

# LASER RADAR MONITORING OF THE POLAR MIDDLE ATMOSPHERE

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**Abstract:** The laser radar system used for the monitoring of the polar middle atmosphere (10 km–120 km) is discussed. This system can emit laser pulses at three different wavelengths: 694 nm, 589 nm and 347 nm. The laser radar measurements make an important contribution to clarify the behavior of stratospheric aerosols, the formation of stratospheric “Cist”, and the interaction between noctilucent cloud particles and charged particles in the lower ionosphere of the polar region.

## 1. Introduction

Laser radar is one of the most powerful instruments used to monitor remotely the upper atmosphere. Nowadays, about twenty laser radar systems are located world-wide for monitoring the stratospheric aerosol density and so on. Unfortunately, most of these laser radars are distributed in the middle latitudinal region of the northern hemisphere, and it has been a serious problem to study the behavior of stratospheric aerosols on a global scale. The polar stratosphere has been speculated to be an active sink region of stratospheric aerosols, but this processes is not theoretically or observationally clear. In addition to this, the polar middle atmosphere (10 km–120 km) is an interesting and attractive region from the viewpoint of aerosol science. There are many interesting phenomena which cannot be found in the middle or low latitudinal upper atmosphere; for example, the effect of precipitation of high energy charged molecules or atoms on the nucleation processes of stratospheric particles, estimation of the contribution of polar stratospheric cloud, so-called “Stratospheric Cist” the water vapor budget or large scale circulation in the stratosphere, the formation of noctilucent clouds and so on. Laser radar measurements will certainly contribute to research on these problems.

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In this paper, we should like to describe the laser radar system which is being prepared for use during the "Middle Atmosphere Program (1982–1985)" at Syowa Station (Antarctica).

## 2. Laser Rader System for Monitoring of the Polar Middle Atmosphere

The main characteristics of the laser radar system which is planned to be used at Syowa Station, Antarctica to monitor the polar middle atmosphere during the "Middle Atmosphere Program" are listed in Table 1. This system can emit laser pulses at three different wavelengths: 694 nm, 347 nm and 586 nm.

Table 1. Characteristics of laser radar.

Laser	Ruby (694.3 nm, 347.2 nm) Dye
Power of laser pulse	1 J/pulse (Max.)
Pulse repetition rate	1 Hz (Ruby) 0.25 Hz (Dye)
Receiving telescope	50 cm $\phi$
Data display	A-scope, and photoncounter (100 channels and 2 series)

The laser radar equation is given in following form,

$$P_{r,i}(Z) = P_{e,i} [K_i T_i^2(Z) B_i(Z)] / Z^2, \quad (1)$$

where

- $Z$ : altitude,
- $i$ : suffix showing  $i$ -th wavelength,
  - $i=1$ : wavelength=694 nm
  - $i=2$ : wavelength=347 nm
  - $i=3$ : wavelength=589 nm

- $P_{r,i}(Z)$ : receiver power,
- $P_{e,i}$ : output power of laser pulse,
- $K_i$ : system constant,
- $B_i$ : backscattering coefficient,
- $T_i$ : atmospheric transmittance.

The backscattering coefficient is easily estimated using the assumption that the atmospheric transmittance is constant above the tropospheric region. The backscattering coefficient  $B_i(Z)$  contains both the Rayleigh scattering component due to molecular scattering and the Mie scattering component due to aerosols (or other particulate matter). Then, total backscattering coefficient  $B_i(Z)$  is given in the following form:

$$B_i(Z) = B_i(\text{Aerosol}; Z) + B_i(\text{Molecule}; Z), \quad (2)$$

where  $B_i(\text{Aerosol}; Z)$  and  $B_i(\text{Molecule}; Z)$  are the coefficient of Mie backscattering by particulate matter and the coefficient of Rayleigh backscattering by air molecules respectively.

The ratio of the signal determined from eq. (1) to the return signal estimated on the basis of a dust free atmosphere (for example, U.S. Standard Atmosphere, 1976) is given by,

$$\begin{aligned} R_s(Z) &= B_i(Z)/B_i(\text{Molecule}; Z) \\ &= 1 + B_i(\text{Aerosol}; Z)/B_i(\text{Molecule}; Z). \end{aligned} \quad (3)$$

Equation (3) can be rearranged and used to determine the aerosol backscattering coefficient:

$$\begin{aligned} B_i(\text{Aerosol}; Z) &= (R_s(Z) - 1)B_i(\text{Molecule}; Z) \\ &= (R_s(Z) - 1)\sigma_{m,i}N_m(Z) \end{aligned} \quad (4)$$

where

- $\sigma_{m,i}$ : Rayleigh scattering cross section at  $i$ -th wavelength,
- $N_m(Z)$ : density of air molecules at altitude  $Z$ .

