RECONSTRUCTION OF THE SCATTERING COEFFICIENT IN THE LOWER TROPOSPHERE USING GROUND–BASED MEASUREMENTS

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Abstract

The model proposed provides for reconstruction of the aerosol scattering coefficient profile. In so doing one needs for, as input data, the scattering coefficient value of the dry aerosol substance at the near ground level measured at the wavelength of 0.52 μm, relative air humidity, aerosol optical thickness, and mean temperature of air in the low troposphere. The model assumes the account for correlation between the values of scattering coefficient at different altitudes, as well as the dependence of the height of mixing layer on the heating of the lower atmospheric layers. The errors in reconstructing the vertical profiles of the aerosol scattering coefficient are analyzed for different ways of taking into account the external factors and input parameters. It is shown that the use of such an approach provides a decrease in the rms error in reconstruction even at this stage approximately by 30% for winter and by 3 to 4 times for summer in comparison with the rms deviation of this parameter in the initial data sets.

1. **Introduction.**

It is known that the basic principles of any model are closely related to the model purpose and depend on the bulk and quality of information, either experimental or theoretical, available. The main purpose of this paper is creation of a model of aerosol optical characteristics.

Vast experimental data on the aerosol characteristics were obtained using airborne sounding of the atmosphere. The factors affecting the variability of the submicron aerosol characteristics in the lower troposphere were analyzed (Panchenko and Terpugova, 1994, 1995, 1996). It allows us to start the development of an empirical regional model of aerosol optical parameters in the altitude range up to 5 km.

Based on the practice of aerosol studies (Pandis and Seinfeld, 1998) one can isolate two basic approaches to the development of empirical aerosol models that may conditionally be called the optical and microphysical ones (Veretennikov et al 1986; Kabanov et al 1988). Naturally, each of the approaches has its own advantages and drawbacks, as applied to estimation of optical parameters. Thus, the microphysical approach provides for a possibility of calculating any of the optical parameter of aerosol using some theoretical grounds. However, in this case the question on possible errors in calculation of the optical characteristics remains to be addressed. Actually, when setting empirically the size-distribution function of aerosol particles and optical constants of the particulate matter (that is a difficult task alone) it is hard to assess the contribution of the particles that have not been measured. As a rule, there are no data on the most fine aerosol (nucleation) mode as well as a significant part of the coarse aerosol fraction because of the instrumental limitations (especially in airborne conditions).

The optical approach is, in principle, free of these problems, but, at the same time, it is restricted only by the optical parameters measured and the spectral region used in the experiments.

The widely used models of the aerosol optical properties LOWTRAN (Kneizys et al, 1988) allow for the stratification of aerosol microphysical parameters with height, as well as admit an important part the relative humidity of air plays in the transformation of the aerosol optical properties. However, these models assume only two contrast seasons, the winter and summer. The models provide rough regard for the height of mixing layer and meteorological visual range, and they completely ignore the possibility of taking into account the factors that may vary within a season. The development of a more dynamical model can only be done at the expense of increased number of input parameters that, as a result, would lead to a decrease in the scale of averaging. However, too fine fractioning would make the model to be of no climate and geophysical significance. For these reasons, we believe that the scale of an air mass could be most appropriate because of the integrity of the aerosol processes within the same air mass. It is also very important that any particular geographical region has its own sources of aerosol, the statistics of occurrence of the pressure systems and air mass, as well as the air mass trajectories that are characteristic only of this region. In other words, the models that could be more dynamical and correctly describing the aerosol situation are those that are applicable on a regional scale.

2. Strategy of the experiment.

Our approach is based on the following considerations.

The scattering coefficient is one of the aerosol characteristics that is most sensitive to variations in microphysics of the sub-micron aerosol (that, according to G.V. Rosenberg et al (1980) are «memorized» by the atmosphere).

The variations of the aerosol scattering coefficient in the atmosphere are governed by two basic processes:

- the first one is natural variability in the content of dry aerosol substance (or in other words by the relatively long processes of generation, aging, accumulation, and sinking of the particulate matter in the atmosphere),
- the second one is the transformation of aerosol microphysical properties under the action of the air relative humidity, the main manifestation of which is diurnal behavior. It is evident that those two types of processes are regulated by the geophysical factors of different spatio-temporal scales.

Hence, in order to correctly assess the role of one or another factors in the variability of aerosol properties (and for a more efficient incorporation of those variations into the dynamic aerosol models) one should separately measure the characteristics of dry aerosol matter and its response to variations of the air relative humidity.

Of course, it would be an ideal situation if one can measure angular behavior of the scattering phase matrix elements in different spectral ranges, but unfortunately we had no sufficient funding and equipment to design such an instrument that could withstand the long-term exploitation onboard an aircraft. Therefore, in our studies of the aerosol scattering coefficient by the nephelometric method, we have used an airborne instrument equipped with an attachment capable of artificially regulating the relative humidity of air samples and heating the aerosol particles (for details see Zuev et al, 1992).

