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In situ measurements of aerosol extinction profiles and their spectral dependencies at tropospheric levels

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ABSTRACT

Aerosol extinction coefficients are measured using rocket and balloon-borne photometers from Thumba (8.5°N) and Hyderabad (17.5°N) respectively, in southern India. The spectral bands are selected within the 300 to 900 nm wavelength region. The aerosol extinction coefficients are found to exhibit large temporal and spatial variations. The spectral dependence of aerosol extinction, derived using a simple Angström power law shows that the wavelength exponent parameter α is altitude dependent. The data are grouped into 2 categories, to study the seasonal variations, in aerosol parameters before and after the Indian monsoon. In general, the premonsoon tropospheric aerosol extinction coefficients are found to be higher compared to the postmonsoon values. Furthermore, a layer characterized by a low α value, situated around 13 km in most of the premonsoon data, is found to come down to about 11 km after the monsoon with a corresponding increase in α.

1. Introduction

Aerosols play a significant rôle in the atmospheric radiation budget because of their direct interactions with the solar and terrestrial radiations as well as through their influence in cloud processes. The tropical lower stratosphere and tropical tropopause are more sensitive to perturbation in radiative heating rates than at other latitudes (Ramaswamy, 1987). Hence, it is important to monitor the altitude and seasonal variations in the aerosol amount and their optical properties, in order to understand the rôle of aerosols in atmospheric radiative processes and climate studies. At the Physical Research Laboratory, a programme of monitoring aerosols began in 1980 using rocket and balloon-borne photometers launched from Thumba (8.5°N 76.9°E) and Hyderabad (17.5°N 78.6°E) respectively. Details of the experiments and data analysis procedure have been reported earlier (Subbaraya and Jayaraman, 1982 and Jayaraman et al., 1987). In the present work, an attempt has been made to study the wavelength dependence of aerosol extinction coefficients and its variation with altitude and season.

2. Experiment

Table 1 shows the list of rocket and balloon measurements made and the center wavelengths used. The bandwidths (full width at half maximum) of the photometers are restricted to less than 10 nm using optical interference filters. In the case of rocket experiments, the photometers are mounted to look at the sun directly during the flight duration. However, in the case of balloon experiments, the photometers are made to track the sun using 2 motors and photodiode assembly in order to compensate for the balloon undulations and to accommodate the changes in solar elevation during the balloon ascent period of about 2 hours. Apart from aerosol studies, in some of the experiments, measurements of ozone and

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Table 1. *List of PRL rocket and balloon measurements made from Thumba and Hyderabad, respectively, and used in the present study*

Sr. no.	Rocket/ balloon	Launch date	Wavelength used (nm) (typical bandwidths 10 nm)
1.	balloon	18 April 1984	280, 310, 450, 800
2.	balloon	22 October 1985	214, 290, 310, 450, 600, 800
3.	rocket	3 October 1986	250, 280, 310, 750
4.	balloon	11 April 1988	302, 310, 496, 670, 860, 950
5.	rocket	12 April 1988	310, 496, 668, 859
6.	balloon	5 November 1988	302, 311, 497, 669, 797, 950
7.	rocket	6 November 1988	309, 495, 668, 795

The center wavelengths of the photometers used are also shown. Data from photometers operated in the UV region ($\lambda < 300$ nm) and at 950 nm are used for ozone and water-vapor measurements whereas the other photometers are used for aerosol studies.

water vapour concentrations are also made using photometers operating at wavelengths less than 310 nm for ozone and at 950 nm for water vapour. Aerosol data are obtained upto an altitude of about 30 km both in the case of rocket and balloon experiments (Jayaraman and Subbaraya, 1988). Above about 30 km, the aerosol extinction in general becomes very small and inseparable from the scattering by air molecules. However, as the aim of the present work is to study the seasonal behaviour of aerosols, results corresponding to tropospheric altitude levels (below 17 km) only are presented. At stratospheric altitudes, the aerosol amount and their size distributions are controlled mainly by volcanic eruptions which are sporadic and are not considered in the present study.

3. Data Analysis

If I is the intensity of the solar radiation at altitude z, then the atmospheric extinction coefficient β, at that altitude can be expressed as:

$$\beta(z) = \mathrm{d}I/[I(z)\,\mathrm{d}z \sec \chi] \quad (1/\mathrm{km}), \qquad (1)$$

where χ is the solar zenith angle at the time of observation. The total extinction coefficient is made up of,

$$\beta = \beta_{ma} + \beta_{rs} + \beta_{ae}, \qquad (2)$$

where β_{ma} is the absorption coefficient due to molecular gases such as ozone and is given as,

$$\beta_{ma} = \sigma_{ma} \times n(z) \times 10^5 \quad (1/\mathrm{km}) \qquad (3)$$

where σ_{ma} is the absorption cross section value (cm^2) and $n(z)$ is the number density per cm^3. β_{rs} is the Rayleigh scattering coefficient and β_{ae} is the aerosol extinction coefficient.