Obviously it is necessary to know the profile of  $N_m(Z)$  or  $B_i(\text{Molecule}; Z)$  in order to distinguish the backscattering coefficient due to particulate matter from total backscattering coefficient  $B_i(Z)$  measured by laser radar. Usually it is assumed that the clean level where aerosol density is so low that the backscattering coefficient of aerosol particles does not contribute to the total backscattering coefficient is near 10 km or above 30 km, and that the profile of  $B_i(Z)$  measured by laser radar can be fitted to the calculated  $B_i(\text{Molecule}; Z)$  at this level. After that, the residue between  $B_i(Z)$  and  $B_i(\text{Molecule}; Z)$  found in the stratosphere is recognized to be due to the scattering component by aerosols or other particulate matter (the so-called "Matching Method"). Many investigators have pointed out that serious error in laser radar measurements can be caused by this method of analysis (*e.g.*, RUSSELL *et al.*, 1976).

Laser pulses of wavelength = 347.2 nm used in this system are very useful in solving the above-mentioned problem. As is described in the section "Molecular scattering and Mie scattering", the measurement of  $B_i(Z)$  is very sensitive to the molecular scattering, since the Rayleigh scattering cross-section is in inverse proportion to the fourth power of the wavelength,  $\lambda^4$ , and then the Rayleigh cross-section at  $\lambda = 347.2$  nm is 16 times the cross-section at  $\lambda = 694.3$  nm. On the other hand, the difference between the Mie scattering cross-sections at these wavelengths is not so serious, and the major contribution to  $B_2(Z)$  is the Rayleigh scattering component. In other words,  $B_2(Z)$  can be approximated by  $B_2(\text{Molecule}; Z)$ , and we can easily estimate, approximately, the profile of air molecule density from the measurement of  $B_2(Z)$ .

The laser pulse of 694.3 nm is used for the detection of particulate matter, and

589.0 nm laser (Dye laser) for monitoring the sodium vapor layer, which is usually found in the region 87 km–96 km.

Additionally, this system contains a polarizer and monochrometer. The polarization properties of scattering material are very useful in determining whether the particulate matter is crystalized or not. In Fig. 1, typical results measured at Nagoya (34°N, 137°E) are shown. The sudden increase of the backscattering coefficient,  $B_1$  (Aerosol; Z) was found in the stratosphere at the end of May 1980. After that, very intense backscattered light from the stratospheric aerosol layer was measured on June 3, 1980. This sudden increase of backscattered light from the stratospheric aerosol layer must be due to the volcanic eruption of Mt. St. Helens (the first erup-

