REMOTE OPTICAL SENSING IN THE ULTRAVIOLET REGION OF THE AEROSOL LAYER NEAR THE STRATOPAUSE FROM ONBOARD THE ASTROPHYSICAL SPACE STATION "ASTRON"

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We present results of investigations carried out in 1984 and 1985 using an ultraviolet telescope of the "Astron" Astrophysical Space Station, of the aerosol layer near low boundary of the mesosphere at midlatitudes and in the equatorial zone. Data were obtained by the technique of tangent sensing in the ultraviolet wavelength range (λ = 273 and 280 nm, $\Delta\lambda$ = 3 nm). *The problem of determination of aerosol scattering parameters at altitudes near* 50 km *at the wavelengths specified is the problem of remote sensing under conditions of significant optical depths. Analysis of the method sensitivity is carried out. The data obtained show the presence of aerosol layers with high values of the turbidity coefficient S at altitudes* 65–100 km. *The atmosphere at altitudes* 55–65 km *is relatively clear, while at attitudes near* 50 km *there is observed a powerful aerosol peak. The peaks of aerosol scattering occur at altitudes* 49–52 km, *the mean value of the turbidity coefficient maxima is equal to* 0.8.

The aerosol layer in the region near the boundary between the stratosphere and mesosphere has been being investigated by different methods. Nevertheless, no clarity has yet been achieved in quantitatively characterizing aerosol scattering at these altitudes, Refs. 1 and 2.

Interpretation of data obtained from ground-based twilight observations[3–9] carried out in the visible region essentially depend on the methods used to estimate the contribution of multiple light scattering to the twilight sky brightness. The observations of daytime and twilight horizon of the Earth in the visible region from onboard the manned space vehicles[10–13] confirm the existence of aerosol layer near the altitude of 50 km. However, the interpretation of these observations also faces the problem on a proper account for multiple light scattering. The results of laser sensing of the atmosphere at altitudes from 40 and up to 120 km are well described by the Rayleigh molecular scattering[14,15] ($S < 0.1$). According to data obtained in a series of rocket measurements,[16,17] there exists an aerosol layer at altitudes near 50 km with the $S_{\max}$ value of the order of unity. The models describing the optical properties of aerosol at these altitudes[1,10,18,19] also differ from each other.

In Ref. 19 it has been assumed, that a solution to the problem on background aerosol in the upper atmosphere can most likely be obtained using methods, that provide for the inverse problem solution in most simple way. In our opinion, the method of tangent sensing in the ultraviolet wavelength region from 200 to 300 nm, that involves the Hartley absorption band of ozone may be referred as such. Under these conditions the atmospheric brightness is mainly formed due to single scattering of solar radiation with only small contribution coming from multiple scattering (see Ref. 20).

In our earlier papers[20,21] we have presented some results on aerosol layers, of natural and anthropogenic origin, obtained from investigation by the method of tangent sensing in the ultraviolet region (λ = 273 and 280 nm) from onboard an astrophysical space station (ASS) "Astron" at altitudes 65–100 km over the equatorial zone and midlatitudes. Besides, we have estimated vertical distribution of ozone concentration over the height range from 55 to 65 km.

The primary goal of this paper was to estimate scattering characteristics of aerosol in the height region near the low boundary of the mesosphere at about 50 km altitude using the same experimental data obtained using the tangent sensing method. In so doing, we used the techniques from our previous paper.[20] The techniques of conducting measurements in the sessions of tangent sensing from onboard the ASS "Astron", as well as the altitude referencing of the experimental brightness curves, and absolute calibration of a spectrometer have been described in Ref. 20. The background state of the atmosphere has been

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investigated over the equatorial zone (Central Africa) and over midlatitudes (the Atlantic) in 1983–1985. The limb scanning has been being carried out at the fixed wavelengths $\lambda = 273$ and 280 nm. The spectral resolution used equaled to 3 nm. The readouts have been being taken every 0.61 s with the vertical step varying from 0.4 to 1.2 km in different scans. The half-width of the instrumental function that characterizes the scale of the vertical averaging of measurements over midlatitudes was 8 km, and 5 km in equatorial zone.

