A Study on Aerosol Optical Depth at a Coastal Station, Trivandrum

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Using a three-wavelength solar radiometer in the visible band, the characteristics of atmospheric aerosol optical depth at the coastal station, Trivandrum, are studied. The aerosol optical depth peaks up in summer months with a deep trough in winter. Its wavelength dependence shows that columnar aerosol size distribution changes from a bimodal type in winter to a relatively monomodal type in summer. The optical depth is found to increase with relative humidity at high humidity values.

1 Introduction

Atmospheric aerosols scatter and absorb optical radiation depending upon their size distribution, refractive index, and total atmospheric loading. This results in attenuation (or extinction) of solar radiation reaching the earth’s surface. The attenuation can be characterized by the parameter, optical depth. Aerosols in the size range 0.1-1 μm are most effective in attenuating sunlight\(^1\). Normally most of the aerosol optical depth is contributed by tropospheric aerosols and a small fraction by stratospheric aerosols. However, major volcanic eruptions can lead to significant increase in stratospheric contribution. Aerosol optical depth varies temporally and spatially, mainly due to variability of tropospheric aerosols.

Atmospheric turbidity measurements with a sun photometer using a wideband filter (300 nm to 630 nm) have been made for a number of years at some selected locations in India\(^2\). Mani et al.\(^3\) carried out measurements on solar flux using filters with cut-off values at 525, 630 and 710 nm at a number of Indian stations. They obtained the Ångström’s turbidity coefficient and wavelength exponent assuming a power law size distribution for aerosols. Though these measurements are very useful in the study of atmospheric turbidity, they are not adequate to obtain information on size distribution except in cases when the distribution follows a simple power law. That this in general is not the case has been borne out by studies using multiwavelength radiometer with narrow-band filters\(^4\). In view of the increasing additional aerosol inputs into the atmosphere, it is necessary to characterize the size distribution of aerosols at different locations in order to understand their interaction with radiation and consequent effects on the earth-atmosphere radiation system.

At Trivandrum (8°33′N, 76°57′E), a solar radiometer was operated during the year 1984 to study the characteristics of aerosol optical depth. Using the radiometer data, the monthly variations and dependence on wavelength and relative humidity (RH) of aerosol optical depth were studied. In this paper, we present the results of this study.

2 Experimental Method and Data Analysis

The radiometer system employed three narrow band interference filters, centred at 420, 620 and 700 nm, each with a bandwidth (full width half maximum) of 10 nm. Solar radiation after passing through the filters was directed on to a linear photodetector by means of suitable optics and the resultant electrical output (\(V\)) was recorded using a digital voltmeter. The overall field of view of the system was 3°. Measurements at the three wavelengths were made at regular intervals of time, each time pointing the system towards the sun to obtain \(V\) as a function of solar zenith angle (\(\chi\)). During morning (0800 to 1030 hrs) and evening (1430 to 1700 hrs), when the variation of sec \(\chi\) with time was rapid, measurements were made at 15-min intervals and in the intervening period, at 30-min intervals. Measurements were made only on days with clear sky condition. During the months March, June to August, and November, measurements could not be made due to cloudy sky conditions.

The system being linear, the recorded signal is directly proportional to the solar flux \(F_\lambda\) reaching the ground at wavelength \(\lambda\). Solar flux \(F_\lambda\) is related to the extraterrestrial solar flux \(F_{\text{0,\lambda}}\) corresponding to mean sun-earth distance \(R_0\), through Lambert-Beer
law. Thus, the recorded solar flux can be represented by

\[ \ln V(\lambda, x) = \ln C_i + \ln F_{0\lambda} + 2 \ln \frac{R_0}{R} - \tau_i \sec x \ldots (1) \]

where \( C_i \) is the system constant at \( \lambda \), \( R \) the sun-earth distance corresponding to the day of observation, and \( \tau_i \) the total atmospheric optical depth; \( \tau_i \) comprises contributions due to Rayleigh scattering (\( \tau_r\)), molecular absorption (\( \tau_{\alpha\lambda} \)), and aerosol scattering and absorption (\( \tau_{a\lambda} \)). By subtracting \( \tau_r \), \( \tau_{\alpha\lambda} \) and \( \tau_{a\lambda} \) from \( \tau_i \), \( \tau_{a\lambda} \) can be obtained. The optical depth is related to relevant cross-section (scattering/absorption) by