The accuracy of the obtained aerosol extinction profiles depends on the accuracy with which the solar radiation intensity is measured. As the attitude of the rocket does not show any significant deviation in the low altitude region (during its ascent) and due to the sun-tracking mechanism employed in the case of balloon experiments, I-values are measured with an accuracy better than 1%. As the analogue values are digitised and recorded on magnetic tapes for data analysis, there is no further error introduced in the retrieval of I-values. However, as the aerosol extinction coefficients are obtained from the total extinction after subtracting the contribution due to scattering and absorption by the molecular species, the accuracy of the final results depends on the accuracy of the input parameters. In the absence of meteorological radiosonde data on the air density profile, model values (Sasi and Sengupta, 1986; US Standard Atmosphere, 1976) are used. It is found that the uncertainty in the model values are within 2% at tropospheric altitudes. However, the individual contributions from the Rayleigh scattering, absorption due to molecular species and the aerosol extinction vary with respect to wavelength as well as with altitude. For example, at 5 km, the contributions due to Rayleigh scattering to the total extinction is about 74% at 300 nm which decreases to about 12% at 1060 nm because of the λ^{-4} dependence of Rayleigh scattering. Similarly,

the absorption due to ozone molecules is significant below about 330 nm as well as in the 500 to 700 nm region. Absorptions due to nitrogen dioxide in the 400 to 450 nm region, as well as by the water vapour at 800 nm, are less than 1% of the total extinction coefficient at these wavelengths. It is estimated that any variation, of the order of 2%, in the computed Rayleigh scattering coefficient will introduce a maximum error of about 10% in the derived aerosol extinction coefficient at 400 nm, but reducing to about 1.5% at 1000 nm at lower altitudes. Due to appreciable contributions of ozone absorption around 310 and 600 nm, the accuracy of the derived aerosol extinction coefficient values around these wavelength regions is affected by anywhere from 10% to 30% at the upper tropospheric altitudes due to the uncertainties in the ozone concentration values. However, in most of the aerosol measurements, ozone number densities are also measured by employing 2 other photometers operating in the wavelength regions centered around 250 and 290 nm, which are used to correct the aerosol data. In the absence of the simultaneously measured ozone density profile or in case of data gaps, model ozone values (Subbaraya and Jayaraman, 1987) are used. In general, it can be said that the derived aerosol extinction coefficient profiles have a maximum uncertainty of 10% in the lower altitudes which increases to a maximum of about 30% at upper tropospheric levels.

It is also known that the spectral dependence of the aerosol extinction coefficient can be fitted using an analytical expression of the form:

$$\beta_{ae} = K\lambda^{-\alpha} \tag{4}$$

the so called Ångström law. Another useful way of fitting the extinction coefficients is using the expression:

$$\ln \sigma(\lambda) = \ln \sigma(\lambda_0) - \alpha \ln(\lambda/\lambda_0) - \beta[\ln(\lambda/\lambda_0)]^2 \tag{5}$$

proposed by Brogniez and Lenoble (1988) for analyzing the SAGE data. However in the present work, the former expression (eq. (3)) is preferred as the wavelengths used are different in different experiments which prevent the use of a more specific wavelength-dependent expression like the latter. Assuming that the absorption due to aerosols is negligible, aerosol extinctions are studied as a function of wavelength, λ on a log–log plot for all altitudes where the slopes of the fitted lines give the wavelength exponent α. While for Rayleigh scattering, $\alpha = 4$, for aerosol scattering $\alpha < 4$, and is inversely proportional to the size of the particles.

4. Results and discussions

Fig. 1 shows the typical results of aerosol extinction profiles obtained over Thumba and Hyderabad. In general, large variations are seen in aerosol extinction coefficients both with respect to location and time of the measurements. Also, in the case of Thumba measurements, aerosol extinction values are generally high at 310 nm compared to that of Hyderabad. This may be attributed to the difference in the types of particles found over these regions. In general, the aerosol extinction coefficients are found to be high during the March/April period, before the onset of Indian southwest monsoon, compared to in October/November period, after the monsoon at all wavelengths.