The wavelengths of 273 and 280 nm are not optimum for remote sensing in the altitude region considered. Actually, this problem is a problem of remotely sensing at large optical depths τ. Correct account for multiple scattering contribution into the return signals is also a problem in laser sensing of the atmosphere both with ground-based[22] and spaceborne[23] instrumentation. In our case the single scattering approximation is sufficient, while the real problem is to isolate a weak signal from lower atmospheric layers against the background of higher signals from scattering layers in the upper atmosphere. Besides, under these conditions the influence of such noises, as spurious light in the spectrometer used, increases. Therefore, we have developed a technique to isolate signal, that, in our opinion, enables one to essentially decrease the error due to subtraction of the contribution from the spurious light and to better understand the data processing.

Figure 1 presents the ratio of simultaneously observed signals $I^*_{1\lambda}/I_2$ in the first operating channel and in the second UV channel of the spectrometer as a function of height.

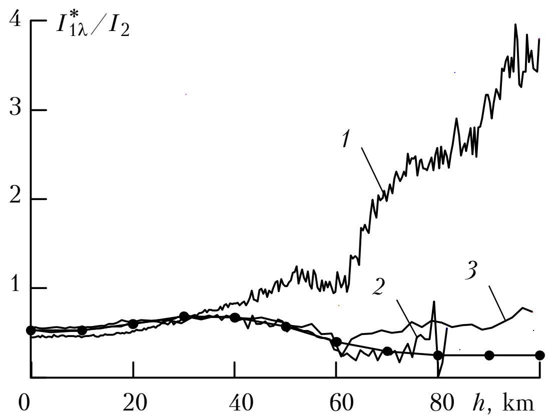

FIG. 1. The ratio, $I^*_{1\lambda}/I_2$, of signals in the first and second UV channels of the spectrometer, dots denote the curve showing the contribution from spurios light.

The subscript λ denotes both the operating wavelength of the first UV channel (the region from $\lambda = 245$ to 353 nm) and synchronous position on the wavelength scale of the output slit of the spectrometer in the second channel. The value $I^*_{1\lambda}$ is defined as the difference between the experimentally measured signal and signal $I_{r,\lambda}$ calculated for the case of molecular scattering of solar radiation, with the radiation attenuation being due to absorption by the ozone. The values of spectral density of solar radiation flux have been taken from Ref. 24. The molecular scattering has been calculated using cross sections of Rayleigh scattering from Ref. 25 and the atmospheric model that allows for seasonal, latitudinal, and diurnal variations of the atmospheric air density.[26,27] We also used the model of ozone height distribution from Ref. 28. The cross sections of light absorption by the ozone molecule we took from Ref. 29.

It may be noted that at the above wavelengths and altitude region considered the main contribution to optical depth of a measurement path is due to the absorption of light by the ozone while the molecular scattering adds only about 1% of it and the contribution from scattering on atmospheric aerosols is mostly below the molecular one, as it follows from the *a posteriori* analysis.

The ratio $I^*_{1\lambda}/I_2$ shown in Fig. 1 can be presented as the ratio $(I_{a,\lambda} + I_{1\lambda,\mathrm{scat.}})/I_{2,\mathrm{scat.}}$. Here $I_{a,\lambda}$ determines the aerosol component of the atmospheric brightness, thus the sum $I_{a,\lambda} + I_{r,\lambda}$ in the single scattering approximation determines the total energy brightness of the atmosphere I_λ. The values $I_{1\lambda,\mathrm{scat}}$ and $I_{2,\mathrm{scat}}$ are the contributions from spurious light scattered in the spectrometer to the signal both in the first and second operating UV channels of the spectrometer. It is important to note that analysis of data acquired in 1983–1985 from onboard the ASS "Astron" on the UV spectra of solar radiation reflected from the Earth's surface during daytime, has revealed high correlation between $I_{1\lambda,\mathrm{scat}}$ and $I_{2,\mathrm{scat}}$ (the correlation coefficient appeared to be between 0.96 and 0.98). According to this analysis, the signal in the second UV channel I_2 is mainly, about 97%, determined by spurious light scattering in the spectrometer that means that $I_{2\mathrm{scat}} \approx I_2$.

