SPECTRAL EXTINCTION MEASUREMENT BY SUNPHOTOMETER AT SYOWA STATION, ANTARCTICA

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Abstract: Spectral measurement of direct solar radiation was made by sunphotometer (λ=368, 500, 675, 778, 862 nm) at Syowa Station (69°00'S, 39°35'E, 21 m a.s.l.), Antarctica in the period January 1980 through January 1981. Average aerosol optical thickness (λ=500 nm) was 0.023 ± 0.010 and its maximum appeared in October, minimum in November. Monthly mean Ångström turbidity coefficient β lies in the range from 0.0045 to 0.0207 and α from 1.27 to 2.15.

The present result was in good accordance with the recent measurement of optical thickness in Antarctica made by other reseachers.

The most important finding is that turbidity at Syowa Station was fairly constant throughout a year, spectral extinction exhibit marked seasonal variation. It can be also found that the seasonal variation of spectral extinction is caused by the seasonal change of aerosol which has resulted from the seasonal difference of the origin of air mass.

1. Introduction

Atmospheric turbidity, or aerosol content can be determined through the measurement of intensity of direct solar beam. Recent developments in man's activities have caused considerable changes to the atmospheric environment. An increase of particles in the atmosphere is one of the noticeable changes of atmospheric environment. Increase in atmospheric particles causes a change of radiation field in the atmosphere, which necessarily leads to a change of the heat budget of the earth-atmosphere system (YAMAMOTO and TANAKA, 1972).

Since Antarctica is remote from man's activity, the antarctic atmosphere has been considered to be a reference for the state of world-wide air pollution. Several studies have been carried out concerning the antarctic aerosols (SHAW, 1978a; PETERSON and SZWARC, 1977; ONO *et al.*, 1981; IWAI *et al.*, 1981; ITO, 1982; ITO *et al.*, 1982).

The observation of the atmospheric turbidity in Antarctica was started from the pioneering work by LILJEQUIST (1956) and compiled by KUHN (1972) and SHAW (1978a). Preliminary measurements with a pyreheliometer and cutoff filters have been made by Japanese Antarctic Research Expedition (JARE) team since 1973 (KAWAGUCHI, 1974; KAWAGUCHI and HAYASHI, 1975; SUZUKI *et al.*, 1977; YAMANOUCHI, 1982).

JARE-21 team successfuly obtained 317 data of the aerosol optical thickness (λ=368, 500, 675, 778, 862 nm) by use of a sunphotometer at Syowa Station (69°00′S, 39°35′E, 21 m a.s.l.) for the sunlit period from January 1980 to January 1981.

From these data, characteristics of spectral extinction of the atmosphere over Antarctica were examined. In this paper, the details of the examination are given.

2. Measurements

Direct solar radiation was measured by an Eko MS-110 sunphotometer. The sunphotometer had an aperture angle of 2.5 degrees. A cover glass of fused quartz was on the aperture. The detector was photo-diode. Spectral measurements were made with interference filters set on the rotating disk. Their nominal pass wavelengths were 368, 500, 675, 778 and 862 nm.

3. Theory of Measurement

The transmittance of solar radiation through the atmosphere is mainly dependent on three factors; the scattering by the molecules (Rayleigh scattering), the selective absorption by gaseous constituents (in the present concern, absorption by ozone) and the scattering (and absorption) by aerosols. If we denote the optical thickness for molecules, ozone and aerosols with $\tau_R(\lambda)$, $\tau_0(\lambda)$ and $\tau_A(\lambda)$ respectively, we may express the solar radiation $I(\lambda)$ at a given wavelength passing through the air mass M from Bouguer-Lambert law as

$$I(\lambda)=\frac{I_0(\lambda)}{S}\exp\{-(\tau_R(\lambda)+\tau_0(\lambda)+\tau_A(\lambda))M\}\ (M\leq 6)\,, \tag{1}$$

where $I_0(\lambda)$ is extraterrestrial solar radiation, and S is the correction factor for mean sun–earth distance.

