LIDAR OBSERVATION OF STRATOSPHERIC AEROSOL INCREASE AFTER THE EL CHICHON ERUPTION: NAGOYA, APRIL TO DECEMBER 1982

Sachiko HAYASHIDA and Yasunobu IWASAKA

Water Research Institute, Nagoya University, Furo-cho, Chikusa-ku, Nagoya 464

Abstract: The volcano El Chichon (Mexico) erupted violently over several days at the end of March and at the beginning of April 1982. After the eruptions a sudden increase of backscattered light from the stratospheric aerosol layer was detected by a lidar (0.6943 μm) at Nagoya (35°N, 137°E) in mid-April 1982. At the end of May the observed scattering ratio reached the maximum value about 44. The effect of the eruption of El Chichon on the stratospheric aerosol layer seems to be larger than that of the volcan de Fuego eruption or Mt. St. Helens eruption.

Immediately after the eruption, an apparent two-layer structure of scattering ratio profiles was observed. According to the radio sonde data over Japan, the aerosols in the upper layer were possibly transported by easterly wind and those in the lower layer by westerly wind, respectively. After September 1982 the centroid height of particulate backscattering and the main peak height decreased gradually due to the effect of sedimentation.

1. Introduction

Lidar systems have been used to monitor the stratospheric aerosol content for over twenty years. One advantage of the lidar technique is that vertical profiles of aerosol distribution can be obtained routinely on clear nights. Therefore, lidar measurements have given a lot of valid information about spreading and transport of volcanic dust. And these measurements also help to know the extent of the variation of the stratospheric aerosol content and to estimate the effect of volcanic eruptions on climate (POLLACK *et al.*, 1976). Increase of the aerosol content and subsequent decline of it after large volcanic eruptions have been reported by many researchers, for example, GRAMS and FIOCCO (1967) after the Mt. Agung eruption in 1963, and RUSSELL and HAKE (1977), HIRONO *et al.* (1977), MCCORMICK *et al.* (1978) and IWASAKA (1980) after the volcan de Fuego eruption in 1974.

After the decay of the volcan de Fuego eruption, from 1977 to 1979 the aerosol content in the stratosphere was nearly of the background level. This "quiet period" in the stratosphere was terminated by the La Soufriere and Sierra Negra eruption in 1979 (KENT and PHILIP, 1980; FUJIWARA *et al.*, 1982) and Mt. St. Helens eruption in 1980 (REITER *et al.*, 1980; D'ALTORIO *et al.*, 1981; HIRONO *et al.*, 1981). However, one year after the eruption of Mt. St. Helens, the aerosol content in the stratosphere was reduced to almost the background level (REITER *et al.*, 1982a). In 1981 another increase of the stratospheric aerosol content due to the Mt. Alaid eruption was observed with balloon-borne measurements (HOFMANN and ROSEN, 1981) and ground-based lidar (REITER *et al.*, 1982a), and in January 1982 so-called "mysterious cloud" of unknown source was reported in Japan (HIRONO, private communication), Germany (REITER *et al.*, 1982b) and other places (Scientific Event Alert Network (SEAN) Bull., 1982).

In March and April 1982 the volcano E1 Chichon in Mexico (17°N, 93°W) erupted violently over the period of several days (SEAN Bull., 1982). These eruptions transported extremely large amounts of gaseous and particulate matter into the stratosphere, and large amounts of volcanic aerosols were superimposed on the "mysterious cloud". In mid-April 1982, a tremendously large sudden increase in backscattered light was observed by ground-based lidars in Hawaii, Japan, U.S.A. and Germany (SEAN Bull., 1982; HIRONO and SHIBATA, 1983; IWASAKA *et al.*, 1983). And the obtained values of scattering ratio were larger than any values ever obtained after large volcanic eruptions in the past two decades.

This paper describes stratospheric lidar measurements obtained at Nagoya, Japan (35°N, 135°E) for a period of 8 months (April 1982–December 1982) showing a sudden tremendous increase in the stratospheric aerosol content after the eruption of E1 Chichon. The dense aerosol cloud with scattering ratio about 5 was observed even in March 1983. Therefore, it seems that the effect of the E1 Chichon eruptions on the stratosphere is long lasting. Further observations and examination are necessary to study the decay time of the volcanic aerosol content.

