Chapter 6

SEASONAL VARIATION AND LONG TERM TRENDS

6.1 Introduction

One of the most important effects of aerosol loading in the atmosphere is the deterioration of air quality and thereby reducing the visibility (Went, 1960). The interaction of aerosols with radiation and the consequent effects are important factors in the global climate system. The non-absorbing sub-micron particles cause an increase in the effective planetary albedo and thereby producing a negative radiative forcing on the earth’s surface (Cookley and Cess, 1985; Charlson et al., 1991; Pilinis et al., 1995a,b). Increased loading of carbonaceous aerosols (absorbing type) can have a different radiative impact than that of sulphate aerosols (Cooke and Wilson, 1996).

The characteristics of atmospheric aerosols presented in the preceding chapters are used to study attenuation (or extinction) introduced by these aerosols on optical radiation in the visible region. Physical properties of surface aerosols obtained from LPI measurements and the same in the atmospheric boundary layer obtained from CW lidar measurements are used for this purpose. In addition to these two experimental systems, a Multiwavelength Solar Radiometer (MWR) is in regular operation at Trivandrum (Krishna Moorothy et al., 1989) since 1985 as a part of the Indian Geosphere Biosphere Program (IGBP). This instrument essentially measures the directly transmitted solar irradiance as a function of time, which is used to estimate the average (over a day) columnar aerosol optical depth employing Langley technique (Shaw et al., 1973). The MWR is being operated regularly at ten wavelengths 400, 450, 500, 550, 600, 650, 750, 850, 935 and 1025 nm with 5 nm (FWHM) bandwidth. A good database of columnar aerosol optical depth in these wavelengths for a period of ten years from 1987 is available for Trivandrum (Krishna Moorothy et al., 1999). Using
this data along with those obtained from LPI and CWL measurement (described above), the influence of aerosols on optical radiation in different regions of the atmosphere, their relative importance, relative contribution of different regions to the columnar aerosol loading, similarities and differences in the seasonal variations etc. are studied. In this chapter the results of this study on aerosol extinction at different regions of the atmosphere are presented.

The three different experiment systems described above essentially measures three different characteristics of atmospheric aerosols. The LPI measures the aerosol concentration (mass or number) near the surface along with the size distribution. The CWL measures the altitude profile of aerosol number density assuming the size distribution to be of a modified power law type with a size index 4.5 and aerosol refractive index 1.45. The MWR provides the integrated aerosol extinction in the entire column of the atmosphere (aerosol optical depth). All these parameters are used to compute integrated aerosol extinction coefficients (or optical depths) in different regions of the atmosphere. The near surface aerosol size distribution from LPI is used to estimate the aerosol extinction coefficient at 500 nm wavelength using Mie theory. The number size distributions (humid) obtained for different months are used to study the month-to-month variation of aerosol extinction. The dry aerosol refractive index is assumed to be 1.5 and the refractive index for the prevailing humidity is obtained by using the equation (Shettle and Fenn, 1979)

\[
n = n_w + (n_0 - n_w) \left( \frac{r_0}{\tau(a_w)} \right)^3
\]

(6.1)

where \(n_0\) is the dry aerosol refractive index and \(n_w\) the refractive index of water (which is taken as 1.33). Variation of aerosol refractive index with RH in this case has been presented by Parameswaran (1996). In calculating the extinction all the size distributions are truncated to an upper aerosol radius limit of 10 \(\mu m\) and the lower size limit is 0.05 \(\mu m\). The values of aerosol extinction, \(\alpha_s\) (m\(^{-1}\)), are obtained in different months during the period 1991-1999. These values of aerosol extinction essentially provide the integrated extinction over a path length of 1 m near the surface, or the aerosol optical depth for the altitude region 0-1 m above the surface (which is referred to as \(\tau_s\)). From the altitude profiles of aerosol number density, the mixing region aerosol optical depth (\(\tau_m\)) is estimated by integrating this profile up to 1.1 km and multiplying it
with the effective aerosol extinction cross section (a power law with size index of 4.5 for aerosol size distribution and aerosol refractive index 1.45 used for inverting the lidar equation yields an effective extinction cross section $3.0374 \times 10^{-10}$ cm$^2$ is adopted for this purpose). The columnar optical depth from MWR at 500 nm wavelength is directly used in conjunction with the above data to assess the trends in aerosol extinction at different regions of the atmosphere (viz. near to the surface, within the mixing region, above the mixing region and in the entire column of the atmosphere).

