VARIATIONS OF THE AEROSOL OPTICAL DEPTH OF THE ATMOSPHERE ABOVE TOMSK FOR A NUMBER OF SEASONS IN 1992–1995

D.M. Kabanov and S.M. Sakerin

Institute of Atmospheric Optics,
Siberian Branch of the Russian Academy of Sciences, Tomsk
Received January 26, 1996

This paper presents first results of studying short-term variability of the aerosol optical depth of the atmosphere mainly for warm seasons over the city of Tomsk. The statistical characteristics of the optical depth are shown to be spectral-dependent, they smoothly decrease as the wavelength increases. The year-to-year variability of the atmospheric turbidity, during the postvolcanic relaxation of the atmosphere to the background state, is evaluated. Considered are also the data on the character of diurnal variability of the aerosol optical depth and the presence of synoptic and global oscillations. It was noticed that the influence of the city manifested itself in higher turbidity (by 28% on the average) and variability.

INTRODUCTION

The results of studying the variability of the atmosphere recently generalized[1–6,etc.] revealed the most general regularities in the spatial distribution and long-term oscillations of the aerosol optical depth (AOD) of the atmosphere, which are typical for the territory of the former USSR. A peculiar feature of these works is in the fact that they were based on many-year observations at the actinometric or ozonometric network of stations, therefore the initial data included only the data on daily average values of AOD and only for the visible wavelength range.

More detailed studies, including the spectral dependence, the day time dependence and the correlation with other optical parameters, are conducted irregularly and in a limited number of regions.[7–9] From this point of view, the systematic cycles of spectral AOD measurements, which were started in Tomsk in 1992 and now are being conducted together with the studies of the physical characteristics of the atmosphere under SATOR Program,[10] are of particular interest.

This paper presents the first results of analysis of the observation series on the spectral AOD of the atmosphere for several seasons in 1992–1995.

INSTRUMENTATION AND MEASUREMENT TECHNIQUES

The spectral transmission of the atmospheric column was measured with the AMSF multiwave solar photometer.[11] During years of its exploitation, it was several times modified in order to improve its performance characteristics and to increase the number of spectral ranges . The short-wave channel of the last version of the AMSF photometer (1995) has 12 interference filters in 0.37–1.06 μm spectral range. One filter is centered at 0.94 μm absorption band to monitor the water vapor content in the atmosphere.

The data were recorded in an unattended mode by a continuous series up to 30 min long each hour under sunshine conditions. The technique for calibration and AOD (τ_λ^{a}) calculation was considered in our previous papers.[12] Let us note that for a more accurate account for the gas (τ_λ^{g}) and Rayleigh (τ_λ^{R}) components of the optical depth, they were computed by integrating within the entire transmission band $\Delta\lambda$ of the light filters by the equation

$$\tau_\lambda^{g+R} = -\ln\left[\frac{\int\limits_{\Delta\lambda} T_f\, I_0\, S\, T_a\, d\lambda}{\int\limits_{\Delta\lambda} T_f\, I_0\, S\, d\lambda}\right],$$

where T_f is the light filter spectral transmittance, I_0 is the spectral solar constant,[13] S is the photodetector spectral sensitivity, T_a is the spectral transmission of the atmosphere (the gas and Rayleigh components) by the LOWTRAN–7 model. The values of optical depths ($\tau_\lambda^{g} + \tau_\lambda^{R}$), computed for the midlatitude summer, are listed in Table I.

To analyze the AOD variability characteristics, individual spectra (spectra recorded for 1–2 minute intervals) were first averaged over hour periods and then over a day. To reveal possible seasonal differences, the whole data arrays were divided into subarrays for spring, summer, and winter periods. The general characterization of the results obtained is given in Table II.

TABLE I. Optical depths ($\tau_\lambda^g + \tau_\lambda^R$) for different AMSF spectral channels in 1995.

λ, nm	370	409	425	439	485	514	553	638	673	870	1061
$\tau^g + \tau^R$	0.498	0.320	0.293	0.252	0.170	0.145	0.127	0.097	0.077	0.067	0.087

TABLE II. Data bulks for different seasons.

