature, pressure, and water vapor profiles (Komarov and Lomakina, 2008), taking into account absorption by all atmospheric gases, which was presented in the AFGL meteorological model (Anderson et al., 1986). The total moisture column content was assumed to be $W = 1.9\,\mathrm{g\,cm^{-2}}$, the total ozone content was equal to 336 DU according to the data of the TOMS satellite instrumentation (ftp://toms.gsfc.nasa.gov), and the total CO_2 content in the atmosphere was 380 ppm. (These values correspond to the summer averages over the territory of West Siberia; Sakerin and Kabanov, 2007; Sakerin et al., 2009; Arshinov et al., 2009.). The data of Fontenla et al. (1999) were used to account for the spectral dependence of the solar constant.

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The incident radiation was reflected from the underlying surface according to the Lambert law. The surface albedo was specified using the MODIS satellite measurement data (Moody et al., 2005) for the region of Tomsk.

4.1 The influence of spectral variations in aerosol optical characteristics

Calculations of the upward ($F^{\uparrow}(z)$) and downward ($F^{\downarrow}(z)$) solar radiation fluxes require data on spectral dependence of the extinction $\sigma(\lambda, z)$ and scattering $\sigma_s(\lambda, z)$ coefficients, as well as on spectral dependence of the scattering phase functions $g(\lambda, \theta, z)$ (θ is the scattering angle) within the entire wavelength range 0.2–5.0 µm considered here. As noted above, our empirical aerosol model provides a correct description of the aerosol extinction and $\Lambda(\lambda, z)$ and $\mathrm{AF}(\lambda, z)$ in the wavelength interval 0.44–0.87 µm. It is therefore relevant to ask how critical it is for the solar radiative flux simulations to take into account the spectral dependence of the aerosol optical characteristics beyond this range.

To answer this question, we performed additional numerical experiments using the OPAC model of continental aerosol (continental average, RH = 70 %; Hess et al., 1998). This model contains all input aerosol parameters necessary for the numerical simulation, both in terms of including the wavelength dependence in the range 0.25–5 µm and the vertical profiles of the optical characteristics within the interval of altitudes 0–35 km.

As a *basic variant*, we took the calculation performed assuming the aerosol extinction coefficients $\sigma(\lambda, z)$ to be constant outside the wavelength interval 0.37–1.02 µm: $\sigma(\lambda < 0.37\,\mu\mathrm{m}, z) = \sigma(\lambda = 0.37\,\mu\mathrm{m}, z)$, $\sigma(\lambda > 1.02\,\mu\mathrm{m}, z) = \sigma(\lambda = 1.02\,\mu\mathrm{m}, z)$, while $g(\lambda, \theta, z)$ and $\Lambda(\lambda, z)$ exactly corresponded to the model data (Fig. 4a and b). The spectral AOD dependence was modified for the following reasons. Based on the data of multiyear ground-based observations on the territory of Siberia, it was shown that AOD hardly varies in the interval $\lambda > 1$ µm; therefore, we can take that $\tau_a(\lambda > 1\,\mu\mathrm{m}) \sim \tau_a(\lambda \sim 1\,\mu\mathrm{m})$ (Sakerin and Kabanov, 2007).

The calculations were performed within the time interval $T_1 \le t \le T_2$, where T_1 and T_2 corresponded to the sunrise and sunset times on July 15 for the latitude of Tomsk

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(56° N), assuming that τ_a (0.55 μm) = 0.262. The simulation results showed that at $z = 0$ and the solar zenith angles SZA ~ 35–60°, the nonscattered radiation fluxes $S(z = 0)$, calculated without the $\tau_a(\lambda)$ modification, were overestimated by ~8–10 W m^{-2}, while the diffuse radiation fluxes $F_s^{\downarrow}(z = 0)$ were underestimated by ~4–6 W m^{-2}. Importantly, the main contribution to the radiative flux discrepancies comes precisely from the region $\lambda > 1$ μm.