6.2 Aerosol Extinction Near the Surface

The number-size distributions of aerosols obtained in the prevailing atmospheric humidity condition ('humid' condition as described in Chapter 3) for different months during the period 1991-1998 are used to estimate the aerosol extinction coefficient near the surface. The aerosol refractive index for the prevailing RH condition (for the respective month) is estimated using equation (6.1). Figure (6.1) shows the values of aerosol extinction obtained for different months from LPI data.

![Graph showing aerosol extinction coefficient](image)

**Figure (6.1)** Month-to-month variation of surface aerosol extinction for the period 1991-1998.

High values of aerosol extinction observed in the year 1991, 1994 and 1998 are due to the increased aerosol loading in these years. On other years the extinction
coefficient generally lies around 0.1 to 0.2 km$^{-1}$, with a mean value of 0.15 km$^{-1}$ (with standard deviation 0.05 km$^{-1}$). Considering the entire values of aerosol extinction ($\tau_z$) during the period 1990-1998, the mean value of $\tau_z$ is 0.34 km$^{-1}$ with a standard deviation of 0.29 km$^{-1}$, which is quite significant, is contributed by high values of $\tau_z$ encountered in the above noted years.

6.3 Mixing Region Aerosol Optical Depth from CW Lidar

Altitude profiles of aerosol number density (monthly mean profiles) obtained using the CW lidar are used to estimate the monthly mean aerosol optical depths ($\tau_m$). The monthly mean aerosol optical depth up to 1 km is obtained by integrating this profile and multiply it by $\alpha_z$. The lowest altitude for which $N_a$ is obtained is 53 m. While calculating the aerosol optical depth, it is assumed that the aerosol extinction coefficient remains the same at altitudes below 53 m. The values of mixing region aerosol optical depth ($\tau_m$) obtained in different months for the period 1989 to 1997 is presented in Figure (6.2). The mean level of aerosol optical depth for this altitude region is around 0.08 and shows large month to month variation (standard deviation ~0.04) indicating that the seasonal variation is quite significant in this region.

![Figure (6.2) Month-to-month variation of mixing region aerosol optical depth for the period 1989-1997.](image)
6.4 Columnar Aerosol Optical Depth from Solar Radiometer

The monthly mean columnar aerosol optical depth ($\tau_c$) at 500 nm wavelength obtained using the Solar Radiometer is presented in Figure (6.3). The vertical bars represent the standard error due to day to day variability. Only clear sky data is used for this purpose. The columnar aerosol optical depth shows a sharp increase in the monsoon season of 1991 followed by an exponential decay lasting for ~18 months.

![Graph showing month-to-month variation of columnar aerosol optical depth](image)

Figure (6.3) Month-to-month variation of columnar aerosol optical depth for the period 1989-1997.

This is due to the increased loading of aerosols in the stratosphere following the eruption of Mt.Pinatubo volcano in Philippines (15.1°N, 120.4°E) in June (Jager, 1992), the effect of which persisted for about 2 years (Defoor et al., 1992). Mt.Pinatubo eruption was the strongest volcanic eruption in the 20th century. Studies before and after its eruption indicates that stratospheric columnar optical depth reached its minimum value (after the eruption of El Chichon) just before the eruption of Mt.Pinatubo. Mt.Pinatubo volcanic eruption caused abundant injection of sulphur bearing compounds (from the earth crust) into the stratosphere causing an order of magnitude increase in backscatter ratio in these altitudes due to the prevailing micro physical processes (Turco et al., 1982) and GPC leading to the formation of $H_2SO_4$. 
droplets. The size distribution of these aerosols can be fairly represented by a monomodal size distribution with a mode radius around 0.35 μm (Krishna Moorthi et al., 1996). This resulted in an increase of stratospheric aerosol extinction by a factor of around 0.25 in the visible region as can be observed from Figure (6.3). The disturbance caused by this eruption lasted for ~18 months (i.e. up to January 1993). But Figure (6.3) shows a small long-term increase in $\tau_c$ after the period also.

Excluding the effect of Mt.Pinatubo eruption, the mean level of $\tau_c$ is ~0.3 with a standard deviation of 0.011. The seasonal variation is less prominent (relative standard deviation about the mean) in $\tau_c$ compared to $\tau_m$.

