SOME RESULTS OF SINGLE-FREQUENCY SOUNDING OF THE STRATOSPHERIC AEROSOL

V.A. Lyadzhin, B.T. Tashenov, and T.P. Toropova

Astrophysical Institute of the Academy of Sciences of the Kazakhstan SSR, Alma-Ata
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The results of single-frequency (532 nm) laser sounding of the stratospheric aerosol, performed in the period 1984–1987 for altitudes of 12–30 km above sea level, are presented.

The lidar was constructed based on a one-meter telescope located at an altitude of 2700 m. The altitude behavior of the backscattering ratio $R(H)$ was determined for different seasons. The values of $R(H)$ at the maximum of the Jung layer $R_{\max}(H)$ were on the average equal to 1.33 during the summer and 1.36 during the fall. Statistical analysis showed that the distribution of the quantities $R_{\max}(H)$ and $R(H)$ is somewhat asymmetric (during the summer the most probable value of $R_{\max}(H)$ was found to be equal to 1.25).

Some lidar data on the altitude stratification of the atmospheric aerosol at altitudes ranging from 12 to 30 km above sea level are presented in this paper. The data were obtained at the Astrophysical Institute of the Academy of Sciences of the Kazakhstan SSR in the period from September 1984 to January 1988. An attempt was made to estimate the altitude behavior of the backscattering ratio $R(H)$ for different seasons and to determine the most probable values of these quantities, which are important for the construction of a regional opto-physical model of the atmosphere.

The lidar[1] was built based on a one-meter telescope and placed in the mountains of Zailiĭskiĭ Alatau. The radiation source was a YAC laser ($\lambda = 532$ nm) with a collimator (the divergence angle of the light beam ~ 1.5′). The monostatic sounding scheme was used in the lidar in this scheme the receiver and the transmitter were separated by a distance of 1 m. The receiver and transmitter were aligned with an accuracy of ~ 0.5′.

Locating the lidar system at an altitude of 2.7 km above sea level in a location far from artificial sources of pollutants and using a noncoaxial sounding scheme reduced the errors arising owing to the effect of the lower layers of the atmosphere on the results of sounding. The absence of significant systematic errors was checked additionally by analyzing the sounding data in different directions and by comparing the altitude profiles obtained using light signals with different powers transmitted into the atmosphere.

The analysis of the results of laser sounding of the atmosphere was based on the solution of the lidar equation in the approximation of single scattering, which for a fixed wavelength of the sounding pulse and detection of signals in the photon-counting mode has the form

$$N(z) = N_0 K z^2 \cdot \sigma_\pi(z) \cdot T^2(z), \qquad (1)$$

where $N(z)$ is the number of photons received from a distance z from the lidar, at which the backscattered signal was formed; N_0 is the number of photons emitted; $K = K_{rec} \cdot K_{trans} \cdot A \cdot \Delta z \cdot \eta$ (where K_{rec} and K_{trans} are the transmittances of the receiving and transmitting channels, respectively; η is the quantum efficiency of the photomultiplier; A is the area of the receiving mirror of the telescope; and, Δz is the averaging interval (spatial resolution); $\sigma_\pi z$ is the backscattering coefficient at a distance z; T is the transmittance of the atmosphere on the path between the lidar and the scattering layer. The source of information on the optical properties of the atmospheric layer studied are the backscattering coefficients

$$\sigma_\pi(z) = \sigma_{\pi m}(z) + \sigma_{\pi a}(z),$$

where $\sigma_{\pi m}(z)$ and $\sigma_{\pi a}(z)$ are the molecular and aerosol backscattering coefficients, respectively. The coefficient $\sigma_{\pi m}(z)$ is related uniquely with the density of the molecular atmosphere at a height H while $\sigma_{\pi a}(z)$ will be determined by the concentration and physical-chemical properties of the aerosol. To solve the lidar equation (1) for $\sigma_\pi(z)$ it is necessary to give *a priori* the transmittance T as well as to study the Instrumental parameters of the lidar setup. Methods for solving the lidar equation and analyzing its limits of applicability are reviewed in detail in Refs. 2 and 3.