The sensitivity of radiative fluxes to the spectral dependence of the scattering phase function and single scattering albedo was estimated by considering the following cases (Fig. 4a and b):

- *Variant 1* (the single scattering albedo is assumed to be constant outside the interval 0.44–0.87 μm): $\Lambda(\lambda < 0.44\ \mu m,\ z) = \Lambda(\lambda = 0.44\ \mu m,\ z)$, $\Lambda(\lambda > 0.87\ \mu m, z) = \Lambda(\lambda = 0.87\ \mu m, z)$;
- *Variant 2* (the scattering phase functions are assumed to be constant outside the interval 0.44–0.87 μm): $g(\lambda < 0.44\ \mu m,\ \theta,\ z) = g(\lambda = 0.44\ \mu m,\ \theta,\ z)$, $g(\lambda > 0.87\ \mu m, \theta, z) = g(\lambda = 0.87\ \mu m, \theta, z)$;
- *Variant 3*: the scattering phase functions and the single scattering albedo are assumed to be constant outside the interval 0.44–0.87 μm and equal to the respective boundary values (see variants 1 and 2).

The results of numerical experiments indicate that when the spectral variations in the scattering phase functions and single scattering albedos are neglected, the biases are no more than 2 W m^{-2} and 1 W m^{-2} in the calculations of instantaneous fluxes of diffuse downward ($z = 0$) and upward radiation ($z = 100$ km) (Fig. 4c). This small discrepancy in the radiative fluxes is most probably due to significant variation in $\Lambda(\lambda, z)$ and AF(λ, z) taking place for $\lambda > 1.25$ μm in the spectral bands of moderate to strong water vapor absorption, as well as in the spectral intervals whose contribution to the broadband radiative flux is small because of low values of the solar radiation incident on the top of the atmosphere. Evidently, in the situations where the spectral behavior

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of the aerosol optical characteristics is more pronounced, especially in the interval $\lambda < 1.25\,\mu m$, a neglect of the spectral dependences of $\Lambda(\lambda, z)$ and $AF(\lambda, z)$ may lead to more substantial errors.

4.2 Testing of the empirical aerosol model for calculating the solar radiative fluxes

To obtain a complete set of the input parameters for the simulation, the empirical model was complemented as follows:

- the optical characteristics in the altitude interval 5–35 km were taken from the OPAC model of continental average RH = 70 % (Sect. 3.4);
- the extinction coefficients were extrapolated to a wider range 0.37–1.02 µm using the Ångström exponent α (0.44–0.87 µm), and outside the interval 0.37–1.02 µm the extinction coefficients were assumed to be constant and equal to the respective boundary values for $\lambda = 0.37\,\mu m$ and $\lambda = 1.02\,\mu m$;
- the scattering phase functions and single scattering albedos of aerosol outside the interval 0.44–0.87 µm were assumed to be constant and equal to the respective boundary values for $\lambda = 0.44\,\mu m$ and $\lambda = 0.87\,\mu m$.

We used this approach because (i) the Ångström exponents α (0.37–1.02 µm) according to the data of airborne and ground-based measurements have close values and (ii) the neglect of the spectral dependences of the scattering phase function and single scattering albedo of aerosol does not significantly affect the solar radiative fluxes, at least at the atmospheric boundaries (Sect. 4.1).

Simulations of the instantaneous solar radiation fluxes using our empirical model of vertical profiles of optical characteristics for the summer season are illustrated in Fig. 5.

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5 Summary

The empirical model of the vertical profiles of the angular scattering coefficients and disperse composition of the atmospheric aerosol, which was created earlier from the multiyear data of airborne studies, has been modified to include the multiyear measurements of the content of absorbing particles, making it possible to estimate the imaginary part of the refractive index. This empirical model allows the vertical profiles of the aerosol optical characteristics to be retrieved as a function of the season, air mass type, and time of the day.

The test results have shown that the available empirical database and the model developed on its basis are sufficient to calculate the solar radiative fluxes taking into account the regional features of West Siberia. The data on the vertical profiles of aerosol optical characteristics would ensure a more exact estimation of the radiative characteristics (in particular, solar radiation cooling rates and absorption) within the troposphere.

