LASER SOUNDING OF STRATOSPHERIC AEROSOL AT 532 AND 1064 nm WAVELENGTHS UNDER SATOR AND LITE PROGRAMS IN 1994

V.V. Zuev, A.V. El'nikov, V.D. Burlakov, M.V. Grishaev, and V.L. Pravdin

Institute of Atmospheric Optics,
Siberian Branch of the Russian Academy of Sciences, Tomsk
Received December 29,1994

We present here some results of lidar studies of the aerosol vertical distribution in the stratosphere, obtained in 1994. These results are a part of a long-term series of observations. The results obtained show that the aerosol content in the stratosphere, polluted in 1991 by volcanic products of Pinatubo eruption, is now approaching its background level. A qualitative assessment of the aerosol particle size–distribution in the stratosphere is carried out based on the results of two-frequency laser sounding at 532 *and* 1064 nm *wavelengths.*

Since 1986 our group regularly performs laser sounding of the stratospheric aerosol stratification at the wavelengths of 532 nm (Refs. 1 and 2). An increase of the bulk of measurement data due to sounding though at two wavelengths enables one to estimate microphysical and optical parameters of the aerosol.

Laser sounding of the stratospheric aerosol layer (SAL) at the wavelengths of 532 and 1064 nm has been performed in summer of 1994 when conducting measurements under the SATOR complex ecological program by the Institute of Atmospheric Optics[3] and in September 1994 under the program of ground-based correlative measurements within the framework of the NASA LITE (Lidar In – Space Technology Experiment) Program.[4] In this case the laser sounding at our lidar station[5] was performed simultaneously with the laser sounding from onboard the Shuttle spacecraft with a spaceborne lidar. The final summary of the results, obtained when performing investigations under the above-mentioned programs, will be published after final processing of all the experimental data. This paper presents only preliminary results on the sounding stratospheric aerosol obtained during this period.

In the experiments discussed we have employed a Nd:YAG laser with energy per pulse of 60 mJ at the wavelength of 532 nm and 120 mJ at a wavelength of 1064 nm, the pulse repetition rate and the beam divergence being 10 Hz and 0.15 mrad.

The lidar returns have been recorded immediately in the focal plane of a receiving mirror 2.2 m in diameter and 10 m in focal length. Spectroscopic device coupled with a PMT is mounted at a special washer. A field stop 3 mm in diameter, matched with a sounding beam image, determines the viewing angle of the receiving antenna to be 0.3 mrad that enables us to strongly decrease the background radiation. After passing through the field stop the collected radiation is collimated by a short-focus lens to a dichroic mirror, then passes through narrow-band interference filters (the maximum transmission is $\tau_{max} = 60\%$; the transmission band width at the level of $0.5\,\tau_{max}$ is $\Delta\lambda_{0.5} = 2$ nm at $\lambda = 532$ nm and $\Delta\lambda_{0.5} = 3.9$ nm at $\lambda = 1064$ nm) and further it is focused by short-focus lenses to photocathodes of PMTs. To record signals at 532 nm, we use a photomultiplier FEU–130. For recording signals at 1064 nm wavelength, where modern photomultipliers have low quantum efficiency, we use a specially selected photomultiplier FEU–83, which is cooled to the temperature –30°C to decrease thermal noise. In this case lidar returns at 1064 nm are reliably recorded at heights up to 25–30 km when accumulating signal during 30 min.

Signals are recorded in the photon counting mode using a two–channel counter with 512 time gates, each 100 m long. For cutting off the near zone of lidar return we use a mechanical chopper in front of the field stop, which operates in synchronism with the laser.

One of the parameters, determined from the laser sounding data, is the scattering ratio $R(H)$

$$R = 1 + \beta_\pi^a(H)/\beta_\pi^m(H), \qquad (1)$$

where $\beta_\pi^a(H)$ and $\beta_\pi^m(H)$ are the coefficients of aerosol and molecular backscattering, respectively. The technique of their determination is presented in detail in Ref. 6. When using the lidar with the above characteristics, for obtaining the scattering ratio with a reasonable accuracy the lidar returns are accumulated within a half-hour interval.