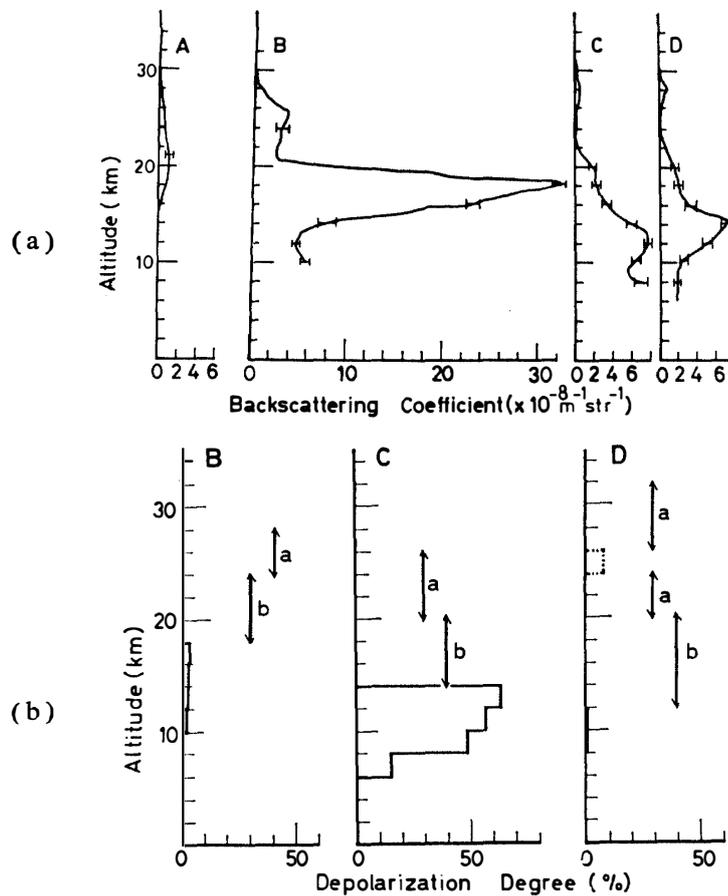


Fig. 1. Laser radar measurements of the stratospheric aerosol layer enhanced by the volcanic eruption of Mt. St. Helens (May 1980). The curves in Fig. 1a show the backscattering coefficient of aerosol particles, and curves in Fig. 1b show the degree of depolarization of stratospheric particles.

A: May 10, 1980, B: June 3, 1980, C: August 24, 1980, D: September 1, 1980, a: degree of depolarization  $< 0.2(\pm 0.2)\%$ , b:  $0.2(+0.2)\% \leq$  degree of depolarization  $< 1.00(+0.2)\%$ .

tion was on May 18, 1980, and eruptions continued after that date), considering the scale of the Mt. St. Helens eruption, and other investigators' preliminary reports (laser radar measurements, direct sampling data and so on). Many investigators suggested that stratospheric aerosols are possibly composed of sulfuric acid (75%  $\text{H}_2\text{SO}_4$  and 25%  $\text{H}_2\text{O}$  in weight) droplets and not crystallized particles under normal stratospheric conditions (*e.g.*, TOON and POLACK, 1976). But, of course, aerosol chemical composition can be transformed through the various interactions between aerosols and the many gases surrounding the aerosols. Particularly, the transformation of stratospheric aerosols from sulfuric acid droplets to ammonium sulfate particles, which proceeds gradually under quiet stratospheric conditions, is one of the most important problems in discussing aerosol chemistry or assessing the climatic effect of the stratospheric aerosol layer (*e.g.*, FARLOW *et al.*, 1979). Before the eruption of Mt. St. Helens, some investigators pointed out the transformation of aerosols from  $\text{H}_2\text{SO}_4$  droplets to  $(\text{NH}_4)_2\text{SO}_4$  crystal particles possibly occurred in the stratospheric aerosol layer (FARLOW *et al.*, 1979; REITER *et al.*, 1979). The very low degree of depolarization of the stratospheric aerosol layer found after the eruption of Mt. St. Helens possibly suggests that most of the particles produced through the injection of a large amount of volcanic matter by the eruption of Mt. St. Helens are amorphous and are composed of a sulfuric acid solution. Then, we can say that the re-transformation from  $(\text{NH}_4)_2\text{SO}_4$  particles to  $\text{H}_2\text{SO}_4$  particles occurred due to the new particle production through the volcanic matter injection due to the eruption of Mt. St. Helens.

The spectrometer in this system can be used to detect some important gases,  $\text{H}_2\text{O}$ ,  $\text{O}_2$ ,  $\text{N}_2$  using the Raman spectrometric technique. Simultaneous measurements of particulate matter and gas components by laser radar must be very useful in dis-

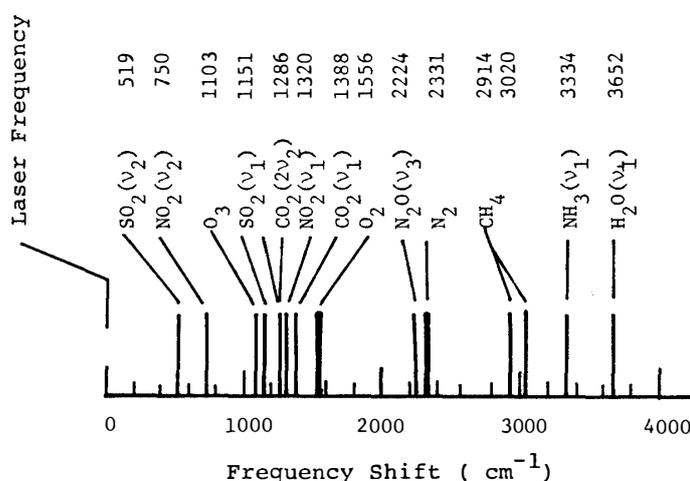


Fig. 2. Frequency shifts of Q-branch of vibrational-rotational Raman spectra of typical molecules presented in standard atmosphere relative to the exciting laser frequency.

cussing the detailed physical process of transformation of particles, and other interaction processes between aerosols and gases surrounding particles. In Fig. 2, frequent shifts of Raman spectra of interesting gases are shown.