3. Brief analysis of the experimental data set.

When flying over West Siberia, we have compiled an array of 602 vertical profiles acquired in 1986–1988 in different seasons and under different meteorological and synoptic situations. In order to avoid the obvious effect of anthropogenic factors, the data measured just near large industrial centers and towns were excluded from consideration. Measurements of the aerosol characteristics were carried out, as a rule, in the height range from 0 up to 5 km. But in some cases, the maximum height was lower because of the weather conditions. So, there are less data obtained at the altitudes above 3 km. Then the data array formed has been divided into seasonal sub-arrays according to climatic criteria of seasons (Koshinskii et al, 1982) characteristic of the region under study. For each season we have analyzed the frequency of occurrence of different situations depending on type of air mass and pressure system. As analysis has shown, the frequencies of occurrence obtained well agree with the data of long–term observations. In a similar way we have also compared the vertical profiles of the meteorological parameters measured with the data on long–term mean profiles measured at the aerological network (Guterman, 1980).

The agreement of synoptic and meteorological characteristics of the atmosphere during the observation period with the climatic mean ones enables us to consider the array of data on aerosol properties compiled to be representative of this geographical region as well.

4. Selection of the input parameters.

Let us firstly note some methodical aspects. Statistical models may only be few-parameter ones. Indeed, the variability of the aerosol optical characteristics that is memorized by a model is capable of representing only the most frequent situations that occur under the action of the variety of external aerosol and weather factors which, in their turn, are interrelated in a complex way.

We took as the first and the basic one the natural parameterization principle of dividing the data arrays into seasonal sub-arrays. Indeed, the conditions of the underlying surface, composition and power of the sources of particles, and the capacity of sinks vary from season to season. Therefore, dividing the data array into seasonal sub-arrays, we can take into account the differences in these factors in the empirical model.

Further analysis of the data obtained showed that no additional subdivision, for example, into monthly sub-arrays is useful. The matter is that the monthly average values of meteorological parameters, synoptic and weather situations in the region under study undergo significant variations from year to year (Koshinskii et al 1982). It is worth mentioning in this connection that the time of covering the underlying surface with snow and rivers with ice as well as the reverse processes occur in different calendar times every year.

As analysis of the variability of the vertical profile of the scattering coefficient of the dry submicron aerosol matter showed, the account for types of air masses and meteorological parameters can be used for the model parameterization (Panchenko and Terpugova 1994, 1995, 1996).

At the same time, our knowledge reached to date about the «aerosol weather» in the specific region does not enable us to increase the accuracy of estimation of the optical characteristics at the use of only synoptic or meteorological parameters. The increase of accuracy is possible only at attracting the measured data on the aerosol characteristics.

Actually, the «aerosol weather» at a concrete point of observations and moment is determined not only by synoptic and meteorological factors, but also by the whole complex of global and local, inherently aerosol, factors of natural and anthropogenic origin (Jaenicke, 1980; Kabanov and Panchenko, 1984; Pandis and Seinfeld, 1998).

The development of an aerosol model that could provide for the account and forecasting of the whole variety of factors, would require many-year experiments and a wide measurement network for monitoring a large number of atmospheric parameters including the aerosol ones as well as advanced climate and weather models.

From this point of view, it seems to be reasonable that the development of an aerosol model uses, at the first stages, as most important input parameters, those aerosol parameters that are measured *in situ* and which carry qualitative and quantitative information on the state of a particular aerosol.

5. Scheme for reconstructing values of the scattering coefficient in the lower troposphere.

The general scheme for reconstructing vertical profiles of the scattering coefficient at the wavelength $\lambda = 0.52$ μm is based on the principles of the hierarchy of external factors and many ways for taking into account the *a priori* information and input parameters measured. The general scheme of reconstruction is shown in Fig. 1.

One can divide it into three blocks. The first block makes it possible to obtain the information on the mean values and variances of the parameters determined in the altitude range up to 5 km, only by setting the external factors. The following external factors (input parameters of this block) have been selected: season, type of air mass, and time of a day. This block contains the information on the mean values and variance of the vertical profile of the scattering coefficient $\sigma_d(H)$ of the dry matter of aerosol particles, *in situ* scattering coefficient $\sigma_m(H)$, temperature $T(H)$, relative humidity of air