\[ \tau = \int_0^h n \sigma d h \ldots (2) \]

where \( n \) is the number of molecules/particles per unit volume and \( h \) the altitude. For total optical depth, the upper limit of integration can be taken as \( h > 30 \text{ km} \) as contributions from \( h > 30 \text{ km} \) to \( \tau \) are negligible. In the case of aerosols, the optical depth \( \tau_{a\lambda} \) can be expressed as

\[ \tau_{a\lambda} = \int_n \pi r^2 Q(\lambda, m, r) N_c(r) \, dr \ldots (3) \]

where

- \( r \) Aerosol radius
- \( Q(\lambda, m, r) \) Mie extinction function
- \( m \) Aerosol refractive index
- \( N_c(r) \) Columnar size distribution

From a plot of \( \ln V(\lambda, x) \) versus \( \sec x \) (Langley plot) and Eq. (1), \( \tau_i \) and \( \ln C_i + \ln F_{0\lambda} = \ln V_0(\lambda) \), where \( \ln V_0(\lambda) \) is the value of \( F_{0\lambda} \) measured by the system, can be obtained. A typical Langley plot obtained on 16 Oct. 1984 is shown in Fig. 1. The measurement values at each wavelength are shown as points and regression fitted lines are drawn through them. In each case the line is extended to meet the Y-axis corresponding to zero air mass. The slope of the line yields \( \tau_i \) and the Y-axis intercept gives \( \ln (V_0) + 2 \ln (R_0/R) \), from which \( \ln (V_0) \) is deduced.

Since \( F_{0\lambda} \) is a constant, any day-to-day changes in \( \ln V_0(\lambda) \) reflect changes in the experimental-system characteristics. In order to find out the system changes, if any, the average of the \( \ln V_0(\lambda) \) values on all the days of observation was obtained along with the standard error for each wavelength (Table 1). As can be seen from Table 1, the standard errors are quite small indicating that changes in system characteristics are insignificant. In view of this, the average value of \( \ln V_0(\lambda) \) has been used to determine the total optical depth for each observation. For the field of view of the present system (3°), the errors in measurement of \( \tau_i \) (due to forward scattering effects) would be less than about 2 per cent.  

**3 Optical Depth due to Molecular (Rayleigh) Scattering and Absorption**

Calculation of \( \tau_{R\lambda} \) was done using the atmospheric model (for obtaining the number of air molecules/m\(^3\)) pertaining to Trivandrum, and that of Rayleigh scattering cross-section taking into account the depolarization correction. For contribution of \( \tau_{\alpha\lambda} \), absorptions due to ozone and water vapour alone were considered at the wavelengths of present interest. The value of \( \tau_{\alpha\lambda} \) due to ozone was calculated using the absorption coefficients given by Kneizys et al., and mean ozone altitude profile for Trivandrum was obtained by combining the balloonsonde measurements and rocketsonde measurements.

The optical depth due to absorption by ozone was obtained as 0.021 and 0.0043 at 420, 620 and 700 nm, respectively. Total ozone data of Kodikanal (a station close to Trivandrum) shows a seasonal variation of about 11.5%. Taking this as applicable to Trivandrum as well, the seasonal variation in optical depth due to ozone was found to be

<p>| Table 1 — Average Values of ( \ln V_0(\lambda) ) at Different Wavelengths |
|----------------------|-----------------|-----------------|</p>
<table>
<thead>
<tr>
<th>( \lambda ) (nm)</th>
<th>( \ln V_0(\lambda) )</th>
<th>Standard error</th>
</tr>
</thead>
<tbody>
<tr>
<td>420</td>
<td>-2.51</td>
<td>0.1</td>
</tr>
<tr>
<td>620</td>
<td>-3.25</td>
<td>0.06</td>
</tr>
<tr>
<td>700</td>
<td>-0.22</td>
<td>0.01</td>
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</table>

Fig. 1 — A typical Langley plot obtained using radiometer data at Trivandrum
± 0.001 and ± 0.0003 at 620 and 700 nm, respectively. As these are quite insignificant, the seasonal variation in ozone absorption was neglected in the present analysis.

