REMOTE SENSING OF VERTICAL DISTRIBUTIONS OF AEROSOL AND OZONE IN THE UPPER ATMOSPHERE FROM A BOARD OF THE ASTROPHYSICAL SPACE STATION ASTRON IN THE ULTRAVIOLET REGION

A.A. Cheremisin, L.N. Granitskii, V.M. Myasnikov, N.V. Vetchinkin, and V.V. Slabko

Scientific-Research Physical-Technical Institute at the Krasnoyarsk State University
Krasnoyarsk State Technical University
E.K. Fedorov Institute of Applied Geophysics, Moscow
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The results of investigations of the upper atmosphere in the equatorial zone and at middle latitudes are presented obtained by the method of tangent sensing in the ultraviolet wavelength region (λ = 280, 273, and 255 nm, $\Delta\lambda$ = 3 nm) *with the help of an ultraviolet telescope of the Astrophysical Space Station Astron in* 1984–1985. *As a result of analysis of experimental vertical profiles of spectral brightness of the atmospheric limb of the Earth, the vertical profiles of the aerosol scattering parameters have been reconstructed between* 65 *and* 100 km *and the vertical profiles of the ozone concentration between* 55 *and* 65 km. *According to our data in the equatorial zone and at middle latitudes, a thick aerosol layer is observed between* 65 *and* 100 km. *The maxima of the differential light scattering coefficients are located between* 72 *and* 75 km. *On the turbidity curves S, two peaks are revealed. The first peaks have their maxima at* ~80 km *(S* = 1.0–1.5*). The second peaks have the maxima between* 90 *and* 95 km *with even higher values of the coefficient S. The data on the ozone between* 55 *and* 65 km *agree fairly well with the data of the reference ozone model.*

At present the aerosol of the mesosphere and lower troposphere is poorly known in comparison with the aerosol of the troposphere and stratosphere and the data about it are contradictory.[1,2] The data of a few direct contact measurements obtained with the help of geophysical rockets differ by the value of the measured concentration by several orders of magnitude.

Systematic ground-based twilight observations reveal a complex layered structure of the mesospheric aerosol with the maxima often observed near altitudes[3] 50–60, 60–70, 75, 85, and 95 km and high values of the turbidity coefficient S in the regions of maxima.[3–4] Under normal conditions, the lidar data on backscattering of the laser radiation from altitudes 30–90 km agree very well with pure Rayleigh molecular scattering and the noticeable increase of lidar signals is associated with the intrusion of significant masses of meteoric matter.[5] Models based on the rocket and lidar data[6] yield lower values of turbidity than models based on the data of ground-based twilight observations. A large number of works are devoted to investigation of aerosol formations near 75–95 km named noctilucent clouds,[7–10] including investigations in the ultraviolet region.[11] We believe that to study aerosol layers in the upper atmosphere, it is expedient to use the ultraviolet wavelength region. Efficiency of the UV region including the Hartley band of ozone absorption is connected with the fact that atmospheric brightness is formed primarily due to a single scattering of the solar radiation, the background illumination from lower-lying layers of the atmosphere is greatly suppressed, cloudiness has no effect, and there are no variations of the albedo of the underlying surface. According to the results of Monte-Carlo calculations,[12] the contribution from double scattering is 0.0–0.7%, and according to the results of our calculations it is 0.6–1% of the intensity of single scattering near altitudes 50–90 km at wavelengths 250–280 nm.

In the present paper, the results of investigations of the upper Earth's atmosphere are presented obtained by the method of tangent sensing in the ultraviolet wavelength region with the help of an ultraviolet telescope of the Astrophysical Space Station (ASS) Astron.