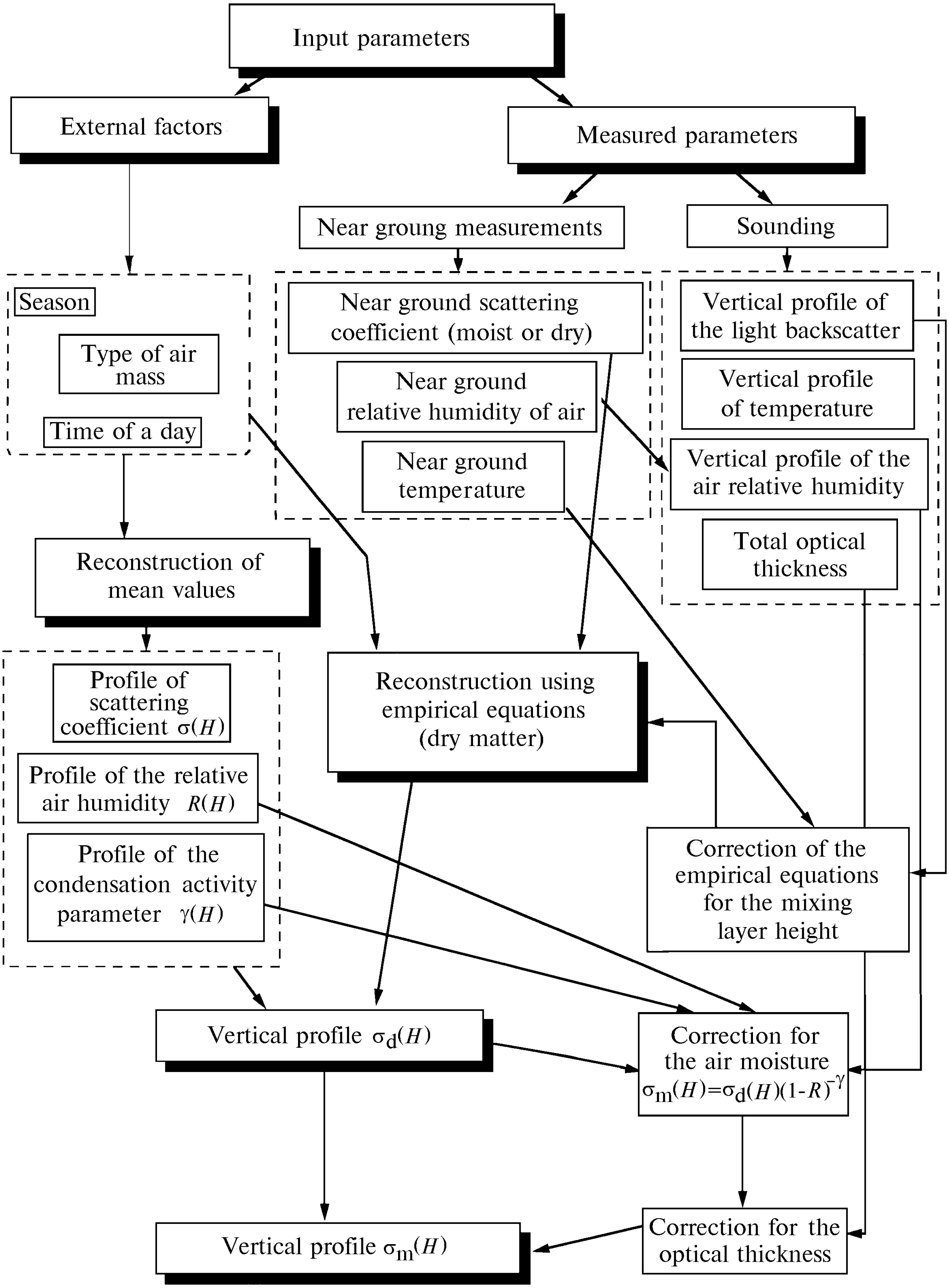

Fig. 1. Scheme of reconstructing of the vertical profile of the scattering coefficient.

$R(H)$, specific humidity $Q(H)$, and the parameter of condensation activity $\gamma(H)$. We have not revealed any significant differences in the parameter $\gamma(H)$ in different air masses and different time of a day due to statistically poor bulk of data, so for the input parameter we used only seasons.

As it was shown earlier, (Panchenko and Terpugova, 1994) natural variations of the scattering coefficient in seasonal data arrays are approximately 100% of the mean value. So, the use of any seasonal mean characteristics is hardly acceptable for more accurate estimates. At the same time, the data on mean vertical profile of aerosol characteristics can be useful for different kind of climatic calculations. As it was mentioned above, our results are in good agreement with the long-term average synoptic and meteorological data for the region under investigation (Panchenko et al, 1994).

The second block of the scheme proposed is connected with the parameters to be measured.

One can conventionally select two groups, among the great amount of optical and meteorological characteristics, that can be measured (meteorological parameters can be calculated in the case of climatic modeling) and used as input parameters. These are the data that can be obtained in the near-ground atmospheric layer and the data on vertical profiles or the total optical thickness obtained with the ground based systems (they are united in the group «sounding» in Fig. 1).

At this stage of creating the reconstruction scheme, we tried to include and test, as input parameters, only the characteristics that are most easy for measurements. In the group of "ground based measurements" there are the scattering coefficients ("dry" or "moist"), temperature and relative humidity of air. For further development of the model, one should include measurements of the atmospheric transmission in a wide wavelength range, as well as the data on the scattering phase matrix components. In the group "sounding" we consider the possibilities of obtaining data on the vertical profiles of meteorological (aerological sounding), use of photometers for measuring the optical thickness, as well as the lidars. As the analysis of autocorrelation matrices (Panchenko and Terpugova, 1994) of the scattering coefficients $\sigma_d(H)$ and $\sigma_m(H)$ show, the measurement data on σ at some altitudes, for example, above and below the mixing layer, can be used for reconstruction of the entire vertical profile most effectively. But to obtain such data one should have a mobile carrier such as aircraft-laboratory, helicopter or a balloon with the instruments onboard it, i.e. the use of quite expensive tools, that limits the possibility of their routine use in practice of aerosol observations. It is clear, that to develop a model for a wide wavelength range it is necessary to include the data on the spectral optical thickness and solar aureole measurements.