2. Lidar System and Data Analysis

The lidar system used here consists of Q-switched Ruby laser ($\lambda=0.6943\ \mu$m) which produces an output of 0.1–1.0 J/pulse and a 40 cm Newtonian type telescope. A mechanical shutter is used to eliminate laser fluorescence. Parameters of this lidar system are described in detail by IWASAKA *et al.* (1976), and IWASAKA and ISONO (1977).

The profile of backscattering coefficient or scattering ratio is obtained from the average value of the returned signal of a few hundred laser pulse firings. The scattering ratio $R(z)$ is the ratio of the total atmospheric backscatter to the backscatter expected from a molecular atmosphere;

$$R(z)=\frac{\beta_m(z)+\beta_r(z)}{\beta_r(z)}\,, \tag{1}$$

where

β_m: backscattering coefficient due to Mie scattering of aerosol,

β_r: backscattering coefficient due to Rayleigh scattering of atmospheric molecules,

and z represents the altitude. $R(z)-1$ shows the ratio of β_m/β_r, which corresponds to the optical mixing ratio of aerosols.

The vertical resolution of the averaged profiles is 1.5 km. To distinguish the backscatter due to aerosol particles from the backscatter due to air molecules, the profile was normalized to the backscatter from an "aerosol free" region either above or below the main aerosol layer (RUSSELL *et al.*, 1976) and the backscatter from a molecular atmosphere was calculated by using "U.S. Standard Atmosphere" (U.S. Government Office, Washington, D.C., 227 p., 1976).

3. Lidar Measurements

In Fig. 1b, 36 profiles of $R(z)$ measured at Nagoya during the period from April 10 to December 27 in 1982 are shown (IWASAKA *et al.*, 1983). One sample profile (Fig. 1a) before the eruption of Mt. St. Helens (April 2, 1980) which shows the background aerosol content level was compared with those profiles.

The increase of scattering was measured first in the height range of 27.0–28.5 km on April 19 and the scattering ratio (R) was 1.6 (IWASAKA *et al.*, 1983). From April

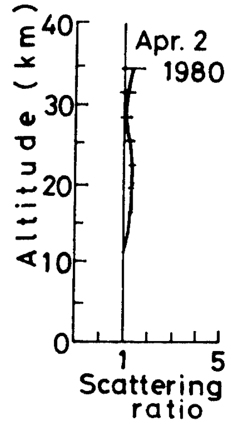

Fig. 1a. The scattering ratio profile before the Mt. St. Helens eruption (April 2, 1980).