The ASS Astron, put into satellite orbit on March 23, 1983, had an apogee of 200 000 km and a perigee of 2 000 km. Its orbit was tilted at an angle of 56° and its rotation period was 96 h. The ASS was equipped with an ultraviolet telescope[13,14] constructed on the basis of the Ricci-Chretien scheme with a diameter of the principal mirror of 80 cm. The telescope was coupled with a spectrometer constructed on the basis of the Rowland scheme with a toroidal diffraction grating. With the help of three PMTs the radiation was

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simultaneously recorded in two spectral channels between 245 and 353 nm and between 157 and 263 nm in the zero order of the spectrum. The PMTs operated in the photon counting mode with a time resolution of 40 ns.

The curves of the spectral brightness of the Earth's atmospheric limb were recorded in the following way. By turning of the entire ASS, the telescope optical axis was first pointed at the given point of the Earth's surface. Because of the ASS orbiting with unchanged orientation about stars, its optical axis was moved through the Earth's surface with subsequent intersection of the Earth's edge and the atmospheric limb. In analogous way, the intersection of the Earth's disk was organized. As a basic working diaphragm, a 12″ offset input diaphragm of the spectrometer was used. The limb was scanned at fixed wavelengths λ = 255, 273, and 280 nm with spectral resolution of 3 nm. Readings were taken every 0.61 s with vertical resolution changing from 0.4 to 1.2 km for different scans. The background state of the atmosphere was investigated in the equatorial zone (Central Africa) and at middle latitudes (the Atlantic) in 1984–1985. Information about the sessions of tangent sensing is presented in Table I. The halfwidth of the instrumental function characterizing the vertical averaging for the measurements at middle latitudes was 8 km. According to our estimates, this leads to the increase of the measured signal intensity by ~6% for altitudes above 70 km, the decrease of the signal intensity by 4% in the vicinity of the maxima in the curves, and to shifts of the maxima by less than 1 km. For the equatorial measurements, for a halfwidth of the instrumental function of 5 km, these errors are estimated to be much less: 3%, 2%, and 0.6 km, respectively.

TABLE I. Information about the conditions of tangent sensing of the atmosphere from a board of the ASS Astron

No.	Session date	Distance to ASS, th.km	Coordinates of the tangential point		Local time, h:min	Solar zenith angle, deg	Scattering angle, deg	Wavelength, nm
			N latitude	E longitude				
1	4.6.84	181	34.5	282.6	8:12	59.8	94.1	280
2	4.6.84	182	32.2	290.5	9:09	46.9	94.6	280
3	4.12.85	181	40.5	289.6	9:10	49.4	95.9	273
4	8.25.85	105	−1.4	32.1	10:00	32.8	84.5	273
5	8.25.85	102	−1.1	32.3	9:25	41.0	83.0	255

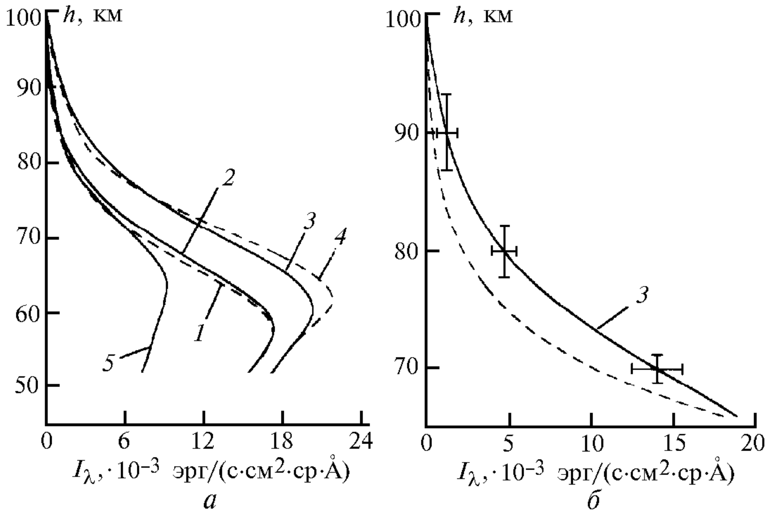

FIG. 1. Vertical profiles of the spectral brightness of the Earth's atmospheric limb from the data of observations from a board of the ASS Astron in the UV region: a) smoothed experimental curves, b) fragment of experimental curve 3 (the solid line) and the model curve of the Rayleigh molecular scattering brightness (the dashed line).