Inclusion of the lidar data into the list of measured parameters, we had in our minds the following ideas. The development of lidar facilities is aimed at obtaining a complete data set on the vertical profiles of different atmospheric parameters, including the aerosol characteristics, directly from the data of sounding. At the same time, known problems in solving the lidar sounding equation determine the need for *a priori* information for interpretation of the lidar data. In this case, the use of the reconstruction scheme proposed will be useful. Depending on the information available, the lidar data can be interpreted by using the scheme in different ways. For example, when using Raman lidars that are capable of simultaneously obtaining the data on humidity, temperature and backscattering coefficient profiles, the scheme proposed is fully applicable. In the case of single-frequency lidars operating in a monostatic mode, it is quite useful to have information on the near-ground values of optical characteristics and on the total optical thickness, and then to reconstruct the profile in the entire altitude range.

Detailed description of the techniques for reconstructing the profile from the data of lidar sounding is an independent problem that is out of the scope we concern with in this paper. So let us limit ourselves only by the aforementioned ideas, and let us use the parameters that were determined in our airborne experiments when testing the scheme proposed.

The third, of the principal blocks of the scheme, is directly connected with the reconstruction of the vertical profile of aerosol scattering coefficients. Let us note that at the stages of testing the operation ability of the scheme the principal block "reconstruction using empirical equations" and the blocks of reconstructed meteorological parameters contain the empirical functions, the parameters of which are determined in the frameworks of linear empirical equations. A more detailed analysis of the relations and selection of an optimal combination of the input parameters are the problem at the next stage of creating the model, and are not considered in this paper.

The block of reconstruction assumes the following sequence of procedures:

By setting the external factors, i.e. season, air mass and time of a day (in a necessary combination), we enter into the system of empirical equations. If no information is available, additionally only the mean values of the vertical profile of the scattering coefficient of the dry matter of aerosol particles $\sigma_d(H)$ are the outcome of this block.

If data on the near-ground values of the scattering coefficient (moist *"m"* or dry *"d"*) are available, the input into this block is done through the value of the "dry" scattering coefficient $\sigma_d(0)$ (if the near-ground measurements have been carried out for σ_m, the value $\sigma_d(0)$ is first calculated using the Kasten-Hanel formula (Hanel, 1976)

$$\sigma_m = \sigma_d\,(1 - R)^{-\gamma}, \tag{1}$$

where R is the relative humidity of air, and γ is the parameter of the condensation activity.

Then the vertical profile $\sigma_d(H)$ is reconstructed using linear empirical equations of the form

$$\sigma_d(H) = K(H)\,\sigma_d(0) + C(H)\,, \tag{2}$$

where $K(H)$ and $C(H)$ are the empirical coefficients for corresponding (by the external factors) data arrays.

Table I gives the values of the coefficients of the empirical equation (2) for different seasons.

TABLE I. Coefficients of the empirical equation.

Winter			Spring			Summer			Fall		
H, km	*K*(*H*)	*C*(*H*)	*H*, km	*K*(*H*)	*C*(*H*)	*H*, km	*K*(*H*)	*C*(*H*)	*H*, km	*K*(*H*)	*C*(*H*)
0	1	0	0	1	0	0	1	0	0	1	0
0.4	0	0.072	0.4	0.57	0.024	0.6	0.58	0.011	0.4	0.32	0.016
1.2	0	0.028	1.4	0.23	0.014	2.8	0.30	–0.0054	1.2	0.10	0.022
2.6	0	0.0077	2.4	– 0.06	0.026	3.4	0.081	0.003	2.4	0	0.0065
5	0	0.004	5	– 0.094	0.024	5	0	0.0026	5	0	0.0028

If the data on air temperature are available (either near-ground values $T(0)$ or a vertical profile $T(H)$), the empirical equations (2) are corrected for the height of the mixing layer. In the case when only the near-ground value $T(0)$ has been measured, the daily average value $\overline{T}(0)$ is determined taking into account the mean daily behavior of temperature in the corresponding array of data. If the measurements of the vertical profile of temperature $T(H)$ have been performed in some layer at the height $\tilde{H}$, the integral value $\overline{T}_{\tilde{H}}$ is calculated. Then, the height of the mixing layer H_c ("correlation" layer in our case) is determined from these data (Panchenko and Terpugova, 1994):

$$H_c = 0.75\exp[(4.5\times10^{-3}\tilde{H} + 6.5\times10^{-2})\overline{T}_{\tilde{H}} + 0.2\tilde{H}]\,, \tag{3}$$

where $\overline{T}_{\tilde{H}}$ is the average temperature of the atmospheric layer of the height $\tilde{H}$.

Then, the coefficients $K(H)$ and $C(H)$ of Eq. (2) are corrected. Correction is carried out as follows. If $H_c < \overline{H}_c$ ($\overline{H}_c$ is the seasonal mean value of the "correlation" layer height), then:

– $K(H)$ and $C(H)$ are assumed constant up to the height $H = H_c - 200$ m;

– the "upper" parts of the dependencies $K(H)$ and $C(H)$ are extrapolated from the height $\overline{H}_c$ down to the height $H_c + 200$ m;

– linear interpolation between "lower" and "upper" parts of the dependencies $K(H)$ and $C(H)$ is carried out on the height interval ΔH = 400 m (i.e. from H_c – 200 m to H_c + 200 m).