The extraterrestrial irradiance ($I_0(\lambda)$) is determined by Langley method. The output voltages ($J(\lambda)$) of sunphotometer are read for the various values of air mass (M) during periods of stable turbidity. These readings are plotted against air mass (M) in order to obtain $J_0(\lambda)$ corresponding to $I_0(\lambda)$.

Starting with eq. (1), knowing $J_0(\lambda)$ for the instrument, and using the observed $J(\lambda)$, the aerosol optical thickness can be calculated by:

$$\tau_A(\lambda)=\frac{\ln(J_0(\lambda)/(J(\lambda)\cdot S))}{M}-(\tau_R(\lambda)+\tau_0(\lambda))\,. \tag{2}$$

$\tau_R(\lambda)$ obtained by ELTERMAN (1964) was thought to give overestimation of the optical thickness (FLÖHLICH and SHAW, 1980) and rather small values of $\tau_R(\lambda)$ were given by FRÖHLICH and SHAW:

$$\tau_R(\lambda)=\frac{P}{1013.26}\cdot 0.00838\cdot\lambda^{-(3.916+0.074\lambda+0.050/\lambda)}\,, \tag{3}$$

where P is the station pressure. $\tau_0(\lambda)$ was determined by use of absorption coefficients given by Vigroux and daily ozone amount observed by Dobson ozone spectrophotom-

eter. Air mass M was calculated by means of the empirical formula derived by KASTEN (1966):

$$M=\frac{1}{\sin h+0.15(h+3.885)^{-1.253}}, \tag{4}$$

with h as solar altitude in degrees.

The aerosol optical thickness was approximately expressed by a power law function of the wavelength λ as

$$\tau_A(\lambda)=\beta\lambda^{-\alpha}, \tag{5}$$

where β and α are the Ångström turbidity coefficient at $\lambda=1000$ nm and the wavelength exponent, respectively (ÅNGSTRÖM, 1961, 1964).

The aerosol optical thickness is related to the size distribution of aerosol particles as follows:

$$\tau_A(\lambda)=\int_r \pi r^2 Q\left[\frac{2\pi r}{\lambda}, m^*(\lambda)\right]\cdot n(r)\cdot dr, \tag{6}$$

where r is particle radius, $m^*(\lambda)$ refractive index of particle, $n(r)$ aerosol number concentration in vertical column per unit radius interval at radius r, and Q the ratio of the scattering cross section to the geometrical cross section of a particle. From the measured values of $\tau_A(\lambda)$, we can infer the size distribution of the particles by means of the inversion method. In practice, we used the technique developed by YAMAMOTO and TANAKA (1969), which is based on the method of PHILLIPS (1962). In this paper, $m^*(\lambda)$ is assumed to be 1.49 and the size range of radius r is assumed to be 0.03 μm to 7 μm. Using all of size distribution obtained by inversion, spectral extinctions were calculated in order to check the performance of inversion procedure. According to the comparison of the reconstructed spectral extinction with the original spectral extinction, a fairly good agreement among them was confirmed.

4. Results and Discussion

4.1. *Errors in determination of* $\tau_A(\lambda)$

The determination of $J_0(\lambda)$ is essentially important for the evaluation of aerosol optical thickness. Eighteen determinations of $J_0(\lambda)$ value by Langley method were made thirteen times at Syowa Station and five aboard. There were no remarkable changes in $J_0(\lambda)$ during the observation period. Table 1 shows the errors expected in the determination of $\tau_A(\lambda)$ in our data. According to Table 1, these errors of $J_0(\lambda)$, however, do not cause serious deformation of determined spectral extinction.

Table 1. The accuracies of the extrapolation process at each wavelength.