Water vapour did not have any absorption band at 420 and 620 nm. However, it had significant absorption at 700 nm. For the calculation of optical depth due to water vapour absorption at 700 nm, the absorption cross-sections as given in Kneizys et al. were used. The total water vapour content in the atmosphere was obtained by the following method.

Altitude profiles of dew point temperature from radiosonde measurements at Trivandrum (from which water vapour partial pressure can be obtained) are available only at 0530 and 1730 hrs IST, but not at other times of day corresponding to the radiometer observations. However, surface relative humidity and temperature data (from which surface water vapour partial pressure can be obtained) are available as functions of time of the day. Using the altitude profiles of water vapour partial pressure at 1730 hrs from radiosonde data at Trivandrum over a period of eight years, an empirical relation between the surface water vapour partial pressure and the total water vapour content was obtained for each month separately. The relationship is of the form

\[ P(h) = P_0 \exp (-h/H) \]  

where \( P(h) \) and \( P_0 \) are the partial pressures of water vapour at altitude \( h \) and surface respectively, and \( H \) is a constant. The value of \( H \) was determined for each month separately using the data of eight years. The total atmospheric water vapour content was obtained as \( P_0 H \) on integration of Eq. (4). It was observed that these monthly mean \( H \) values did not differ in any significant manner from those obtained using data at 0530 hrs IST, indicating that there were no morning to evening changes in \( H \). Assuming that the same monthly \( H \) values were applicable at other times of the day as well, the water vapour content at any desired time was obtained from the surface meteorological data. Using the water vapour content value thus obtained and the absorption cross-section as mentioned above, the optical depth due to water vapour absorption at 700 nm for each of the radiometer observations was obtained.

Finally, the optical depth due to aerosols, \( \tau_{p,k} \), was obtained by subtracting \( \tau_{a,k} \) and \( \tau_{r,k} \) from \( \tau_{p} \).

4 Results and Discussion

4.1 Seasonal Variation of \( \tau_{p,k} \)

From an examination of the \( \tau_{p,k} \) values, it is found that the diurnal changes are quite small compared to longer-term variations. This is also evident from the very small scatter of the experimental points about the regression fitted lines as seen from Fig. 1. Any significant short-term changes in \( \tau_{a,k} \) (which could essentially be caused by changes in \( \tau_{p,k} \)) will result in enhanced scattering of the points with respect to the regression line. For studying the seasonal variations, \( \tau_{p,k} \) values in each month are averaged and plotted against the month as shown in Fig. 2 for the three wavelengths. Though there is a continuous gap of three months in the data (June to August), the trend of seasonal variation is easily discernible; \( \tau_{p,k} \) is high in April and May and low in December and January at all the three wavelengths, indicating a summer maximum and winter minimum. Maximum to minimum ratio is about 4 at 420 and 700 nm, and is much greater at 620 nm. This aspect is discussed in detail in Sec. 4.2. The standard error in the monthly variation of \( \tau_{p,k} \) is ~ 0.02, which is quite small. In view of this, the optical depths in Fig. 2 can be taken to be representative values for the respective months.

Mani et al. reported a summer maximum of Angström turbidity coefficient at a number of Indian stations including both inland and coastal stations. However, for Trivandrum the maximum to minimum ratio as observed by Mani et al. is not as high as in the present investigation. This may be due to the wideband filters employed in their measure-
ments which would smooth out some of the wavelength-dependent variations. It is reported that, in general, in the northern hemisphere, high turbidities are observed in summer\(^\text{11}\).

The observed monthly variations in \(\tau_{\lambda}\) can be attributed mainly to changes in the lower tropospheric aerosols, in view of longer residence times\(^\text{12}\) and smaller contributions from upper tropospheric and stratospheric aerosols. Pueschel \textit{et al.}\(^\text{13}\) attributed the high turbidities in the northern hemisphere summer to (i) increase in atmospheric water vapour content, (ii) increase of tropospheric vertical mixing caused by increased surface heating, and (iii) possible increase of worldwide photochemical aerosol formation caused by the oxidation of volatile organic materials of plant origin. All these three processes could be contributing to the observed seasonal variations at Trivandrum. We have not attempted to estimate the relative importance of these three processes. However, the role of atmospheric water vapour increase is examined in Sec. 4.3.