If $H_c > \overline{H}_c$:

– the "lower" parts of the dependencies $K(H)$ and $C(H)$ are extrapolated from the height of the seasonal mean mixing layer $\overline{H}_c$ up to the height H_c – 200 m;

– linear interpolation is performed on the portion from H_c – 200 m up to H_c + 200 m;

– $K(H)$ and $C(H)$ are not changed on the interval from H_c + 200 up to 5 km.

Vertical profile $\sigma_d(H)$ is calculated after the correction of the equation coefficients.

Then the scattering coefficient values calculated for the dry aerosol base are reduced to relative humidity values at the corresponding height by formula (1). There are three options to do this. If no information on $R(H)$ profiles is available one may use the corresponding season mean profile. In the case when a researcher has at his disposal only the data on near-ground values of the relative air humidity then it is possible to reconstruct, using some empirical relations like equation (2), the $R(H)$ profile by making use of the correlation between $R(H)$ and $R(0)$. And, finally, if data on the $R(H)$ profile are available, for instance as measured with a radiosonde, one uses the true profile of the relative air humidity. The value of the condensation activity parameter γ is either the set based on the external factors or its mean value $\overline{\gamma}$ is used (in particular, we took the value $\overline{\gamma}$ = 0.5 in all examples presented below).

At the final stage, the profile of scattering coefficient is to be corrected using the optical thickness value. In the case when only τ is known, and measurements of the parameters are impossible, the seasonal mean profile $\overline{\sigma}(H)$ is recalculated proportionally to the ratio $\overline{\tau}/\tau_{meas}$ (where $\overline{\tau}$ and τ_{meas} are the seasonal mean and measured values of the optical thickness, respectively). At the presence of other measured parameters in the subsequence of operation we have chosen the value τ_{rec} is calculated from the reconstructed profile σ_m(H) in the height range up to 5 km.

The values of $\overline{\Delta\tau}_5$ of the remainder of the optical thickness of the height range above 5 km was estimated using the LOWTRAN–7 model for continental mid-latitudes (Kneizys et al, 1988). Three variants can be considered here: background stratosphere, extreme and moderate volcanic eruptions. The optical thickness was calculated by means of integrating the profile of the scattering coefficient $\sigma(H)$ from 5 to 100 km. The respective values of the remainder of the optical thickness $\overline{\Delta\tau}_5$ are shown in Table 2.

TABLE 2. The values of the remainder of the optical thickness of the layer H > 5 km for different stratospheric conditions (calculated from the data of Kneysis et al, 1988)

Fall–winter		Spring–summer	
Stratospheric conditions	$\Delta\tau_5$	Stratospheric conditions	$\Delta\tau_5$
background	0.0236	background	0.040
moderate volcanic	0.0463	moderate volcanic	0.0635
extreme volcanic	0.100	extreme volcanic	0.140

Then the value of the optical thickness of the molecular atmosphere τ_M, seasonal mean value, the remainder $\overline{\Delta\tau}_5$, and the value of the optical thickness reconstructed using our scheme τ_{rec} are subtracted from the measured value τ_{meas}, i.e.

$$\Delta\tau = \tau_{\mathrm{meas}} - \tau_{\mathrm{M}} - \overline{\Delta\tau}_{5} - \tau_{\mathrm{rec}}\,. \tag{4}$$

Vertical profile of the coefficient of correction for τ is calculated in the following form:

$$K_{\tau}(H) = \frac{1}{\tau_{\mathrm{rec}}} \int\limits_{H-\Delta h}^{H+\Delta h} \sigma_{\mathrm{rec}}(h)dh\,. \tag{5}$$

Then the vertical profile of the scattering coefficient is corrected:

$$\sigma(H) = \sigma_{\mathrm{rec}}(H) + K_{\tau}(H)\,\Delta\tau\,/\,2\Delta h\,. \tag{6}$$

Unfortunately, our data array was obtained without regular parallel measurements of the optical thickness. So for testing the model, the "quasi-measured" values τ^{*}_{meas} of the optical thickness were considered, which were obtained by integration of the measured profile of the scattering coefficient :

$$\tau^{*}_{\mathrm{meas}} = \int\limits_{0}^{5km} \sigma_{\mathrm{meas}}(H)dH\,. \tag{7}$$

The correction was calculated as follows:

$$\Delta\tau = \tau^{*}_{\mathrm{meas}} - \tau_{\mathrm{rec}} \tag{8}$$

Then the profile of the scattering coefficient was corrected by formulas (5) and (6).

Let us note that selection of the optimal way of using τ is a different problem to be discussed in the future papers, and its statement needs for certain experimental verification.

6. Analysis of the reconstruction errors.

For estimation of the error in reconstructing the scattering coefficient, the absolute and rms errors were calculated. Absolute error was calculated by the formula:

$$\Delta\sigma = \sigma_{\mathrm{rec}} - \sigma_{\mathrm{meas}}\,, \tag{7}$$

where σ_{rec} and σ_{meas} are the reconstructed and measured values of the scattering coefficient, respectively.

