DETERMINATION OF THE STRUCTURE OF ATMOSPHERIC AEROSOL BY THE METHOD OF SPECTRAL TRANSPARENCY

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Abstract

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DETERMINATION OF THE STRUCTURE OF ATMOSPHERIC AEROSOL BY THE METHOD OF SPECTRAL TRANSPARENCY

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This article describes the results of processing data on the spectral transparency of the aerosol component of the thickness of a moist atmosphere, obtained under conditions of high transparency. The processing was carried out by the transparency method proposed in ([1]). This method makes it possible to determine the particle-size distribution without any hypotheses concerning the structure of the disperse system under study.

The experimental part of the work was carried out near Leningrad, in the settlement of Sosnovo, in the summer–autumn of 1951 over a period of 50 days. However, the results presented below pertain only to 15 days distinguished by comparatively high transparency during the prevalence of arctic air masses in the observation region. The actinometric transparency coefficient p, reduced to an optical mass $m = 2$ according to the tables ([2]), varies in the interval 0.81-0.84. Such high values of p occur rather rarely under the conditions of Leningrad Oblast.

Studies of spectral transparency were performed with a Feussner thermoelectric actinometer operating together with a galvanometer of sensitivity 10^{-9} a, and a set of narrow-band Zeiss interference light filters with $\lambda_{\max}$ equal to 372, 406, 464, 532, 650, 715, 760, 805, and 1013 μ. In processing the data, temperature effects on the apparatus were taken into account; the sensitivity and linearity of the actinometer-galvanometer system were monitored, and the transmission curves of the light filters were studied over a broad range before and after the experiment. In processing by the transparency method, absorption by ozone and oxygen was taken into account ([3],[4]). The measurement procedure is described in greater detail in ([5],[6]).

The aerosol optical thickness τ^*_λ was calculated from Bouguer' s formula

$$\tau^*_\lambda = \frac{1}{m} \ln \frac{I_{0\lambda}}{I_\lambda} - \tau^R_\lambda, \tag{1}$$

Fig. 1. Aerosol components of the atmospheric optical thickness, measured on 5 VIII (a, b) and 9 IX (c, d, e)

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where τ_λ^R is the Rayleigh optical thickness; $I_{0\lambda}$ is the intensity of light outside the atmosphere; I_λ is the observed light intensity. The values of $I_{0\lambda}$ were determined by extrapolating Bouguer straight lines beyond the atmosphere on stable days of observation (measurements were made up to $m \simeq 5$, in all spectral regions). The criterion of stability was the coincidence of Bouguer straight lines obtained before and after noon, with 30-40 series of observations per day. The use of the results of repeated measurements made it possible to reduce the influence of random error in determining τ_λ^* (the coefficient of variation for τ_λ^* was 0.5-0.7%).

In contrast to the monotonic increase of τ_λ^* with decreasing wavelength λ usually observed under conditions of a moist atmosphere, during all 15 days distinguished by high transparency a distinct maximum is found in the curve $\tau_\lambda^* = g^*(1/\lambda)$, varying in the range 0.4—

0.65 μ^*. Two typical curves $\tau_\lambda^* = g^*(1/\lambda)$ are shown in Fig. 1. They were obtained on 5 VIII and 9 IX before noon at $m = 4$.

It should be noted that the use of an actinometer with a field-of-view angle of 10° requires a careful analysis of measurement errors caused by the influence of the aureole, especially since the correction here has a spectral behavior. For this purpose, control measurements were carried out with special attachments having a variable field-of-view angle (from 10° to 3°). It turned out that, for the cases of high transparency used in the present work, the correction for the aureole is insignificant. Moreover, with a noticeable influence of the aureole there should have been a violation of the linearity of the Bouguer straight lines in the region of large m. In all the cases used by us, this effect was not detected.

Fig. 1. Aerosol components of the thickness of the atmosphere, measured on 5 VIII (a, b) and 9 IX (c, d, e)

