Geophysical factors of aerosol weather formation in the Western Siberia

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This paper generalizes the available data on basic geophysical factors which determine variability of aerosol properties (optical, first of all) in the Western Siberia. They are described following the hierarchy of the processes usually used for description of the weather and climate in a specific geographical region. The seasonal variability of aerosol optical characteristics, their variations caused by change of air masses in the region, and diurnal behavior are considered based on data of airborne and ground-based measurements. Inter-annual periodic variations of the annually mean aerosol number density are revealed. The hypothesis is put forward that this periodicity is caused by circulation processes on the scale more than the regional one. It is shown that the effect of practically all significant geophysical factors can be seen in the aerosol weather. These factors are the inter-annual cycles, annual behavior, processes of synoptic scale, and diurnal transformation.

1. Introduction

At present a great bulk of data is available on aerosol optical and microphysical characteristics, principal processes of aerosol generation and transformation in all the altitude ranges where the presence of aerosol particles can affect optical properties of the atmosphere.[1–9]

At the same time, the wide spatial and temporal variability of aerosol properties and their connection with all the atmospheric processes cause the necessity of a more detailed study of all the variety of states in order to reveal the role of atmospheric particles in the change of the global climate and their effect on geospheric and biospheric processes, as well as to improve the dynamic models of optical characteristics that are necessary for solving problems of weather and climate forecast and for estimating the efficiency of systems operating through the atmosphere in the optical wavelength region.

The doubtless progress achieved in development of numerical models of the general circulation of the atmosphere (the radiation block in these models is one of the most important elements) dictates the necessity to achieve the corresponding level in the basic data on the principal optical characteristics of aerosol. Regardless of how perfect could be the analytical and numerical methods for solution of radiative problems, the success in forecasting climate change is mainly determined by the reliability of experimental data on optical parameters of the atmosphere and correctness of the description of their variability under the effect of external factors. In particular, we cannot ignore the possibility that correct allowance for optical properties of the particulate matter in the radiation balance will show (especially, taking into account the increase of the aerosol content due to anthropogenic sources) that influence of aerosol can compensate for the warming due to the greenhouse effect that is predicted by modern models of the atmosphere.[10]

Even the simple enumeration of such atmospheric situations as hazes, fogs, different kinds of clouds (liquid-droplet, crystal, mixed), forest fire smokes, emissions of industrial enterprises, dust storms, volcanic eruptions, precipitation in the form of rain and snow, and many others, in which the presence of particles is visible, gives an insight into the variety of forms and physico-chemical properties, as well as very wide size spectrum of atmospheric particles.

The most typical situation in the atmosphere is the aerosol state that, according to the classification given in Refs. 11 and 12, falls in the class of haze (occurs in more than 90% of events). Atmospheric hazes are observed when relative humidity of air is below 100% (Refs. 8 and 13).

The term "atmospheric aerosol" in this paper corresponds just the particles being in this state.

Since we deal with the real atmosphere, practically any problem concerning the study of aerosol is a typical multidisciplinary problem, the qualitative solution of which can hardly be obtained without using the methodology accepted in meteorological and climatic investigations.

In this paper we try to generalize the available data on the principal geophysical factors which determine the variability of the aerosol properties (optical, first of all) in the Western Siberia. To do this, we have chosen a sequence according to the certain hierarchy of processes, which are usually used for description of the weather and climate in a specific geographical region. We believe that this approach not only helps to construct the formal scheme of presentation of the available data, but also has a deeper meaning inherent of the object under study itself. That

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is why we use the term "aerosol weather" in the title of this paper.

" y definition, weather is "the physical state of the atmosphere in a certain time or limited time period, which is determined by the totality of meteorological parameters and atmospheric phenomena."[14]

Atmospheric aerosol is an inherent component of the atmosphere; it mainly causes its optical state and, in the general case, should be considered as one of the elements determining the weather. At the same time, when considering the "life" of aerosol particles in the atmosphere and analyzing the factors causing their transformation, one can use, in our opinion, the term "aerosol weather."

The expediency of using this term is caused by the following reasons: the specific state of aerosol (which can be represented as the totality of its characteristics) in the atmosphere depends on a great number of aerosol processes: power of sources and capacity of sinks, transformations during the process of particle origination and life in the atmosphere, i.e., the processes which have their own scales of variability. Some of these processes can be not related to traditional weather factors (it is especially true for the anthropogenic component).

In its turn, it is known that most aerosol processes in the atmosphere depend on the combination of practically all geophysical, synoptic, and meteorological factors (i.e., just the processes which determine the weather).

Obviously, for thorough description of the "aerosol weather" it is necessary to know the sufficiently large number of values which characterize microphysical parameters of particles (size distribution, shape, structure, electric charge, etc.) and their chemical composition. At present different characteristics of aerosol are studied to different extent, so it seems impossible to characterize the variability of the whole set of its parameters under the effect of external geophysical factors with equal reliability. Keeping in mind the principal fields of research carried out in the Institute of Atmospheric Optics, in this paper we pay the principal attention to optical characteristics and some parameters of aerosol microstructure, which noticeably determine the transformation of the optical state of the atmosphere in the Western Siberia (taking into account the fact that the limited space of this paper does not allow us to present a detailed description for all the processes, we restrict ourselves to the principal aspects of manifestation of geophysical factors, omitting the proofs which can be found in our earlier papers).

2. Annual behavior of the aerosol variability

Figure 1 shows the monthly average values of the scattering coefficient $\sigma_d(H)$ of the dry matter of aerosol particles (i.e., the values reduced to the zero relative humidity of air) obtained from the data of airborne sounding over the Western Siberia.[17] To estimate the impact of local sources on formation of the vertical profile, we have compared the monthly mean values of σ_d obtained at different altitudes separately over Tomsk and other sites of the Western Siberia.[17] The data differ markedly only in the near-ground atmospheric layer. Above this layer there are no pronounced differences. This is indicative of the homogeneity of aerosol properties over the entire region (for monthly mean values).

High near-ground values of σ_d in winter can be explained by intensification of anthropogenic sources of aerosol in the cold period and by the fact that the inverse temperature profile, which is typical of this period,[16] in the near-ground atmospheric layer prevents penetration of aerosol into upper layers of the atmosphere.

High values of the aerosol content that are distinctly seen in summer (the maximum in our data is observed in August) can be explained by the following factors. The underlying surface is most strongly heated in this season. As a result, aerosol particles from the surface reach high altitudes and, therefore, their lifetime in the atmosphere increases. Other factors are high insolation and long daytime, which determine intensity and duration of photochemical processes, as well as prevalence of slow small-gradient pressure fields[16] that favor accumulation of tropospheric aerosol. However, the maximum aerosol content is observed in August rather than July, although the above-listed factors manifest themselves most strongly in July. This is likely caused by smaller amount of precipitation in August.[16] Forest fires that arise in our region, as a rule, in this season can be additional sources of aerosol.

In fall the total aerosol content decreases likely due to the weaker role of the factors that are important in summer, as well as due to more frequent, compared to other seasons, fogs, drizzle, and steady rain[16] that favor clearing of the atmosphere and switch off the action of the underlying surface.

Thus, for the three seasons analyzed here the main processes determining the variability and the effect of aerosol sources on the regional scale are similar to those for large areas (at least, for midlatitudes of the Northern Hemisphere). " ut some peculiarities observed in spring require particular consideration.

The well pronounced maximum of the scattering coefficient is observed in April at the altitudes above 2 km (Figs. 1*c*, *d*, and *e*).[17] The appearance of the enhanced aerosol content at these altitudes is related to, first, the enhanced dynamic of atmospheric circulation caused by the great contrast of albedo due to the snow thawing boundary in spring.[18] " y this reason, a significant spatial contrast must occur also in the state and power of particle sources and aerosol producing vapor.