### 3. Molecular Scattering and Mie Scattering

The molecular scattering (Rayleigh scattering) coefficient is written as;

$$B_i(\text{Molecule}; Z) = k_i^4 A^2 N_m(Z) p, \quad (5)$$

where

$k_i$ : the wavenumber of incident radiation with wavelength of  $\lambda_i$ ,  $(2\pi/\lambda_i)$ ,

$A$ : the polarizability

and  $p$  is the depolarization factor defined by the following formula,

$$p = 3(2 + \Delta)/(6 - 7\Delta). \quad (6)$$

Usually the depolarization factor is on the order of  $10^{-2}$ .

The treatment of aerosol scattering given here is similar to the general treatment given by BULLRICH (1964). Assuming that aerosols in the atmosphere may be considered as a polydispersed collection of homogeneous spheres of average refractive index  $\eta$ , the aerosol backscattering coefficient is given by:

$$B_i(\text{Aerosol}; Z) = \int_{R_2}^{R_1} [I_1(A, \eta, \theta) + I_2(A, \eta, \theta)] n(R) / (2k_i^2) dR, \quad (7)$$

where

$I_1(A, \eta, \theta)$ ,  $I_2(A, \eta, \theta)$ : the Mie intensity function for light with electric vector perpendicular and parallel, respectively, to the plane through the direction of propagation of the incident and scattered radiation,

$R_1$ ,  $R_2$ : the radius of the aerosol particles at the lower and upper limits of the size distribution,

$n(R)$ : the derivative of size distribution function of aerosols and density of aerosols with radius between  $R_1$  and  $R_2$ ; the distribution function is given by

$$\int_{R_2}^{R_1} n(R) dR,$$

$\theta$ : the scattering angle measured between the directions of incident and scattered radiation.

For the backscattering radiation,  $\theta = 180^\circ$ , and  $I_1 = I_2$ .

The values of  $B_i(\text{Aerosol}; Z)$  and  $B_i(\text{Molecule}; Z)$  listed in Table 2 are calculations for the condition that the molecular number densities are given by the U.S. Standard Model Atmosphere (1962), the aerosol size distribution function is given by the Junge type ( $n(R) \propto R^{-4}$ ,  $R_1 = 0.124 \mu\text{m}$  and  $R_2 = 3 \mu\text{m}$ ), the number density profile of aerosol particles is constructed using ELTERMAN's data (1964), and the

Table 2. Backscattering coefficient;  $B_1$  (Aerosol; Z),  $B_1$  (Molecule; Z),  $B_1$  (Z),  $B_2$  (Aerosol; Z),  $B_2$  (Molecule; Z),  $B_2$  (Z).

Height (km)	Aerosol density ( $\text{cm}^{-3}$ )	Molecular density ( $\text{cm}^{-3}$ )	$B_1$ (Aerosol; Z) ( $\text{m}^{-1}\text{str}^{-1}$ )	$B_1$ (Molecule; Z) ( $\text{m}^{-1}\text{str}^{-1}$ )	$B_1$ (Z) ( $\text{m}^{-1}\text{str}^{-1}$ )	$B_2$ (Aerosol; Z) ( $\text{m}^{-1}\text{str}^{-1}$ )	$B_2$ (Molecule; Z) ( $\text{m}^{-1}\text{str}^{-1}$ )	$B_2$ (Z) ( $\text{m}^{-1}\text{str}^{-1}$ )
0	4.50(2)	2.54(19)	5.35(-6)	5.36(-7)	5.89(-6)	1.04(-5)	8.58(-6)	1.90(-5)
2	2.95(1)	2.09(19)	3.51(-7)	4.41(-7)	7.92(-7)	6.84(-7)	7.06(-6)	7.74(-6)
4	2.15	1.70(19)	2.56(-8)	3.59(-7)	3.85(-7)	4.99(-8)	5.74(-6)	5.79(-6)
6	6.50(-1)	1.37(19)	7.73(-9)	2.89(-7)	2.97(-7)	1.51(-8)	4.63(-6)	4.65(-6)
8	3.50(-1)	1.09(19)	4.16(-9)	2.30(-7)	2.34(-7)	8.11(-9)	3.68(-6)	3.69(-6)
10	3.50(-1)	8.60(18)	4.16(-9)	1.81(-7)	1.85(-7)	8.11(-9)	2.90(-6)	2.91(-6)
12	7.50(-1)	6.49(18)	8.92(-9)	1.37(-7)	1.46(-7)	1.74(-8)	2.19(-6)	2.21(-6)
14	1.25	4.74(18)	1.49(-8)	1.00(-7)	1.15(-7)	2.91(-8)	1.60(-6)	1.63(-6)
16	1.70	3.46(18)	2.02(-8)	7.30(-8)	9.32(-8)	3.94(-8)	1.17(-6)	1.21(-6)
18	2.50	2.53(18)	2.97(-8)	5.34(-8)	8.31(-8)	5.79(-8)	8.54(-7)	9.12(-7)
20	1.25	1.85(18)	1.49(-8)	3.90(-8)	5.39(-8)	2.91(-8)	6.24(-7)	6.53(-7)
22	8.00(-1)	1.34(18)	9.51(-9)	2.83(-8)	3.78(-8)	1.85(-8)	4.53(-7)	4.72(-7)
24	6.50(-1)	9.76(17)	7.73(-9)	2.06(-8)	2.83(-8)	1.51(-8)	3.30(-7)	3.42(-7)
26	5.0(-1)	7.12(17)	5.95(-9)	1.50(-8)	2.10(-8)	1.16(-8)	2.40(-7)	2.52(-7)