The rms error $\varepsilon(H)$ at a certain height was calculated by the formula:

$$\varepsilon(H) = \sqrt{\frac{1}{N}\sum_{i}\left(\sigma_{\mathrm{rec},i}(H) - \sigma_{\mathrm{meas},i}(H)\right)^{2}}\,, \tag{8}$$

Then, the rms deviation of the initial data array was calculated by the formula

$$\varepsilon_{0}(H) = \sqrt{\frac{1}{N}\sum_{i}\left(\sigma_{\mathrm{meas},i}(H) - \overline{\sigma}(H)\right)^{2}}\,, \tag{9}$$

where $\overline{\sigma}$ is the mean value of the scattering coefficient at the height H.

Figure 2 presents the ratio profiles $\varepsilon/\varepsilon_0$ between the rms error of $\sigma(H)$ reconstruction and the rms deviation of the initial data array for each season.

The profiles of this ratio are shown in a succession of the increasing number of input parameters. We consider the following versions of the reconstruction scheme. 1) The first one is in reconstructing the profile of dry aerosol base using a near-ground value $\sigma_d(0)$ with the following 'moistening' using a season mean profile of relative humidity $\overline{R}(H)$. 2) The second way assumes reconstructing «dry» aerosol profiles and the profile of relative humidity via the respective near-ground values. 3) The third option assumes, in addition to the procedure under previous point, the account for the mixing layer height based on mean temperature of the layer from the ground and up 3-km height. 4) In the fourth version the aerosol optical thickness is being taken into account, besides the factors mentioned in the third version. 5) The reconstruction scheme in this version is the

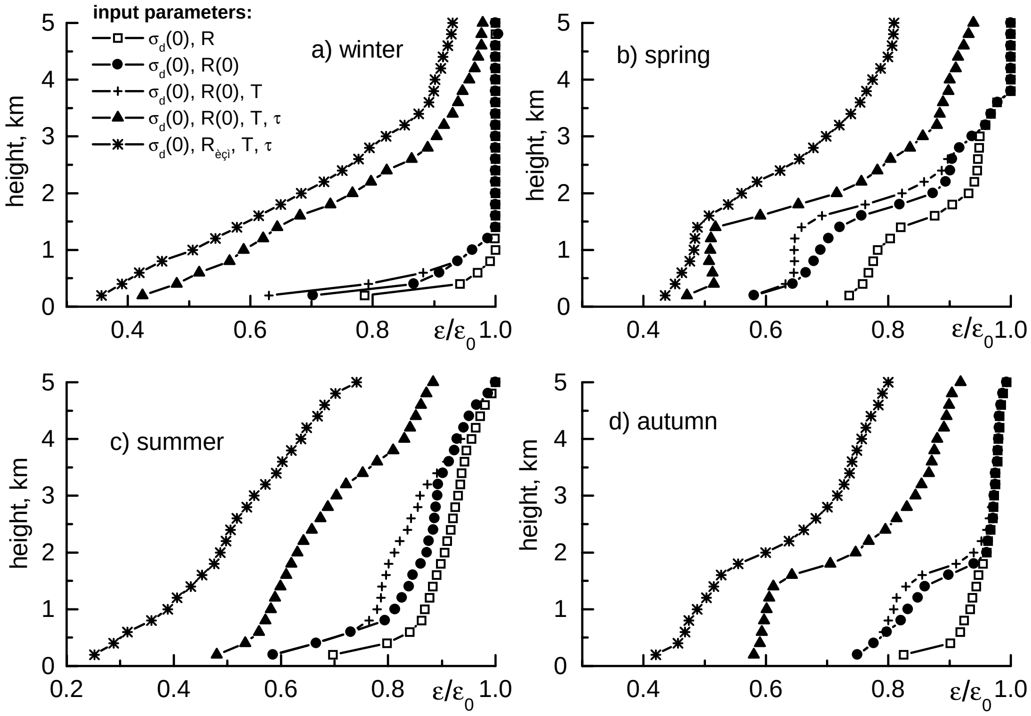

Fig. 2. Ratio of the rms error of reconstruction to the rms deviation of the initial data array.

same as the previous one except that «moistening» of the «dry» scattering coefficients is performed using measured profile of the relative humidity, $R_{\text{meas}}(H)$.

To illustrate the distribution of the absolute reconstruction errors $\Delta\sigma$, we show in Fig. 3 the histograms of these errors that are characteristic of summer season.

For this illustration, we have chosen the atmospheric layers at three specific heights. The layer at 1 km height is taken because it is within the mixing layer, the layer at 3 km is at the height of season mean upper boundary of the mixing layer, and the layer at 4 km height is already in the free atmosphere. Histograms that are shown in the first column represent the errors when no measurement data are used and the season mean profile $\sigma(H)$ is taken as the reconstructed one. Other columns show the error histograms for the $\sigma(H)$ reconstruction versions used to obtain data shown in Fig. 2. It is seen from Fig. 3 that at all heights an increase in the number of input parameters used in reconstruction results in a narrower distribution of the reconstruction errors that becomes more close to the normal one. If reconstruction of the profile is being performed using only one input parameter, namely the near-ground value of the scattering coefficient, we obtain a very asymmetric histogram of errors at the height within the mixing layer. Moreover, at the height near the mixing layer top the error distribution may even become a bimodal one. By introducing into the reconstruction scheme the mean temperature of the low atmospheric layers or the temperature of the near-ground layer, one may calculate the mixing layer height for each individual profile. As a result, the rms error of reconstruction not only falls off but, in addition, becomes closer to the normal view. An increase in the accuracy of reconstructing the aerosol scattering coefficient at the heights in the free atmosphere can only be achieved when taking into account the aerosol optical thickness. Similar behavior of the reconstruction error with the increasing number of input parameters may be demonstrated for other seasons.