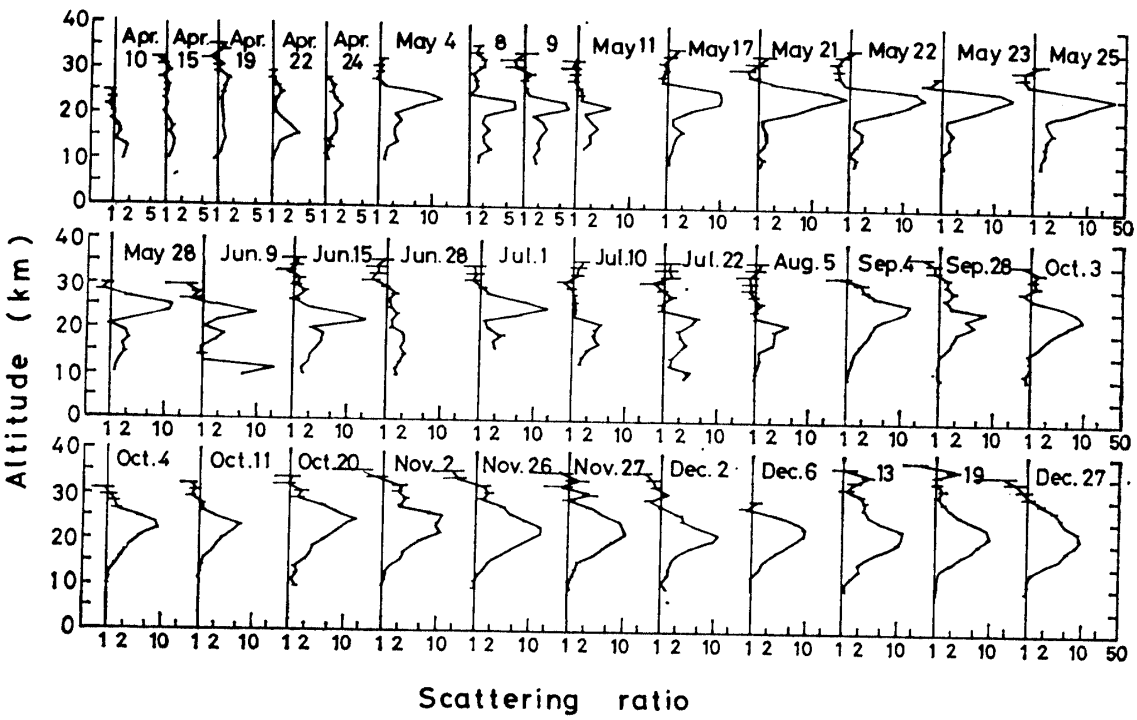

Fig. 1b. The scattering ratio profiles from April to December 1982. Error was shown by "—".

22 to the beginning of July, the apparent two-layer structure was observed except at the end of June. The peak value of R of the upper aerosol layer suddenly increased to 14 on May 4, and after large fluctuations, it reached the maximum value 44 on May 21 at a 24.0–25.5 km level. Afterwards the peak value of R decreased gradually.

From mid-July to early August, nearly all of the increased scattered light returned from the region below 24 km and the peak values of R were 4–5. After September the $R(z)$ profile formed a single broader peak and the height of the peak was about 24 km.

The time-height cross section of the backscattering coefficient is shown in Fig. 2.

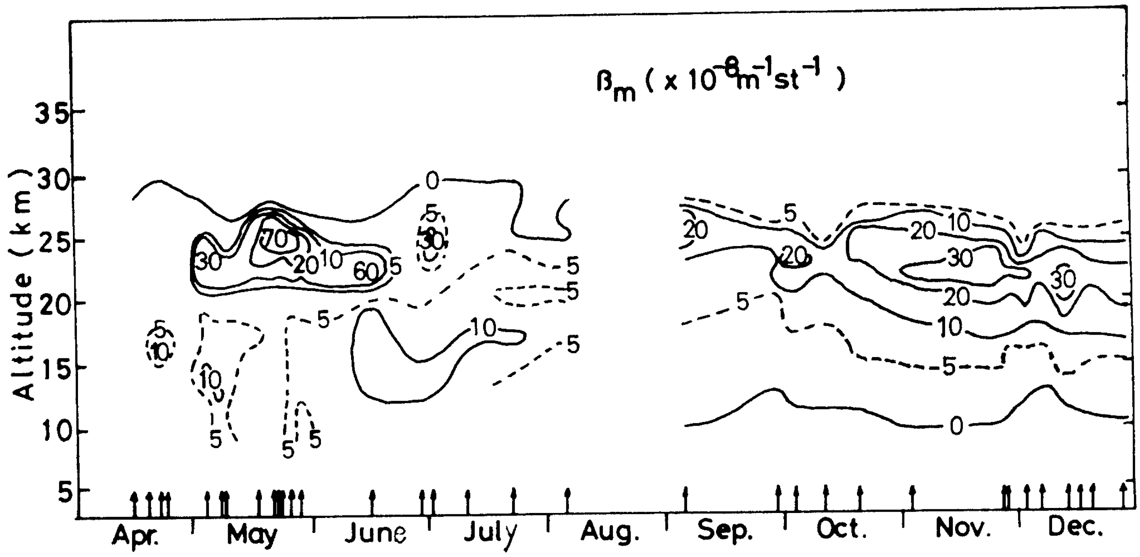

Fig. 2. The time-height cross section of particulate backscattering coefficient during the period of April to December 1982.