### 6.5 Seasonal and Annual Variation at Different Altitude Regions

The data presented in figures (6.1) (6.2) and (6.3) indicate a significant month to month variation in aerosol extinction. To study the seasonal and annual variation in $\tau_s$, $\tau_m$ and $\tau_c$, the values of these parameters for the same month in different years are averaged to obtain the monthly representative value. The month-to-month variation of surface aerosol extinction is presented in Figure (6.4). It shows a prominent maximum in June-July months (southwest monsoon) and minimum in winter. The high aerosol

![Month-to-month variation of aerosol extinction near the surface averaged for the period 1990-1998.](image)
extinction observed during the monsoon period compares favourably with the high in $m_a$ (Figure (3.14a)) during this period. The secondary peak observed in $m_a$ during winter is not observable in aerosol extinction. Thus even though there is an increase in $m_a$ during this period aerosol extinction is lower. This can be attributed to variation in aerosol size distribution. In order to examine this aspect in detail the monthly mean effective aerosol extinction cross section is estimated by normalising the observed size distributions to 1 particle/cm$^3$. Month-to-month variation of this extinction cross section is presented in Figure (6.5). This is independent of the month-to-month variation of aerosol concentration and only depends on the variation in the shape of aerosol size distribution. This shows a maximum in July and minimum in winter. The decrease in effective aerosol extinction cross section in winter overrides the increase in aerosol concentration during this period, which results in a minimum in aerosol extinction Figure (6.4). During the southwest monsoon period both the aerosol concentration (Figure (3.14a)) and aerosol extinction cross section (Figure (6.5)) show an increase resulting in the prominent peak in aerosol extinction as observed in Figure (6.4).

![Graph](image.png)

Figure (6.5) Month-to-month variation of aerosol extinction cross section averaged for the period 1990-1998.

The monthly average values of mixing region and columnar aerosol optical depths ($\bar{\tau}_m$ and $\bar{\tau}_c$) presented in figures (6.6a) and (6.6b). The mixing region aerosol
Figure (6.6) Month-to-month variation of (a) mixing region aerosol optical depth and (b) columnar aerosol optical depth.
optical depth shows two maxima; one during the transition period from winter to summer (during February) and the other during the southwest monsoon period (June-September). The winter season is characterised by lowest mean temperatures, low relative humidity and scanty rainfall (1 to 2% of annual). The observed high \( \overline{\tau}_m \) in February can be attributed to the increased production of continental aerosols from the dry land surface. As seen in Chapter 4, this peak is significantly contributed by particle aloft (above \( \sim 300 \) m). The mixing region aerosol optical depth shows a broad secondary peak during the southwest monsoon period. Increased production of sea spray aerosols (Fairall et al., 1983; Monahan et al., 1983) by the interaction of strong monsoon winds with the water body (sea surface) is an important production mechanism during this period. These aerosols brought over to the land by the favourable wind direction mixes with the already existing continental aerosols and cause the observed increase in optical depth. However, the wet removal of aerosol, cloud activity and monsoon rains is also quite significant during this period. This strong loss process significantly reduces the amplitude of the peak in \( \overline{\tau}_m \). But the continued influx of sea spray aerosols maintains a fairly large concentration of large particles in the lower atmosphere. As described in Chapter 4 this peak is mainly contributed by particles in the lower regions of ABL (the well-mixed region). This is again in conformity with the observed seasonal variation of \( \tau_s \) presented in Figure (6.4).

The columnar aerosol optical depth \( \langle \tau_c \rangle \) is maximum during the local summer months of April and May and shows a minimum during the southwest monsoon months. The summer maximum in \( \tau_c \) is attributed to increased production of aerosols added to strong vertical transport due to increased convection. In situ production of aerosols in the upper troposphere also can be a probable cause for this increase. The months March to May are the hottest months with highest surface temperatures, which induces strong vertical diffusion of aerosol particles. The particles are carried to higher altitudes in the troposphere. But the mixing region aerosol optical depth shown in Figure (6.6a) shows maximum in February. The difference in the seasonal pattern of \( \overline{\tau}_m \) and \( \tau_c \) can be explained on the basis of the fact that while the seasonal variation of
mixing region aerosol optical depth is governed by mesoscale process prevailing in the lower atmosphere, the seasonal pattern of columnar aerosol optical depth is governed by synoptic and global scale variation in the free troposphere and stratosphere. Trivandrum experiences strong summer showers in April-May period (Parameswaran et al., 1995) causing significant aerosol washout in the lower altitudes and thereby reducing the values of \( \tau_m \) in these months. \( \bar{\tau}_c \) shows a significant low during the southwest monsoon months. The convective activity is rather low during this period and the vertical transport is mainly due to the mechanical turbulence caused by wind shears (Parameswaran et al., 1997b). Mechanical mixing though extends to higher altitudes (due to increase in \( h_m \)) is discontinuous in altitude and hence rather weak in transporting surface aerosols to higher altitudes in the troposphere. The sea spray aerosols generated by monsoon winds are mainly confined to altitude below \( \sim 100 \) m (Parameswaran et al., 1998). Thus the monsoon peak in aerosol loading is rather confined to lower altitudes of the mixing region. In addition to direct injection, in free troposphere a significant production of aerosol particles comes through the gas-to-particle conversion processes, which will be more effective during the summer months. Also there will be a significant increase in wet removal processes because of the rain and cloud systems associated with southwest monsoon. About 70 to 75 % of the annual rainfall at Trivandrum occurs during these months. Particles in the middle and upper troposphere take part in cloud cycling and eventually get eliminated. This also can result in the observed decrease in \( \bar{\tau}_c \) during this period.