In addition to the three probable causes listed above, increase in the aerosol input to the atmosphere due to wind-blown dust in summer may also contribute to the observed seasonal variation. Another source of increased aerosol input at a coastal station like Trivandrum is sea surface agitation. The south-west monsoon, setting in generally over south-west coast of India (where Trivandrum is located) between the last week of May and first week of June, is known to cause a change in the surface wind pattern\(^\text{13}\), in addition to heavy rainfall. The differential solar heating and tropographical effects cause pressure gradient with a low pressure over land and high pressure ocean that drive westerly surface winds across the south-west coast of India even before the onset of monsoon\(^\text{13}\), i.e. from April itself. With the progression of summer, the westerlies gain in strength. These strong winds cause considerable agitation of the sea surface (Arabian sea). This continuous strong agitation results in the formation of surfs and white caps on the sea surface close to the coast. It is well known that sea surfs and white caps produce considerable amount of marine aerosols\(^\text{14}\) whose properties depend upon the wind force. The size and number density of the sea spray particles are known to increase with wind speed. These particles are carried by the wind to continental regions due to the favourable wind direction, resulting in increased marine aerosol input into the atmosphere in summer. The premonsoon rains during April-May contribute to aerosol removal processes by wet scavenging. It appears that in summer (April and May), the additional inputs of aerosols prevail over the removal by wet scavenging, resulting in enhanced optical depth as observed. In the months of south-west monsoon and the following north-east monsoon (October-November), heavy rains contribute to aerosol scavenging leading to smaller optical depths. In north-east monsoon and in the winter months, the absence of strong westerly surface winds leaves the sea surface relatively calm (i.e. relatively free from surfs and white caps). However, the surface winds which are north-easterly during the monsoon may carry land aerosols. But this input can be expected to be generally small because the heavy summer rains over the land (in India) would substantially reduce the aerosol loading. In the winter months, because of the much weaker surface winds, the contribution due to wind blown dust to the aerosol input would be very small. The tropospheric vertical mixing would also be weak due to less surface heating in winter as compared to summer. These considerations provide a qualitative explanation to the observed monthly variation pattern of \(\tau_{\lambda}\) at Trivandrum.

4.2 Wavelength Dependence of \(\tau_{\lambda}\)

The wavelength dependence of \(\tau_{\lambda}\) is shown in Fig. 3. In fact, these data are the same as in Fig. 2,
but are presented in this form to bring out clearly the wavelength dependence of $\tau_{pl}$. One remarkable feature of Fig. 3 is the reversal of the trend of wavelength dependence from the winter (December and January) months to the other months. In the winter months, $\tau_{pl}^{-\lambda}$ plots show positive curvature with minimum at 620 nm, while in the other months they show negative curvature with maximum at 620 nm. The changeover in the curvature is gradual. The depth of the minimum decreased from December to January. In February, a weak maximum is shown at 620 nm and in April and May the maximum is very pronounced. It again becomes weak in September and October. The standard error, which is $\sim 0.02$, is quite small, indicating that the plots represent the monthly patterns quite well.

The wavelength dependence of optical depth is governed principally by the aerosol columnar size distribution\(^\text{15}\). The widely used Junge’s inverse power law size distribution yields a monotonically decreasing $\tau_{pl}$ with $\lambda$ (Ref. 4). Obviously, the observed wavelength dependence does not conform to the power law. Using matrix inversion methods, multiwavelength radiometer measurements of $\tau_{pl}$ have been analyzed to deduce the total columnar size distributions. These methods are useful in inverting $\tau_{pl}$ observations even when the wavelength dependence is non-monotonic. King et al.\(^4\) developed a method for such inversion. Using multiwavelength radiometer data in the wavelength range 0.4-0.9 $\mu m$ at Tucson (USA), they deduced the columnar size distributions. They classified the distributions into three types, Types I, II and III. A monotonically decreasing $\tau_{pl}^{-\lambda}$ variation gives an inverse power law size distribution as

$$\frac{dN_c(r)}{d\log r} = Kr^{-\nu} \quad \ldots \quad (5)$$

where $N_c(r)$ is the total columnar aerosol content/unit area, $r$ the aerosol dimension, $K$ a constant, and $\nu$ the power law index. The value of $\nu$ depends upon the rate of decrease of $\tau_{pl}$ with $\lambda$. This distribution is referred to as Type I. Type II size distribution is a monomodal-type distribution with a distinct maximum in the number density at a particular value of $r$. Type II results from a $\tau_{pl}^{-\lambda}$ variation with a negative curvature. The size where $|dN_c(r)/d\log r|$ maximizes depends upon the maximum in the $\tau_{pl}^{-\lambda}$ variation. Type III can be considered to be a combination of types I and II, and results from a $\tau_{pl}^{-\lambda}$ variation with a positive curvature. It resembles a power law-type distribution at smaller sizes, and at greater sizes it shows a maximum. This can be referred to as a bimodal-type distribution with two peaks, one in the smaller-size range and the other in the greater-size range.