In Fig. 1*a*, the vertical profiles of the atmospheric spectral brightness I_λ are shown measured under unperturbed atmospheric conditions. Experimental dependences were smoothed by the Gaussian point filter with halfwidths 2.5–6 km. In Fig. 1*b*, the fragment of the experimental curve of the atmospheric brightness is shown in more detail. The errors in intensity measuring and corresponding errors of altitude referencing are also shown on the experimental curve. As can be seen from Fig. 1*b*, the difference between the experimental brightness curve and the model curve of brightness due to pure molecular Rayleigh's scattering calculated for the model of the atmosphere that considers seasonal, latitudinal, and diurnal variations of the atmospheric density,[15,16] significantly exceeds the experimental errors. The difference may be explained by the effect of aerosol scattering.

The altitude referencing of the curves was made as follows. Altitudes of the maxima of the experimental curves were affixed to those of the maxima of the model brightness distributions. In model calculations, seasonal and latitudinal variations of ozone concentrations[17] and seasonal, latitudinal, and diurnal variations of the atmospheric density[15,16] were taken into account. Consideration of the aerosol scattering for the model proposed by Krekov and Zvenigorodskii[1] did not result in significant shifts of the maxima. Total accuracy of referencing was no worse than ±1 km. Determination of the vertical scale (rate of change of the altitude of the telescope line of sight) was carried out when in addition to the working 12″ diaphragm the second 1″ diaphragm of the spectrometer was simultaneously opened. In this case, registergram was the superposition of two analogous patterns that significantly differ in the intensities, but had well-pronounced vertical shift (time delay). From the known angular distance between the diaphragms, ballistic data, and parameters of the tangential point, the vertical shift of images was calculated. Accuracy of determination of the vertical scale was 2–10%.

Absolute calibration for the energetic brightness was made so that with the consideration of the relative spectral sensitivity of the spectrometer determined from the measurements of the spectral flux of the star α Lyra, the data of investigations of the atmospheric spectra from a board of the ASS Astron (1983–1985) were fitted to the corresponding data of observations from Nimbus–7 (see Ref. 18) in the following way. The average (over the spectra) signal value at $\lambda = 302$ nm (minus the correction for the scattered light) was fitted to the spectral brightness averaged over the observational data from Nimbus–7. In so doing, the correction factor was introduced calculated for the model data on ozone[17] and atmospheric density[15,16] that differed by ~26% from unity and allowed us to consider the seasonal, latitudinal, and geometrical differences in the observations. According to the data of Frederick and Serafino,[18] the Earth's spectral albedo at 302 nm is fairly stable quantity – its standard deviation does not exceed 10%. Analogous spread of data was obtained by us. Change of the relative spectral sensitivity of the first channel of the spectrometer with the wavelength was determined from the change of the spectral flux of the star α Lyra. According to the star calibration curve, the relative spectral sensitivity is 100% at a wavelength of 310 nm, 99% at 302 nm, 85% at 280 nm, 72% at 273 nm, and 26% at 255 nm. By our estimates, the total error of the absolute calibration with consideration of the statistical averaging must be less than 10%.