Of course, it is quite clear that the errors of reconstruction are too high and, as a consequence, thus reconstructed aerosol characteristics can hardly be used for some operative and accurate assessments. Moreover, the absence of knowledge of the condensation activity of the aerosol at different heights and at any concrete time is one more source of errors that may be essential in magnitude. In this study we have used only its average value $\gamma = 0.5$. As our earlier airborne studies

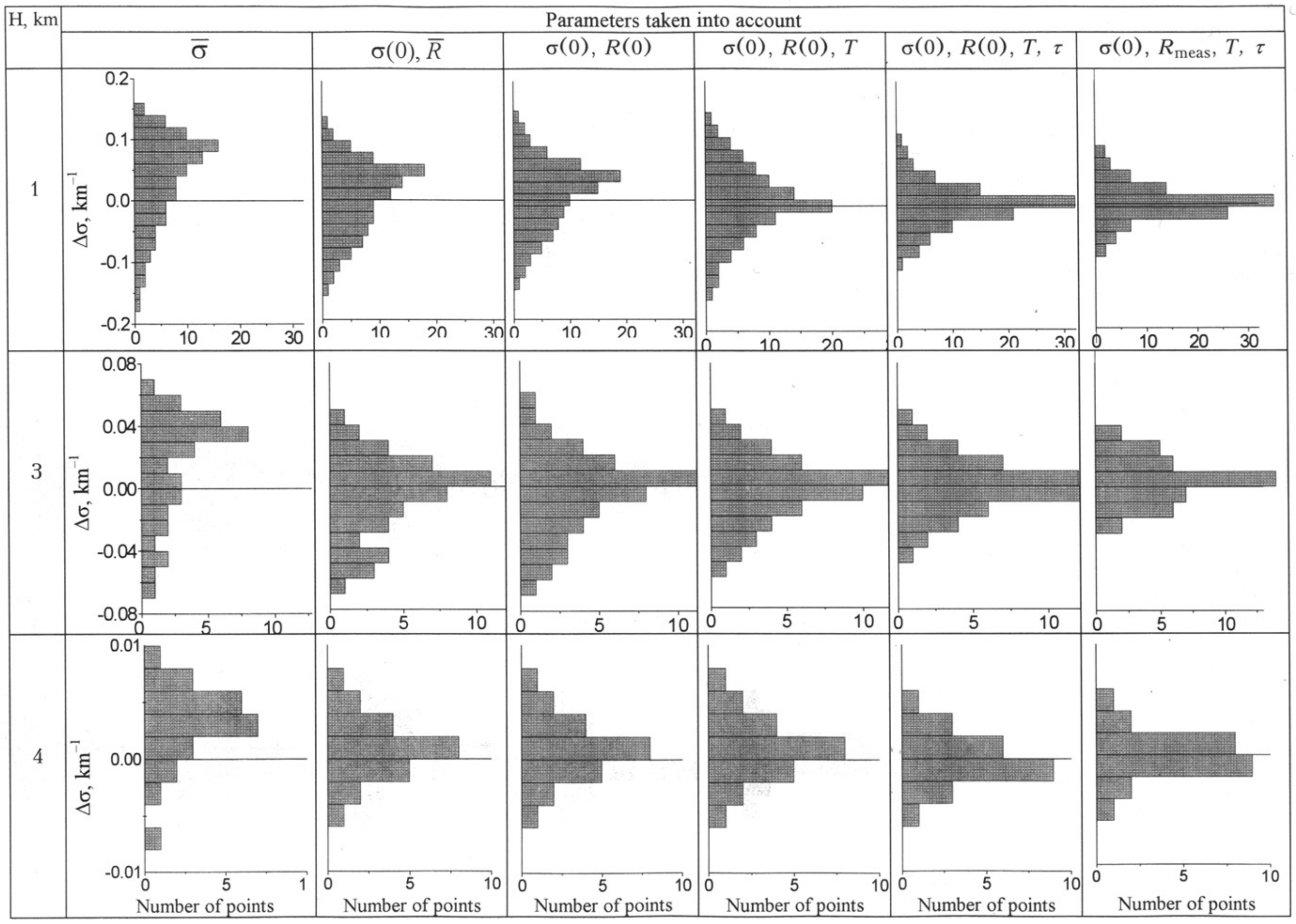

Fig. 3. Histograms of the absolute errors of reconstruction.

have shown, the value γ may vary from case to case having also certain seasonal peculiarities and some vertical behavior (Panchenko et al 1996). However, the bulk of experimental material compiled up to now on the behavior of this parameter is yet insufficient for reliably parameterizing it and thus we could not involve it into our scheme of reconstruction.