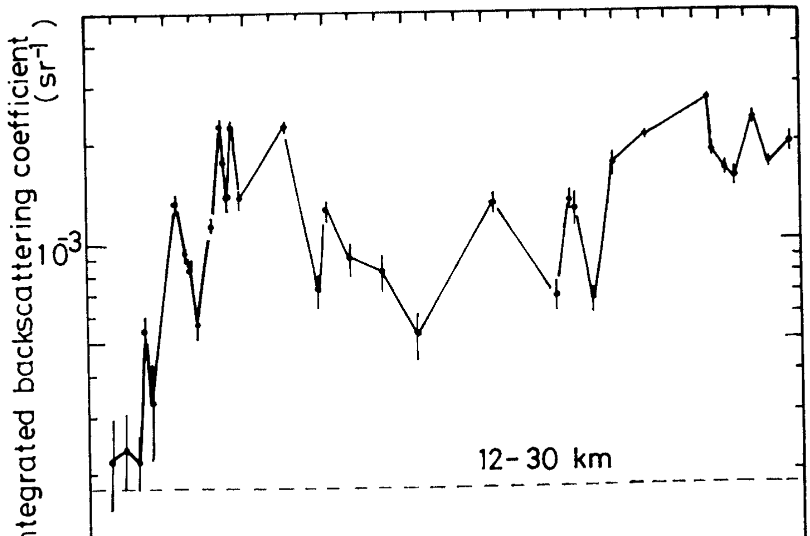

Fig. 3. The integrated particulate backscattering coefficient in the range from 12 to 30 km. The value of April 2, 1980 is shown by dashed line.

The densest aerosol layer appeared at the end of May at about 24 km with the value about $7\times10^{-7}\,m^{-1}sr^{-1}$.

In Fig. 3, the time series of the integrated backscattering coefficient from 12 to 30 km is shown. The level of the aerosol content of April 1980 is indicated by a dashed line. In May 1982 the integrated backscattering coefficient increased to as large as $3\times10^{-3}\,sr^{-1}$ and afterwards gradually decreased until August. However, after September, it increased again as large as that of May.

4. Aerosol Transport

As mentioned in Section 3, the two-layer structure of the scattering ratio profiles was observed continuously from the end of April till the beginning of June. According to the radio sonde data, wind was easterly at the altitude of the upper aerosol layer over Japan while it was westerly at the lower layer. In Fig. 4, E to W component of the wind over Shionomisaki (about 200 km west of Nagoya) is shown. The figure reveals that the stratospheric winter regime was still dominating during the

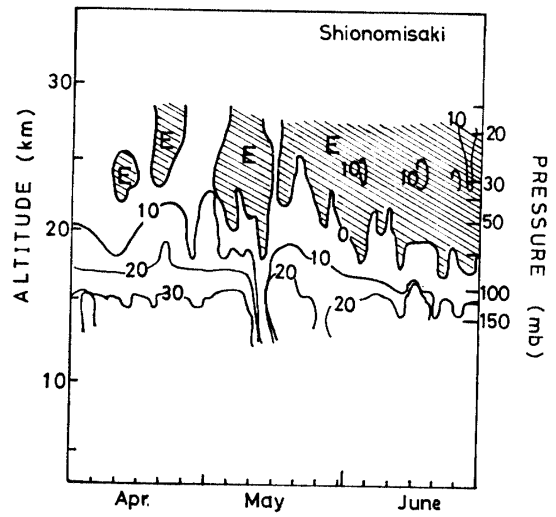

Fig. 4. Westerly-easterly component of the wind over five stations of Japan at 30 and 50 mb levels from April to June 1982.

mid-April. During the period from the end of April to the beginning of May the stratosphere was in a state of change from the winter to summer regime. From the scattering ratio profiles shown in Fig. 1, it is found that the boundary altitude of the two aerosol layers during the period from the end of April until the beginning of June was about 21 km which corresponds to the altitude where wind shifted from westerly to easterly.