Curve *1* in Fig. 1 presents the data of tangent sensing at 273 nm wavelength (scan 3, Table I). One can clearly see three peaks on the curve of aerosol scattering in the height range 50–100 km. Local maxima within these peaks occur at the altitudes about 50, 75, and 95 km. In Fig. 1 the line along big dots shows the altitude dependence of the contribution from scattered light used to correct the initial data. This dependence of light scattering contribution actually coincides with the curve *2*, plotted using the data of the tangent sensing at the wavelength $\lambda = 249.4$ nm (the Atlantic in 1985). The scattering angle in these measurements was $\theta = 95°$, the solar zenith angle at the tangency point was 42°. At this wavelength the sensitivity of the spectrometer is about 15% and the level of solar radiation is about 30% of the corresponding values at the 273 and 280 nm wavelengths. Using curve *2* to retrieve the altitude dependence of spurious light scattering contribution provides for a 5% relative error of isolating the signal

$I_{a,\lambda}$. For a comparison, shown in Fig. 1 is curve *3*, that shows the data of tangent sensing at the wavelength of 255 nm (the Atlantic, 1985) acquired at the scattering angle $\theta = 95.8°$, and the solar zenith angle at the tangency point of 33°. The spectrometer sensitivity at $\lambda = 255$ nm is essentially higher, than at $\lambda = 249.4$ nm, while being several times lower, than that at $\lambda = 273$ nm. As one can see from Fig. 1, the altitude behavior of the ratios $I^*_{1\lambda}/I_2$ at altitudes above 60 km noticeably depends on the contribution from aerosol light scattering.

The complicated behavior of the ratio $I_{1\lambda,\text{scat.}}/I_{2,\text{scat}}$ is qualitatively confirmed by direct modeling. The atmospheric brightness has been calculated in the spectral region from 700 to 200 nm. These calculations used the sensitivity of PMTs and performance parameters of a filter protecting the output window of the first channel in the spectrometer from corresponding specifications. The distortions of radiation spectrum due to reflections from optical surfaces of the telescope used have also been considered. Light scattering simulations involved two extreme cases. In one case we assumed that scattering does not modify the spectrum at all, while in the other case the transformation was assumed to be proportional to a factor of λ^{-2}, that is characteristic of scattering by rough surfaces.[30]

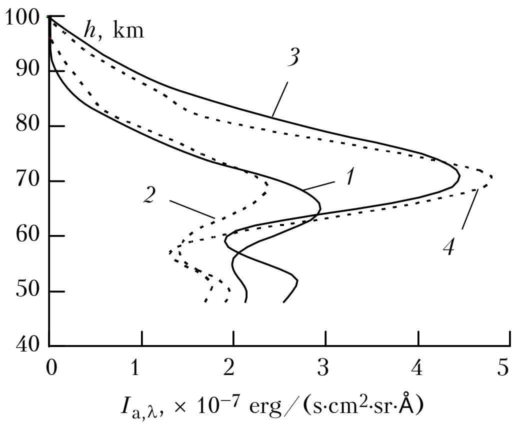

FIG. 2. Altitude behavior of the aerosol component of the spectral brightness of the Earth's atmosphere limb $I_{a,\lambda}$ in the ultraviolet region ($\lambda = 273$ and 280 nm).

TABLE I. Data on the conditions of tangent sensing of the atmosphere from onboard the ASS "Astron"

Number	Session date	Distance to the ASS, 10^3 km	Coordinates of the tangency point		Solar zenith angle, deg	Scattering angle, deg	Wavelength, nm
			N Latitude,	E Longitude,			
1	4.6.84	181	34.5	282.6	59.8	94.1	280
2	4.6.84	182	32.2	290.5	46.9	94.6	280
3	4.12.85	181	40.5	289.6	49.4	95.9	273
4	8.25.85	105	−1.4	32.1	32.8	84.5	273

Figure 2 shows altitude behaviors of the aerosol component of unperturbed atmospheric brightness in the Earth's limb $I_{a,\lambda}$.