However, the spatial resolution of the scattering ratio determined is worse than the potential one and is 400 m. Similar space-time characteristics for R enabled one to reduce the errors of determining the scattering ratio at $\lambda = 1064$ nm to 5% at 25 km altitude and at $\lambda = 532$ nm to less than 3% even at 30 km altitude.

0235-6880/95/10 813-04 $02.00

Figure 1 presents three profiles of the scattering ratio at $\lambda = 532$ (*a*, *c*) and 1064 nm (*b*, *d*) obtained during summer (June 22, 24, 27) and fall (September 15, 17, 19) of 1994. It is evident that aerosol stratification for one and the same day is identical at both sounding wavelengths. The extreme points at the scattering ratio profile at one wavelength correspond to analogous extrema at the another one. Maximum values of the scattering ratio at $\lambda = 1064$ nm are higher ($R \sim 2$ in summer and 2.5 in fall) than those at $\lambda = 532$ nm ($R \sim 1.2$ in summer and 1.25 in fall). In this case the altitude of these maxima in summer ($H \sim 19$ km) at both $\lambda = 532$ and 1064 nm is higher as compared with that in fall ($H \sim 17$ km). Such variations of maximum altitude and maximum value are observed since 1987 and present the regular seasonal process, characteristic of any season[1] and obviously conditioned by the same reason as the seasonal decrease in the tropopause altitude from summer to winter and its increase from winter to summer.

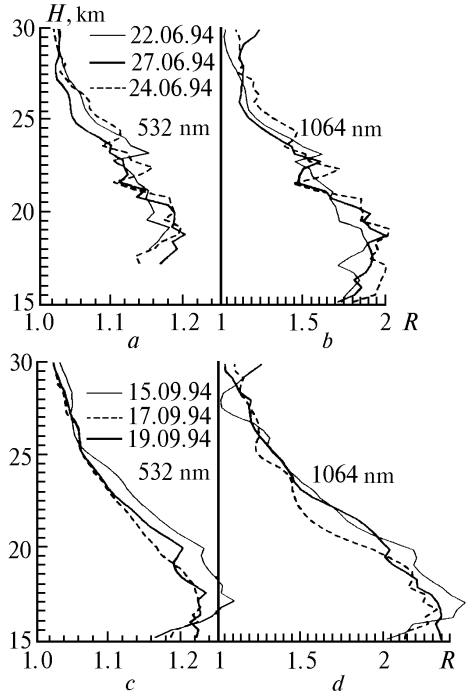

FIG. 1. Altitude profiles of scattering ratio.

Besides, for summer profiles of R in the altitude range from 17 to 24 km a pronounced laminar structure is observed. Similar stratification of the scattering ratio for this season is evidently determined by the remainders of aerosol of volcanic origin emitted to the stratosphere in 1991 during the Mount. Pinatubo eruption and by weak intensity of mixing processes. By fall the aerosol laminar structures in the stratosphere at altitude higher than 20 km disappear due to more intense mixing and circulation processes.

For a qualitative estimate of the aerosol particle size spectrum, we used the parameter X (Ref. 7) representing the exponent in the relationship:

$$\beta_\pi^a(\lambda_1)/\beta_\pi^a(\lambda_2) = (\lambda_2/\lambda_1)^X, \qquad (2)$$

where $\beta_\pi^a(\lambda_1)$ and $\beta_\pi^a(\lambda_2)$ are the backscattering coefficients of one and the same aerosol at the wavelengths λ_1 and λ_2. It is believed that the stratospheric aerosol is of one nature over the entire altitude range. In this case small values of X must characterize the presence of large particles, and vice versa, large values of X characterize the fine aerosol. Since for the coefficients of molecular backscatter the ratio $\beta_\pi^m(\lambda_1)/\beta_\pi^m(\lambda_2) = (\lambda_2/\lambda_1)^4$ is valid and in our case $\lambda_2/\lambda_1 = 2$, we have the final expression for the profile $X(H)$:

$$X(H) = 4 + \frac{1}{\ln 2} \ln \left\{ \frac{R_1(H) - 1}{R_2(H) - 1} \right\} . \qquad (3)$$

The scheme of $X(H)$ calculation is very sensitive to the errors of determination of the scattering ratio profiles $R(H)$.