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Table 1. Parameters of the lognormal distributions (median radius R, distribution half-width v, and volume concentration of fraction V) for the submicron and coarsely dispersed fractions, angular scattering coefficients for the dry submicron fraction μ_0 (0.51 µm, 45°), and relative mass concentrations of the absorbing substance P at different altitudes z.

z,	P	μ_0 (0.51 µm, 45°),	Submicron fraction ($i = 1$)			Coarsely dispersed fraction ($i = 2$)		
km		$km^{-1}\,sr^{-1}$	$\ln v_1$	R_1, µm	V_1	$\ln v_2$	R_2, µm	V_2
0	0.05	9.767E-3	0.8	0.114	1.548E-11	0.65	2.078	1.145E-11
0.5	0.03	7.067E-3	0.8	0.096	1.4138E-11	0.65	1.8035	1.573E-11
1	0.03	6.178E-3	0.8	0.098	1.236E-11	0.65	1.797	1.321E-11
1.5	0.03	5.089E-3	0.8	0.098	1.018E-11	0.65	1.7905	1.044E-11
2	0.04	3.478E-3	0.8	0.096	6.956E-12	0.65	1.784	6.841E-12
3	0.04	2.078E-3	0.8	0.095	4.156E-12	0.65	1.771	3.738E-12
4	0.05	6.067E-4	0.8	0.092	1.213E-12	0.65	1.758	9.914E-13
5	0.05	3.011E-4	0.8	0.089	6.022E-13	0.65	1.745	4.435E-13

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Table 2. Refractive indices for the submicron ($i = 1$) and coarsely dispersed ($i = 2$) fractions for different cases of dry aerosol wetting.

z, km	Only submicron fraction is wetted				Both fractions are wetted			
	n_1	n_2	χ_1	χ_2	n_1	n_2	χ_1	χ_2
0	1.4486	1.5016	0.0171	0.0040	1.4486	1.4428	0.0171	0.0026
0.5	1.4266	1.5007	0.0087	0.0016	1.4266	1.4235	0.0087	0.0009
1	1.4182	1.5007	0.0079	0.0017	1.4182	1.4153	0.0079	0.0008
1.5	1.4200	1.5007	0.0081	0.0018	1.4200	1.4170	0.0081	0.0009
2	1.4261	1.5010	0.0113	0.0024	1.4261	1.4221	0.0113	0.0013
3	1.4294	1.5011	0.0117	0.0027	1.4294	1.4253	0.0117	0.0015
4	1.4413	1.5015	0.0161	0.0037	1.4413	1.4357	0.0161	0.0023
5	1.4538	1.5016	0.0179	0.0041	1.4538	1.4477	0.0179	0.0028

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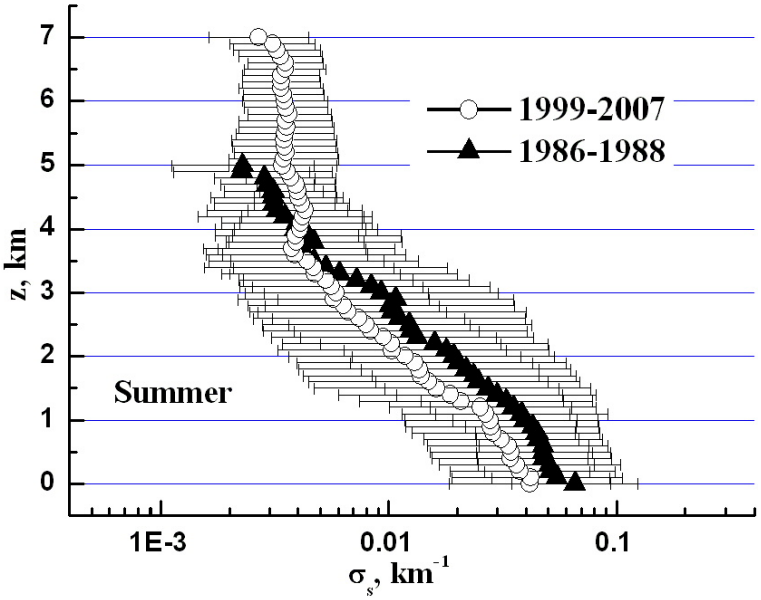

Fig. 1. Comparison of the average altitude profiles of the aerosol scattering coefficient according to data of airborne measurements in 1986–1988 and 1999–2007 (West Siberia).