The contribution to \( \tau_c \) by aerosols in the altitude region above 1 km can be studied by examining the difference between \( \tau_c \) and \( \tau_m \). Grouping individual monthly mean values of \( \delta\tau_c \) (obtained by subtracting monthly mean mixing region aerosol optical depth from monthly mean columnar aerosol optical depth) in the same month but for different years, the monthly average value of \( \delta\tau_c \) for the entire period of analysis (represented by \( \bar{\delta\tau}_c \)) is estimated. As we are interested only in the seasonal pattern of background aerosol system the volcanically disturbed period is excluded for the estimation of \( \delta\tau_c \). The month-to-month variation of \( \bar{\delta\tau}_c \) is presented in Figure (6.7). This shows a prominent semi-annual variation with peak amplitudes in April-May and
November months. This matches favourably with the month-to-month variation of $\tau_c$. This shows that the seasonal variation of $\tau_c$ is governed mostly by altitude above 1 km (i.e. above ABL). The peak in summer is quite prominent. This can mainly be attributed to increased aerosol loading in the troposphere which is partially due to increased flux from atmospheric boundary layer and partly due to increased winter production due to gas-to-particle conversion which is more efficient in summer when temperature in this region is relatively high. The November peak also can be partly attributed to the increased gas-to-particle conversion. In winter the photochemical processes leading to gas-to-particle conversion becomes weaker because of the associated lower temperatures (Masuda et al., 1988) in this region causing a minimum in $\delta\tau_c$. 

![Month-to-month variation of columnar aerosol optical depth from altitudes above 1 km](image)

Figure (6.7) Month-to-month variation of columnar aerosol optical depth from altitudes above 1 km.

Figures (6.6) and (6.7) indicate that the prominent periodicity associated with $\tau_m$, $\tau_c$ and $\tau_c-\tau_m$ is semi-annual. But the position of the peaks in aerosol optical depth changes and their prominence also varies with altitude. The winter peak in aerosol optical depth observed in the mixing region shifts towards summer at higher altitudes. The monsoon peak observed in the lower altitude changes to a minimum in free troposphere.
In order to examine the presence of an annual component in aerosol extinction (or optical depth) a seven month running mean of these parameters are taken and its month-to-month variation is presented in Figure (6.8). A seven monthly running mean effectively suppresses the semi-annual component. The annual component is very prominent in mixing region aerosol optical depth (well above the error bars) with a peak during summer and minimum in October. But for the other two parameters, $\tau_c$ and $\delta \tau_c$, the annual component is not quite significant (considering the error bars). This shows that annual and semi-annual components are quite significant for aerosol extinction in the lower altitudes (below 1 km) which is significantly controlled by surface meteorological features while as at higher altitudes the annual component is less significant.

![Figure (6.8) Month-to-month variation of seven month running mean of (a) mixing region aerosol optical depth (b) columnar aerosol optical depth and (c) columnar aerosol optical depth above 1 km.](image)