For single-frequency sounding the method of reconstruction from the results of measurements of the profile of the backscattering ratio

$$R(z) = \frac{\sigma_{\pi m}(z) + \sigma_{\pi a}(z)}{\sigma_{\pi m}(z)}, \qquad (2)$$

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is widely employed. This ratio characterizes the amount by which the experimentally observed backscattering coefficient exceeds the molecular backscattering.

This method is based on the calibration based on the computed molecular-scattering signal. The method is based on the assumption that at least one of the local minima in the profile of the photocounts in the section of the sounding path studied is caused by scattering in the molecular atmosphere free of aerosol.[4] The calibration consists of combining the real measured profile of the function $S(z) = N(z) \cdot z^2/N_0 \cdot k$ at the point of one z_0 or several local minima with the profile of the computed normalizing signal $S_m(z) = \sigma_{\pi m}(z) \cdot T^2(z)$. Here $T^2(z) = T_m^2(z) \cdot T_a^2(z)$ where $T_m(z)$ is the atmospheric transmittance determined by molecular attenuation only, and is calculated just like $\sigma_{\pi m}(z)$ based on model tables or data from aerological sounding; $T_a(z)$ is the transmittance owing to the presence of aerosols in the atmosphere and is determined from model tables, for example, Ref. 5. Then the calibration constant is given by

$$K = \frac{N(z_0) \cdot z_0^2}{N_0 \cdot \sigma_{\pi m}(z_0) \cdot T^2(z_0)} .$$

The backscattering ratio is determined from the expression

$$P(z) = \frac{S(z) \cdot T^2(z_0)}{S_m(z) \cdot T^2(z)} = \frac{N(z) \cdot z^2\, \sigma_{\pi m}(z_0) \cdot R(z_0)}{N(z_0) \cdot z_0^2 \cdot \sigma_{\pi m}(z)\, T^2(z + z_0)}, \tag{3}$$

where $R_0(z) = S(z_0)/S_m(z_0)$ (we assume that $R(z_0) = 1$ and $T(z{-}z_0) = T(z)/T(z_0)$) is the transmittance of the atmosphere on the path from z to z_0 (for the altitude range 12–30 km we assume that $T(z - z_0) = 1$, which according to the estimates of Ref. 3 can lead to an error of about 10% in $R(z)$ at the far end of the path, when calibrating according to the near end). The calibration procedure was performed by graphically combining at points of local minima the plot of the function with the plot of the function $S_m(z)$, as done in Ref. 4.

Analysis of the errors in determining $R(z)$, performed by the method of Ref. 6, gave the following expression for the mean square error:

$$\left[\frac{\delta R(z)}{R(z)}\right]^2 = \left[\frac{\sqrt{N(z) + N_b}}{N(z)}\right]^2 + \left[\frac{\sqrt{N(z) = N_b}}{N(z_0)}\right]^2 + (0.037)^2, \tag{4}$$

where N_b is the sum of the background signal from the section of the sky visible by the photometer and the characteristic noise signal of the photomultiplier. This estimate of the error was obtained under the assumption that the photon flux at the receiving antenna of the lidar is a Poisson flux.

With the help of the lidar built based on the telescope it is possible to sound the atmosphere in different directions. This makes it possible to determine the transmittance of the atmosphere and the horizontal nonuniformity of the atmospheric layers and to determine the effect of systematic errors on the observations. To estimate the systematic errors in determining the altitude profiles of the echo signals we employed data obtained by sounding the stratosphere at two angles.

