The Vertical Change of Stratospheric Aerosol Size Measured by Two-Color Laser Radar Technique

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(Received June 11, 1979; Revised November 30, 1979)

The stratospheric sulfate aerosol layer has been a subject of considerable interest since it has been discovered some 20 years ago by JUNGE and his co-workers (JUNGE, 1961; JUNGE and MANSON, 1961; JUNGE *et al.*, 1961). Concern that man's activities might cause significant perturbations of the stratospheric aerosol has led to a recent intensification in research efforts to measure properties of the layer and characteristics of the particle. Consequently, there presently exist various observational data on the stratospheric aerosol. Nevertheless, many fundamental questions about microphysical processes which control the formation and morphology of the aerosol layer have remained unanswered. The knowledge about the vertical change of aerosol size distribution is one of the most important information to understand these processes, and to assess more accurately the optical effect to aerosols on stratospheric environment.

It may be fully interesting to present the vertical change of the stratospheric aerosol size measured remotely by laser radar technique, although the observational results concerning this have been already presented by other investigators, because all previous information was due to the direct measurements such as aerosol sampling method, balloon-borne photoelectric aerosol counter technique and so on. The results described here are due to measurements at Nagoya, Japan (35°N, 137°E) during winter, January–March 1977, using two-color laser radar system which has an advantage point that this can emit laser pulse at two different wavelengths, 0.6943 μm (Ruby laser) and 1.06 μm (YAG laser) by turns, maximum repetition rate is 0.2/sec. The detail and discussion for the determination of size distribution function of stratospheric aerosol by two-color laser radar technique has been already described in other papers (IWASAKA, 1977; IWASAKA and ISONO, 1977), then we will not repeat it here.

The backscattering function, B_i , is given by following equation

$$B_i(Z) = \int n(R, Z) \sigma_i(R) dR, \quad (1)$$

where

Z =altitude

R =radius of aerosol

i =suffix for the wavelength of laser radar, and $i=1$ and 2 are corre-

sponding to wavelength=0.6943 and 1.06 μm respectively.

$n(R, Z)$ =size distribution function at altitude Z , and aerosol density within the radius range R_1 – R_2 is calculated by $\int_{R_2}^{R_1} n(R, Z) dR$.

$\sigma_i(R)$ =Mie differential back scattering cross-section at wavelength= i .

Here, the backscattering functions of aerosols were decided using “Matching Method,” and the altitude of matching point used in this report was found near 9 km. The meteorological data measured by sounding of sonde at Hamamatsu meteorological observatory (100 km SW of the laser-radar site) are used to decide the dust-free atmosphere model. It should be pointed out here that we cannot gain adequate backscattered photons per small number of transmitted pulses owing to the low quantum efficiency of photomultiplier used for the detection of scattering light at YAG laser wavelength (1.06 μm). Accordingly, it is necessary to integrate some received power to achieve adequate signal-noise ratio. We received a few photons due to backscattering of aerosols and atmospheric molecules per 50 YAG transmitted laser pulses from upper region of stratospheric aerosol layer.

Fig. 1. The profiles of aerosol backscattering coefficients B_1 and B_2 . B_1 and B_2 are coefficients measured at wavelength=0.6943 and 1.06 μm respectively.

Curves in Fig. 1 are average profiles of aerosol backscattering coefficient in winter season 1977 (9 Jan., 24 Jan., 19 Feb., 10 March, 27 March, and 28 March). The symbol in figure means the range of dispersion due to both the component of day to day variation and statistical error of photon counting method.

The profile of backscattering function ratio, B_1/B_2 shown in Fig. 2 is the results due to laser radar measurements in winter 1977 during which we emitted 1,200 laser pulses with average energy 0.6 J/pulse at each wavelengths, 0.6943 and 1.06 μm respectively. The feature of profile of B_1/B_2 is very interesting, and it can be easily found

that the minimum height of vertical profile of B_1/B_2 is in the region where we frequently observed the density maximum of stratospheric aerosols. Curves in Fig. 3 are atmospheric temperatures measured at Hamamatsu meteorological observatory which date are all corresponding to date of laser radar measurements. The temperature profiles are indicating that the tropopause heights containing first and second tropopause levels are in the range from 8 to 16 km, and average height of all is found at 14 km. According to IWASAKA *et al.* (1976), backscattering phenomena of aerosols at YAG laser wavelength are more sensitive for the larger size aerosols in comparison with that at Ruby wavelength, and we therefore can conclude that the region from 18 to 21 km is noticeably rich in the larger size particles. The measurements by PINNICK *et al.* (1976) shown in Fig. 4 is one of the most interesting results to compare with present results. Both observations clearly show that the relative density of the larger size particle has maximum near the center of layer.