In Fig. 5, the wind data at 30 and 50 mb levels in Japan and the scattering ratio at the corresponding altitude are compared. The five stations (Wakkanai, Sendai, Shionomisaki, Kagoshima and Naha) are located at nearly equal intervals of latitude and they cover all over Japan. It is found that the scattering ratio increased at both levels when wind direction changed from westerly to easterly. That is, the scattering

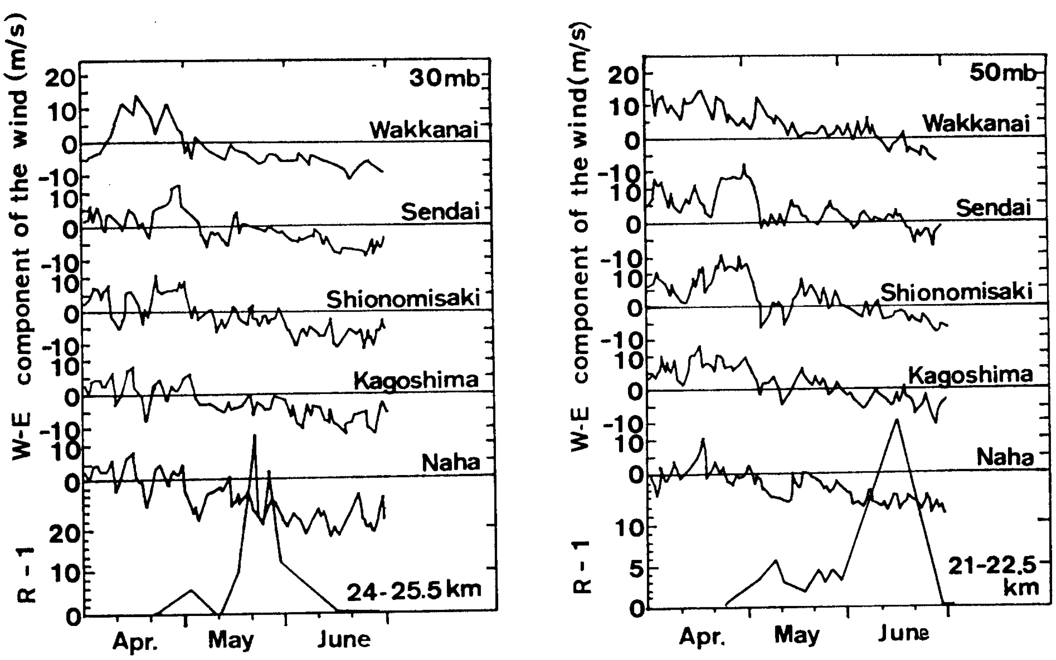

Fig. 5. Time-height cross section of westerly-easterly component of the wind over Shionomisaki from April to June 1982.

ratio increased when easterly stream line of air containing highly concentrated aerosols came over Japan. At 100 mb level (about 17 km) corresponding to the lower aerosol layer, wind was continuously westerly. Therefore, volcanic aerosols in the upper aerosol layer above 21 km would be transported by easterly wind while those in the lower layer would be transported by westerly wind and they reached almost at the same time over Japan coincidently. As a matter of fact, it is reported that the dense aerosol cloud with scattering ratio of 222 in the height range 24.5–26.2 km was detected on April 10 and 11 by the lidar in Hawaii (20°N, 155°W) (SEAN Bull., 1982).

5. The Effect of Sedimentation

In Fig. 6, the time variation of centroid height of particulate backscattering (β_m) is shown. The centroid calculations were restricted to the altitude range 12–30 km. That is,

$$h_c=\int_{z_1}^{z_2}\beta_m(z)\cdot Z dz\Big/\int_{z_1}^{z_2}\beta_m(z)dz, \tag{2}$$

where h_c presents the centroid height of backscattering coefficient, $z_1=12$ km and $z_2=30$ km. However, during the period from the end of April to mid-June when the two-layer structure was observed, the calculations were performed both for the centroid upper than about 21 km and for the centroid below 21 km separately. These two regions almost correspond to easterly wind region and westerly wind region as mentioned in Section 4.