If the sounding is performed in two different directions Θ_1 and Θ_2, then under the condition that the atmosphere is uniform in the horizontal direction the transmittance T for the layer located at the altitude H can be determined from the following expression:

$$T^2(H) = \left[\frac{\sec^2\Theta_1}{\sec^2\Theta_2} \cdot \frac{N(H, \Theta_1)}{N(H, \Theta_2)}\right]^{\frac{1}{\sec\Theta_1 - \sec\Theta_2}}, \tag{5}$$

where $N(H, \Theta_1)$ is the echo signal from the altitude H in the direction of the zenith angle Θ_1. If $\Theta_1 = 60°$ and $\Theta_2 = 0°$, then

$$T^2(H) = \frac{4N(H, 60°)}{N(H, 0°)} . \tag{6}$$

The main interference in determining the transmittance by the method of multiangle sounding is the horizontal nonuniformity of the atmosphere and different errors in measurements of the amplitudes of the lidar signals, which significantly affect the results. This makes it difficult to use the lidar method for routine determination of atmospheric transmittance. However, by smoothing the altitude profiles of the echo signals and averaging the data from a series of measurements for separate nights, it is possible to estimate T quite reliably by the method of dual-beam sounding, and then. by comparing with the transmittance data obtained by an independent method based on the stars, to estimate the maximum error in the determination of T and then the systematic error in the amplitudes of the lidar signals. To reduce the error in the two-angle method of determining the atmospheric transmittance the values of $N(H, \Theta_1)$ were determined after smoothing the experimental profiles of the backscattering signals with a curve drawn through two or more local minima, i.e., the experimental profile of the lidar signal was actually replaced by the normalizing molecular–scattering profile or a curve close to it. This method makes it possible to avoid errors associated with the spatio-temporal nonuniformity of the atmosphere. The atmospheric transmittance determined in this manner from the location of the lidar up to altitudes of 16–18 km, averaged over 20 series of measurements in September 1984, was found to be equal to

0.68±0.02. Taking into account the attenuation of light by the higher layers according to the model of Ref. 3 reduces the transmittance approximately by 0.02 and gives the value $T = 0.84$.

According to data on the spectral transmittance according to the stars,[7] obtained in 1983–1984 in the region of the lidar setup, the average transmittance $T = 0.87$ during the winter and 0.83±0.03 in the summer-fall. (Here the maximum deviation from the average is indicated.) It is close to the result presented above; this makes it possible to take the maximum absolute error in determining the transmittance $\varepsilon_T \approx 0.03$ and the maximum relative error as $\varepsilon_T/T \approx 0.04$.

The correctness of the transmittance determined by the method of sounding in two directions indicates that there are no significant systematic errors in the altitude profiles of the echo signals associated with the possible parasitic irradiations (reflection from nearby objects, walls and surfaces in the pavilion, light from the laser pumping lamp at the moment of the pulse, scattering from the near zone, afterglow effect, etc.). Indeed, the transmittance T is a continuous differentiable function of the echo signals, measured in the directions Θ_1 and Θ_2: $x = N(H, \Theta_1)$ and $y = N(H, \Theta_2)$, respectively. For this reason the maximum absolute error in the measurement of the transmittance is determined as follows:[6]

$$\varepsilon_T = \left|\frac{\partial T}{\partial x}\right|_{\substack{x=N(\Theta_1)\\ y=N(\Theta_2)}} \cdot \varepsilon_x + \left|\frac{\partial T}{\partial y}\right|_{\substack{x=N(\Theta_1)\\ y=N(\Theta_2)}} \cdot \varepsilon_y,$$

where ε_x and ε_y are the maximum absolute errors in the arguments. For $\Theta_1 = 60°$ and $\Theta_2 = 0°$ we obtain from Eq. (4)

$$\varepsilon_T = (x\cdot y)^{-1/2}\cdot\varepsilon_x + x^{1/2}y^{3/2}\cdot\varepsilon_y,$$

$$\frac{\varepsilon_T}{T} = \frac{1}{2}\left[\frac{\varepsilon_x}{x} + \frac{\varepsilon_y}{y}\right].$$

If the relative errors in the measurements of the echo signals from the altitude H are identical, then $\varepsilon_T/T = \varepsilon_x/x = \varepsilon_y/y$. Since $\varepsilon_T/T \le 4\%$ the relative systematic error in the average profiles of the echo signals also does not exceed 4%. In the general case the sum of the maximum relative errors of the echo signals measured in two directions does not exceed 8%.