The main problem for the ASS Astron was the investigation of rather hot blue star objects.[13,14] For colder objects, such as the Sun, the structure of their UV spectra was complicated by a significant contribution from visible light scattered in the spectrometer.[14] In this case, the contribution of the scattered light estimated from synchronous readings of the PMT in the second UV channel was subtracted from a PMT signal in the first (working) channel of the spectrometer. According to the results of investigations of the atmospheric spectra in the second UV channel (λ shorter than 200 nm), the useful signal was 3% from the total signal and thereby the scattered light was registered in fact in this channel. From this, the contribution from the scattered light in the first UV channel was correlated with the synchronous readings of the PMT of the second UV channel with fairly high correlation coefficient lying in the limits 0.96–0.98. During limb measurements, the portion of the subtracted scattered light was about 10% of the total signal. In addition to the contribution from the scattered light, the contribution from the dark current of the PMT (~0.34% of the signal value corresponding to the brightness maximum in the limb) and also the contribution from practically constant value of the signal of the opened 1″ diaphragm (4.4% of the maximum value) were subtracted from the recorded signal. Beginning with an altitude of ~55 km, the portion of the scattered light is exponentially increased with the decrease of the altitude and near ~30 km, the scattered light contribution becomes comparable to the useful signal, and when the line of sight of the telescope moves along the day Earth's surface, it is nearly twice as large as the useful signal. The maximum contribution of the scattered light to the signal under these conditions gives the radiation with a wavelength of ~350 nm.

First, based on the model for optical characteristics of the aerosol[9] and on the data published in Ref. 6 we concluded that the aerosol effect at altitudes higher than 50 km can be neglected and the vertical profiles of the ozone concentration between 50 and 70 km can be reconstructed from the obtained data. Analogous conclusion about insignificant effect of the aerosol at altitudes higher than 50 km on the information content of the indirect methods for ozone reconstruction from measurements of the UV radiation scattered by the atmosphere in fact can be drawn from the results of Ref. 21 in which the same models of the mesospheric aerosol were used. More detailed analysis allowed us to reconstruct the vertical profiles of the aerosol scattering characteristics at altitudes higher than 70 km and to estimate the influence of the aerosol layer observed at these altitudes on the accuracy of the ozone reconstruction between 50 and 70 km.

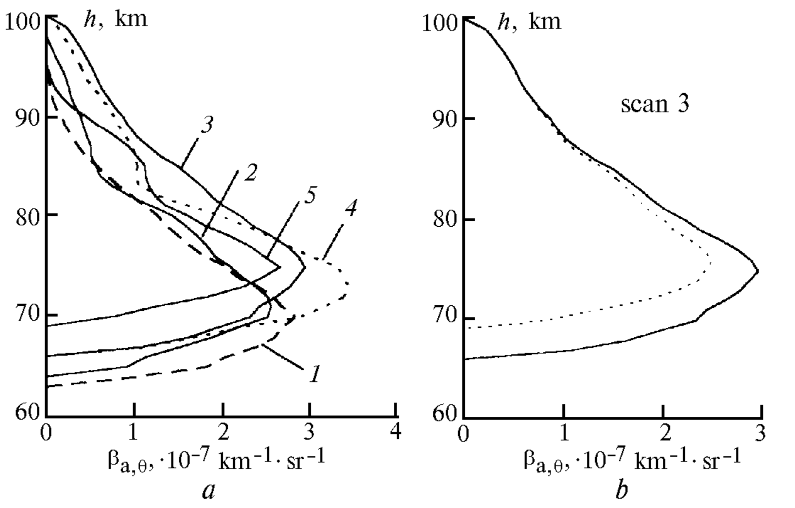

FIG. 2. Reconstructed vertical profiles of the differential light scattering coefficient of the aerosol $\beta_{a,\theta}$ for the ultraviolet region of wavelenghts (λ = 280, 273, and 255 nm) at mid-latitudes (1–3) and in the equatorial zone (4–5): a) curves 1–5 were reconstructed with the correction for radiation absorption by ozone; b) results of reconstruction with (the solid line) and without correction (the dashed line) for ozone contribution.