Serial number	Observation period	The number of days with observations	The number of series	The number of hourly averaged measurements
1	Summer/92 (May 23 – July 30)	52	1388	340
2	Winter/92 (December 5–December 22)	8	105	21
3	Spring/93 (April 4 – May 19)	21	400	117
4	Summer/93 (May 22.05 – June 13)	20	503	176
5	Summer /94 (June 19 – July 27)	24	1126	148
6	Summer /95 (June 2 – July 17)	30	4272	187
Total number of observation		155	7864	989

The major part of researches was carried out in Akademgorodok, the suburb of Tomsk. To evaluate possible influence of the city, in 1992 and 1994 short series of measurements were conducted in the forest zone – at the scientific station of the Institute of Atmospheric Optics (Kireevsk village, about 60 km from Tomsk).

AOD STATISTICS FOR DIFFERENT PERIODS

A general idea of the variability of the atmospheric turbidity with aerosol is given in Fig. 1 using $\tau_{0.48}^d$ as an example. To obtain the quantitative information about the character of AOD variability, we have calculated the average, maximal (τ_{max}), and minimal (τ_{min}) values, rms deviations σ_τ, and variation coefficients V_τ. These values are presented in Table III. The results presented indicate that τ_λ^d varies within one order of magnitude: from 0.04 to 0.5, whereas the value of AOD averaged over the entire period of observations is about 0.2 in the 0.48–0.55 μm spectral range. This spread of AOD and its average value are in good agreement with the results of observations in continental areas.[4–8]

An important feature of τ_λ^d statistical characteristics is their spectral dependence: they decrease smoothly as the wavelength increases (see Fig. 2). The spectral dependence of the variation coefficient is not so well pronounced. Because of similar spectral dependence of τ_λ^d and σ_τ, the variation coefficient V_τ is stable: whatever a season or a wavelength is, the AOD relative variability is within 30–50%.

The histograms of τ_λ^d frequency of occurrence for different seasons are, as a rule, unimodal and asymmetric with a long fall off toward large atmospheric turbidity. This is illustrated in Fig. 3, where set out is the histogram for all summer periods of the observations. As follows from Fig. 3, the most probable value of $\tau_{0.48}^d$ is 0.21 (average $\tau_{0.48}^d$ is 0.212) and more that 50% of AOD value are within 0.12 to 0.24. For a comparison, many-year data for Moscow in summer periods[4] are: the mode and the average $\tau_{0.55}^d \sim 0.22$.

SEASONAL AND YEAR-TO-YEAR VARIABILITY

The observation series obtained does not yet allow us to reveal the characteristics of long-term AOD oscillations.

The presence of a year-to-year trend in AOD value is already evident from the data for summer periods of 1992–1995 presented in Table III. However, to be more correct, we have selected one month, June, because for this month there is the observation series including all years from 1992 to 1995. The results presented in Table IV are indicative of the presence of a year-to-year trend toward a smooth reduction of the atmospheric turbidity, that reaches 70% for three years.

To a certain extend the decrease in AOD can be explained by the reduction of industrial activity beginning from the late 1980s. But, taking into account the fact that in summer of 1991 strong eruption of Mt.Pinatubo took place, there are grounds for believing that the trend noticed is the relaxation of the atmosphere to the background state. The AOD decrease can be traced over the entire spectral range. As to the absolute value, a somewhat more sharp decrease is observed in a short-wave range. If the atmosphere is assumed to reach its background state (as before the eruption) by 1995, then the relative contribution of the volcanic component to AOD was 42% in 1992, 29% in 1993, and 20% in 1994. It should be noted that in the year-to-year change of the AOD variability no its regular decrease is observed. It supports indirectly the presence of the volcanic factor, because the short-term AOD variations are caused by the processes in the troposphere, whereas the trend is connected with evolution of the volcanic layer in the stratosphere. Thus, the year-to-year variation of AOD for the period under study was significant and caused mainly by progressive spread and sedimentation of aerosol produced in the volcanic eruption.

