Typical and Anomaly Spectral Behavior of Aerosol Optical Thickness of the Atmosphere in Western Siberia

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Introduction

Despite the variety of dependencies of aerosol optical thickness (AOT) in the atmosphere, its general peculiarity is its gradual decrease as the wavelength increases (Angstrom formula): $\tau^A(\lambda) = \beta \cdot \lambda^{-\alpha}$. Anomalous situations are rare when the dependence $\tau^A(\lambda)$ in the range 0.35 μm to 1 μm has a quasi-neutral shape with one or more extremes (Rodionov 1970; Barteneva and Nikitinskaya 1991; Krauklis et al. 1990). In this paper, we analyze the typical dependencies $\tau^A(\lambda)$ in Western Siberia (Tomsk) and one example of anomalous transparency of the atmosphere observed at the intrusion of Arctic air. Observations were carried out in the wavelength range 0.35 ÷ 4 μm, but the principal portion of the data were obtained in the range up to 1.06 μm. The mean (hourly and daily) values were used for the analysis, so the random error does not exceed ~0.005, and the systematic error is estimated as 0,005 ÷ 0,01.

Typical Dependencies $\tau^A(\lambda)$

The results obtained have confirmed that the spectral dependence of AOT up to ~1 μm in the majority of events is well described by the Angstrom formula and has a neutral shape (Figure 1). That means, the parameters α and β allow us to quantitatively characterize the change of the spectral behavior of AOT (Table 1). The parameter β, close to the value τ^A(1.06 μm), is approximately representative of the content of coarse particles. The parameter α is related to the value of the relative content of fine fraction in comparison with the coarse one. The enhanced values β and low α observed in 1992 can be explained by the additional effect of volcanic aerosol (Mt. Pinatubo eruption). If the postvolcanic 1992 is removed, the mean value β varies within the limits 0.04-0.1, and the maximum values reach ~0.2.

The total range of the parameter α variability is -0.27 ÷ 2.53, and the mean values α changed from 0.4 (winter 1992) to 1.43 (spring 1993). This fact is evidence of the seasonal differences in the content of the particles of microdisperse fraction. The mean summer values α (without 1992) are in the limits 1.05-1.43, which is in good agreement with the results of investigations in other mid-latitude continental regions (Barteneva and Nikitinskaya 1991; King et al. 1980; Smirnov et al. 1994).

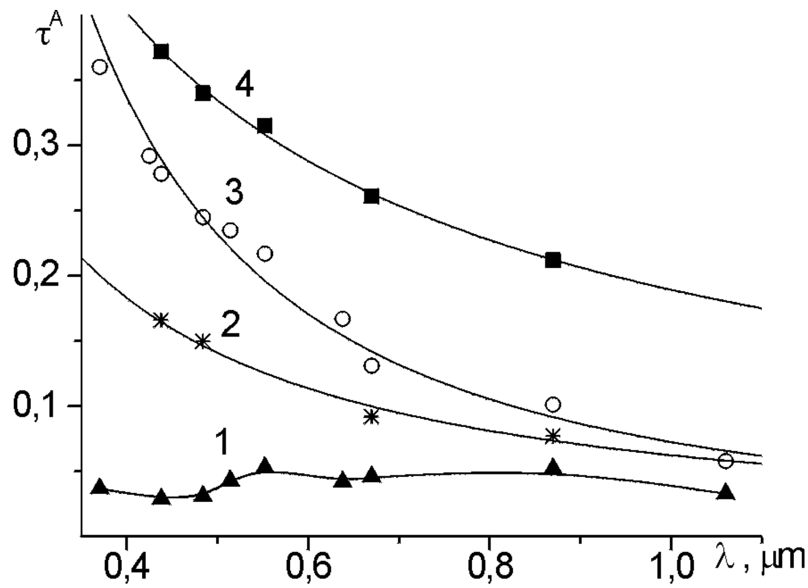

Figure 1. Examples of the spectral behavior of AOT: 1) anomalous dependence τ^A; 2) mean dependence for the total bulk of data (1992-1997, the volcanic effect is removed); 3) at the big content of fine aerosol; and 4) in conditions of maximum turbidity of the atmosphere.

Table 1. Statistics of α and (β); V = mean/SD.