Fig. 2

Fig. 3

Fig. 2. The ratio of backscattering function, B_1/B_2 . The bar --- means error range of measurements. Average tropopause height during Jan.-March, 1977 is indicated by the arrow.

Fig. 3. Temperature profiles of 9 Jan., 24 Jan., 19 Feb., 10 March, 27 March, and 28 March, measured at Hamamatsu.

In Table 1, we listed values of parameter γ which are decided by the analytical procedure of IWASAKA and ISONO (1977) on assumption that the size distribution function are described as the power law type of aerosol radius R , $n(R) = AR^{-\gamma}$ in the radius range from 0.1 to 1.0 μm , and compared with the measurements by MIRANDA and FENN (1974). The parameter γ becomes minimum in the region from 18 to 21 km, and it is corresponding to the fact that this region contains relatively more the larger size particles. Measurements by MIRANDA and FENN (1974) also suggest that parameter γ becomes larger according to increase of altitude in the upper region of layer. In addition to these, similar feature can be found in other investigators' measurements (BIGG, 1975, 1976; HOFMANN *et al.*, 1975; HOFMANN and ROSEN, 1977; FARLOW

Fig. 4

Fig. 4. Aerosol concentration ratio determined by photoelectric particle counter [number (radius > 0.15 μm)/number (radius > 0.25 μm)] versus pressure for 22 balloon flights over Laramie from December 1971 through February 1974. The bar $\overline{\quad}$ is the range of the observed tropopause heights (after PINNICK *et al.*, 1976).

Fig. 5

Fig. 5. The size distribution function, $n(R)$ determined on the basis of laser radar measurements. Curve J means Junge type function with $\gamma=4$ which is fitted to the curve of region, $Z=18-21$ km at $R=1.0$ μm to compare with present results.

Table 1. Parameter of size distribution function.

Altitude (km)	γ on laser radar measurements
10.5-12	5.4
12 -15	5.2
15 -18	4.8
18 -21	4.1
21 -24	5
	γ by MIRANDA and FENN (1974)
20.7-22.5	4.4
22.5-24.7	7.0
24.7-26.5	16.5

γ is the parameter of size distribution function $n(R) \propto R^{-\gamma}$, and aerosol number density in the radius range R_1-R_2 is given by $\int_{R_2}^{R_1} n(R) dR$.

et al., 1979). In Fig. 5, we show the size distribution function, $n(R)$ estimated on the basis of laser radar measurements in winter 1977. The curve of region from 18 to 21 km is very close by Junge type function with $\gamma=4$, and all other curves have more

steep gradient ($\gamma > 4$). Recently, HAMILL *et al.* (1977) examined various and chemical processes controlling the behaviour of stratospheric aerosol particles such as nucleation, condensation, evaporation, and sedimentation, and pointed out that the upper altitude limit of layer was due to the evaporation of sulfuric acid from aerosol particles, while the lower limit was due to mixing of Aitken size particles (smaller size than $0.1 \mu\text{m}$ radius) across the tropopause. This concept seems to be very reasonable to interpret our observational results, in other words, the active evaporation of H_2SO_4 gas from sulfate particle near the top of layer makes the average size of particle to be smaller and the particle found near the bottom of layer must be smaller size aerosols since larger size particle is easily removed out through the strong rain-out process in the troposphere and not transported to stratosphere. Near the center of layer, both condensation and coagulation processes make the average size of aerosol to be larger one. Consequently, the region near center of layer is rich in the larger size particles compared with other region. Of course our knowledge of stratospheric aerosol layer is far from complete and we therefore cannot definitely conclude this. There is a need for more observational and experimental work on the nature of the stratospheric particles.

The author should like to express many thanks to Professor A. Ono and Emeritus Professor K. Isono, Nagoya University for their continuous encouragement.

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