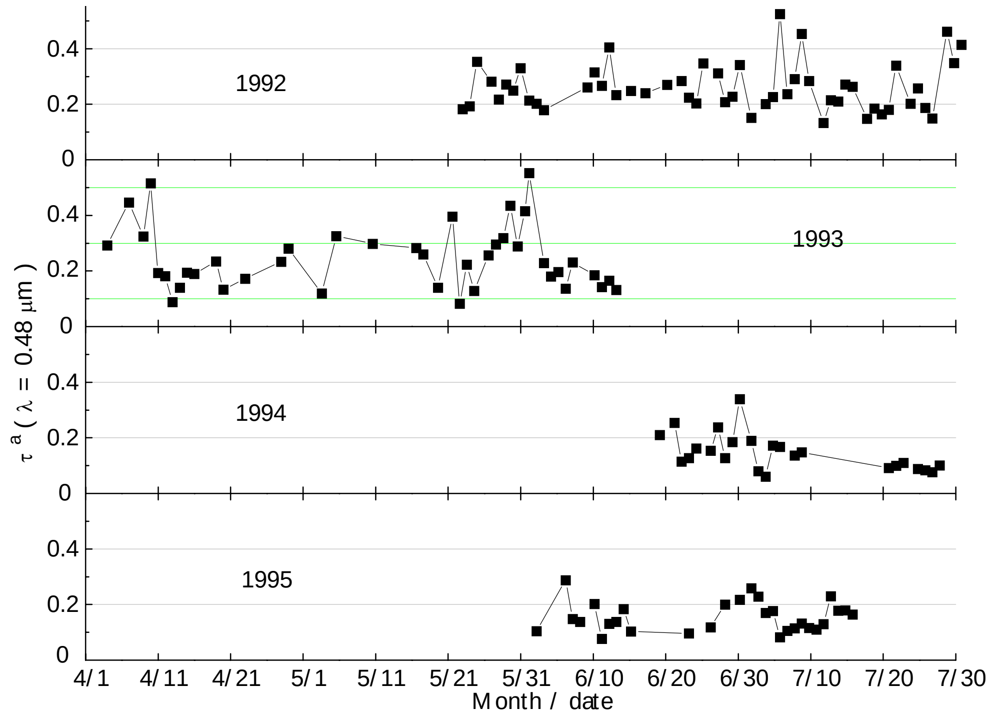

FIG. 1. Time variation of daily average values of the atmospheric AOD in the 0.48 μm *spectral range.*

TABLE III. Statistical characteristics of spectral components of the atmospheric AOD.

	λ, nm	0.370	0.425	0.44	0.48	0.514	0.55	0.638	0.67	0.87	1.061
Summer/92	Average			0.284	0.260		0.215		0.163	0.117	
	σ			0.094	0.085		0.068		0.052	0.040	
	V			0.329	0.327		0.314		0.321	0.341	
	Max			0.553	0.524		0.428		0.301	0.214	
	Min			0.138	0.133		0.105		0.066	0.045	
Winter/92	Average			0.233	0.238		0.222		0.189	0.173	
	σ			0.083	0.081		0.075		0.061	0.065	
	V			0.355	0.340		0.337		0.321	0.376	
	Max			0.359	0.359		0.326		0.266	0.295	
	Min			0.095	0.101		0.092		0.082	0.079	
Spring/93	Average			0.260	0.239		0.204		0.147	0.110	
	σ			0.118	0.107		0.095		0.074	0.065	
	V			0.453	0.445		0.465		0.506	0.593	
	Max			0.557	0.515		0.449		0.326	0.270	
	Min			0.091	0.087		0.063		0.048	0.024	
Summer/93	Average			0.282	0.249		0.190		0.147	0.129	
	σ			0.136	0.122		0.090		0.067	0.045	
	V			0.482	0.490		0.474		0.456	0.439	
	Max			0.612	0.551		0.401		0.279	0.209	
	Min			0.092	0.082		0.057		0.037	0.028	
Summer/94	Average	0.195	0.172	0.164	0.146	0.131			0.086	0.064	0.060
	σ	0.097	0.081	0.076	0.066	0.060			0.033	0.024	0.025
	V	0.496	0.473	0.467	0.453	0.461			0.390	0.368	0.414
	Max	0.493	0.417	0.393	0.339	0.307			0.177	0.111	0.1
	Min	0.082	0.069	0.073	0.06	0.055			0.034	0.027	0.025
Summer/95	Average	0.212	0.181	0.176	0.155	0.146	0.133	0.116	0.100	0.074	0.058
	σ	0.077	0.061	0.065	0.054	0.054	0.041	0.055	0.039	0.053	0.030
	V	0.363	0.338	0.368	0.348	0.371	0.310	0.471	0.387	0.720	0.516
	Max	0.399	0.320	0.347	0.287	0.306	0.228	0.271	0.212	0.248	0.135
	Min	0.097	0.098	0.088	0.076	0.077	0.029	0.034	0.046		0.007