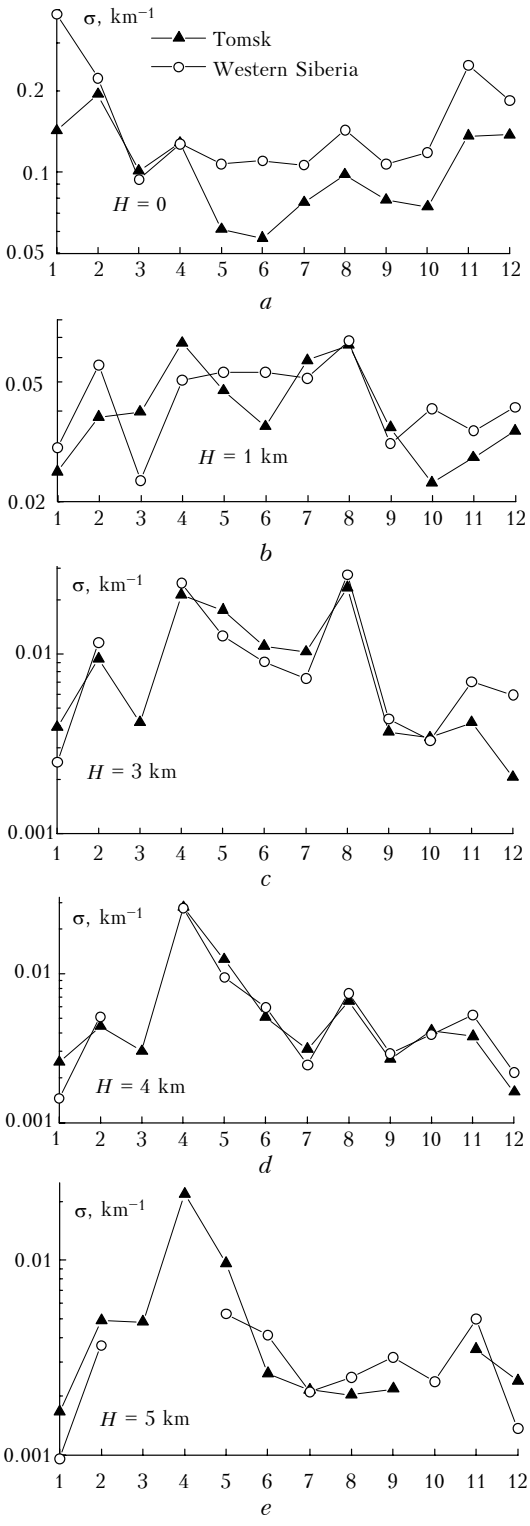


Fig. 1. Monthly mean values of aerosol scattering coefficient at different altitudes over Tomsk and other sites of the Western Siberia.

Second, since the westward transfer is characteristic of midlatitudes, we can suppose that aerosol particles, produced from various chemical compounds accumulated in snow during winter and emitted into the atmosphere because of snow melting in spring in west regions, come into the atmosphere of Western Siberia. Possibly, the products of the aroused vegetation make some contribution to aerosol generation (in April most conifers are free of snow and can emit the aerosol producing vapor under exposure to solar radiation[12]).

Seasons. Evidently, averaging the experimental data over months cannot give the characteristic pattern of the principal peculiarities of formation of the vertical profile and other optical characteristics, because the monthly mean values of even traditionally measured meteorological parameters of the atmosphere noticeably vary from year to year. At the same time, seasonal averaging is rather suitable for analysis.[17]

As is seen from Fig. 2, which shows the seasonal mean profiles normalized to the near-ground value $\sigma_d(0)$, the vertical stratification of aerosol characteristics varies markedly from season to season.

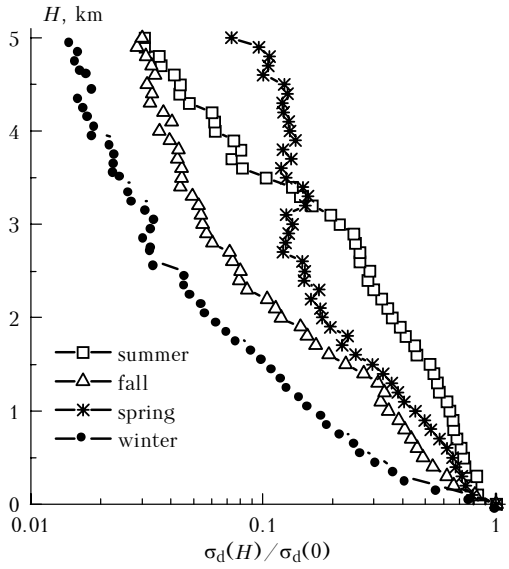


Fig. 2. Normalized vertical profiles of the aerosol scattering coefficients.

In *winter* the inverse temperature profile[16,17] prevents vertical motion of air, and the principal amount of aerosol particles is concentrated in the lower 200–500-m layer.

In *summer* the high temperature of the underlying surface and low atmospheric layers provides for mixing of air and aerosol particles along the vertical. A small decrease is observed in the near-ground layer (~100 m); then the aerosol scattering coefficient decreases insignificantly up to 3 km, decreases sharply at the altitudes from 300 to 400 m, and achieves the values characteristic of the free atmosphere.[17]

The vertical profiles of $\sigma_d(H)$ for the *spring* and *fall* seasons are close to each other up to 2.5 km and occupy the intermediate position between the winter and summer profiles. Above 2.5 km in spring the above-mentioned enhanced filling of the atmosphere with aerosol is observed. It is caused by the action of remote sources instead of temperature stratification.

The results of numerous ground-based measurements we conducted in the region do not contradict principal manifestations of the annual behavior and seasonal peculiarities, which are described above based on the data of airborne measurements, and significantly expand our knowledge of the optical characteristics in a wide spectral region.

3. Seasonal variability of aerosol

3.1. Statistical characteristics of the aerosol extinction coefficients in the near-ground atmospheric layer for different seasons

Figures 3–6 show the results of statistical processing of data on the aerosol extinction coefficient $\alpha(\lambda)$ in the wavelength range 0.4 to 12 μm. The content of all figures is identical: the mean spectral behavior of the coefficients $\alpha(\lambda)$, rms deviation $\sigma_\alpha(\lambda)$, and the normalized correlation coefficients between the aerosol extinction of radiation in the visible and IR wavelength regions $\rho_{\alpha(0.48),\alpha(\lambda)}$.

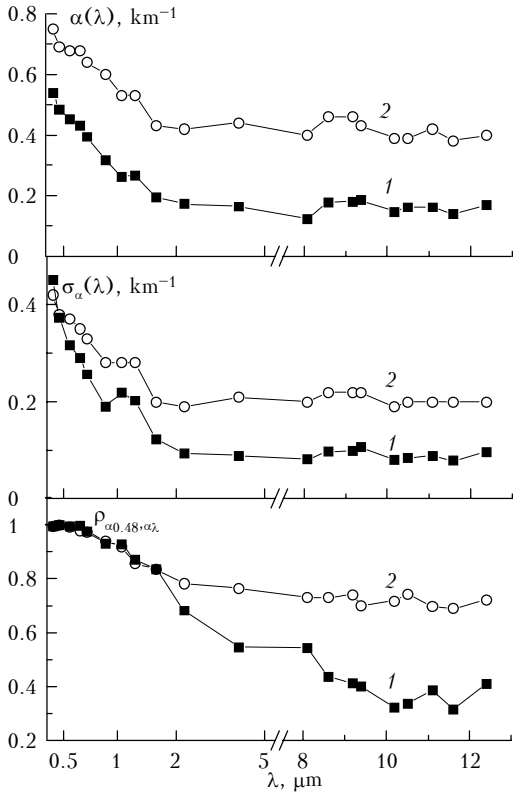


Fig. 3. Statistical characteristics of the aerosol extinction coefficients obtained in winter under conditions of haze (curves *1*) and ice fog (curves *2*). Measurements of 1992.