Fig. 2 shows the spectral dependencies of the measured aerosol extinction values at 8 and 12 km. No attempt has been made to group the results in terms of the location or the period of observation, but rather, they are shown to bring out the general pattern in the obtained spectral dependencies as a function of altitude. The α-values derived following the procedure described above for the individual cases are shown in paranthesis in the figure. In general, low α-values are found at 12 km compared to those at 8 km, impling that the ratio of the number of bigger particles to smaller particles is higher at 12 km compared to that at 8 km.

Fig. 3a shows the α-profiles obtained for all the individual measurement as a function of altitude. Fig. 3b shows the mean α-value. The horizontal bars correspond to 1σ of the mean. In general, α-values are found to vary between 0.2 and 3 and show altitude variations. Low α-values are found near the ground level as well as in the 11 to 13 km altitude region. As discussed earlier, low α-values indicate a higher mean radius for the particles. It should be noted here that a low α-value and hence a higher mean radius need not always necessarily lead to larger extinction values as the aerosol extinction coefficient also depends on the total

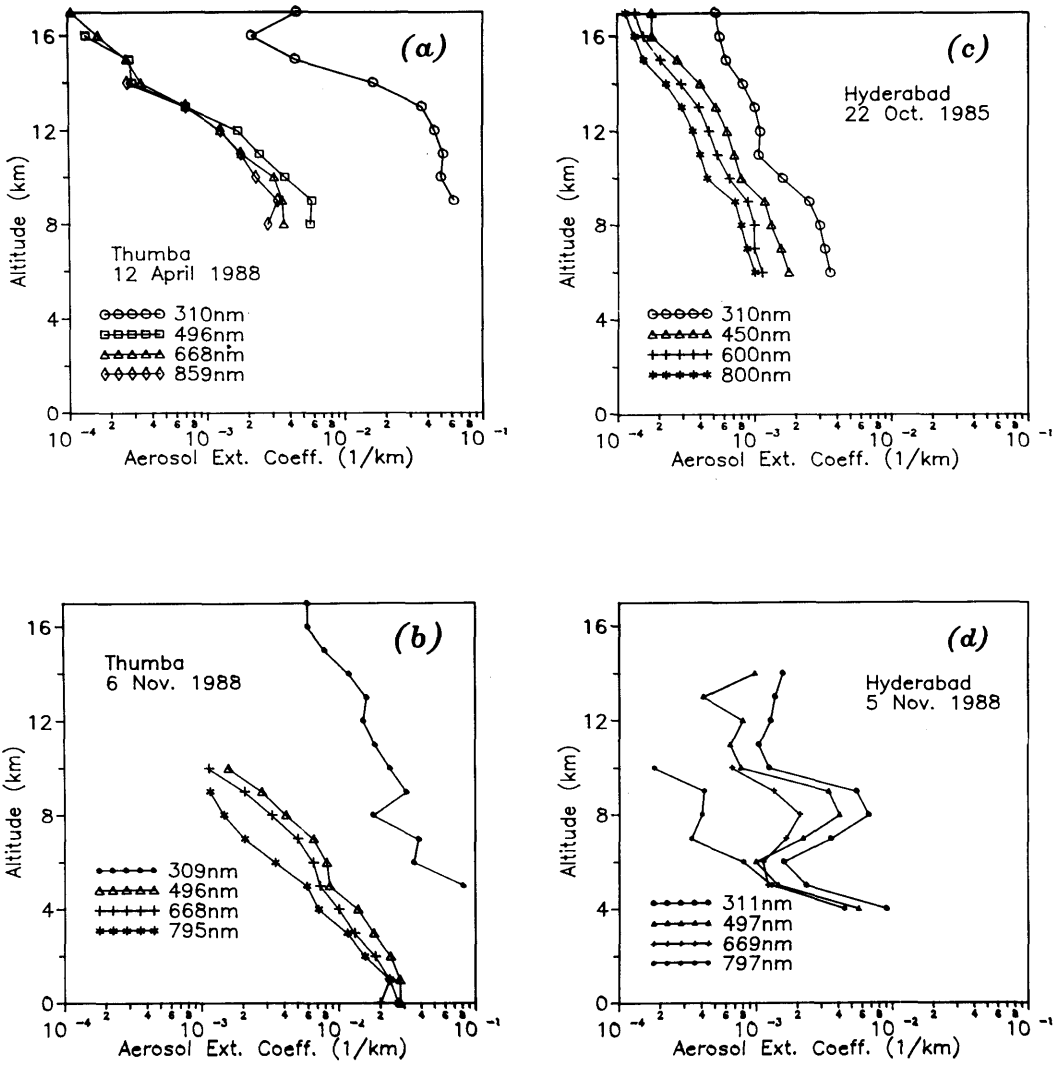

Fig. 1. Profiles of aerosol extinction coefficients measured over Hyderabad and Thumba at the wavelengths and dates given in the figure.

number of particles. From Mie scattering computations, it can be shown that when large numbers of small particles coagulate to form a smaller number of bigger particles, the net extinction coefficient of the ensemble of bigger particles can be less than the sum of extinction coefficients of all small particles (Jayaraman and Subbaraya, 1987).