	Wavelength				
	862 nm	778 nm	675 nm	500 nm	368 nm
$\Delta J_0/J_0$	0.005	0.005	0.006	0.007	0.014
$\Delta\tau_A$	0.003	0.003	0.003	0.004	0.007

$\Delta J_0/J_0$ means uncertainty (r.m.s) in the case of extrapolation by Langley method. $\Delta\tau_A$ means the error (r.m.s) of the optical thickness at $M=2$.

4.2. *Mean aerosol optical thickness during the observation period*

Table 2 shows the mean τ_A as 5 wavelengths during the observation period. It can be seen that τ_A ($\lambda=500$ nm) at Syowa Station were quite comparable to those obtained at McMurdo Station (77.85°S, 166.67°E 0 m a.s.l.), Antarctica (SHAW, 1982). These values seem to be representative in the coastal region of Antarctica.

Table 2. Mean aerosol optical thickness during the observation period.

	Wavelength				
	862 nm	778 nm	675 nm	500 nm	368 nm
τ_A	0.012	0.011	0.012	0.023	0.047
$\Delta\tau_A$	0.008	0.008	0.010	0.010	0.017

$\Delta\tau_A$: standard deviation.

4.3. *Seasonal variation of* $\tau_A(\lambda)$

Figure 1 gives the time series of $\tau_A(\lambda)$ at 5 wavelengths. In the figure, it can be seen that $\tau_A(\lambda)$ at Syowa Station is rather constant for every wavelength.

Figure 2 shows a comparison of seasonal variation of τ_A ($\lambda=500$ nm) at Syowa Station with that at Barrow (71.25°N, 156.67°W 0 m a.s.l.), Arctic (SHAW, 1982). $\tau_A(\lambda=500$ nm) at Barrow shows a rather high value and remarkable seasonal variation, whereas τ_A ($\lambda=500$ nm) at Syowa Station is characterized by extremely low and relatively uniform values throughout the year.

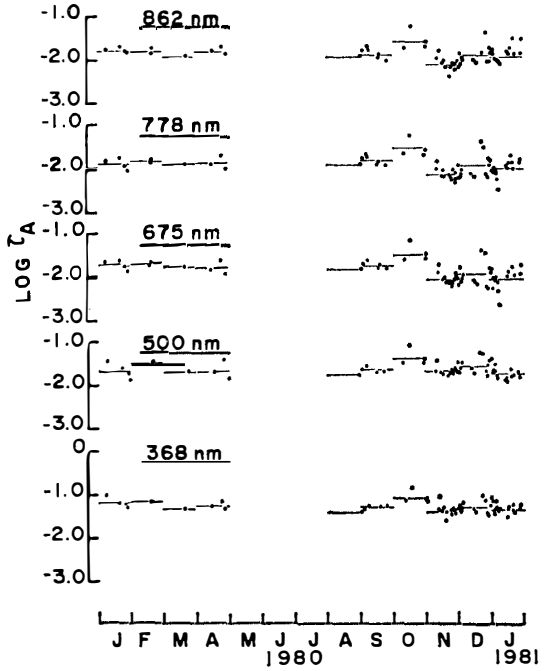

Fig. 1. Time series of $\tau_A(\lambda)$ *at 5 wavelengths.*

4.4. *Ångström wavelength exponent and turbidity coefficient*

Table 3 shows Ångström wavelength exponent α and turbidity coefficient β evaluated from eq. (5) by use of the least square fit method. During the observation

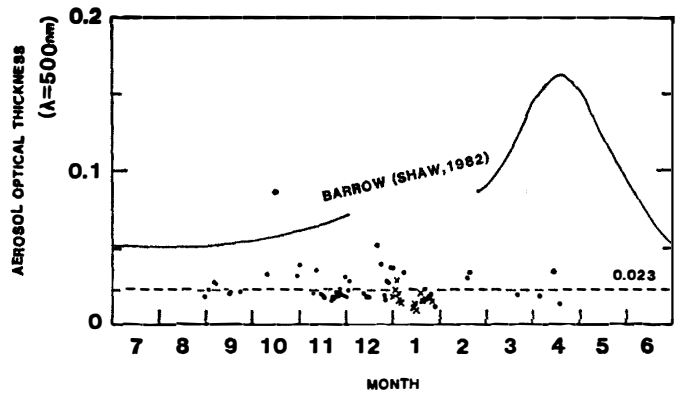

Fig. 2. A comparison of seasonal variation of aerosol optical thickness between Syowa Station and Barrow, Arctic (SHAW, 1982). Solid circles and crosses denote the data in the period from January to December 1980 and the data in January 1981 observed at Syowa Station, respectively.