The conditions under which the sensing session has been performed are given in Table I. Numbers at the curves in Fig. 2 correspond to the scan numbers shown in Table I.

The reconstructed altitude behaviors of the volume directional scattering coefficient of aerosol, $\beta_{a,\theta}$, are presented in Fig. 3*a*, and the turbidity coefficient S in Fig. 3*b*. The turbidity coefficient was calculated with the use of the mean (typical for the atmosphere) scattering phase function of aerosol taken from Ref. 31. At the scattering angle $\theta = 90°$ the ratio between the molecular Rayleigh scattering phase function to that of the aerosol is equal to 1.7.

The technique of the inverse problem solution is similar to that described in Ref. 20. We used the step-by-step iteration method that minimized the discrepancy between the calculated and experimental values of brightness. The values of the most uncertain function have been reconstructed using a 1-km step. To eliminate the solution instability the initial data on brightness $I_{a,\lambda}$ were smoothed using a truncated point Gaussian filter with the half-width less, than the scale of the instrumental smoothing of the brightness over the altitude.

The altitude behaviors of the scattering coefficient $\beta_{a,\theta}$ and the factor S at altitudes 65 to 100 km practically coincide with the dependences obtained in Ref. 20 for the same sensing sessions. The total characteristic of the aerosol scattering at these altitudes may also be found in Ref. 20.

As seen from Fig. 3*b*, in the range 65–55 km the values of the turbidity coefficient, S, obtained are small being on the average of the order of 0.1 for all scans, that is compared with the reconstruction errors. The reconstructed values S vary in the range from 0.05 up to 0.2. That low level of the atmospheric turbidity can be considered as a ground for estimating the altitude behavior of the ozone concentration in this altitude region without the account of light scattering by aerosols.[20]

As seen from Fig. 3*a*, there is a peak in the altitude behavior of the directional scattering coefficient near the altitude of 50 km. As the altitude decreases, a sharp increase in the aerosol light scattering is observed starting from approximately 56 km and reaches its maximum at altitudes of 50–52 km. The power of aerosol scattering from this layer is an order of magnitude higher than from the aerosol layers at altitudes from 65 to 100 km. The peak of aerosol scattering at 50–52 km altitudes also manifests itself on the turbidity curves (Fig. 3*b*) with the maximum values of S being from 0.6 to 0.9. The mean value S_{max} is equal to 0.8. It should be noted, that the sensing sessions involved in our analysis have been carried out in different seasons and over different latitudes being also separated by long time intervals. It is also important that the scales of a horizontal averaging characteristic of the tangent sensing were on the order of thousand kilometers.