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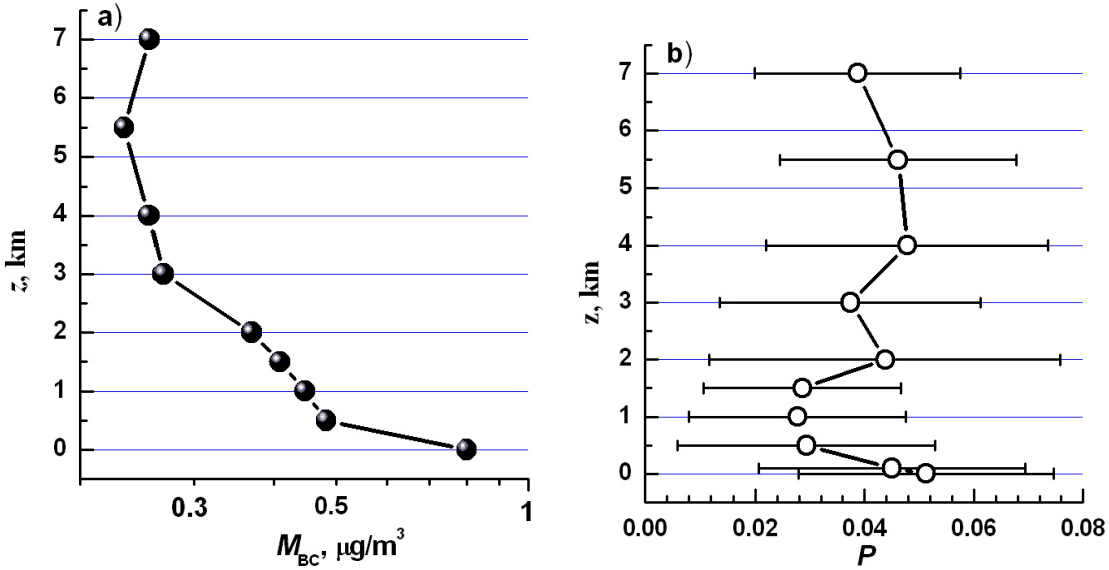

Fig. 2. Average altitude profiles of mass concentration **(a)** M_{BC} and relative content of black carbon in submicron aerosol P **(b)**. Results are given for West Siberia in the summertime of 1999–2007.

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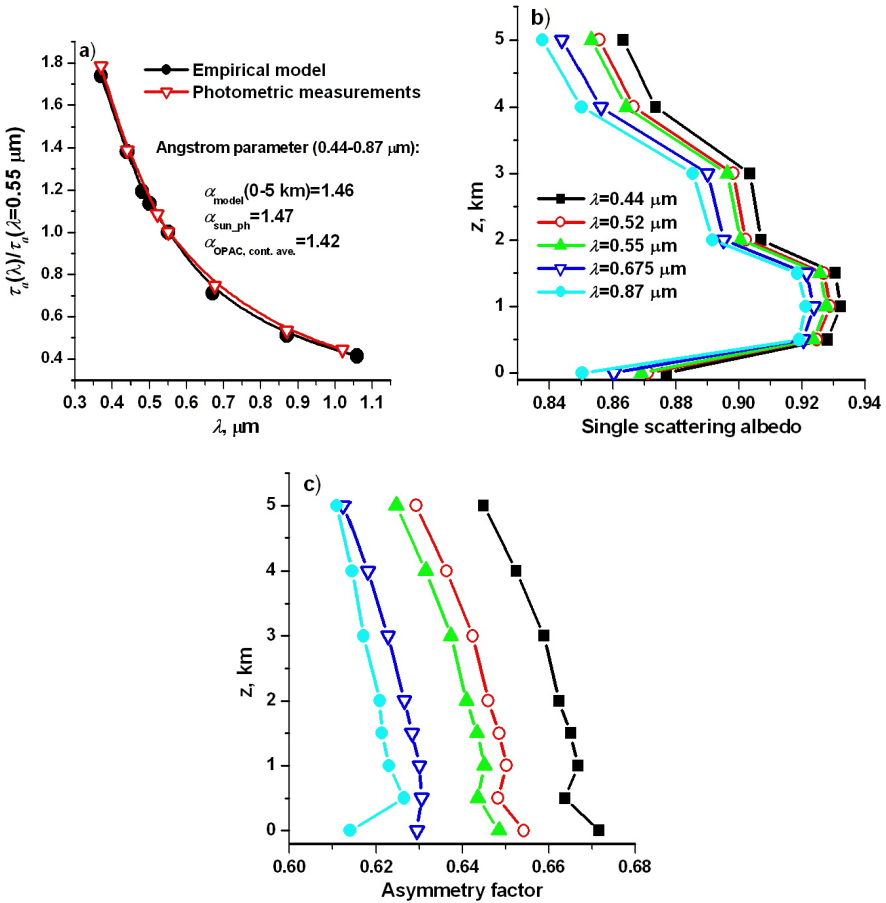

Fig. 3. Spectral behavior of the normalized AOD according to data of multiyear airborne and ground-based photometric measurements **(a)** and vertical profiles of single scattering albedo **(b)** and asymmetry factor **(c)** in the altitude range 0–5 km according to data of airborne measurements in the summer period.