In Ref. 8 the case when the absolute values of the systematic errors in the directions 60° and 0° are equal ($\varepsilon = \varepsilon_x = \varepsilon_y$) is studied. In practice this can happen if the background is not taken into account correctly, parasitic irradiations are present (including the possible effect of light from the pumping lamp and reflected from nearby objects), the afterglow of the photomultiplier, etc. For the case when $\varepsilon = \varepsilon_x = \varepsilon_y$ the true transmittance $T(H)$ and the value determined from the measurements $T_{meas}(H)$ are related as follows:[8]

$$T^2_{meas}(H) = T^2(H) \pm \frac{[4 - T^2(H)]\cdot\varepsilon}{N(H, 0°)}. \qquad (7)$$

The plus sign on the right side of Eq. (7) corresponds to a systematic overestimation while the minus sign corresponds to systematic underestimation of the measured signal $N(H, 0°)$. One can see from Eq. (7), for example, that for $\varepsilon = 0.1N(H, 0°)$ and $T^2 = 0.7$ ($T = 0.837$), $T^2_{meas} = 0.33$, i.e., the error in estimating the squared transmittance is equal to about 50% while the error in determining T is equal to about 30%, which is significantly greater than the maximum absolute error in the determination of the transmittance ($\varepsilon_T = 3\%$) and suggests that such systematic errors are negligibly small.

For the analysis we used data from about 50 nights of observations over the period from September 1984 to January 1988. The number of altitude profiles obtained during each night fluctuated from several profiles to several tens of profiles with a spatial resolution for most of them equal to 0.96 km. (The number of laser pulses in obtaining each profile $\approx 10^3$–10^4). In the analysis the data for each separate night were first averaged, after which the results were analyzed and the seasonal and temporal dependences were determined (Table I). For the summer and fall seasons, aside from the average behavior $R(H)$, the rms error of a separate measurement δ is presented.

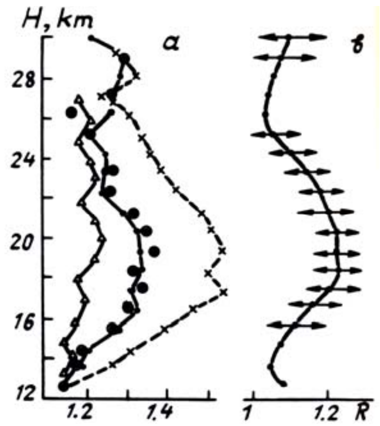

FIG. 1. The average altitude behavior of the backscattering ratio R(H): a) summer (dots), fall (circles) and winter (crosses). The triangles correspond to the winter of 1985. 6) August 19, 1986.

Analysis of the data shows that the summer altitude profiles of the backscattering coefficients $R(H)$ are close to the fall profiles (Fig. la). A layer of high turbidity (Jung's layer) can be distinguished in the profiles. Averaging significant smooths the fine structure of the profiles in separate observations. However the average profiles contain, together with the main maximum, an additional maximum at an altitude of 26-28 km. In the main maximum there are also two and sometimes three peaks, corresponding to the often observed layered nature of the stratosphere on the path section studied.

The values of the backscattering ratio at the maximum of Jung's layer $R_{max}(H)$ on the average are equal to 1.33 in the summer and 1.36 in the fall. The spring and winter data are sparse. The variations of $R(H)$ during these seasons are maximum. The rms deviation for the same number of nights during the winter, for example, is two to three times greater than the analogous value for the fall.

The altitude behavior of the winter and spring curves $R(H)$ for different years is different. In this connection Table I gives for these seasons separate results, obtained during different years. In 1985 and 1988 the values of $R(H)$ are less than the analogous average values for the entire period for the summer season in the entire section of the path (12–30 km). At the same time in the winter-spring period of 1986 and partially in 1987 stronger scattering of light by the stratospheric aerosol was observed. A thicker Jung layer was observed on February 27, 1986. The value of $R_{max}(H)$ on this day was found to be equal to 2.5–2.7. This could be connected with the effect of the eruption of the volcano Nevada-del-Ruis on November 13, 1985. The high turbidity of the atmosphere in the winter of 1985–1986 was also noted in other works.[11] However the values of $R(H)$ during the summer of 1986 are already close to the data for other years. The statistical sample for the winter and spring results is inadequate, and the question of the seasonal variations in the scattering properties of the stratospheric aerosol requires further Investigation.