In Fig. 2 the results are presented of reconstruction of the vertical profiles of the volume differential aerosol light scattering coefficient $\beta_{a,\theta}$ between 65 and 100 km. The energetic brightness of the atmosphere was calculated in the well-known single scattering approximation. To solve the inverse problem, we used the step-by-step iteration method that minimized the discrepancy between calculated and experimental values of brightness. The functions themselves were reconstructed with vertical resolution of ~1 km. The initial data on the brightness were smoothed by the point truncated Gaussian filter. The problem refers to the class of weakly incorrect problems – the solution is unstable toward the initial data error. In connection with this, we compared the obtained solutions with the solutions found with the help of regularizing algorithms. At altitudes higher than 70 km, where we can neglect the ozone absorption and the optical thicknesses are small, the equation being solved is reduced to the well-known Abel equation. We obtained the solutions to the problem by the inversion method based on the use of smoothing cube splines.[22] The difference between the vertical profiles of the aerosol characteristics reconstructed by this method and those obtained by the step-by-step iteration method with smoothing of the initial curves by the truncated Gaussian filter was insignificant for altitudes below 95 km.

The profiles shown in Fig. 2*a* were reconstructed with the correction for the ozone absorption. In so doing, the model ozone distribution obtained in Ref. 17 was used. Cross sections of the ozone absorption were borrowed from Ref. 34. The spectral energy flux density of the solar radiation incident on the upper atmospheric boundary was borrowed from Ref. 24. Molecular scattering characteristics were calculated with the use of the Rayleigh scattering cross sections calculated in Ref. 25. In Fig. 2*b*, the results of reconstruction of the differential aerosol light scattering coefficients, which were obtained with the correction for the ozone absorption and without it, are shown for scan 3. It can be seen that at altitudes above 75 km the differential aerosol light scattering coefficients reconstructed neglecting completely the ozone absorption are close to the coefficients reconstructed with the correction for the ozone absorption. The difference between the coefficients is that with the consideration of the ozone absorption, pronounced aerosol scattering begins at altitudes above 65 km, whereas without consideration of the ozone absorption – at altitudes above 70 km.

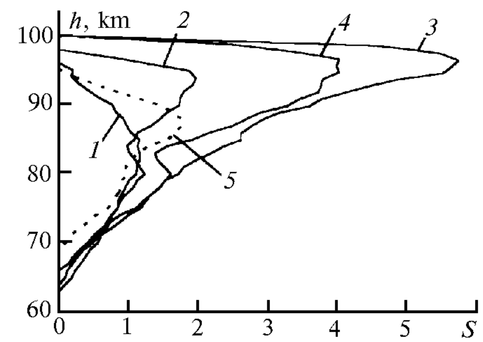

FIG. 3. Vertical profiles of the turbidity coefficient for the ultraviolet region of wavelengths (λ = 280, 273, and 255 nm) at middle latitudes (1–3) and in the equatorial zone (4–5).

In Fig. 3 the vertical profiles of the turbidity coefficient S are shown. In the construction of the turbidity coefficient curves, the ratio of the scattering phase functions of the Rayleigh molecular scattering and aerosol scattering was taken to be 1.7 (for the scattering angle $\theta = 90°$), in accordance with the value of the mean (typical) atmospheric phase function of the aerosol scattering from the data of Ref. 26.

According to the obtained results, in the equatorial zone and at middle latitudes a thick aerosol layer was observed. The curves of the aerosol scattering for the equatorial zone and middle latitudes are to some extent similar to each other. The curves of the differential light scattering coefficient are bell-shaped. The maxima on the curves are between 72 and 75 km. The intensities of the aerosol scattering in the maxima differ from each other by 50%. Here it should be noted that for the employed method of tangent sensing the horizontal scale of averaging was equal to 10^3 km. The turbidity curves S have two-peak structure. The first peaks have the maxima near ~80 km with high values of the turbidity coefficient $S = 1.0–1.5$. The second peaks have the maxima between 90 and 95 km with even greater values of the coefficient S. Here it should be noted that at altitudes greater than 92–95 km, the signal-to-noise ratio for the initial data was less than 1. The thickness of the observed aerosol layer may be characterized as follows: the calculated vertical optical thickness for the aerosol scattering in this layer is estimated as $1–2\cdot10^{-4}$ (λ = 255–280 nm), which is comparable with the thickness of the noctilucent clouds observed primarily between 80 and 83 km in the polar regions. The typical vertical optical thicknesses of noctilucent clouds are $1\cdot10^{-5}–3\cdot10^{-3}$ (in the visible range, see Ref. 1).