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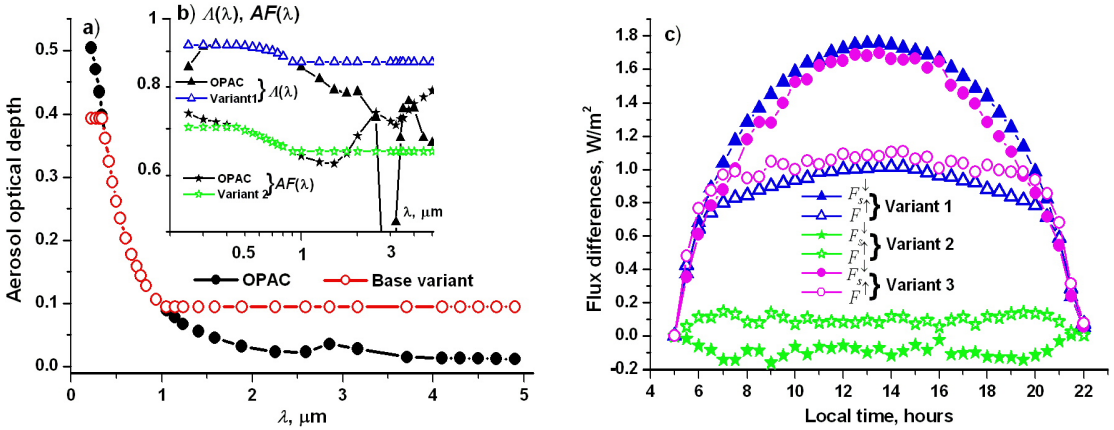

Fig. 4. Spectral dependence of aerosol optical depth **(a)** and single scattering albedo and asymmetry factor **(b)** in the OPAC model (continental average RH = 70%) in the basic variant and in variants 1 and 2, and the flux difference between the basic variant and variants 1–3 **(c)**; τ_a (0.55 μm) = 0.244.

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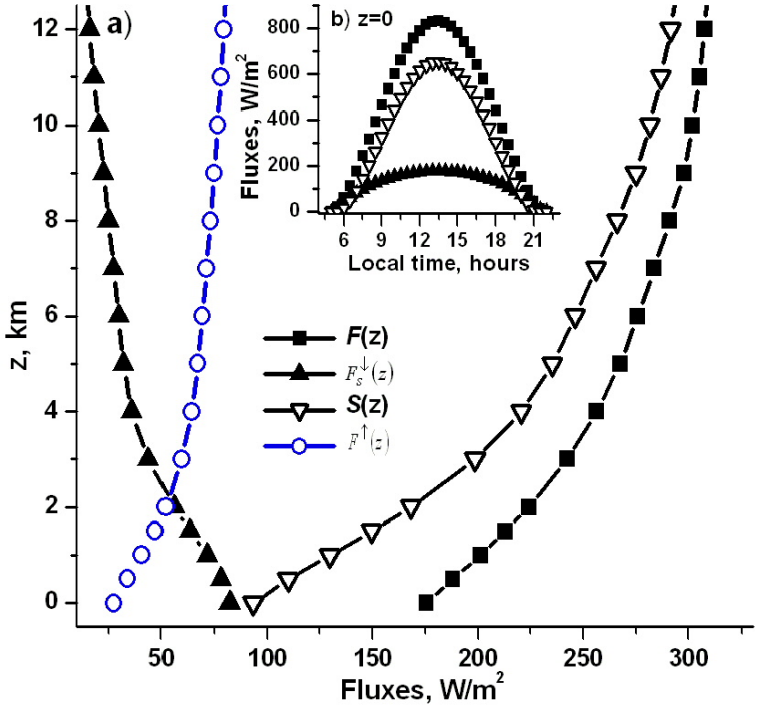

Fig. 5. Vertical profiles of the solar radiative fluxes calculated using our empirical model at 07:00 LT on 15 July (SZA = 75.618°) **(a)** and time variations in the downward radiation fluxes at the surface level **(b)**.

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