To decrease the error of the profile of X determined by Eq. (3) we use the linear smoothing by the method of moving average. In this case the smoothing intervals are given from the 5% error of X determination. As to the spatial resolution in the X profiles, it becomes worse, being in the range of 16, 20–21, and 24 km the resolution is about ~1.2, ~2, and ~ 2.8 km, respectively.

The profiles of $X(H)$, shown in Fig. 2, are determined from sounding data on the same days of June (Fig. 2*a*) and September (Fig. 2*b*) as the data shown in Fig. 1. On the whole, the behavior of the $X(H)$ profiles both in summer and in fall should characterize the process of aerosol particle sedimentation. As a result of this process, near the tropopause the large-mass and large-size particles are accumulated. This results in the situation when the major contribution to the total aerosol content in the atmosphere, at the upper levels of the altitude range studied, comes from fine aerosol particles (larger X values) in contrast to the situation at the low levels (Fig. 2). For the summer profiles (Fig. 2*a*) a stable behavior of $X(H)$ profiles is typical in the altitude range from 19 to 23 km (with the growth rate of the parameter $X \sim 0.04$/km) as well as their large difference from the fall ones. The summer values of $X(H)$ over the entire altitude range are larger than the fall values. In this case the latter ones have different altitude behavior (Fig. 2*b*). On September 15 the value of X increases with the rate of ~0.05/km, and on September 19, 1994, the value of X increases with the rate of ~0.02/km. While on September 17, 1994, the altitude behavior of X has three features, which differ in the growth rate: 15–21 km – ~0.02/km, 21–22.5 km – ~0.18/km, and 22.5–25 km – ~0.05/km. With the altitude growth this results in a significant difference in the X values, although at the beginning of the 15 km altitude range the X values for these days are

practically the same and equal to ~1.4. Thus, on September 15 we have observed the maximum values of X (1.4–1.9) of three days of observations, and on September 19 the minimum values (1.4–1.6) occurred.

As to the altitude behavior of X on September 17, it coincides, up to 21 km, with the altitude behavior of X on September 19, 1994. Its value sharply increases at altitudes above 21 km and from 22.5 km altitude it practically coincides with $X(H)$ on September 15. From the viewpoint of the aerosol particle size, such a behavior of $X(H)$ on these three days can be treated as the replacement of the fine aerosol typical for stratospheric aerosol layer on September 15, 1994. This process was completed by September 19, 1994. Displacement of aerosol occurred gradually covering the higher altitude levels. This fact is demonstrated by the $X(H)$ altitude behavior on September 17, 1994, occupying the intermediate position between the dates of beginning and completion of the process and having the transition region (21–22.5 km) from fine aerosol on September 15, 1994, to the coarse one on September 19, 1994. The recorded process is evidently connected with the unstable synoptic situation during this period of lidar observations.

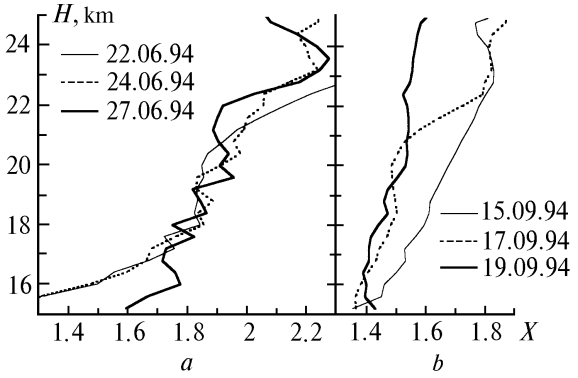

FIG. 2 Altitude behavior of the parameter X.