Figure 3 shows the data of measurements in winter (December 1992).[19] Curves *1* in this figure correspond to atmospheric haze, and curves *2* are for ice fog. It is appropriate to mention here that the term "ice fog" means the presence of very small ice crystals formed at temperature lower than −12°C under conditions of clear anticyclonic weather.[20] The largest particles of ice fog are well seen in Sun rays.

It is seen from Fig. 3 that the spectral behavior of the coefficient $\alpha(\lambda)$ characteristic of winter haze is the same as that of ice fog. It is indicative of the stable particle size spectrum and number density of the haze fraction in winter. Appearance of ice fog only forms the enhanced aerosol extinction (approximately by 0.2–0.25 km^{-1}) caused by ice crystals, but absolutely does not affect the optical properties of the fine fraction of atmospheric aerosol.

The variability of the coefficients $\alpha(\lambda)$ characterized by the values $\sigma_\alpha(\lambda)$ in the visible spectral range is more pronounced in winter haze than in ice fog, and vice versa for the IR wavelength range.

The comparison of the statistical characteristics of aerosol extinction obtained in December 1992 (curves *2*) and June 1995 (curves *1*) is shown in Fig. 4. It is seen that the aerosol extinction and its variations are some times greater in winter than in the warm season in the entire wavelength range. The correlation $\rho_{\alpha(0.48),\alpha(\lambda)}$ in spring decreases to the level of 0.5–0.4 in the near IR range (wavelengths of 0.87 to 1.06 μm), and in winter this correlation is observed only for $\lambda > 8$ μm.

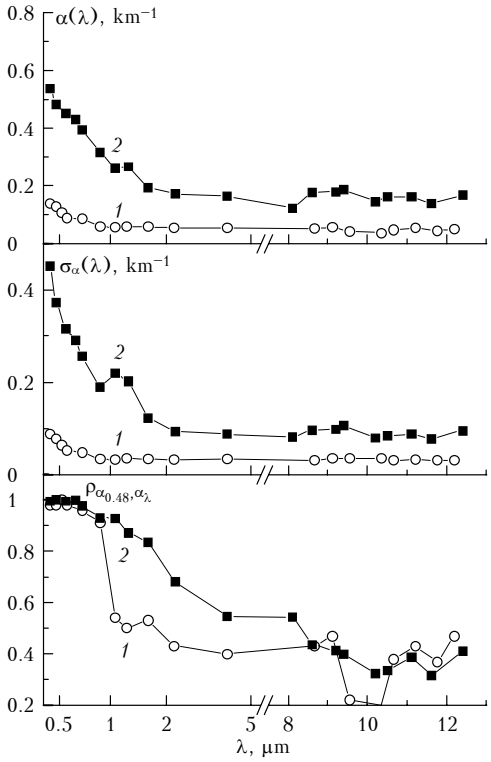


Fig. 4. Statistical characteristics of the aerosol extinction coefficients obtained under conditions of spring (curves *1*) and winter (curves *2*) haze. Measurements of 1992 and 1995.

The comparison of the statistical characteristics of aerosol extinction obtained in summer (curves *2*) and spring–fall (curves *1*) hazes is shown in Figs. 5 and 6.

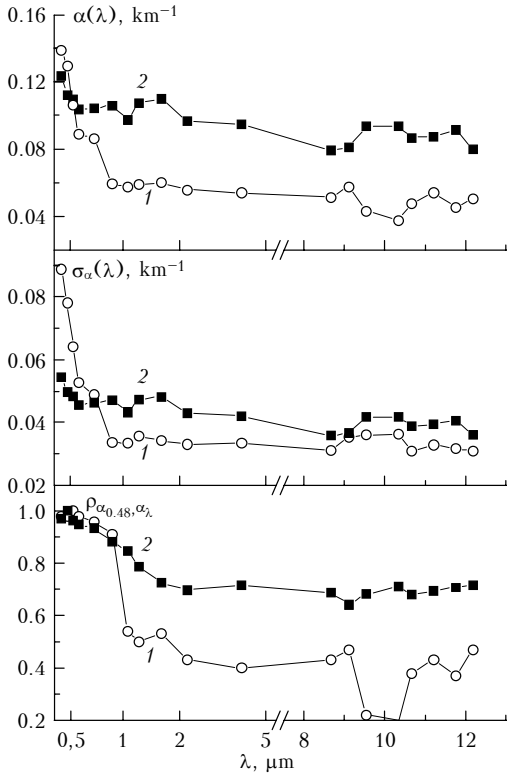


Fig. 5. Statistical characteristics of the aerosol extinction coefficients obtained under conditions of spring (curves *1*) and summer (curves *2*) haze. Measurements of 1995.

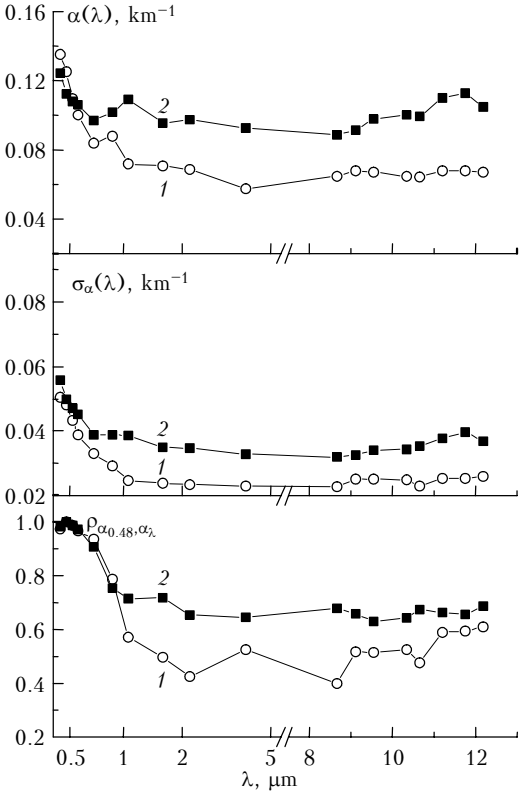


Fig. 6. Statistical characteristics of the aerosol extinction coefficients obtained under conditions of fall (curves *1*) and summer (curves *2*) haze. Measurements of 1998.

It is seen that in summer the stable increase of the aerosol extinction and its variations in the IR wavelength range is observed in comparison with the measurements of June 1995 (about two times) and with the data of fall 1998 (about 1.5 times). "esides, a more flat spectral behavior of the coefficients $\alpha(\lambda)$ is observed in summer haze in the visible range. This is likely caused by the decrease of the role of submicron aerosol under conditions of high temperature and low relative humidity.

3.2 Effect of the type of air mass

The main factors that essentially affect the variability of the aerosol characteristics for each season are synoptic-scale processes,[21] mainly, changes of air masses. Actually, at the very beginning of regular observations, meteorologists noticed that air masses differ in air color and meteorological visibility range.[22] However, by now, except for results of optical measurements in the ground atmospheric layer[23] and measurements of the spectral optical depth $\tau(\lambda)$ of the atmosphere,[24] the data on aerosol characteristics for different air masses are obviously insufficient.

The weather in the Western Siberia is determined by air masses of two types: continental Arctic and continental midlatitudinal ones.[16]

The results of comparison of the vertical profile of σ_d in the midlatitudinal and Arctic air masses are shown in Fig. 7 for every season.[25] The generally accepted idea that Arctic air masses bring the clear transparent air into the region of observations proves to be correct only for summer conditions (Fig. 7*c*), under which the aerosol loading of the midlatitudinal air mass is greater than that of the Arctic one at almost all altitudes. In fall (Fig. 7*d*) there are no reliable differences between the Arctic and midlatitudinal air masses.