Analysis of the individual flight results (Fig. 3a) shows that a layer characterised by a low-α value which is situated at around 13 km during the premonsoon period (profiles 1, 4 and 5) is found to come down to about 10.5 km after the monsoon (profiles 2, 3 and 6). Also, there is a definite rise in α, from a mean premonsoon value of 0.7 to a mean value of 1.0 after the monsoon as shown in Fig. 4. The aerosol size distribution can undergo

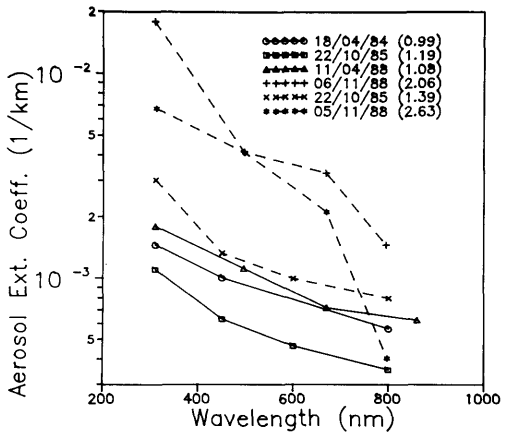

Fig. 2. Spectral dependencies of the measured aerosol extinction coefficients at 8 km (dashed lines) and 12 km (solid lines). The slope of the spectral variation α (eq. (4)) for the individual curves are shown along the dates on which the measurements are made.

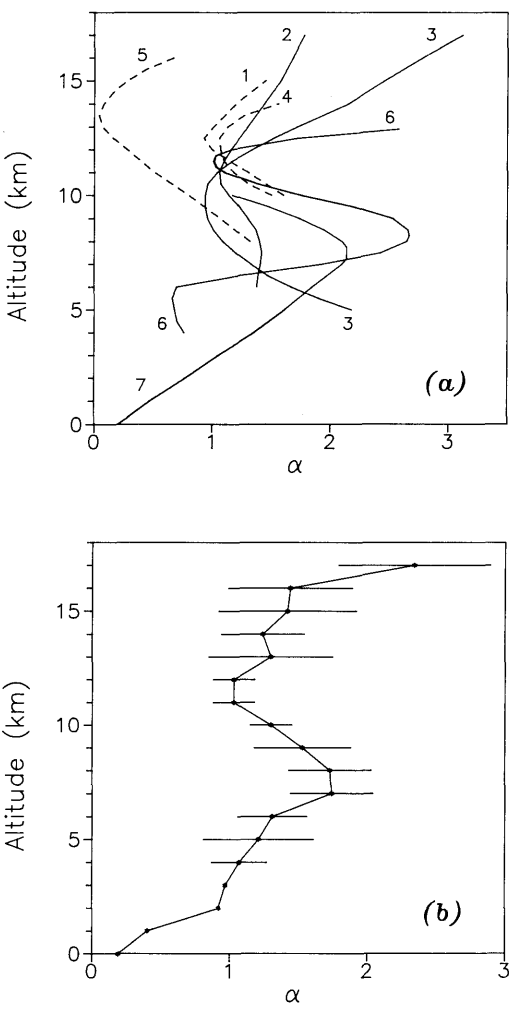

Fig. 3. Altitude variations of the spectral dependence factor α, (a) for all individual experiments numbered 1 to 7 listed in Table 1; (b) the mean value with horizontal bars representing one standard deviation of the mean.

modifications due to various physical and chemical processes (Turco et al., 1982). However these processes are more predominant at stratospheric altitudes. At tropospheric levels, the geophysical and meteorological parameters are important, which influences the aerosol characteristics.

During January to March, dust storms are common in the arid and semiarid regions of western and central India which inject lot of dust particles into the atmosphere up to an altitude level of about 9 km (Middleton, 1986). However, after the monsoon, while the aerosols at low altitudes are rain-washed, the further input of soil-derived particles is also considerably reduced. The columnar aerosol size distribution derived from the ground-based multi-wavelength radiometer observations (Krishna Moorthy et al., 1991) over Thumba also exhibits variations before and after the monsoon. Relatively larger numbers of bigger particles are found during the premonsoon and monsoon periods, which are attributed to the increased production of larger numbers of maritime particles due to strong monsoon winds over the ocean surface. No other measurements on the altitude variation of aerosol size distribution are available over these regions. The observed variations in the aerosol size distribution and the layer height before and after the monsoon indicate that the residence time of these particles in the troposphere should be not more than a few months and is mainly controlled by gravitational settlings and rainout.