Figure 3 presents the mean profile of aerosol backscattering coefficients obtained from the sounding data on June 9, 15, 16, 20, 22, 24, and 27, 1994, at 532 nm wavelength. This profile is shown in comparison with the mean summer profiles of aerosol backscattering coefficients obtained earlier in 1987, 1988, and 1989 (Ref. 1). At altitudes below 20 km the mean June profile of $\beta_\pi^a(H)$ exceeds the values of $\beta_\pi^a(H)$ for 1987, 1988, 1989. This is due to residual phenomena of the stratospheric pollution with the products from Mount. Pinatubo eruption. In this case the shape of the $\beta_\pi^a(H)$ profile in 1994 is analogous to the profile in 1987 as concerning the distinct maximum at the altitudes of the Junge layer. Recalling that in December of 1985 the Ruis volcano eruption took place,[8] being less strong than the Pinatubo eruption, we can assume that the Junge layer is of volcanic origin. The lack of maximum in $\beta_\pi^a(H)$, i.e., the Junge layer in the profiles of 1988 and 1989, characterizing the background unperturbed state of the stratosphere, also counts in favor of this assumption.

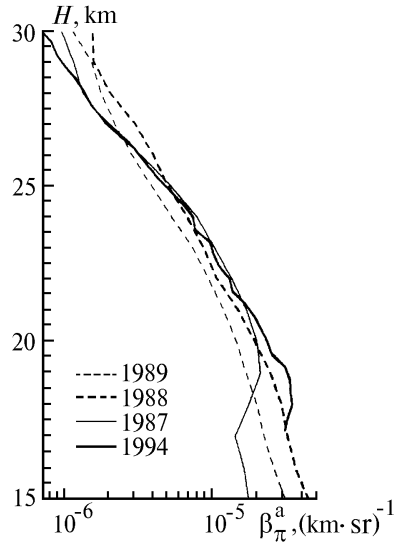

FIG. 3 Mean profile of the aerosol backscattering coefficients.

Thus, the above fragments from the measurement series of the laser sounding of stratospheric aerosol in summer and fall of 1994 indicate that although there are some features of residual influence of the Mount. Pinatubo eruption in 1991 (laminar structure of the scattering ratio profiles, increased values of aerosol backscattering coefficients) as a whole the stratosphere in its aerosol content is close to its background state. In this case simultaneous sounding at two wavelengths has made it possible to study qualitatively the aerosol particle size distribution and to determine the sedimentation effect on the above distribution. Against the background of this effect, the seasonal difference in aerosol particle size can be determined (in summer in the stratosphere the aerosol particle size is less than in fall).

ACKNOWLEDGMENTS

The authors would like to express their gratitude to M.I. Andreev, A.V. Nevzorov, and E.V. Sharabarin for their active participation in preparation and performance of the experiments.

The work was supported in part by the Russian Fundamental Research Foundation Grant No. 93-05-9383.

REFERENCES

1. A.V. El'nikov, V.V. Zuev, and V.N. Marichev, Atm. Opt. **4**, No. 6, 462–465 (1991).
2. V.D. Burlakov, A.V. El'nikov, V.V. Zuev, et al., Atmos. Oceanic Opt. **5**, No. 10, 664–667 (1992).

3. V.V. Zuev, Atmos. Oceanic Opt. **5**, No. 6, 354–359 (1992).
4. M.P. McCormick, D.M. Winker, E.V. Browell, et al., Bulletin of the American Meteorological Society **74**, No. 2, 205–214 (1993).
5. V.V. Zuev, V.D. Burlakov, M.V. Grishaev, et al., in: *Abstracts of Papers of 17th ILRC*, Sendai, Japan (1994), pp. 489–490.
6. A.V. El'nikov, V.V. Zuev, and V.N. Marichev, Atm. Opt. **4**, No. 2, 175–183 (1991).
7. E.D. Hinkly, ed., *Laser Monitoring of the Atmosphere* (Springer Verlag, New York, 1976).
8. A.V. El'nikov, G.M. Krekov, and V.N. Marichev, Izv. Akad. Nauk SSSR, Fiz. Atmos. Okeana **24**, No. 8, 818–823 (1988).