In winter (Fig. 7*a*) at altitudes above ~ 500 m the Arctic air contains, on average, more aerosol particles than the midlatitudinal air masses. In spring at altitudes above 2.5 km the differences characteristic of the winter conditions retain, whereas in the lower 1 km layer the character of differences becomes similar to the summer one. These facts[25] are, at first sight, unexpected, so let us consider them in greater detail.

Systematic study of the Arctic aerosol has been started in 1972, though the phenomenon of "arctic haze"[26,27] has been observed already in the 50's. Now this problem attracts considerable attention of different specialists; the papers devoted to it are generalized in Refs. 28 and 29. "ased on the data from Ref. 29, the reasons for winter and spring accumulation of aerosol in the Arctic air become clear. The Arctic air masses are separated from the warmer midlatitudinal air by the wide Arctic front, which is a zone of precipitation and intense mixing, and hence it is impenetrable for aerosol from the south.

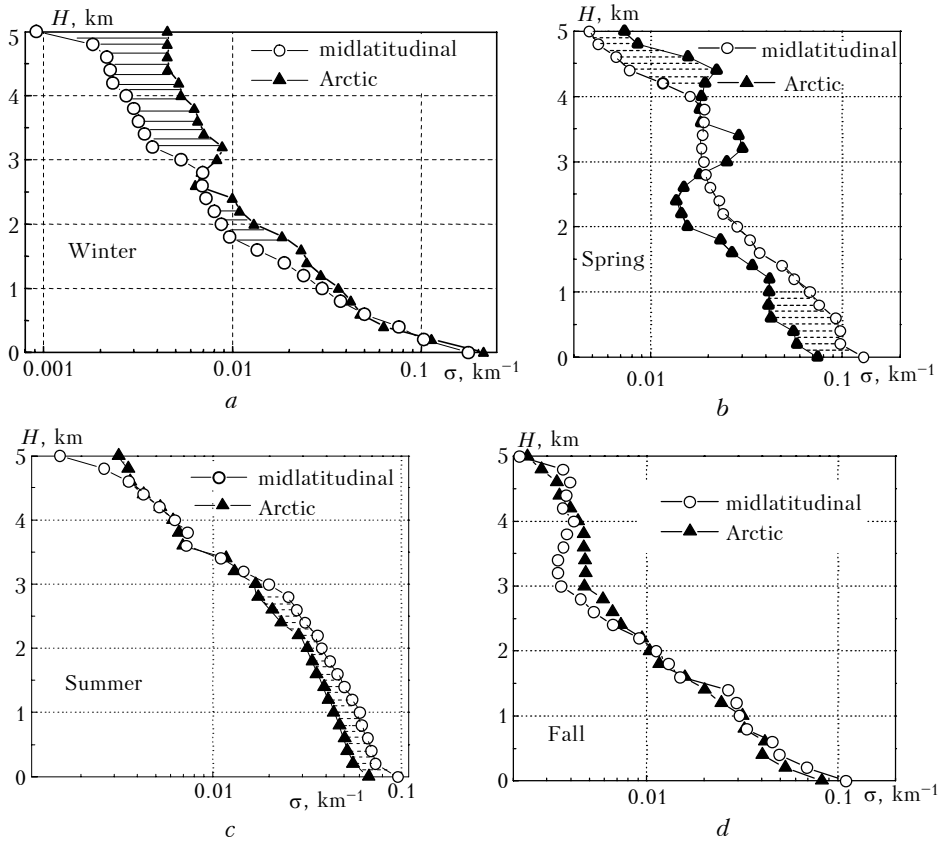


Fig. 7. Comparison of the vertical profiles of σ_d in the midlatitudinal and Arctic air masses for different seasons.

Since fall, as the underlying surface and atmosphere become cooler, the Arctic front shifts southwards. In winter and early spring the Arctic air masses cover the large area over North America and Eurasia.[30] As the atmosphere and underlying surface warm in spring, the Arctic front shifts to the north, and in late summer the Arctic air masses are concentrated over the near-pole areas.[28,30] The state of the underlying surface and atmosphere in Arctic in winter favors accumulation of aerosol, because chemical reactions are slow in the dark and cool atmosphere of high latitudes, photochemical reactions are almost absent, and frequent near-ground temperature inversions prevent aerosol sink onto the surface.[28]

One can see this process in Fig. 7, when comparing the mean vertical profiles $\sigma_d(H)$ for the Arctic air masses in different seasons. Starting from fall, the aerosol load increases at the altitudes above the mixing layer and reaches its maximum in spring.

In winter continental air masses bring aerosol to Arctic. Then it returns to the Western Siberia with Arctic air. So the Arctic air masses are more loaded with the aerosol than the midlatitudinal ones practically at all altitudes above the mixing layer. In spring, as the temperature increases, the effect of the underlying surface as an aerosol source grows, and the altitude of the mixing layer increases, so the enhanced aerosol content in the Arctic air in comparison with the midlatitudinal air is observed at the altitudes above 3 km. Below this level the differences in $\sigma_d(H)$ for different air masses acquire the summer character. In summer, as it was mentioned above, entrainment of the continental air into Arctic is less probable, and so the Arctic air masses contain fewer aerosol particles at all altitudes from the ground to 5 km than the midlatitudinal air masses. In fall, when it often rains and the underlying surface is wet, there are no differences between the Arctic and midlatitudinal air masses.[16]

The data of airborne observations and the principal conclusions on the character of differences in the optical state of the atmosphere at change of air masses are confirmed by the results of ground-based measurements of the spectral extinction coefficients and the aerosol optical depth in different seasons.

Using the representative data arrays of 1992 and 1995, it was shown that the spectral structure of the aerosol extinction coefficients in summer hazes depends significantly on the type of the present air mass. For example, the Arctic air mass coming to Siberia in June delayed flowering of conifers and deciduous trees and caused the low aerosol extinction coefficients in the IR wavelength range. But in July, as the air masses came from midlatitudes and subtropics, the level of the

extinction coefficients in the IR range significantly increased under conditions of the high atmospheric transparency. It was also found that, in contrast to summer conditions, the large values of the aerosol extinction coefficients are observed in winter hazes under conditions of the Arctic air mass. To illustrate the results discussed, the temperature dependence (actually, between seasons) of the mean aerosol extinction coefficients in the IR wavelength range at $\lambda = 3.9$ μm and the parameter $\Delta\alpha = \alpha_{0.44} - \alpha_{3.9}$ in the near-ground hazes is shown in the upper part of Fig. 8.

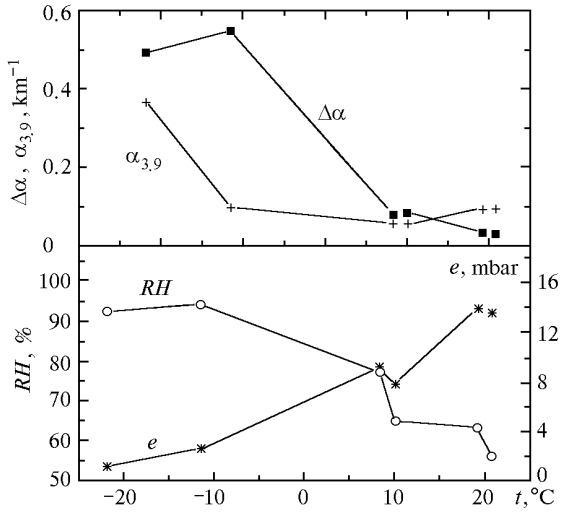


Fig. 8. Relative humidity of air (*RH*), partial pressure of water vapor (*e*), aerosol extinction coefficient $\alpha_{3.9}$, and parameter $\Delta\alpha = \alpha_{0.44} - \alpha_{3.9}$ as functions of air temperature.