On April 2, 1980 the height of the centroid was about 18 km. At the beginning of April 1982, however, it was as low as 13–15 km possibly because of the effect of

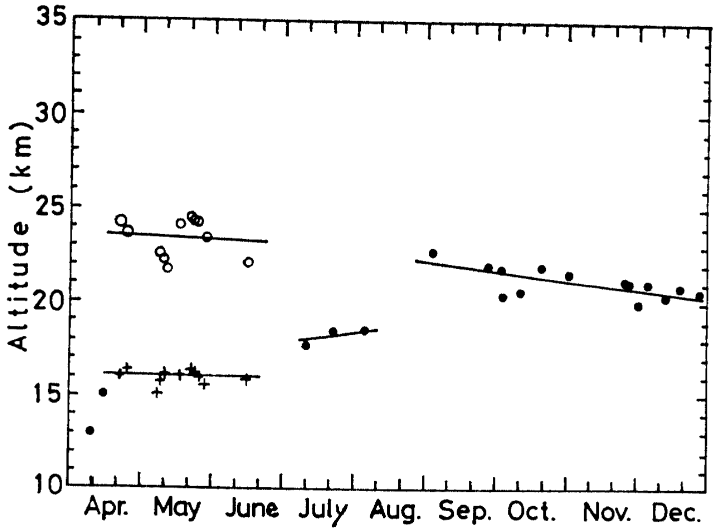

Fig. 6. *The time variation of the centroid height. The calculations were restricted to the 12–30 km range (black dots). During the period from April 22 to June 15, calculations were performed for the upper aerosol layer (above 21 km) and the lower aerosol layer (below about 21 km) separately (white dots and crosses).*

"mysterious cloud" (see Section 3). After April 19, the height of the centroid rose suddenly. In May the centroid of the upper layer almost corresponds to the main peak height, because highly concentrated aerosol existed around the main peak. In July and August, though only a few observations were performed because of cloudy weather, it is found that the centroid height was as low as 18–19 km. At the beginning of September the height of centroid rose suddenly again and afterwards it descended gradually possibly due to sedimentation. The mean descending speed of the centroid height from September to December is about 14 m/day.

In Fig. 7, the time variation of the main peak height of scattering ratio (R) and particulate backscattering coefficient (β_m) from September to December is presented. The mean descending speed of the main peak height is about 30 m/day, that is 3.5×

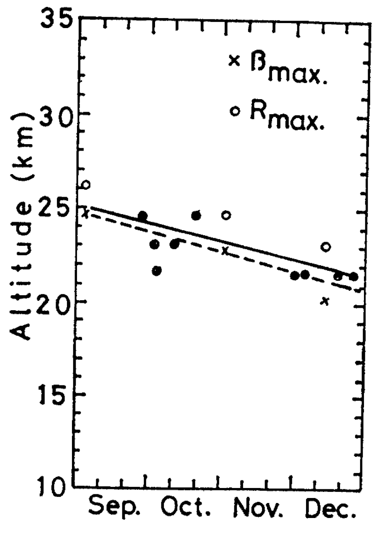

Fig. 7. *The time variation of the height of maximum scattering ratio R_{max} (solid line) and the height of maximum back-scattering coefficient β_{max} (dashed line).*

10^{-3} cm/s. From a rough estimation (KASTEN, 1968), it corresponds to the falling speed of a particle of 0.5 μm radius around 20 km. Falling speed of a particle depends on altitude, therefore the radius of a particle cannot be estimated from the falling speed of centroid. And removal of aerosols from the stratosphere through the tropopause due to sedimentation can also reduce the falling speed of centroid because these calculations are restricted to the altitude range higher than 12 km.

As shown in Fig. 3, the integrated backscattering coefficient increased after September. From Fig. 1, it is found that in mid-July and August the dense aerosol layer which was observed persistently at about 24 km until the beginning of June disappeared and it appeared again after September. This appearance of the dense aerosol cloud therefore caused the rise of the height of centroid and caused the increase of integrated backscattering coefficient in September. The low values of integrated β_m obtained from mid-July to August would be due to the inhomogeneity of horizontal distribution of aerosols. To interprete the increase of the integrated β_m from the beginning of September to the end of November, it might be necessary to take into account microphysical processes (gas to particle conversion and so on) as well as the inhomogeneity of horizontal distribution of aerosols.