Time	Mean	SD	Max	Min	V
Summer 1992	0,81 (0,14)	0,28 (0,14)	1,37 (0,22)	0,27 (0,07)	0,35 (0,26)
Winter 1992	0,40 (0,17)	0,26 (0,06)	0,71 (0,29)	0 (0,08)	0,65 (0,35)
Spring 1993	1,43 (0,09)	0,42 (0,06)	2,52 (0,23)	0,57 (0,02)	0,30 (0,63)
Summer 1993	1,14 (0,10)	0,39 (0,040)	1,76 (0,17)	0,48 (0,03)	0,34 (0,37)
Summer 1994	1,28 (0,05)	0,43 (0,02)	1,84 (0,09)	0,34 (0,02)	0,34 (0,36)
Summer 1995	1,08 (0,07)	0,44 (0,04)	1,76 (0,16)	0 (0,04)	0,41 (0,51)
Summer 1997	1,05 (0,04)	0,69 (0,01)	2,02 (0,07)	-0,27 (0,01)	0,66 (0,31)

The different effects of different aerosol fractions on scattering at different wavelengths are seen in the correlation of τ_λ with α and β (Table 2). The parameter β closely related to the content of coarse particles correlates well with τ_λ in all wavelength range. Optical activity and relative contribution of fine particles increases as the wavelength decreases. So the correlation of τ_λ with α is significant only in the visible range, and disappears or is even negative in the infrared (IR) range.

Table 2. Correlation coefficients τ_λ with α, β.

	0,44	**0,48**	**0,55**	**0,67**	**0,87**	**α**	**β**
0,44	1	0,99	0,94	0,91	0,75	0,31	0,70
0,48		1	0,96	0,93	0,73	0,32	0,70
0,55			1	0,92	0,72	0,27	0,71
0,67				1	0,86	0,06	0,88
0,87					1	-0,26	0,96
α						1	-0,34
β							1

The Anomalous Transparency

The shape of the anomalous dependence $\tau^A(\lambda)$ is shown in Figure 2, and the quantitative data on the dynamics of conditions are shown in Table 3. The initial period was characterized by the sharp decrease of temperature T, absolute humidity e, and columnar water vapor W. These conditions were kept for three days. Then the increase of T, e, and W was observed, but relative humidity (RH) remained practically at the same level ~65%, and exceeded 70% only after July 29. The monotonic decrease of $\tau^A(\lambda)$ observed on July 23 changed under these conditions to the spectral behavior with a depression near 0.44 μm and higher values in the long-wave range. The mean «anomaly depth» $[\tau^A_{0,56}–\tau^A_{0,44}]$ = 0,02 is 2 to 3 times greater than the error. Subsequent transformation of the air mass was expressed in the predominant increase of aerosol turbidity in the «violet» spectral range, and, for 3 days, the dependence $\tau^A(\lambda)$ again became usual for the continental mid-latitudes.

Similar dependencies $\tau^A(\lambda)$ were episodically observed in the Leningrad and Ryazan regions and in the Arctic (Barteneva and Nikitinskaya 1991). The spread maximum of AOT in the wavelength range 0.5-0.7 μm is typical for the majority of events in the Antarctic.

Analysis of the experimental conditions and the character of the spectral dependence gives reason to exclude the idea of instrumental and methodic errors and to discount the gas absorption. First, the absorption was taken into account exactly with examination of the real variability of the water vapor and ozone contents (Kabanov and Sakerin 1997). Second, the degree of selectivity of the absorption is not comparable with smoother dependencies $\tau^A(\lambda)$. However, one cannot exclude the effect of the continuum absorption or the total contribution of very weak water vapor absorption lines (Bikov et al. 1999). There are no exact data concerning this problem, but, according to the rough estimates, continuum absorption does not exceed ~0.002 in conditions of low humidity. Thus, there are reasons to consider the possibility of the anomaly because of the peculiarities of the aerosol composition.

Before describing microphysical simulation, let us present the qualitative ideas on the physical prerequisites of the appearance of the anomalous dependence $\tau^A(\lambda)$. It follows from the weak variability of AOT in the long-wave range, that the content of coarse particles was the same and close to the mean

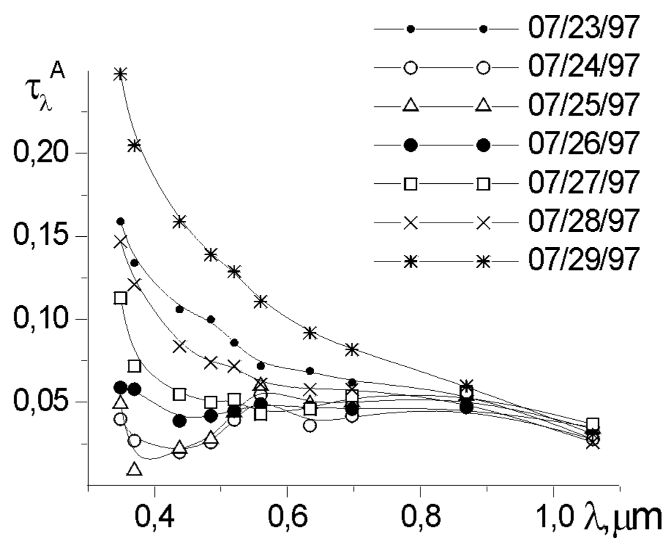

Figure 2. An example of anomalous spectral behavior of AOT at the intrusion of Arctic air and its change during the transformation of the air mass.