TABLE I.

Altitude profiles $\bar{R}(H)$

Altitude a. s. l., km	IX. 1984-VII. 1987		Winter				Spring		
	Summer	Fall	1985	1985-1986	1986-1987	January 1988	1985	1986	1987
12.62	1.14±0.02	1.16±0.02	1.04	1.03	1.21	—	1.16	1.28.	1.07
13.58	1.17±0.02	1.20±0.03	1.07	1.35	1.35	—	1.20	1.48	1.18
14.54	1.19±0.03	1.26±0.03	1.11	1.39	1.39	—	1.12	1.69	1.30
15.50	1.28±0.05	1.30±0.04	1.06	1.50	1.50	—	1.16	1.88	1.44
16.46	1.32	1.32±0.05	1.12	1.51	1.51	1.03	1.28	1.44	1.51
17.42	1.30±0.04	1.34±0.04	1.18	1.56	1.56	1.06	1.18	1.35	1.67
18.38	1.33±0.03	1.36±0.02	1.13	1.55	1.55	1.08	1.16	1.56	1.68
19.34	1.32±0.04	1.34±0.04	1.23	1.50	1.50	1.12	1.22	1.43	1.73
20.30	1.33±0.04	1.36±0.06	1.26	1.48	1.48	1.15	1.26	1.36	1.70
21.26	1.27±0.04	1.34±0.03	1.25	1.43	1.43	1.17	1.20	1.84	1.68
22.22	1.23±0.02	1.25±0.04	1.18	1.35	1.35	1.11	1.19	1.34	1.67
23.18	1.25±0.04	1.25±0.07	1.26	1.30	1.30	1.10	1.19	1.93	1.57
24.14	1.21±0.03	1.32±0.06	1.26	1.26	1.26	1.08	1.12	1.84	1.50
25.10	1.19±0.03	1.22±0.04	1.17	1.16	1.16	1.08	1.08	2.04	1.50
26.06	1.24±0.04	1.29±0.05	1.22	1.16	1.16	1.05	1.24	1.56	1.42
27.02	1.24±0.06	1.28±0.04	1.18	1.12	1.12	1.03	1.22	1.12	1.33
27.98	1.33±0.06	1.23±0.05	—	1.14	1.14	1.05	1.15	1.06	1.27
28.94	1.33±0.05	1.25±0.08	—	1.09	1.09	1.05	1.23	2.43	1.92
29.90	1.17±0.05	1.31±0.05	—	1.10		1.06	1.24	1.07	1.18

Figure 2 shows a histogram of the distribution of the backscattering ratio at the maximum of the Jung layer $R_{\max}(H)$ in the summer-fall period based on observations performed in 1984–1987. The curve is asymmetric. The most probable value (1.25) does not agree with the average value (1.34). For a large statistical sample, in constructing histograms the processing included all values of the backscattering ratio obtained at altitudes from 17.4 to 21.6 km above sea level; these values are close to the maximum.

In order to eliminate the seasonal variations in $R(H)$ in determining the temporal behavior of the altitude profiles of the stratospheric aerosol in the period 1984–1987, the data obtained in August and partially (for 1984–1986) in September were compared. The results are presented in Table II and indicate that there is a small difference in the profiles $R(H)$ obtained in different years during the summer.