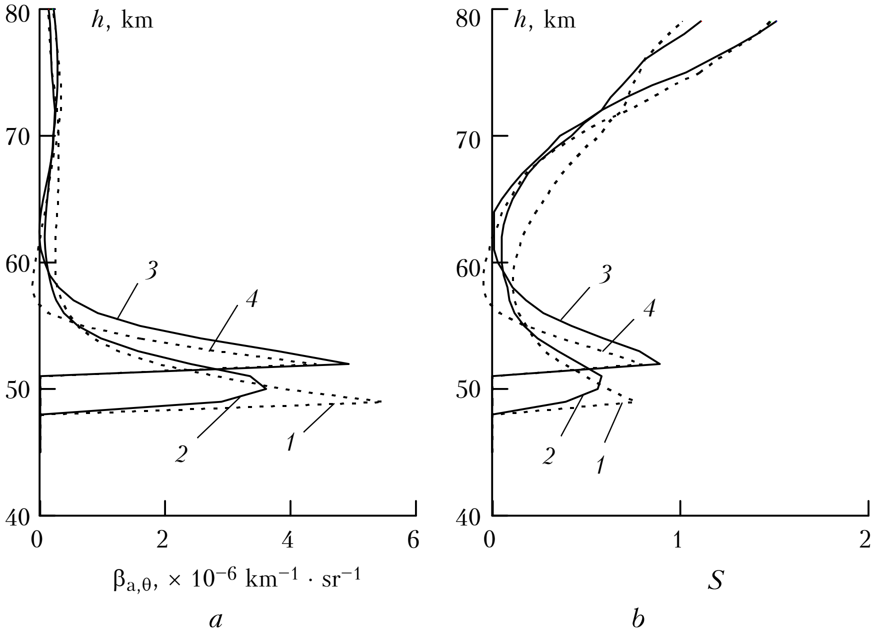

FIG. 3. Altitude behavior of the aerosol scattering characteristics: the coefficient of directional light scattering $\beta_{a,\theta}$ *(a); and the turbidity factor S at the wavelengths* λ = 273 *and* 280 nm *in the ultraviolet region (b); curves 1–3 present data for midlatitudes and curve 4 for the equatorial zone.*

In Ref. 21 we have arrived at a conclusion, based on observational data on the Earth's limb spectral brightness acquired using the method of tangent sensing in the ultraviolet range, that no perturbations occurred in the atmospheric background parameters at the observation points along the active trajectory of Space Shuttle at altitudes below 80 km. Therefore, the reconstructed values of aerosol parameters at altitudes below 80 km may be referred to as the background ones.

TABLE II. Data of tangent sensing of the aerosol layer in the UV region (λ = 280 nm) *at a height of* 50 km *following in time the Space Shuttle launch on April 6, 1984.*

Scan number	The position of maxima, km	The turbidity coefficient	$\beta_{a,\theta}$, km^{-1}sr^{-1}
3	49	1.0	$7.4 \cdot 10^{-6}$
4	50	0.7	$4.8 \cdot 10^{-6}$
5	50	1.1	$7.0 \cdot 10^{-6}$
6	49	0.6	$4.5 \cdot 10^{-6}$
7	49	1.0	$7.4 \cdot 10^{-6}$

Table II gives the parameters of aerosol scattering near the altitude of 50 km obtained by the method of tangent sensing in the ultraviolet region (λ = 280 nm) after the Space Shuttle launch on April 6, 1984. The characteristics listed in Table II are as follows: the altitudes of the aerosol scattering peaks, values of the turbidity coefficients, and of the directional light scattering at the layer maxima. The scan numbers in Table II correspond to the numbers of scans in Ref. 21.

The mean value S_{max}, according to the data in Table II, is equal to 0.9. This value well agrees with the mean value S_{max} according to data presented in Fig. 3*b*. The maximum spread of average values does not exceed 30%. On the whole, the mean, over the entire data set we have compiled, value of S_{max} for the peak of aerosol scattering at the altitude about 50 km is 0.8.

This mean value of S_{max} well agrees with the turbidity values in the altitude range of 50 km obtained by the method of twilight sensing, after eliminating the contribution from multiple scattering using the method of Divary and Fesenkov,[6] and from the rocket sensing data.[16,17] However, according to our data, this peak is more pronounced (a larger difference in the turbidity values at the altitudes 50 and 60 km) as compared with the data of rocket measurements to say nothing about the data from Ref. 6, where S is presented as approximately constant in the altitude range from 40 to 60 km. It is also worth noting, that the turbidity referred has been estimated in the visible region. The mean atmospheric turbidity, as a function of height, has, according to models from Refs. 3 and 18, a pronounced peak near 50 km height, but the mean value of S_{max} is equal to 3 (at λ = 550 nm).

Now we should like to attract attention of the reader that the height dependences of the aerosol characteristics shown in Figure 3 become quantitatively meaningless at heights below about 49 km. The matter is that the method used becomes insensitive in this height region or, what is more correct, according to numerical simulations performed, there is no any physical sense in these measurements, for the atmospheric regions below the corresponding maximum in the aerosol height distribution.