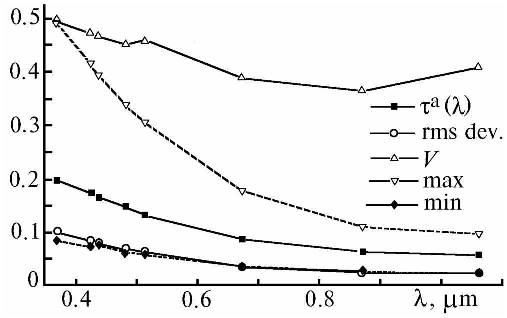

FIG. 2. Spectral dependence of the AOD statistical characteristics for summer of 1994.

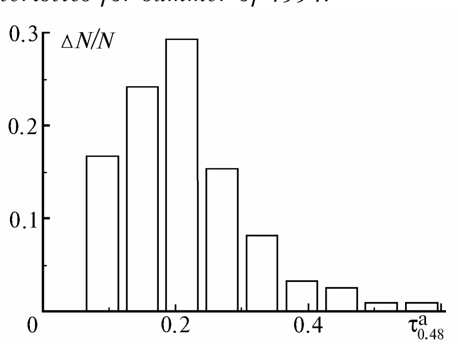

FIG. 3. Repetition histogram of $\tau^{D}_{0.48}$ for summer.

TABLE IV. Statistics for $\tau^a_{0.48}$ characteristics of the spectral components of the atmospheric AOD.

Statistical characteristics for $\tau^a_{0.48}$	Average	Rms deviation	V	Max	Min	N
1992	0.261	0.059	0.226	0.404	0.178	19
1993	0.214	0.123	0.574	0.511	0.131	10
1994	0.191	0.071	0.372	0.339	0.114	10
1995	0.152	0.058	0.382	0.287	0.076	14

It was difficult to reveal seasonal oscillations using the data array for the period from May, 1992 to June, 1993 because of a large year-to-year change in the atmospheric turbidity. In addition, winter observations in 1992 covered only short period and there were no observations during the fall at all. The AOD minimum in cold season and maximum in the warm one, well known from the literature,[3–6] in our case were insignificant and only in the short-wave spectral range (0.44 and 0.48 μm). Total amplitude of the seasonal variation, for example, for $\overline{\tau^d_{0.44}}$ was 0.05. As follows from many-year data,[3–5] the amplitude of seasonal oscillations for different regions was, on the average, about 0.1 and did not exceed 0.15.

SHORT-TERM VARIATIONS OF AOD

To some extent the short-term AOD variability can be estimated from the average diurnal values (Table III and Fig. 1). Synoptic and global-scale oscillations with periods from several to 10–20 days are clearly seen in the figure. Despite of diurnal averaging, the amplitude of these variations is significant (it reaches 0.3 and greater) and exceeds the seasonal variations of AOD.

To determine the main scales of the variability of atmospheric turbidity, we have calculated the amplitude spectrum C^d_λ (see Fig. 4). Due to short duration and discontinuity of observation series, the data presented should be considered only as a rough estimation. In the spectrum shown one can separate out the maxima corresponding to the periods of synoptic (~ 3, 4–5, 7–8 days) and global (~ 12, 20 days) oscillations. Close values for the periods were obtained in Ref. 15 from the many-year observation series from the ozonometric network. A comparison with the main periods of meteorological fields variability shows that the AOD oscillations separated out correspond to the lifetime of synoptic objects and the cycles of zonal circulation for midlatitudes.