This pattern of the turbidity corresponds qualitatively to the results of observations at the Abastuman Observatory by the twilight sensing method[3] and to the data of the rocket investigations at the middle latitudes.[27] It does not contradict with the observations from space vehicles[38] and to the conclusions of Ref. 29 about high dust content in the atmosphere at the examined altitudes.

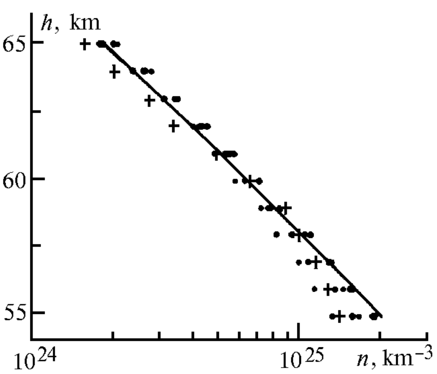

FIG. 4. Vertical profiles of the ozone concentration for the middle latitudes and equatorial zone (•) reconstructed from scans 1–5. The solid curve is the model borrowed from Ref. 17; +, the curve reconstructed from the data of Ref. 30.

In Fig.4 the results of reconstruction of the vertical profiles of the ozone concentration between 55 and 65 km from the measured vertical profiles of the atmospheric spectral brightness of the Earth's limb are shown. In the reconstruction, the aerosol layer between 65 and 100 km was considered. The neglect of the aerosol influence introduces the error Δn in the ozone reconstruction.

TABLE II. Average relative error of the reconstruction of the ozone concentration due to the neglect of the aerosol scattering between 70 *and* 100 km.

h, km	55	57	59	61	63	65	67
$\Delta n/n$, %	0.8	1.6	2.0	4.3	1.8	–33.5	–81.5

In Table II, the relative average errors are given as functions of the altitude. It can be seen from the table that if we neglect the influence of this aerosol layer, the reconstructed values of the ozone concentration will change by 1–2% at altitudes between 55 and 63 km. The aerosol scattering at altitudes 55–65 km was neglected. This can be substantiated as follows. First, according to the model of optical characteristics proposed in Ref. 1, which is in good agreement with the rocket data reported in Ref. 6, the aerosol scattering at these altitudes is relatively small in comparison with the molecular scattering (3 – 6%). Second, the effect of the aerosol scattering is masked by the error in measuring the vertical profiles of the atmospheric spectral brightness. As can be seen from Fig. 4, our data are in good agreement with the model data: relative deviation is less than 20%. Observed deviation is somewhat greater than the standard deviation, which for the examined latitude and season, according to the model proposed in Ref. 17, is ±4% for an altitude of 55 km. On the one hand, the deviation of our profiles of the ozone concentration from the model profile may be connected with the error of measuring the brightness. However, on the other hand, in Fig. 4 the profile of the ozone concentration obtained with the help of the rocket optical ozonesonde is shown reconstructed from the data of measurements of the extinction of the directly transmitted solar radiation at λ = 255 nm and a solar zenith angle of 90° (see Ref. 30), which also significantly differs from the corresponding model profile.

From Table II it is also seen that strong aerosol scattering may significantly influence the results of reconstruction of the ozone concentration. Good agreement of the ozone distribution reconstructed by us at altitudes 55–65 km without consideration of the aerosol scattering with the model distribution of ozone may confirm the idea about comparatively small level of the aerosol scattering at the examined altitudes.

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