The parameter $\Delta\alpha$ characterizes the extinction of radiation by submicron aerosol of the size approximately from 0.1 to 1 μm, while the parameter $\alpha_{3.9}$ mostly characterizes the extinction of optical radiation by coarse-disperse aerosol. It is seen from the figure that the contribution of submicron aerosol to extinction of optical radiation is maximum in winter and minimum in summer.

3.3. Variations of the aerosol optical depth at change of air masses

The aerosol optical depth τ^A is one of the main optical characteristics which directly affects the income of the direct and scattered solar radiation. As a rule, the dependence $\tau^A(\lambda)$ manifests itself at the wavelengths less than 1 μm as a monotonic decrease with the wavelength; and it is described by the Angströ m formula: $\tau^A(\lambda) = \beta\lambda^{-\chi}$ (χ is the Angström parameter characterizing the relative selectivity of the spectral behavior; β is the turbidity coefficient, whose value is close to τ^A near 1 μm). More rare the spectral behavior in the range $\lambda < 1$ μm has a quasi-neutral character. The spectral behavior of $\tau^A(\lambda)$ in the long-wave range is close to the neutral one.

Let us consider the peculiarities of $\tau^A(\lambda)$ variability based on the data obtained near Tomsk in 1992–1997.[31–33]

The analysis of the available data shows that the largest variations of $\tau^A(\lambda)$ and the parameter α occur on the synoptic scale. The change of air masses formed in the regions with different aerosol sources and climatic conditions affects mainly the atmospheric transparency. As a result of the change of air masses typical for a region, the variations of $\tau^A(\lambda)$ from day to day can reach 0.2 and even more. Even the mean amplitude of synoptic oscillations (the doubled value of σ_τ can be taken as an analog) is comparable with or greater than the seasonal and inter-annual variations.

Not only $\tau^A(\lambda)$, but also the Angström parameter changes, as an air mass changes. The area of co-values of $\tau^A(\lambda)$ and χ is illustrated by Fig. 9.

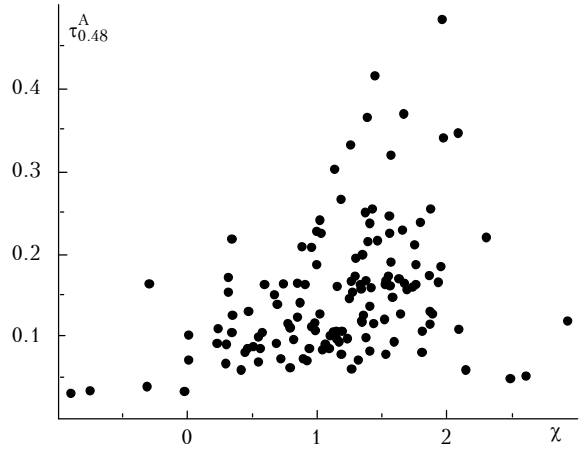


Fig. 9. Correlation diagram of daily mean values of χ and $\tau^A_{0.48}$

The small values $\chi < 0.5$ (quasi-neutral dependence $\tau^A(\lambda)$) occur, as a rule, under conditions of the Arctic air, and the value of $\tau^A_{0.48}$ does not exceed 0.2. The total number of such situations is ~ 15%. In some cases even anomalous dependence $\tau^A(\lambda)$ is observed (negative χ) with maxima in the red and UV spectral ranges.[41] Only high selectivity of the spectral behavior ($\chi = 1.2 - 2$) is characteristic of the high atmospheric turbidity $\tau^A_{0.48} > 0.3$, i.e., fine aerosol prevails in formation of high turbidity, and its content in summer is greater in the midlatitudinal air masses.

4. Diurnal behavior of aerosol

As was shown above, along with the variability of aerosol characteristics caused by the processes of synoptic scale, the aerosol state is also subject to quick variations during a day.

Of the processes that govern the aerosol variability, the main physical processes determining the diurnal behavior of the aerosol content and transformation of its optical properties are understood best of all.[34–36]

At the same time, when trying to quantitatively describe the diurnal behavior of aerosol characteristics under some specific conditions, the geophysical nature of aerosol is in the forefront. So it is necessary to take into account all the variety of external synoptic, geophysical, and local factors, prehistory of a specific air mass, the state of the underlying surface, insolation, etc. As a result, at present it is rather problematic to create a theoretical (quantitative) model describing all the variety of factors and their relationships even for describing the diurnal behavior. So it seems more realistic to describe it based on the data of experimental observations.

4.1. Vertical profile of the scattering coefficient

The data arrays were divided into "morning,B "day,B "evening,B and "nightB subarrays depending on the time of sunrise and sunset at the latitude of a particular geographic site for each month,[37] and then the data were averaged for each season.

agreement of the data obtained at our observation site and in the background region[38]). We plan such experiments to be conducted in the near future.

Spring and *fall* are characterized by the enhanced dynamics of the day-to-day variability of all atmospheric processes (frequent changes of air mass, great number of days with precipitation, and so on). This makes the analysis difficult and hides the peculiarities of the diurnal behavior. The diurnal transformation of the vertical profile $\sigma_d(H)$ in spring and fall is similar to the summer diurnal behavior, but the amplitude of variations is much less.

In summer the diurnal behavior of the aerosol vertical profile is most pronounced. The mean summer profile $\sigma_d(H)$ is shown in Fig. 11.

In *summer*, from the evening and during the nighttime, formation of the temperature inversion is usually observed. Its altitude can reach 400–500 m in the morning. This leads to the decrease in the total aerosol content in the layer under the inversion (100–400 m). The minimum values at the altitude ~ 300 m are observed in the morning.

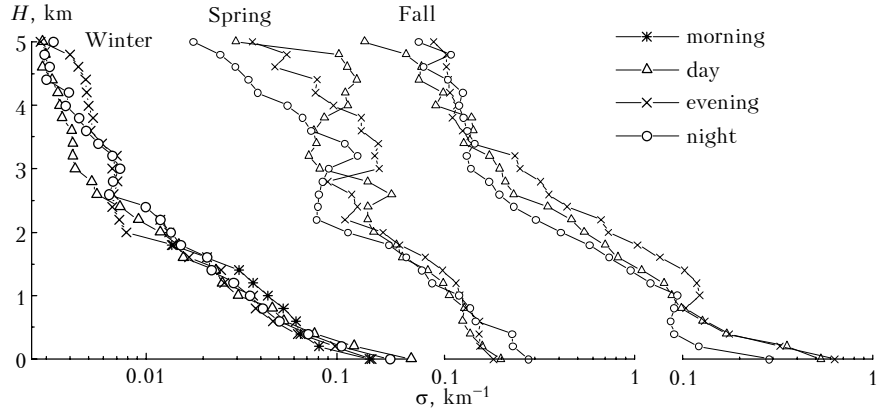


Fig. 10. Diurnal behavior of the vertical profile of the scattering coefficient of the dry matter of aerosol particles.

The thus obtained vertical profiles of the scattering coefficient $\sigma_d(H)$ of the dry matter of aerosol particles for winter, spring, and fall are shown in Fig. 10.