Note: The numbers in parentheses are exponents; for example,  $1.00 \times 10^3$  is written 1.00(s).

refractive index  $\eta=1.5$ . Of course, the choice of parameters used here by no means unique, but it is felt that they are representative of the clear atmosphere. It can be easily found in Table 2 that the contribution of  $B_2(\text{Aerosol}; Z)$  to the total back-scattering function  $B_2(Z)$  is not so effective that we can estimate, even if it is limited, the temporary profile of air molecule density. This must be an important advantage of the present system in comparison with previous laser radars.

#### 4. Particulate Matter in the Polar Middle Atmosphere

In this section, we should like to focus on a discussion about some interesting phenomena which are related to the behavior of particulate matter of the polar middle atmosphere, and show what an important contribution the laser radar measurements can make in research on these phenomena.

##### 4.1. Stratospheric "Cist"

The existence of stratospheric Cist was first emphasized by STANFORD (1973) on the basis of the cloud observation report at Amundsen-Scott Base (located at the South Pole) during the observational periods, 1957, 1960 and 1962–70. There are so many unknown factors (vertical profile of Cist particle density, and its variations and so on) that we cannot definitely say anything, but STANFORD suggested, even if it was speculative to some degree, that this cloud had serious effects on the water vapor budget or circulation in the stratosphere (1973, 1977). The vertical profile of stratospheric Cist particles and of its variation are easily monitored by laser radar, and these measurements are very useful in clarifying not only the formation mechanism of stratospheric Cist but also the stratospheric water vapor budget or circulation on a global scale.

##### 4.2. Stratospheric aerosol transformation

From the viewpoint of stratospheric aerosol transportation on a global scale, measurements of the vertical profile of aerosol density and of its time variation in high latitudes are very important, since the high latitude stratosphere has been considered to be an active sink region. The density maps of the  $\text{SO}_4^{--}$  ion which is contained in stratospheric aerosol were presented by LAZRUS *et al.* (1979), and show very suggestive patterns. The highly concentrated region due to the strong volcanic eruption of Mt. Fuego (October 1974) is found in the measurements in the spring of 1975. After one or two years, the highest concentration region of  $\text{SO}_4^{--}$  gradually shifted to the polar stratosphere, which suggests to us the existence of a transportation mechanism from the equatorial to the high latitude region in the stratosphere. The laser radar measurement of stratospheric aerosol at Syowa Station will contribute to research on the transportation mechanism of stratospheric aerosols.

## 5. Discussion and Summary

Various interesting facts on the aeronomic processes in the polar middle atmosphere can potentially be discovered by laser radar measurements at Syowa Station, Antarctica. The laser radar system described here has some advantages in comparison with previous laser radar system; measurements using this system should make an important contribution to obtain comprehensive and quantitative understanding of the dynamics, and the distribution or transportation of minor constituents (aerosols and gases) of the polar middle atmosphere.

Cooperative experiments using both laser radar and other kinds of measurements, such as following should be useful:

- 1) balloon experiments for the study of the electron and ion density, aerosols, electric field, minor gases, and auroral X-rays,
- 2) aircraft measurements of minor species (gases and aerosols), and solar and atmospheric radiation, and
- 3) ground-based spectroscopic observations of the densities of CH<sub>4</sub>, N<sub>2</sub>O, CO, O<sub>3</sub> and NO<sub>x</sub>.

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