6. Error Calculation

For the background aerosol content level, extinction of the lidar beam in the stratosphere does not affect the backscattering coefficient seriously. However, as for measurements reported here the effect of extinction is not negligible.

Attenuated backscattering coefficient is written as follows;

$$\beta'(z)=\beta(z)\cdot \exp\left(-2\int_{z_0}^{z}\sigma_e(z)\mathrm{d}z\right), \tag{3}$$

where

β': measured (attenuated) backscattering coefficient,

β : real backscattering coefficient,

σ_e: extinction coefficient.

Here we assume the ratio of the volume backscattering coefficient to the extinction coefficient (k) is 1/65 sr^{-1} from the estimation by PINNICK *et al.* (1980). This estimation is based on the assumption of sphericity of particles. The propriety of this assumption is not confirmed because our results of depolarization measurements indicate the existence of non-spherical particles. However, now we assume $k=1/65$ as the first approximation.

From PLATT (1979),

$$\int_{z_0}^{z_T}\beta'(z)\mathrm{d}z=\frac{k}{2}\left\{1-\exp\left[-\frac{2}{k}\int_{z_0}^{z_T}\beta(z)\mathrm{d}z\right]\right\}. \tag{4}$$

Rearrangement of eq. (4) gives

$$\frac{\gamma'}{\gamma}=-\frac{2\gamma'}{k\ln[1-2\gamma'/k]}, \tag{5}$$

where γ' and γ present integrated particulate and molecular backscattering coefficient from z_0 to z_T for attenuated and non-attenuated cases.

The results calculated from eq. (5) show that extinction is only a few percent for $\gamma'<10^{-3}\ sr^{-1}$, but it causes serious underestimation for $\gamma'>10^{-3}\ sr^{-1}$. From Fig. 2, underestimation of integrated particulate backscattering coefficient would amount to at most 30%. The altitude of normalization is mainly taken near tropopause, but for some cases it is taken above the main aerosol layer. The scattering ratios depend also on the altitude of normalization. The effect of normalization altitude on the scattering ratio at the aerosol main peak height amounts to 30% at most and the effect on integrated backscattering coefficient amounts to 20% at most. When the altitude of normalization is taken above the main aerosol layer (about 28 km), the error caused by the extinction is comparable with that caused by the lack of photon number at the normalization height.

Therefore, to discuss the change of scattering ratio profiles in detail, it is necessary to take extinction into account and also to choose the altitude of normalization carefully.

Acknowledgments

We would like to thank Prof. M. Hirono and Mr. T. Shibata of Kyushu University, Dr. K. Naito and Mr. K. Takahashi of Meteorological Research Institute, Prof. I. Hirota of Kyoto University, Mr. K. Takase of Meteorological Agency and Mr. M. Yamanaka of Nagoya University who gave us valuable and up-to-date information. We also thank Prof. T. Takeda and Prof. A. Ono of Nagoya University for helpful discussion.

References

D'Altorio, A., Visconti, G. and Fiocco, G. (1981): Lidar detection of volcanic aerosols in the atmosphere. Geophys. Res. Lett., **8**, 63–65.

Fujiwara, M., Shibata, T. and Hirono, M. (1982): Lidar observation of sudden increase of aerosols in the stratosphere caused by volcanic injections—II. Sierra Negra event. J. Atmos. Terr. Phys., **44**, 811–818.

Grams, G. and Fiocco, G. (1967): The stratospheric aerosol layer during 1964 and 1965. J. Geophys. Res., **72**, 3523–3543.

Hirono, M. and Shibata, T. (1983): Enormous increase of stratospheric aerosols over Fukuoka due to volcanic eruption of El Chichon in 1982. Geophys. Res. Lett., **10**, 152–154.

Hirono, M., Fujiwara, M., Itabe, T. and Nagasawa, C. (1977): On the long term variations of stratospheric aerosol content after the eruption of Fuego volcano observed by lider. J. Geomagn. Geoelectr., **29**, 541–556.