Table 3. The behavior of meteorological parameters and τ^A in the period July 23-29, 1997 (synoptic conditions: 23 – continental polar air (Cp), center of the cyclone, cold front with rain in the night; 24 ÷ 26 –arctic air, anticyclone; 27 – warm arctic front, Cp).

Days/Hours	$\tau^A_{0,52}$	T°C	e, mb	RH, %	W, g/cm^2
23/16	0,086	27,1	23,3	65,0	2,61
24/12	0,040	18,3	13,0	62,3	1,32
25/12	0,044	17,1	12,7	65,5	1,44
26/11	0,045	16,9	12,5	64,9	1,55
27/14	0,052	21	16,1	64,9	1,74
28/12	0,072	24,7	20	64,4	2,53
29/12	0,129	27,4	25,4	69,6	3,31

value. A significant decrease of AOT at wavelengths less than 0.55 µm is evidence of the same significant decrease of the content of the fine fraction, because small particles with a radius less than 0.3 µm (i.e., in the range before the first maximum of the efficiency of scattering K_ρ) play a principal role in the extinction of the short-wave radiation and the power decrease of $\tau^A(\lambda)$. As the effect of small particles decreases, the spectral peculiarities of $\tau^A(\lambda)$ are determined by the middle-disperse fraction. Its «optical image» can be displayed as a spread maximum $\tau^A(\lambda)$ in the long-wave range caused by the first maximum of K_ρ.

Microphysical Simulation

When analyzing the effect of different types of aerosol disperse composition, we used the hypothesis resulting from the interaction of primary and secondary generations of aerosol, the areas of effect which are separate in the size scale. So the aerosol particle size distribution function f(r) was simulated in the calculations as a superposition of lognormal modes (three modes in the range 0.03 ÷ 0.4 μm, two modes in the range 0.4 ÷ 1.5 μm, and one mode supplementing the content of particles in both ranges).

The results of simulative calculations show that the spread maximum τ^A in the wavelength range 0.6 ÷ 0.8 μm is the consequence of the prevalent contribution of the particles with r ~0.4 ÷ 0.75 μm into the aerosol extinction. Some curves $\beta(\lambda)$ in Figure 3 illustrate the dynamics of the spectral behavior when changing the modal radius from 0.36 to 0.54 μm. The sharp increase of $\tau^A(\lambda)$ in the ultraviolet spectral range is caused by the effect of the fine fractions (see Figure 3 C, D). At least, the minimum of AOT of the atmosphere at λ ~0.44 μm is first related to the anomalously low content of the accumulative fraction (r ~0.09 ÷ 0.3 μm).