Table 3. Ångström wavelength exponent α and turbidity coefficient β.

	1980									1981	
	Jan.	Feb.	Mar.	Apr.	Aug.	Sep.	Oct.	Nov.	Dec.	Jan.	Mean
α	1.61	1.41	1.48	1.49	1.30	1.46	1.27	2.15	1.79	1.67	1.56
β	0.0102	0.0119	0.0092	0.0099	0.0090	0.0103	0.0207	0.0045	0.0075	0.0064	0.0100

period, monthly mean α lies in the range from 1.27 to 2.15 and β from 0.0045 to 0.0207. It can be seen that the maximum value of α appeared in November and the minimum in October, and also the maximum value of β in October and the minimum in November. Both α and β are approximately constant from February to August. This result shows a similar tendency to the result obtained through the broad band measurement which was carried out in the period from October to December in 1979 at Mizuho Station, Antarctica, using cutoff filters, by YAMANOUCHI (1982).

4.5. The seasonal variation of spectral extinction of aerosol optical thickness

Monthly variation of wavelength exponent α suggests the seasonal variation in the size distribution of aerosols. In Fig. 3, three typical examples of spectral extinction of aerosol optical thickness obtained in the present observation are shown. Type 1 (white circles) is close to the power law distribution, type 2 (solid circles) is smaller than the power law distribution in larger wavelength and type 3 (white triangles) is larger than the power law distribution in the larger wavelength. All of the spectral extinction obtained in the observation period were roughly classified into these 3 types.

Figure 4 represents the seasonal distribution of each type mentioned above, deduced from the present results. It can be seen that each type tends to appear in the definite period. In the period from January to April, both type 1 and type 3 are present, in the period from late August to October, only type 1, and in the period from November to December, only type 2 is present. Since the spectral extinction of aerosol optical thickness is largely controlled by the size distribution of aerosols in

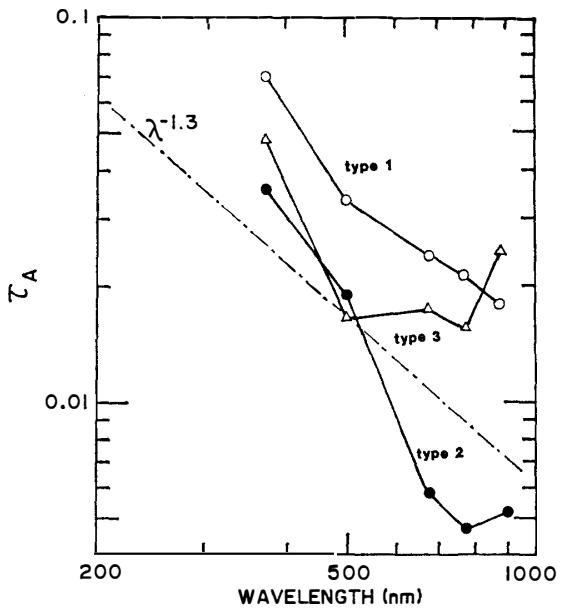

Fig. 3. Examples of aerosol extinction spectra in three types obtained at Syowa Station. A dot-dash-line is evaluated by the power law spectrum.