Hirono, M., Fujiwara, M., Shibata, T. and Kugumiya, N. (1981): Lider observations of volcanic clouds in the stratosphere over Fukuoka, caused by eruptions of Mt. St. Helens in May 1980. Geophys. Res. Lett., **8**, 1019–1022.

Hofmann, D. J. and Rosen, J. M. (1981): Stratospheric aerosol and condensation nuclei enhancements following the eruption of Alaid in April 1981. Geophys. Res. Lett., **9**, 858–859.

Iwasaka, Y. (1980): Variation of stratospheric aerosol content measured by laser radar. Part II. Long term trend of post-Fuego eruption. J. Meteorol. Soc. Jpn., **59**, 446–451.

Iwasaka, Y. and Isono, K. (1977): Lidar observation of the stratospheric aerosols at two different wavelength, 0.6943 μm and 1.06 μm. J. Atmos. Terr. Phys., **39**, 117–120.

IWASAKA, Y., MITA, A. and ISONO, K. (1976): Measurements of the stratospheric aerosols by two-color lidar. Rep. Ionos. Space Res. Jpn., **30**, 51–55.

IWASAKA, Y., HAYASHIDA, S. and ONO, A. (1982): Laser radar measurements of the stratospheric aerosol (I); January-October 1982, Nagoya, Japan. Water Research Institute. MAP Res. Note, No. 1.

IWASAKA, Y., HAYASHIDA, S. and ONO, A. (1983): Increasing backscattered light from the stratospheric aerosol layer after Mt. El Chichon—Laser radar measurement at Nagoya (35°N, 137°E). Geophys. Res. Lett., **10**, 440–442.

KASTEN, F. (1968): Falling speed of aerosol particles. J. Appl. Meteorol., **7**, 944–947.

KENT, G. S. and PHILIP, M. T. (1980): Lidar observations of dust from the Soufrière volcanic eruptions of April 1979. J. Atmos. Sci., **37**, 1358–1362.

MCCORMICK, M. P., SWISSLER, T. J., CHU, W. P. and FULLER, W. H., Jr. (1978): Post-volcanic stratospheric aerosol decay as measured by lidar. J. Atmos. Sci., **35**, 1296–1303.

PINNICK, R. G., JENNING, S. G. and CHYLEK, P. (1980): Relationship between extinction, absorption, backscattering, and mass content of sulfuric acid aerosols. J. Geophys. Res., **85**, 4059–4066.

PLATT, C. M. R. (1979): Remote sounding of high clouds; I. Calculation of visible and infrared optical properties from lidar and radiometer measurements. J. Appl. Meteorol., **18**, 1130–1143.

POLLACK, J. B., TOON, O. B., SAGAN, C., SUMMERS, A., BALDWIN, B. and VAN CAMP, W. (1976): Volcanic explosions and climate change; A theoretical assessment. J. Geophys. Res., **81**, 1071–1083.

REITER, R., JÄGER, H., CARNUTH, W. and FUNK, W. (1980): Lidar observations of the mount St. Helens eruption clouds over mid-Europe, May to July 1980. Geophys. Res. Lett., **7**, 1099–1101.

REITER, R., JÄGER, H., CARNUTH, W. and FUNK, W. (1982a): A stratospheric aerosols increase during 1981, observed by lidar over mid-Europe. Geophys. Res. Lett., **9**, 469–472.

REITER, R., JÄGER, H., CARNUTH, W. and FUNK, W. (1982b): The stratospheric aerosol increase in February and March 1982. Geophys. Res. Lett., **9**, 858–859.

RUSSELL, P. B. and HAKE, R. D., Jr. (1977): The post-Fuego stratospheric aerosol; Lidar measurements with radiative and thermal implications. J. Atmos. Sci., **34**, 163–177.

RUSSELL, P. B., VIEZEE, W., HAKE, R. D., Jr. and COLLIS, T. H. (1976): Lidar observations of the stratospheric aerosol; California, October 1972 to March 1974. Q. J. R. Meteorol. Soc., **102**, 675–693.

SCIENTIFIC EVENT ALERT NETWORK BULLETIN (1982): **8**(1–12).

(Received April 7, 1983; Revised manuscript received June 24, 1983)