In *winter* no transformation of aerosol content along the vertical is observed during a day. Despite of the existence of the diurnal temperature variation in the lower (< 1 km) atmospheric layers, the mean temperature profile is inverse, and thus it suppresses the inter-level aerosol exchange. As for the data of near-ground measurements, it is too early to judge the diurnal behavior in winter. Let us only note that the daytime maximum of the dry aerosol content is often observed. At the same time, since our ground-based observation stations are situated near the city, it is not improbable that the daytime maximum in our case is caused by income of the urban air to the observation site. Hence, to answer this question, it is necessary to conduct a multi-site experiment in winter (such an experiment conducted in summer has shown the good

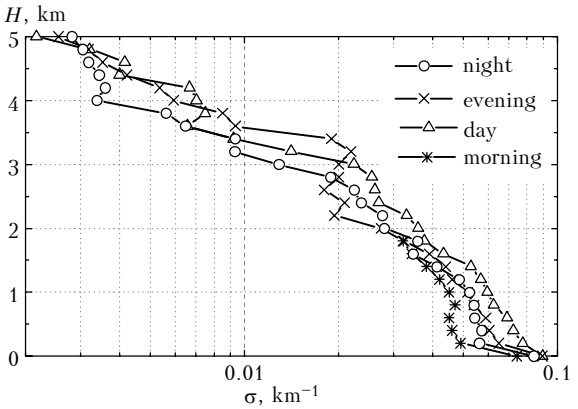


Fig. 11. Diurnal behavior of the vertical profile of the scattering coefficient of the dry matter of aerosol particles in summer.

Because of warming of the underlying surface and the atmosphere during a day, the altitude of the mixing

layer increases, and it is filled with aerosol. In the evening the aerosol emission from the near-ground layer stops, and the lower atmospheric layers (below $H \sim 1.5$ km, on the average) loose aerosol. Above these layers up to $H \sim 3.5$ km the altitude of the mixing layer continues to increase.

4.2. Diurnal behavior of the aerosol optical depth of the atmosphere in summer

To determine the diurnal behavior of the aerosol turbidity, we used the hourly mean values of $\tau^A_{0.48}$ normalized to the daily mean values for each day of measurements.[32,33]

It is seen from Fig. 12*a* that three parts can be separated in the diurnal behavior of $\tau^A_{0.48}$: (1) the morning period till ~ 11:00 a.m. of the mean solar time is characterized by small values of $\tau^A_{0.48}$ and small variations; (2) the continuous increase of turbidity about 3% an hour is observed in the day period till ~ 4:00 p.m.; (3) the decrease of τ^A up to the daily mean level occurs in the evening period. The amplitude of diurnal variations can be estimated as ~ 0.03, and the relative variation is no less than 15%.

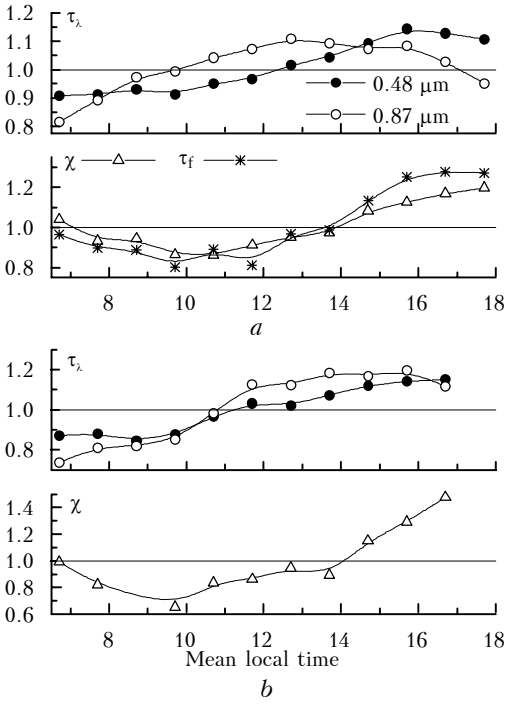


Fig. 12. Diurnal behavior of τ^A_λ, χ, and "fine-disperse" component of the aerosol optical depth $\tau^A_f = \tau^A_{0.48} - \tau^A_{0.87}$ in the region of Tomsk (*a*) and the forest zone (*b*).

The similar peculiarities were revealed for other wavelengths, but in the IR range (0.87 μm) the maximum is observed earlier and is better pronounced. The consequence of the spectral differences is the diurnal behavior of the parameter χ with the characteristic minimum at noon (diurnal variation of χ is about 25%). The statistical distinguishability of the revealed extrema is confirmed by the confidence probabilities calculated by the Student criterion (its value is no less than 0.999 in most cases).

This behavior of the aerosol optical depth is explained by different diurnal transformation of the fine and coarse aerosol fractions. The change of $\tau^A_{0.87}$ occurs mainly due to the effect of lifting of coarse particles from the underlying surface resulting from the diurnal dynamics of convection and turbulence. The change of $\tau^A_{0.48}$ is determined, to a greater extent, by the fine fraction, whose behavior depends on other factors. The daytime increase of the aerosol content begins after sunrise due to increasing industrial activity and switching-on of the mechanism of photochemical generation. At the same time, aerosol "dryingB occurs due to decreasing relative humidity. As a result of the opposite action of these two processes, before noon the fine fraction affects only slightly the change of the optical depth. Later the effect of humidity becomes weaker, and the relative contribution of fine particles increases. In the evening, as convection decreases, the daytime filling of the atmosphere with aerosol stops, and the sink process becomes to prevail. Coarse particles settle faster, thus causing the evening increase of the Angström parameter χ.

To estimate the effect of urban conditions on formation of the diurnal behavior of τ^A, we have additionally analyzed the results obtained in forest zone. It follows from Fig. 12*b* that, in spite of some differences appearing due to different time and duration of observations at the two sites, the diurnal behavior of τ^A is generally the same.

4.3. Diurnal behavior of the aerosol extinction coefficients at near-ground paths

The diurnal behavior of the optical characteristics $\Delta\alpha$ and $\alpha_{3.9}$, as well as the main meteorological parameters of the atmosphere averaged over about 50 days is shown in Figs. 13 and 14.

It is seen from the figures that the parameter $\Delta\alpha$ is mostly related to the relative humidity, while the coefficient $\alpha_{3.9}$ synchronically repeats the diurnal behavior of the air temperature. This is more pronounced in summer. Intense heating of the soil in a summer day favors formation of convective air flows and turbulent diffusion, what leads to the increase in emission of fine aerosol to the upper layers of the atmosphere. The number density of coarse particles simultaneously increases in the near-ground layer. These particles come from the ground and cannot be carried to high altitudes because of their big size. As a result, these particles form the more flat spectral behavior of the aerosol extinction coefficients in the near-ground haze in summer.

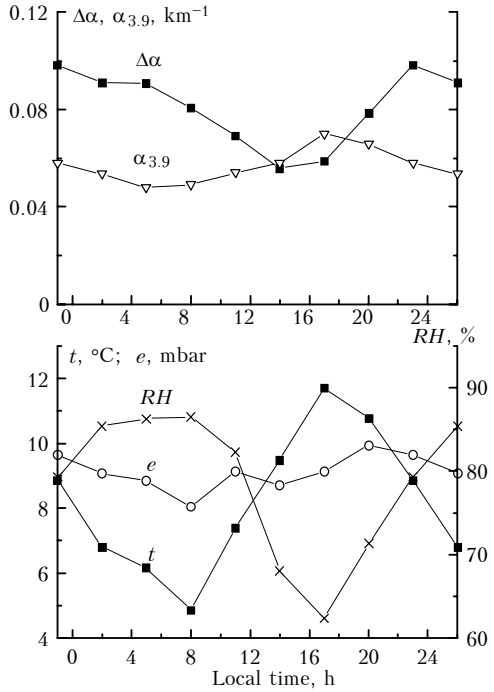


Fig. 13. Diurnal variations of the aerosol extinction coefficients caused by the coarse aerosol fraction (curve $\alpha_{3.9}$), submicron fraction (curve $\Delta\alpha$), and meteorological parameters of the atmosphere (fall 1998).