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As described in Chapter 4 vertical transport of aerosols from the surface strongly depends on the strength of vertical mixing. Convective mixing is more efficient in vertical transport of aerosols from the surface. Mechanical mixing though elevates the height of well-mixed region, being discontinuous in altitude, is not as efficient as the convective mixing for vertical transport of aerosols. The leakage flux across the mixing height, which is, an important parameter governing the transport of boundary layer aerosols to higher altitudes in troposphere, strongly depends on the eddy diffusion in ABL. It is also seen from section (4.5.4) that above 30% mixing region aerosol concentration is contributed by the well-mixed region. Similarly a significant portion of the columnar aerosol content in volcanically quiescent period is contributed by the lower troposphere (Krishna Murthy, 1998). In this context it will be worth examining the contribution of mixing region to columnar aerosol optical depth based on the above data sets. Using the monthly mean values of $\tau_m$ and $\tau_c$ for the period 1989-1997, the average contribution of $\tau_m$ to $\tau_c$ is estimated by taking the ratio $\tau_m/\tau_c$ which is nothing but the relative contribution of $\tau_m$ to $\tau_c$. Figure (6.9) shows the month to month variation average $\tau_m/\tau_c$ thus estimated for the period 1989-1997.

![Graph showing month to month variation of $\tau_m/\tau_c$.](image)
ratio is highest (~0.3) in February and July and lowest (~0.1) in May and October. The peak during the southwest monsoon period is rather broad (extends from June to September). During the winter and southwest monsoon period about 30% of columnar aerosol optical depth is contributed by the mixing region aerosols (upto 1 km) whereas the contribution decreases to ~10% in April -May (local summer) and October (autumn). This shows that in winter and southwest monsoon period significant component of aerosols produced from the surface remains confined to lower altitudes and vertical transport is less effective. In summer and autumn the vertical transport is more efficient and more aerosols are transported to higher altitudes in the troposphere or there is an increased wash out mechanism in troposphere. The latter may be partially true for the autumn minima but for the minimum in summer is contributed by increased vertical transport across ABL. This is also borne out from the shape of aerosol number density profile discussed in section (4.4.2).

6.7 Long Term Trends

The long term trends in aerosol extinction at different regions have been examined by analysing the 13 month running means of the parameters τ_m, τ_c and δτ_c. A 13 month running mean effectively smoothes out all the short period variations less than one year (seasonal and annual). For near surface aerosol extinction, the data is rather discontinuous. From Figure (6.1), no significant long-term trend in this parameter is observable. The available database on surface aerosol does not depict any long term trend. The 13 month running of τ_m, τ_c and τ_c-τ_m represented by (τ_m)_13; (τ_c)_13 and (δτ_c)_13 are presented in Figure (6.10) as a scatter plot. In taking the running means the volcanically disturbed period from June 1991 to November 1992 has been eliminated from τ_c and δτ_c. All these parameters show an increasing trend from 1990 to 1997. In order to quantify the rate of increase, a linear trend (regression line) is fixed on this data which is presented by a continuous line. The increasing trend in (τ_m)_13 is quite significant in Figure (6.10a). The columnar optical depth shows a levelling off in later years. The slope of the respective regression lines gives the rate of increase. The rate of increase in (τ_m)_13 is 0.0020±0.0003 per year and that in (τ_c)_13 is 0.0105±0.00146 per year. The rate of increase in (δτ_c)_13 is 0.01289±0.001176 per year.
Figure (6.10) Long term trends in (a) mixing region aerosol optical depth (b) columnar aerosol optical depth (c) columnar aerosol optical depth above 1 km.
Taking the year 1989 as reference these increases amount to 2.8% per year in the mixing region ($\tau$$_m$$)$ and 4% per year in the entire column of the atmosphere ($\tau$$_J$$). The increase in free troposphere ($\delta\tau$) is about 7%, which is significantly larger than that in the mixing region. This shows that the observed increase in columnar aerosol optical depth is significantly contributed by the aerosol loading at higher altitudes in the troposphere than that in the mixing region. Thus, there is an increasing trend in aerosol optical depth at Trivandrum from 1990 to 1997 and is significantly contributed by aerosols above the mixing region. It may also be noted in this context that the surface flux estimated by establishing the analytical model (Chapter 5) also shows a general increasing trend (Figure (5.7c)) from 1989 to 1994. The observed increasing trend in $\tau$$_m$ is significantly contributed by increase in surface source strength (aerosol flux from the surface).

6.8 Summary

Seasonal and long term trends in aerosol extinction (at 0.5 $\mu$m wavelength) at Trivandrum during the period 1989-1997 have been studied. The annual variation shows a maximum during the local summer months. A semi annual seasonal variation is seen in lower altitude (in the mixing region) with a prominent winter peak and small secondary peak during the southwest monsoon period. This annual variation becomes insignificant at higher altitudes (in troposphere). A long term increasing trend in aerosol optical depth is observed in the mixing region and aloft. The rate of increase is more at higher altitudes.