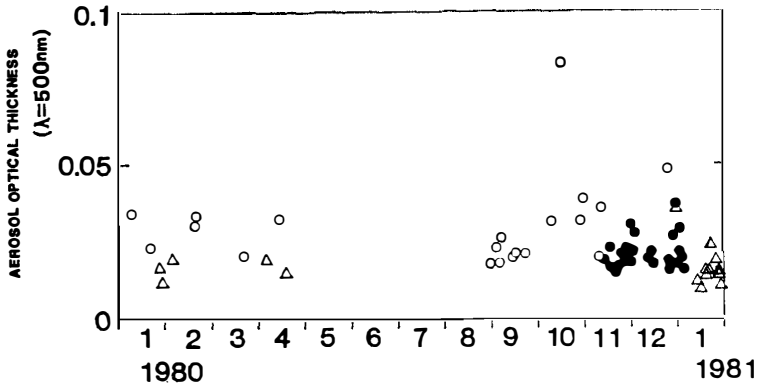

Fig. 4. Seasonal distribution of three types. White circles, solid circles and white triangles denote type 1, type 2 and type 3, respectively.

an air column, this result suggests the seasonal variation of size distribution of aerosol particles in the Antarctic atmosphere.

Figures 5, 6 and 7 show the number size distribution $n(r)$ and volume size distribution $(4/3\pi r^3 n(r))$ obtained by inversion of each type of spectral extinction. Type 1 gives the size distribution which is close to the power law distribution as described before. In type 2, relatively large concentration of small particles can be seen as compared with the power law distribution. Type 3 is the same as that described for type 2 in the range of $0.03 \leq r \leq 0.5$ μm, but in the range of $0.5 \leq r \leq 7$ μm, the size distribution and also the volume size distribution increase as the radii of particles increase.

According to the surface observation of aerosol at Syowa Station (ITO, 1982; IWAI *et al.*, 1981; ONO *et al.*, 1981), in the winter months, aerosols contained originally

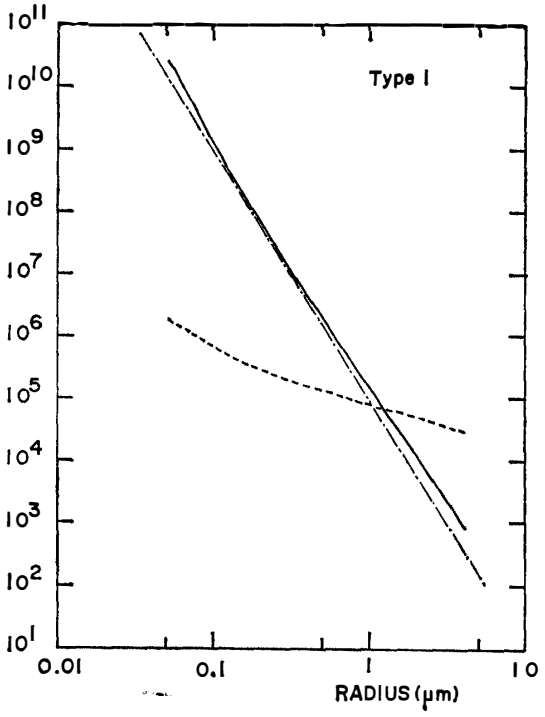

Fig. 5. Size distribution ($n(r)$) (solid line) and volume size distribution ($4/3\pi r^3 \cdot n(r)$) (broken line) of type 1. A dot-dash-line is evaluated by the power law spectrum (unit $n(r)$: $cm^{-2}\ \mu m^{-1}$, $4/3\pi r^3 \cdot n(r)$: $cm^{-2}\ \mu m^{-1}\ \mu m^3$).