At the same time, it is clearly seen that even at this stage of the model development this approach enables reconstructing the aerosol scattering coefficient in the height region from 0 to 5 km with the accuracy that can hardly be achieved, at the present time, by other available models. Thus, for summer conditions the use of a scheme that accounts for the near-ground values $\sigma_d(0)$, temperature profile $T(H)$, and the aerosol optical thickness τ allows the uncertainty in $\sigma(H)$ estimates to be decreased by two to three times, as compared to the case of using the season mean value $\overline{\sigma}(H)$.

7. Conclusion.

Based on data of airborne sounding of the atmosphere we have developed an empirical model of optical properties of aerosol in the low troposphere over West Siberia. Since in so doing we have revealed certain leading factors that cause the variability of aerosol particles content along vertical direction, the applicability of this approach most likely is not bounded by the West Siberian region only. In our opinion the basic empirical relationships proposed in this paper for estimating the optical characteristics of aerosol can be used to interpret data acquired over similar geographic regions.

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REFERENCES

Guterman, I.G. (Ed.) (1980) *New Aeroclimatic Handbook of the Free Atmosphere over USSR.* Gidrometeoizdat, Moscow (in Russian).

Hänel, G. (1976) The properties of atmospheric aerosol as function of the relative humidity at the thermodynamic equilibrium with the surrounding moist air. *Adv. in Geophys.* **19**, 74–183.

Jaenicke, R. (1980) Atmospheric aerosol and global climate. *J. Aeros. Sci.* **11**, 577–588.

Kabanov, M.V. and Panchenko, M.V. *Scattering of Optical Waves by Disperse Media. Part 3. Atmospheric Aerosol*. Tomsk Affiliate of the Academy of Sciences of USSR, Tomsk (in Russian).

Kabanov, M.V., Panchenko, M.V., Pkhalagov, Yu.A et al. (1988) *Optical Properties Of The Coastal Atmospheric Hazes*. Nauka, Novosibirsk.

Kneizys, F.X., Shettle, E.P., Abreu, L.W. et al. (1988) *User Guide to LOWTRAN-7.* AFGL-TR-88-0177, Air Force Geophysics Laboratory, Environmental Research Papers No. 1010.

Kondrat'yev, K.Ya., Moskalenko, N.I. and Pozdnyakov, D.V. (1983) *Atmospheric Aerosol.* Gidrometeoizdat, Leningrad (in Russian).

Koshinskii, S.D., Trifonova, L.I. and Shver, T.A. (Eds) (1982) *Climate of Tomsk.* Gidrometeoizdat, Leningrad (in Russian).

Panchenko, M.V., Terpugova, S.A., Tumakov, A.G., Belan, B.D. and Rasskazchikova, T.M. (1994) Some aspects of a technique for airborne nephelometric studies of the tropospheric aerosol on a regional scale. *Atmos. Oceanic Optics*, **7**, 546–551.

Panchenko, M.V. and Terpugova, S.A. (1994) Annual behavior of the submicron aerosol fraction in the troposphere over West Siberia. *Atmos. Oceanic Optics*, **7**, 552–557.

Panchenko, M.V. and Terpugova, S.A. (1995) Seasonal factors of the variability of the submicron aerosol characteristics. I. Air masses. *Atmos. Oceanic Optics*, **8**, 977–980.

Panchenko, M.V. and Terpugova, S.A. (1996) Seasonal factors of the variability of the submicron aerosol characteristic. II. Diurnal behavior (vertical profile). *Atmos. Oceanic Optics*, **9**, 464–468.

Panchenko, M.V., Terpugova, S.A. and Tumakov, A.G. (1996) Annual Variations of Submicron Aerosol Fraction as Assessed from the Data of Airborne Nephelometric Measurements. *Atmos. Res*. **41**, 203–215.

Pandis, S.N., and Seinfeld, J.H. (1998) *Atmospheric Chemistry and Physics*. Willey, New York

Rosenberg, G.V., Gorchakov G.I., Lyubovtseva, Yu.S. and Georgievskii, Yu.S. (1980) *Optical characteristics of atmospheric aerosol.* In: Atmospheric Physics and Climatic Problems, pp. 216-257. Nauka, Moscow (in Russian).

Veretennikov, V.V., Kabanov M.V. and Panchenko, M.V. (1986) Microphysical interpretation of single parametric model of polarization scattering phase function (coastal haze). *Izv. USSR Acad. Sci., Atmos. Oceanic Phys.* **22**, 1042-1049.

Zuev, V.E. and Krekov, G.M. (1986) *Atmospheric Optical Models.* Gidrometeoizdat, Leningrad, 256 pp. (in Russian).

Zuev, V.E., Belan, B.D., Kabanov, D.M., Kovalevskii, V.K., Luk'yanov, O.Yu., Meleshkin, V.E., Mikushev, M.K., Panchenko, M.V., Penner, I.E., Pokrovskii, E.V., Sakerin, S.M., Terpugova, S.A., Tolmachev, G.N., Tumakov, A.G., Shamanaev, V.S., and Shcherbatov, A.I. (1992) The AN–30 "OPTIC–E" aircraft-laboratory for ecological investigations. *Atmos. Oceanic Optics*. **5**, 658–663.