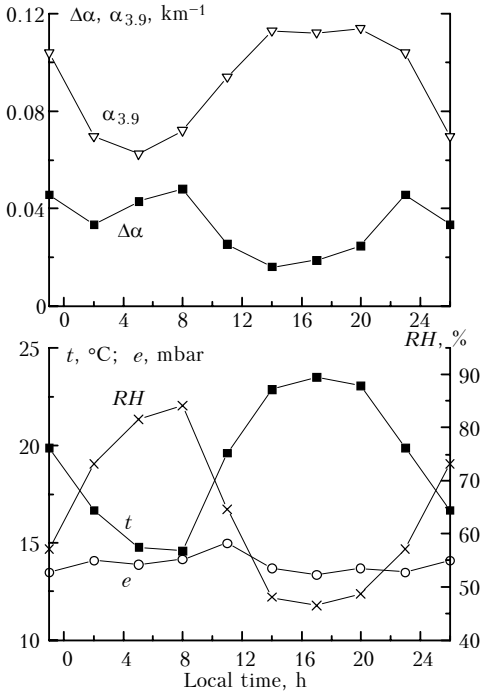


Fig. 14. Diurnal variations of the aerosol extinction coefficients caused by the coarse aerosol fraction (curve $\alpha_{3.9}$), submicron fraction (curve $\Delta\alpha$), and meteorological parameters of the atmosphere (summer 1998).

In fall, as soil heating decreases, convection and turbulent diffusion become weaker, what leads to the decrease in emission of fine aerosol from the near-ground layer in comparison with the summer conditions. The income of coarse soil aerosol decreases simultaneously, because in this season it often rains in the region of observations. As a result of simultaneous effect of these factors, the spectral structure of the coefficients α in fall is formed by both fine and coarse particles, contrary to summer.

5. Interannual variability of aerosol properties

In our opinion, it is too early to draw some climatically significant conclusions on the interannual variability in the region under study because of the limited time of observations. At the same time, let us note some interesting results that could be useful for further investigations. First, let us say a few words about the well-known facts.

The most marked interannual variations of aerosol properties are observed in the stratosphere and related to volcanic eruptions.[39] Leaving aside the question about the specific variations of the stratospheric aerosol that is of separate interest, let us only note that the contribution of stratospheric aerosol at powerful volcanic actions is seen in the integral optical characteristics. Let us illustrate this fact using as an example the interannual variations of the aerosol optical depth τ_λ^A in the region of Tomsk.

Let us consider the peculiarities of the variability of τ_λ^A based on the data obtained near Tomsk in summers of 1992–1997.[31–33] The statistics of spectral dependences of $\tau^A(\lambda)$ and their variations under different atmospheric conditions in Tomsk is presented in Table 1. Note that the statistical characteristics of the parameter β give the tentative concept on variations of the coarse aerosol fraction, and the parameter χ is related to the relative content of the fine fraction. The decrease of aerosol turbidity in the considered period is well seen in the statistical data. To some extent, the decrease of τ_λ^A can be related to the decrease of industrial activity since the early 90's. But, taking into account that the Mt. Pinatubo eruption occurred in 1991, the negative trend of τ_λ^A can be considered as resulting from the return of the atmosphere to the background state. The volcanic addition can be estimated as ~ 0.1, and its relative contribution changed from 40 to 20% for three years.[33]

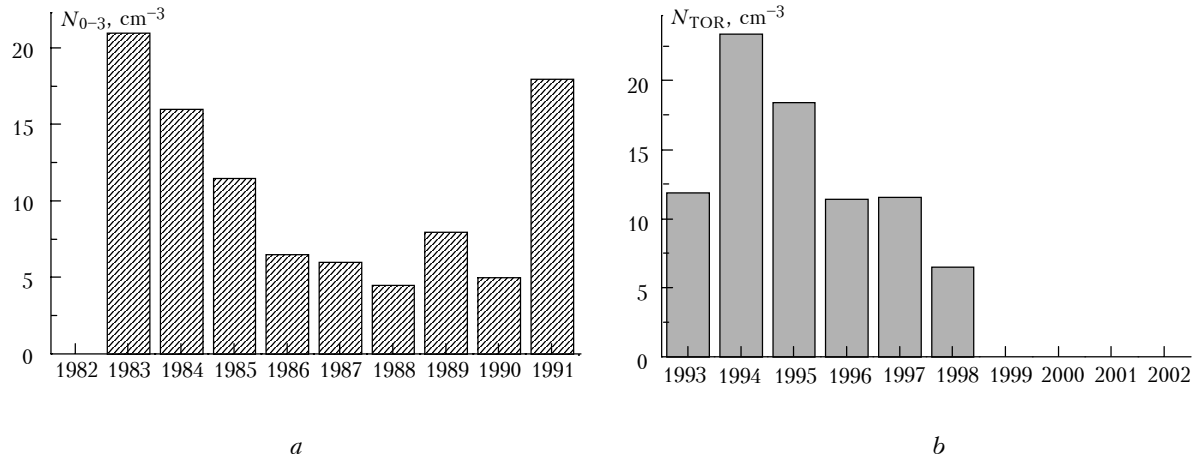
The interannual behavior[42] of the number density of aerosol particles with the radius greater than 0.2 μm observed since 1983 till 1998 is more interesting and requires the explanation. It is shown in Fig. 15.

As is seen from the annual mean values, the aerosol number density changes markedly from year to year, so, to some extent, one can say about the interannual periodicity. As a hypothesis, we can suppose that this periodicity is caused by circulation processes, which are beyond the regional scale[40] (it is not improbable that volcanic eruptions can be a sort of regulator[39]). But, as was mentioned above, more rigorous conclusions can be obtained upon accumulation of sufficient statistics for different regions of the Earth and only by joint efforts of scientists in different fields.

Table 1. Sample statistics of daily mean values of τ_λ^A ($\cdot 10^3$) and the parameters χ and β in the region of Tomsk

Period	Parameter	λ, μm					χ	β
		0.44	0.48	0.55	0.67	0.87		
Summer 1992	$\bar{\tau}$	283	263	232	192	162	0.81	0.14
	σ_τ	094	084	068	053	041	0.28	0.14
	V_τ	333	319	293	276	255	0.35	0.26
	Max	571	528	452	336	259	1.37	0.22
	Min	143	147	129	092	079	0.27	0.07
Winter 1992	$\bar{\tau}$	228	237	224	192	180	0.40	0.17
	σ_τ	082	080	074	060	065	0.26	0.06
	V_τ	360	336	330	312	360	0.65	0.35
	Max	351	356	327	267	303	0.71	0.29
	Min	089	010	094	085	086	0	0.08
Spring 1993	$\bar{\tau}$	261	237	203	149	110	1.43	0.09
	σ_τ	118	107	095	075	066	0.42	0.06
	V_τ	453	450	471	511	602	0.30	0.63
	Max	557	515	449	326	270	2.52	0.23
	Min	091	086	062	047	023	0.57	0.02
Summer 1993	$\bar{\tau}$	284	250	190	148	129	1.14	0.10
	σ_τ	134	121	089	066	044	0.39	0.04
	V_τ	472	484	469	448	341	0.34	0.37
	Max	612	551	401	279	209	1.76	0.17
	Min	114	101	069	052	039	0.48	0.03
Summer 1994	$\bar{\tau}$	164	150	116*	086	064	1.28	0.05
	σ_τ	076	066	051	033	024	0.43	0.02
	V_τ	466	452	440	389	365	0.34	0.36
	Max	393	339	260	177	111	1.84	0.09
	Min	073	060	048	034	027	0.34	0.02
Summer 1995	$\bar{\tau}$	167	152	124	094	088	1.08	0.07
	σ_τ	068	057	041	038	052	0.44	0.04
	V_τ	406	375	328	399	591	0.41	0.51
	Max	356	301	217	197	280	1.76	0.16
	Min	062	070	039	045	041	0	0.04
Summer–fall 1997	$\bar{\tau}$	110	091	079	059	046	1.05	0.04
	σ_τ	060	047	033	022	016	0.69	0.01
	V_τ	543	513	420	369	345	0.66	0.31
	Max	192	172	134	098	079	2.02	0.07
	Min	029	026	027	021	013	−0.27	0.01

* Interpolated value.



a *b*

Fig. 15. Annual mean variability of the number density of aerosol particle with diameter > 0.4 μm from the data of airborne sounding in the layer up to 3 km (*a*) and from the data of aerosol monitoring in the near-ground atmospheric layer at the TOR-station (*b*).