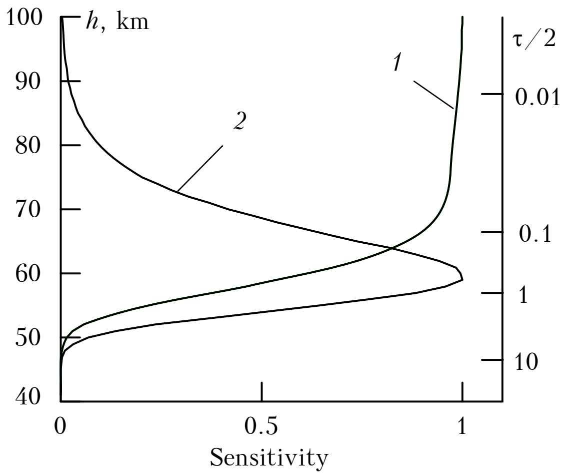

FIG. 4. Sensitivity of the method of tangent sensing at λ = 280 nm *as a function of the depth of an aerosol scattering layer from the top of the atmosphere related to the coefficient of directional light scattering (1); and to the turbidity coefficient (2).*

Figure 4 presents the altitude dependences of relative sensitivity of the tangent sensing method in the ultraviolet range λ = 280 nm. Curve *1* in this figure presents the ratio of the brightness of a 1-km thick spherical aerosol layer at an altitude h above the Earth's surface to the brightness of the layer at 100 km altitude, when observing along the line tangent to the inner surface of the scattering layer. The value of the volume coefficient of aerosol light scattering inside the layer has been taken to be the same for all the layers considered. The meaning of data presented by the curve number *2* in this figure is similar to that of the curve *1*, except that the scattering layers are characterized by the turbidity coefficients of same value, and the normalization condition are selected so that the maximum sensitivity is equal to unity. The right-hand side ordinate axis in the figure shows the halved values, $\tau/2$, of the horizontal optical depth that characterize the optical depth of the atmosphere along the sounding paths. According to data presented by curve *1*, the sensitivity of the method at the altitude 50 km makes up only 1% of its 100%-value at higher altitudes. At the same time curve *2* shows that the sensitivity at the 50-km altitude is still 7% of the maximum value. Thus, the method is yet enough sensitive to detect the aerosol layers of enhanced turbidity at altitudes near and above 50 km while being practically insensitive to those at altitudes below 48 km.

The loss of the method sensitivity at the altitudes below 50 km may result in a shift of maxima in the altitude dependences of the aerosol scattering parameters reconstructed for the height region near 50 km relative to the position of actual maxima. Besides, it can also introduce an additional error into the estimates of maximum values of these parameters. We have assessed these effects using model calculations. The aerosol peak was modeled in the form of Gaussian distribution with a half-width of 10 km. We have added a +5% error to the calculated brightness curves and then the parameters of aerosol scattering were again reconstructed by solving the corresponding inverse problem.

It appeared so, that if the center of the initial peak is successively put at altitudes of 48, 50, and 52 km, then the maxima of reconstructed curves are at the altitudes of 50, 51, and 52 km, and the maximum values of the aerosol scattering coefficient change, accordingly, by −7, +1, and +5 per cent. Similarly, when we add a −5% error, maxima of the reconstructed dependences are at the altitudes of 50, 50, and 52 km, and the maximum values change by −16, −6, and −5%, respectively. Taking into account literature data that the maxima of aerosol scattering are usually recorded at the altitudes above 48 km, one can come to a conclusion, that the loss of the method sensitivity in this case does not give rise to an essential error in the estimated value of the maximum turbidity.

The other problem is connected with the possible dependence of the results obtained on the altitude distribution of the ozone used. The above mentioned results have been obtained using the CIRA–72 model.[28] If, instead, we make use of the height distributions of the ozone reconstructed from our data at altitudes 65–55 km, neglecting the aerosol scattering at these altitudes (see Ref. 20), and then exponentially extrapolated to lower altitudes, we have the maximum S values of the aerosol scattering peak near 50 km increased by approximately 15%. It is worth noting in conclusion that the difference between the altitude profiles of the ozone concentration, we have obtained in our study, from the model ones noticeably exceeds the level of the rms "weather" variations of the ozone, predicted by the model in Ref. 28. Thus, it may be expected that the difference between individual ozone distributions at a certain locality and time of observations and the mean model distribution used does not introduce an essential error.

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