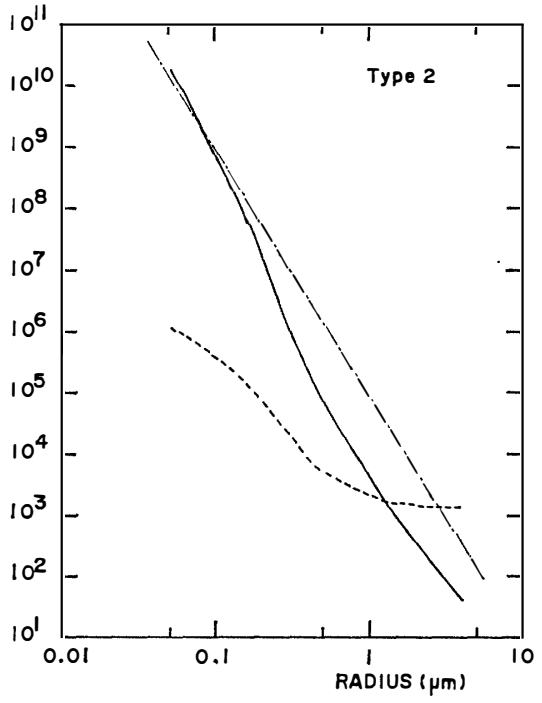

Fig. 6. Same as Fig. 5 but for type 2.

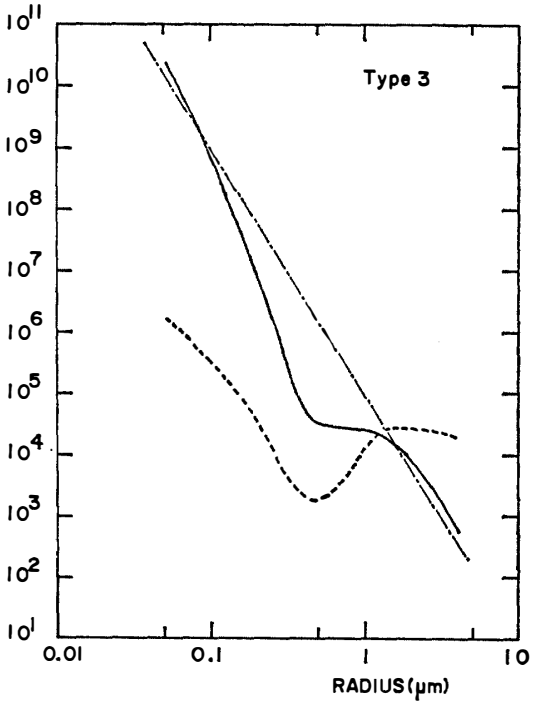

Fig. 7. Same as Fig. 5 but for type 3.

in the "clean" maritime air are dominant, whereas in the summer months the aerosols of sulfic acid droplets produced by gas-to-particle conversion process in the Antarctic atmosphere predominate. From this, it can be roughly said that in the Antarctic atmosphere the aerosols predominate in summer as compared with those in winter.

Since the spectral extinction of type 1, which prevails in winter, gives the size distribution similar to Junge's distribution, the Antarctic aerosols seem to be aged ones. On the other hand, spectral extinctions of type 2 and type 3, which prevail in summer, give the size distribution with relatively high concentration of small particles as compared with Junge's distribution. Therefore, our result also shows that, small particles predominate in the aerosols in summer as compared with those in winter. Thus present results support the inference of the previous work from the observation of the upper atmosphere.

5. Conclusions

Spectral measurements of direct solar radiation were made by a sunphotometer at Syowa Station, Antarctica, for the sunlit period from January 1980 to January 1981. The average aerosol optical thickness ($\lambda=500$ nm) was 0.023 ± 0.010 and its maximum appeared in October and the minimum appeared in November. The monthly mean value of Ångström turbidity coefficient β lies in the range from 0.0045 to 0.0207 and α lies from 1.27 to 2.15. The maximum value of β lies in November and the minimum in October, and also the maximum value of α in October and the minimum in November. Both β and α are approximately constant from February to August.

From the seasonal variation of extinction spectra, seasonal variation of the size distribution of aerosol can be detected. The seasonal variation of extinction spectra is caused by the difference of the origin of Antarctic aerosols.

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Data were analyzed with the aid of HITAC M-200H computer of the Meteorological Research Institute.

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