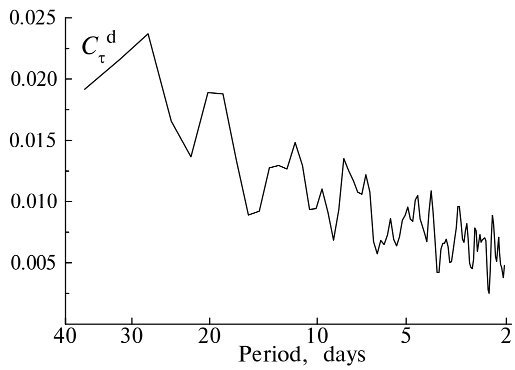

FIG. 4. Periodogram of variations of the atmospheric AOD (λ a 0.48 μm).

Analysis of the literature data on the diurnal run of the atmospheric AOD revealed a diversity of opinions, from its absence[6] to the anomalously large midday maximum reaching more than 50% of the average diurnal value.[4] Such a situation can result from different local conditions and measurement techniques. Among the local peculiarities, affecting the AOD diurnal run, are the closeness to a city and the degree of its anthropogenic impact, a wind–rose, type and state of the underlying surface. Measurement techniques can differ in approaches to the frequency of observations, the data averaging, the work under conditions of broken cloudiness, etc.

To separate out precisely the diurnal run of turbidity of the atmospheric column, we have used the results of calculations of hourly averaged values of AOD ($\overline{\tau^h}$) in two versions: with previous normalization to τ^d and without it (Fig. 5).

It is clearly seen from the figure that in the AOD diurnal run three parts can be separated out: 1) a morning period lasting till midday is characterized by small values of $\overline{\tau^h}$ and hour-to-hour variability; 2) a day period lasting till 18:00 LT demonstrates a continuous increase in turbidity which is more than 3% per hour; 3) an evening period after 18:00 LT shows a less pronounced reduction of $\overline{\tau^h}$. The estimation by Student criterion has shown that $\overline{\tau^h}$ differences at separated parts (or the dynamics of hour-to-hour variability) are significant with the confidence probability α a 0.9–0.995. The amplitude of diurnal variability (difference between the morning values of AOD and the maximum ones at 18:00 LT) is 0.03 – 0.04 with the relative change about 20%.

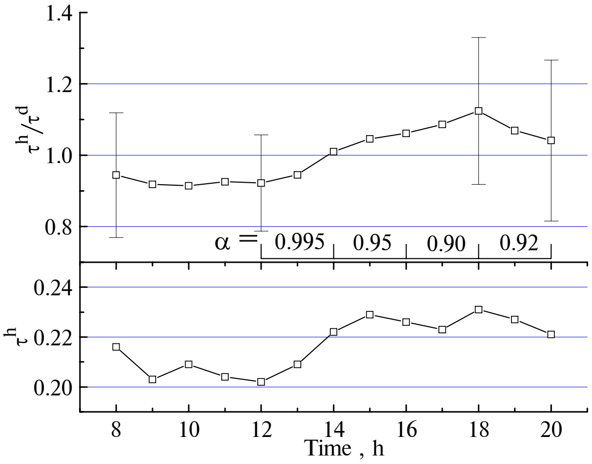

FIG. 5. Diurnal run of the absolute (τ^h) *and normalized* (τ^h/τ^d) *values of the atmospheric AOD.*

The results obtained are in qualitative agreement with the data of aerosol studies in the near-ground layer[14] in spite of their seeming difference in the morning period. One should keep in mind in this case that two main factors affect the atmospheric turbidity, namely, the day increase in aerosol content (sunrise, intensification of turbulence, convection, and industrial activity) and the aerosol "drying up" under the effect of humidity lowering. Before midday the latter factor prevails in the near-ground layer, but in the integral parameter $\xi(\tau)$ the opposing effects from two factors prove to be comparable. In the afternoon the effect of the latter factor becomes insignificant and the dynamics of the aerosol turbidity (the scattering coefficient) becomes the same throughout the atmospheric column.