6. Conclusion

Summarizing the results of this work, we would like to note that the effect of practically all significant geophysical factors, such as interannual cycles, annual behavior, synoptic scale processes, and diurnal transformation, manifests itself in the variability of aerosol weather in a particular region.

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References

1. G.V. Rozenberg and A.B. Sandomirskii, Izv. Akad. Nauk SSSR, Fiz. Atmos. Okeana **7**, No. 7, 737–749 (1971).
2. G.V. Rozenberg, Usp. Fiz. Nauk **71**, issue 2, 173–213 (1960).
3. G.V. Rozenberg. in: *Actinometry and Atmospheric Optics* (Gidrometeoizdat, Leningrad, 1961), pp. 9–15.
4. K. Bullrich, Advances in Geophys. **10**, 99–260 (1964).
5. C. Junge, *Air Chemistry and Radioactivity* (Academic Press, New York &London, 1963).
6. V.E. Zuev, *Atmospheric Transmittance for Visible and Infrared Rays* (Sov. Radio, Moscow, 1966), 317 pp.
7. G.V. Rozenberg, Izv. Akad. Nauk SSSR, Fiz. Atmos. Okeana **3**, No. 9, 936–949 (1967).
8. G.V. Rozenberg, Usp. Fiz. Nauk **95**, issue 1, 159–208 (1968).
9. K.Ya. Kondrat'ev, ed. *Atmospheric Aerosol and Its Effect on Radiative Transfer* (Gidrometeoizdat, Leningrad, 1978), 119 pp.
10. M.I. Budyko, G.S. Golitsyn, and Yu.A. Israel, *Global Climatic Catastrophes* (Gidrometeoizdat, Moscow, 1986), 160 pp.
11. G.I. Gorchakov, Izv. Akad. Nauk SSSR, Fiz. Atmos. Okeana **9**, No. 2, 204–210 (1973).
12. G.V. Rozenberg, Izv. Akad. Nauk SSSR, Fiz. Atmos. Okeana **19**, No. 1, 21–35 (1983).
13. V.A. Golovanov, G.I. Gorchakov, A.S. Emilenko, and V.N. Sidorov, Izv. Akad. Nauk SSSR, Fiz. Atmos. Okeana **7**, No. 12, 1318–1322 (1971).
14. Yu.S. Sedunov, ed., *Atmosphere (Handbook)* (Gidrometeoizdat, Leningrad, 1991), 510 pp.
15. M.V. Panchenko, S.A. Terpugova, A.G. Tumakov, et al. Atmos. Oceanic Opt. **7**, No. 8, 546–551 (1994).
16. S.D. Koshinskii, L.I. Trifonova, and Ts.A. Shver, eds., *Climate of Tomsk* (Gidrometeoizdat, Leningrad, 1982), 176 pp.
17. M.V. Panchenko and S.A. Terpugova, Atmos. Oceanic Opt. **7**, No. 8, 552–557 (1994).
18. V.Ya. Galin, *"Modeling of April circulation,"* Preprint No. 281, Computer Department of the Academy of Sciences of USSR, Moscow (1991).
19. Yu.A. Pkhalagov, V.N. Uzhegov and N.N. Shchelkanov, Atmos. Oceanic Opt. **5**, No. 6, 423–426 (1992).
20. V.L. Filippov, A.S. Makarov and V.P. Ivanov, Dokl. Akad. Nauk SSSR **265**, No. 6, 1353–1356 (1982).
21. L.T. Matveev, *Theory of General Circulation of the Atmosphere and Earth Climate* (Gidrometeoizdat, Leningrad, 1991), 295 pp.
22. I. Bluthgen, *Geography of Climates* [Russian translation] (Progress, Moscow, 1972), Vol. 1, 428 pp.
23. *"Control of the state of the atmosphere over Moscow. Part II,"* Preprint No. 9, Institute of Atmospheric Physics of the Russian Academy of Sciences, Moscow (1991).
24. G.P. Gushchin. *Methods, Instruments, and Results of Atmospheric Spectral Transmission Measurements* (Gidrometeoizdat, Leningrad, 1988), 200 pp.
25. M.V. Panchenko and S.A. Terpugova, Atmos. Oceanic Opt. **8**, No. 12, 977–980 (1995).
26. R.M. Hoff, W.R. Leaitch, P. Fellin, and L.A. Barrie, J. Geophys. Res. **88**, 10947–10956 (1983).
27. A.W. Hogan, S.C. Barnard, K. Kebschull, et al., J. Aeros. Sci. **15**, No. 1, 13–33 (1984).
28. A.A. Vinogradova, *"Arctic aerosol: pollution, microstructure, and optical properties (review),"* Preprint No. 2, Institute of Atmospheric Physics of the Russian Academy of Sciences, Moscow (1992).
29. I.N. Sokolik, Izv. Akad. Nauk SSSR, Fiz. Atmos. Okeana **28**, No. 7, 675–688 (1992).
30. V.I. Vorob'ev, *Synoptic Meteorology* (Gidrometeoizdat, Leningrad, 1991), 616 pp.
31. S.M. Sakerin, D.M. Kabanov, T.A. Eremina, et al., Atmos. Oceanic Opt. **11**, No. 10, 897–901 (1998).
32. V.V. Zuev, M.V. Panchenko, S.M. Sakerin, et al., in: *Proc. of 8th ARM Science Team Meeting*, Tucson (1998), pp. 865–869.
33. D.M. Kabanov and S.M. Sakerin, Atmos. Oceanic Opt. **9**, No. 6, 459–463 (1996).
34. V.N. Sidorov, G.I. Gorchakov, A.S. Emilenko, and M.A. Sviridenkov, Izv. Akad. Nauk SSSR, Fiz. Atmos. Okeana **20**, No. 12, 1156–1164 (1984).
35. L.S. Ivlev, in: *Problems of Atmospheric Physics* (Leningrad State University Publishing House, Leningrad, 1969), issue 8, pp. 125–160.
36. L.S. Ivlev and S.D. Andreev, *Optical Properties of Atmospheric Aerosol* (Leningrad State University Publishing House, Leningrad, 1986), 360 pp.
37. M.V. Panchenko and S.A. Terpugova, Atmos. Oceanic Opt. **9**, No. 6, 464–468 (1996).
38. M.V. Panchenko, V.V. Pol'kin, S.A. Terpugova, et al., Atmos. Oceanic Opt. **8**, No. 7, 579–580 (1995).
39. M.L. Asaturov, M.I. Budyko, K.Ya. Vinnikov, et al., *Volcanoes, Stratospheric Aerosol, and Earth Climate* (Gidrometeoizdat, Leningrad, 1986), 256 pp.
40. G.M. Vinogradova, N.N. Zavalishin, V.I. Kuzin, and P.Yu. Pushistov, Atmos. Oceanic Opt. **12**, No. 6, 475–479 (1999).
41. S.M. Sakerin and D.M. Kabanov, in: *Abstracts of Reports at 5th Workshop on Siberian Aerosols*, Tomsk (1998), pp. 38–39.
42. M.Yu. Arshinov, B.D. Belan, and T.K. Sklyadneva, in: *Abstracts of Reports at 5th Workshop on Siberian Aerosols*, Tomsk (1998), p. 15.