5. Conclusions

Vertical profiles of aerosol extinction coefficients, obtained using rocket and balloon-borne photometric techniques show large altitude as well

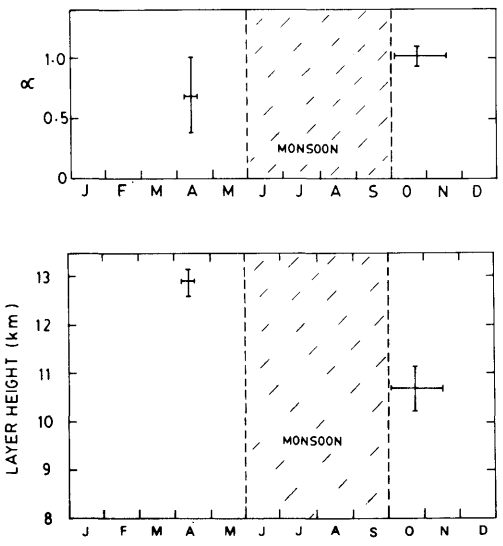

Fig. 4. Variations in the altitude level of the layer having the low α-value and the mean α-value at that altitude, before and after the Indian monsoon.

as seasonal variations. The spectral dependency of aerosol extinctions is found to be different at different altitudes. An interesting feature observed is that a layer characterised by low α (slope of the spectral variation curve), situated at around 13 km in most of the premonsoon data is found to come down to about 10.5 km after the monsoon, with a corresponding increase in α.

6. Future programme

While the PRL rocket and balloon programmes to study tropospheric and stratospheric aerosols are to continue, for the study of the seasonal behaviour in the variation of the aerosol parameters, a systematic monitoring of aerosols is necessary. Recently, a Nd:YAG laser-based LIDAR observatory is being set up and made operational at PRL, which is expected give the important data base for the study of aerosols over the northwest India.

REFERENCES

Brogniez, C. and Lenoble, J. 1988. Size distribution of stratospheric aerosols from SAGE II multiwavelength extinction. In: *Aerosols and climate* (eds. P. V. Hobbs and M. P. McCormick). A. Deepak Pub., 305–311.

Jayaraman, A. and Subbaraya, B. H. 1987. Interaction of EM radiation with aerosols. In: *Atomic and molecular physics* (eds. D. K. Rai and D. N. Tripathi). World Scientific Pub., 511–530.

Jayaraman, A., Subbaraya, B. H. and Acharya, Y. B. 1987. The vertical distribution of aerosol concentration and size distribution function over the tropics and their role in radiation transfer. *Physica Scripta* **36**, 358–361.

Jayaraman, A. and Subbaraya, B. H. 1988. Rocket and balloon measurements of the vertical distribution of aerosols in the Indian tropical region. In: *Aerosols and climate* (eds. P. V. Hobbs and M. P. McCormick). A. Deepak Pub., 117–124.

Krishna Moorthy, K., Prabha Nair and Krishna Murthy, B. V. 1991. Size distribution of coastal aerosols: Effects of local sources and sinks. *J. Appl. Meteorology* **30**, 844–852.

Middleton, N. J. 1986. A geography of dust storms in south-west Asia. *J. Climatology* **6**, 183–196.

Ramaswamy, V. 1987. Aerosol radiative forcing and model responses. In: *Aerosols and climate* (eds. P. V. Hobbs and M. P. McCormick). A. Deepak Publ., 349–372.

Sasi, M. N. and Sengupta, K. 1986. *A Reference atmosphere for the Indian equatorial zone from surface to 80 km, 1985*, Scientific Report: SPL:SR:06:85, Space Physics Laboratory, VSSC, Trivandrum, India.

Subbaraya, B. H. and Jayaraman, A. 1982. Aerosol concentrations and size distributions in the troposphere and lower stratosphere over Thumba. *Pure Appl. Geophys.* **120**, 407–421.

Subbaraya, B. H. and Jayaraman, A. 1987. The vertical distribution of ozone in the equtorial zone. *Adv. Space Res.* **7**, 119–122.

Turco, R. P., Whitten, R. C. and Toon, O. B. 1982. Stratospheric aerosols and theory. *Rev. Geophys. and Space Phys.* **20**, 233–279.

US Standard Atmosphere, 1976. NOAA-S/T76-1562, US Govt. Printing Office, Washington DC, USA.