For application of the transparency method it is necessary to have a reliable estimate of the quantity

$$c_0^* = \lim_{\lambda \to 0} g^*(1/\lambda). \tag{2}$$

At the same time, when measuring τ_λ^* in the atmosphere it is usually impossible to advance sufficiently far into the short-wavelength region because of the presence of ozone absorption bands. Therefore one has to extrapolate the curves τ_λ^* into the region $1/\lambda \to \infty$. The extensions of the curves τ_λ^* covering the range of possible extrapolations are indicated in Fig. 1 by dashed lines. By the transparency method we have (^{1,7})

$$m(a) \simeq \widetilde{m}(a) = -\frac{1}{\pi}\left\{\frac{\tau}{k}\sum_{j=1}^{k} g\left(\frac{x_j}{2}\right)\omega(ax_j) + \tau c_0 \omega_0(a\tau)\right\}, \tag{3}$$

$$\omega(y) = y\sin y + \cos y - 1, \qquad \omega_0(y) = \cos y - 2\sin y/y + 1, \tag{4}$$

$$r = ar_0, \qquad r_0 = \tau\lambda_{\min}/4\pi(n-1), \tag{5}$$

$$m(a) = \pi a^2 f(a), \qquad m^*(r) = r_0^{-2} m(a), \tag{6}$$

where $f(a)$ is the density of the particle distribution; a is the dimensionless radius; r_0 is the length scale; n is the refractive index; $\lambda_{\min}$ is the wavelength corresponding to the emergence of $g^*(1/\lambda)$ onto the asymptote; $g(x/2)$ and c_0 are the dimensionless analogues of $g^*(1/\lambda)$ and c_0^*. We took $\tau = 8$, $k = 40$, and $n = 1.5$. Figure 2 presents the area-normalized functions $m^*(r)$—the results of inverting the curves τ_λ^* shown in Fig. 1; moreover, all variants of the curves $m^*(r)$ corresponding to different extrapolations of τ_λ^* are indicated. Processing of optical information obtained on other days with high transparency leads to analogous results.

Very similar curves of the particle size distribution were obtained in the USA in April–June 1962 by R. Fenn (^{8}). In this work, [[unclear: sentence continues on next page]]

measurements of aerosol particles were carried out directly, with the aid of an impactor of the most modern design (ˆ9). A comparison of the distribution curves from (ˆ8) and in Fig. 2 reveals a noticeable similarity. This can be seen from the following data, for typical ranges of particle radii corresponding to the extrema of the distribution curves.

r, μ	$r_{1\,\max}$	$r_{1\,\min}$	$r_{2\,\max}$	$r_{2\,\min}$	$r_{3\,\max}$	$r_{3\,\min}$	$r_{4\,\max}$
From (ˆ8)	0.1–0.15	0.2–0.3	0.3–0.37	0.45–0.5	0.55–0.65	–	–
From Fig. 2	–	–	0.3–0.35	0.5–0.55	0.6–0.7	0.75–0.85	1.0–1.3

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* In (5), neutral scattering of light was ascribed to the aerosol component of Arctic air masses, which is too crude an approximation.

Fig. 2. Distribution of aerosol particles on 5 VIII and 9 IX

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Thus a stable position of the extrema is obtained for small particle radii and a less stable one for large radii. The values $r_{2\,\max}$, $r_{2\,\min}$, and $r_{3\,\max}$ from (⁀8) and from Fig. 2 agree well with one another. At the same time, we did not detect $r_{1\,\min}$ and $r_{1\,\max}$. In addition, processing by the transparency method gives the extrema $r_{3\,\min}$ and $r_{4\,\max}$ (particles with such radii were not found in (⁀8)). It should be taken into account that the measurements in the two cases were performed at different points, at different times, and referred to the lower layers of the atmosphere (in (⁀8)) and to the entire thickness of the atmosphere (in the present work). Let us add that the author in (⁀8) shows that the data he obtained on the multimodality of the aerosol may be referred to the entire thickness of the atmosphere as a whole.

Fig. 2. Distribution of aerosol particles on 5 VIII and 9 IX

The multimodality of the distributions of atmospheric-aerosol particles established by us is confirmed by direct studies of the chemical composition and sizes of aerosol particles, recently carried out by É. Mészáros (⁀10). As a result of systematic measurements, both ground-based and aircraft, he established that in the region of Budapest the aerosol consists of three substantially different groups: 1) insoluble Aitken nuclei ($r \lesssim 0.1\,\mu$), 2) soluble chloride and sulfate nuclei (with a clear maximum near $r \simeq 0.4\,\mu$), 3) soluble calcium particles (also with a maximum around $r \simeq 0.7$—$1\,\mu$). These results are very close to the data given above.

The use of the transparency method made it possible to obtain important information on the distribution of the radii of atmospheric-aerosol particles by means of simple apparatus. The types of distribution of atmospheric-aerosol particles obtained under different conditions, by substantially different methods, proved to be very close.

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Note: Figure translations are in progress. See original paper for figures.

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