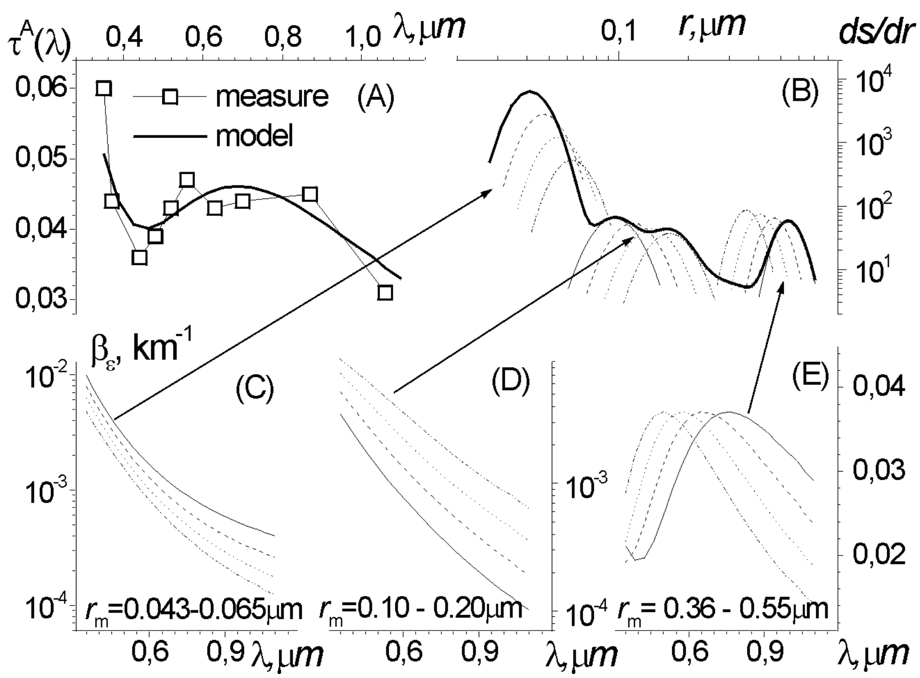

Figure 3. Comparison of the empirical dependence τ^A_λ (A) with the results of simulation f(r) (B); the curves (C) and (E) characterize the contribution of different fractions.

Keeping the long-term anomalous spectral dependence of AOT during the period July 24-26 is possible only if the particle size spectrum is conservative. That means, the stability of the size spectrum of accumulative fraction (averaged over the atmospheric column) in this period was caused by the low efficiency of the formation of new particles from the microdisperse fraction. Also, the initial number density of microdisperse particles in Arctic air was not sufficient for the effective coagulation growth of particles up to the range of resonance scattering.

The particle size spectrum in «primary» aerosol size range has one more specific peculiarity, namely the relatively small content of coarse particles in the column but narrow and stable mode in the range 0.5-0.6 μm. The noted fact can be the consequence of the effective gravitational sink and quick decrease of the content of coarse particles as the height increases. Other results are also evidence of the possibility of the existence of such a particle size spectrum.

So, the estimates of the effect of the Stokes sedimentation in the stratospheric layer (Rakhimov 1992) show that the coarse fraction displays the tendency to monodispersization (Figure 4C). It follows from the results of simulation that, at weak turbulent mixing, the quite narrow mode with r_m = 0.5-0.55 μm is formed at some heights from the initial wide coarse fraction. The particles of the aforementioned size range dominate in the composition of so-called “noncondensation clouds” (Zuev et al. 1990) and are seen in the results of microphysical measurements in the ground layer of the atmosphere (Figure 4B, A, respectively). The constant presence of these kinds of particles in different layers of the atmosphere is the initial reason for the stable maximum of $\tau^A(\lambda)$ in the range 0.6-0.8 μm and the weak mobility of this mode during subsequent days of measurements.

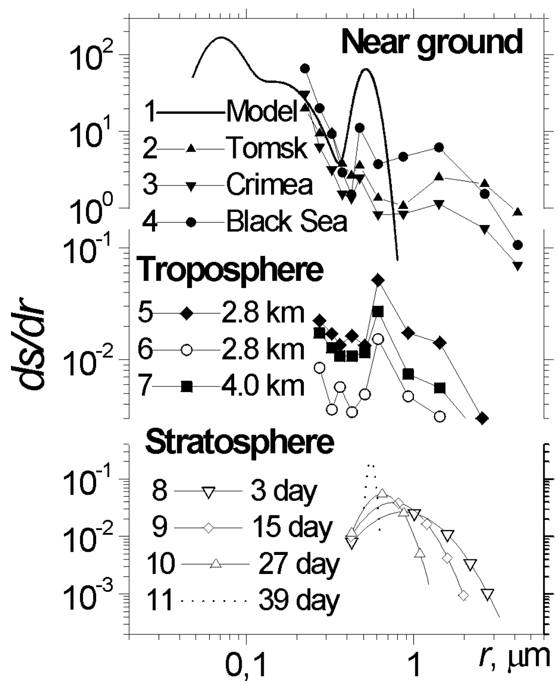

Figure 4. Illustration of the existence of the narrow fraction (r ~0.5-0.6 μm) in different conditions: the results of ground-based measurements (A), the model spectrum for the anomalous dependence τ^A_λ is shown by solid line; airborne measurements inside a “noncondensation cloud” (B); temporal dynamics of the size spectrum of the coarse particle fraction at the height 21 km at gravitational sedimentation (C).

The considered microphysical scenario is in agreement with the atmospheric conditions (Table 3). It is natural to assume that the sharp decrease of the fine particles number density occurred when the Arctic air mass had come. Indeed, air in the near-polar regions is characterized by the small content of both Itken particles and submicron aerosol (e.g., Kondratyev 1991). For far distances from the aerosol sources, low values occur for humidity, temperature and irradiation. If the air mass was transported from the zone of its formation along the short trajectory and was accompanied by wash out of aerosol at the front, the disperse composition would remain initial. Actually, at such intrusions we have a rare possibility to study the proper aerosol thickness of the Arctic zone.

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