A more detailed analysis of the diurnal transformation of the near-ground and integral optical characteristics of aerosol, including those in different spectral ranges, is interesting to be carried out in the future (in collaboration with the authors of Ref. 14).

The excess of the AOD afternoon values over the morning ones could be explained by the influence of industrial enterprises since the sun azimuths after 14:00 LT were directed toward the city. To estimate this influence we have separately calculated the AOD statistical characteristics (before and after 14:00) for the conditions of Tomsk and Kireevsk (a woodland area). The data presented in Table V show that the effect of the AOD diurnal run in both cases is practically the same: a 4–7% excess of AOD afternoon values is observed. As to the difference, the influence of the city manifests itself in higher atmospheric turbidity (~28%) and its variations. To obtain more accurate estimate, simultaneous measurements are needed.

TABLE V.

Statistical characteristics for $\tau^a_{0.48}$	Average	rms dev.	V	N
Tomsk, before 14:00 LT	0.213	0.107	0.502	380
after 14:00 LT	0.228	0.108	0.478	378
Kireevsk, before 14:00 LT	0.169	0.073	0.431	63
after 14:00 LT	0.176	0.077	0.437	54

ACKNOWLEDGMENT

The authors would like to acknowledge Doctor M.V. Panchenko for fruitful discussions of the mechanism of diurnal transformation of different optical characteristics of the aerosol and the statement of the problem of their simultaneous analysis.

REFERENCES

1. B.D. Belan, G.O. Zadde, and A.I. Kuskov, Atmos. Oceanic Opt. **7**, No. 10, 721–724 (1994).
2. B.D. Belan and G.O. Zadde, *Spectral Transmission and Aerosol Extinction over the Territory of the USSR* (Tomsk Affiliate of the Siberian Branch of the USSR Academy of Sciences, Tomsk, 1988), 180 pp.
3. E.V. Yarkho, Atmos. Oceanic Opt. **8**, No. 7, 553–557 (1995).
4. T.M. Abakumova and E.V. Yarkho, Fiz. Atmos. Okeana **30**, No. 11, 107–113 (1994).
5. E.V. Yarkho, Izv. Ros. Akad. Nauk., Fiz. Atmos. Okeana **30**, No. 3, 417–424 (1994).
6. G.P. Gushchin, *Methods, Instrumentation, and Results of Measurements of the Atmospheric Spectral Transmittance* (Gidrometeoizdat, Leningrad, 1988), 200 pp.
7. O.D. Barteneva, N.I. Nikitinskaya, et al., *Transmittance of the Atmospheric Column in the Visible and Near Infrared* (Gidrometeoizdat, Leningrad, 1991), 223 pp.
8. V.N. Glushko, A.I. Ivanov, et al., *Scattering of IR Radiation in the Cloudless Atmosphere* (Nauka, Alma–Ata, 1974), 210 pp.
9. V.V. Lukshin, G.I. Gorchakov, and A.S. Smirnov, in: *Results of Complex Aerosol Experiment ODAEX–87* (Tomsk Scientific Center, Tomsk, 1989), pp. 70–76.
10. V.V. Zuev, Atmos. Oceanic Opt. **5**, No. 6, 354–359 (1992).
11. D.M. Kabanov, S.M. Sakerin, A.M. Sutormin, and S.A. Turchinovich, Atmos. Oceanic Opt. **6**, No. 4, 270–273 (1993).
12. A. Ignatov, I. Dergileva, Yu. Ratner, S. Sakerin, and D. Kabanov, in: *Proc. IGARSS'94*, California, USA (1994), pp. 1497–1499.
13. *Precision Radiation Measurement in Metrology* [Russian translation ed. by K.Ya. Kondrat'ev and L.B. Krasil'shchikov] (Gidrometeoizdat, Leningrad, 1972), 272 pp.
14. M.V. Panchenko and S.A. Terpugova, Atmos. Oceanic Opt. **9**, No. 6, 464–468 (1996).
15. B.D. Belan, G.O. Zadde, and A.I. Kuskov, Atmos. Oceanic Opt. **8**, No. 4, 323–325 (1995).