TABLE II

The altitude profiles R(H) in different years

No	H, km	August			September	
		1985	1986	1987	1984	1985
1	12.62	1.06	1.09	1.17	1.08	1.20
2	13.56	1.07	1.14	1.16	1.16	1.22
3	14.54	1.12	1.18	1.18	1.16	1.31
4	15.50	1.18	1.14	1.22	1.18	1.36
5	16.46	1.23	1.19	1.22	1.17	1.40
6	17.42	1.23	1.22	1.26	1.21	1.40
7	18.38	1.31	1.27	1.25	1.32	1.38
8	19.34	1.33	1.22	1.23	1.30	1.36
9	20.30	1.19	1.26	1.21	1.21	1.43
10	21.26	1.32	1.21	1.14	1.36	1.34
11	22.22	1.19	1.15	1.15	1.33	1.21
12	23.18	1.20	1.15	1.07	1.27	1.24
13	24.14	1.13	1.17	1.06	1.41	1.28
14	25.10	1.10	1.17	1.10	1.16	1.25
15	26.06	1.27	1.12	1.10	1.36	1.26
16	27.02	1.16	1.10	1.04	1.31	1.26
17	27.98	1.48	1.16	1.07	1.24	1.23
18	9.90	1.42	1.18	1.17	1.20	1.27
19	29.94	1.17	1.08	1.09	1.16	1.38

The minimum values of $R(H)$ were observed during the summer. On separate nights a large number of altitude profiles of echo signals were obtained. This made it possible to study the average profile $R(H)$ and the distribution of the backscattering ratio at the maximum of the Jung layer $R_{max}(H)$ for separate characteristic summer night. Thus, statistically well-founded sounding data were obtained on August 19, 1987. A total of 35 series of measurements were performed. The average altitude profile $R(H)$ for this day is presented in Fig. 1b. The arrows in the figure mark the rms deviation δ.

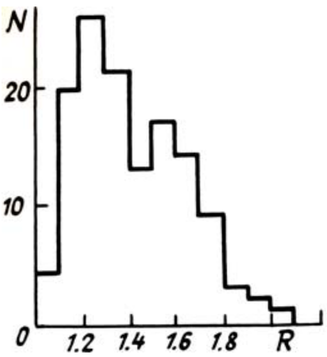

FIG. 2. Histogram of the distribution of the backscattering ratio at the maximum of the Jung layer $R_{max}(H)$ for the summer period of 1984–1987.

Figure 3 shows, a histogram of the distribution $R_{max}(H)$ for August 19, 1987 and the normal distribution (smooth curve), which approximates satisfactorily the experimental histogram. In constructing the histogram for a large statistical sample, as before, all values of $R(H)$ obtained at altitudes from 17.42 to 21.6 km above sea level were included in the analysis.

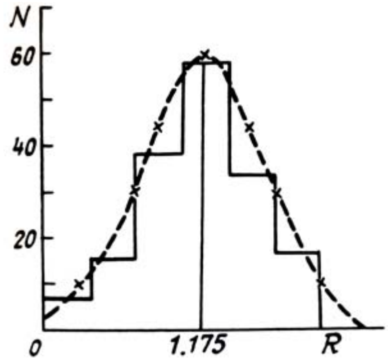

FIG. 3. Histogram of the distribution of the backscattering ratio at the maximum of the Jung layer $R_{max}(H)$ on August 19, 1986.

The confidence intervals of the most probable value $R_{max}(H)$, given with statistical confidence levels of 68, 83, and 95%, were determined with the help of probabilistic methods. They are presented in Table III.

TABLE III.

Confidence probability, %	Most probable $R_{max}(H)$	Confidence interval
68	1.175	1.100-1.250
83	1.175	1.075-1.275
95	1.175	1,050-1.300

The data in Table III show that on good summer nights the peak value of $R(H)$ in the Jung layer is small (1.17) and is only insignificantly greater than the analogous value (1.11) presented in Ref. 12 for a nonvolcanic period (Virginia, 37°N).

It was noted above that there is small difference in the average profiles $R(H)$ obtained in August in different years (1984–1985). The peak value of the average profile $R(H)$ in 1985 (1.35) is only slightly greater than the virtually identical values obtained in 1986 and 1987 (1.27 and 1.26) (see Table II).

Judging from these data, in 1986 and 1987 the optical properties of the stratosphere were significantly stabilized after the eruption of the volcano El Chichon in the spring of 1982. However the comparison of the optical parameters of the stratospheric aerosol with the background values, the determination of the possible temporal behavior of the integral backscattering coefficient, the comparison of our results with data obtained by other authors, and analysis of the variations of the altitude of the main maximum of the Jung layer are subjects of further investigations.

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