As our present observations are available only at three wavelengths, it is not possible to apply the inversion methods which require observations at a greater number of wavelengths (6-8 or even more) to obtain quantitatively reliable size distributions. However, the nature of size distribution functions can be inferred from the foregoing discussion. The $\tau_{pl}^{-\lambda}$ plots with negative curvature observed in the months from February to October indicate relatively monomodal type of size distributions, with a prominent peak at a particular size. The plots with positive curvature observed in the winter months of November and December indicate bimodal type size distribution.

Consideration of aerosol production mechanisms shows that particles with sizes $\leq 0.5 \mu m$ are produced by a combination of nucleation from the gas phase and coagulation processes, whereas particles with sizes $> 1 \mu m$ are mainly the result of mechanical and wind stresses at the earth’s surface\(^4\). The sea salt spray particles extend, in general, in size range from 0.1 to $10 \mu m$ (with a peak in between). A combination of production due to these sources may result in bimodal distribution because of the distinct nature of these aerosol types. On the other hand, a relatively monomodal-type distribution would result when a strong source provides large inputs to dominate over other inputs. From the above discussion, it is clear that the $\tau_{pl}^{-\lambda}$ plots for the months February, April, May and October indicate a strong source of aerosols in these months, leading to a relatively monomodal-type size distribution. This is further corroborated by the fact that the optical depths are generally higher in these months (peaking in April and May). As already discussed, the strong source of aerosols could be from sea surfs and white caps. In the winter months when the wavelength dependence indicates a bimodal-type size distribution, there appears to be no such single dominating source.

4.3 Effect of Water Vapour

Water vapour in the atmosphere can affect optical depth in two ways. One is due to absorption at the particular wavelength under consideration. The other is due to the growth of aerosol particles in size. The growth in particle size affects the size distribution resulting in an increase in the larger size particles and change in the overall refractive index of particles. These changes (in size and refractive index) alter scattering characteristics affecting the optical depth.
In order to investigate the effect of water vapour on $\tau_{\text{a}}$, the optical depth values are grouped according to surface relative humidity (RH) ranges 40-49%, 50-59%, 60-69%, and 70-79% for each wavelength separately. Then these are averaged for each RH group and the resulting plots are drawn (Fig. 4). The vertical bars in Fig. 4 indicate the standard errors. At lower RH values (45% and 55%), $\tau_{\text{a}}$ remains more or less steady but increases rather steeply at higher values of RH. This is observed at all the three wavelengths.

Shettle and Fenn\textsuperscript{16}, using different models of aerosols (different contributions of maritime, continental, rural and urban models), estimated theoretically the effect of water vapour (RH) on aerosol size distribution and refractive index and also the resulting changes in extinction. They found that at high RH values, the extinction increases very significantly. Fig. 5 is obtained from the tabulated values of relative humidity and aerosol extinction in respect of maritime aerosol model by Shettle and Fenn\textsuperscript{16} to show the dependence of extinction on RH at two wavelengths (420 and 700 nm). The maritime model can be considered to be appropriate for a location like Trivandrum. This will be more so for non-winter months for which most of the values in Fig. 4 correspond. A comparison of Figs 4 and 5 shows that the observed dependence on RH is in accordance with the theoretically estimated dependence. It may be noted that the RH values in the present investigation are more or less well distributed in their range in all the months with no significant preference.

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References


9 Kundu N, ISRO-IMAP-SP-09-82 (ISRO, Bangalore), 1982.
13 Rao Y P, Meteorol Monograph No. 1/76 IMD (India Meteor­ological Department, New Delhi), 1976.
14 Junge C, Air chemistry and radioactivity, Int Geophys Ser (USA), 4 (1963) 111.