CHAPTER-5
MODULATIONS IN AEROSOL CHARACTERISTICS
BY ATMOSPHERIC MOTIONS OF VARIOUS TIME
SCALES

5.1 Introduction

Earth’s Atmosphere can support both circulation processes and the atmospheric waves of different time scales, which are governed by the fundamental physical laws of conservation of mass, momentum and energy [Holton, 1972]. Tropical atmosphere is a seat of strong wave activities of varying spatial and temporal scales such as micro-scale, meso-scale, synoptic and planetary scale, and hence these can modulate not only the circulation parameters but the concentration and properties of atmospheric trace species including aerosols also. These result in varying amounts of inter-hemispheric, inter-continental and vertical transport of the species depending on the amplitude and propagation characteristics of these waves. In the earlier two chapters, the role of synoptic scale motions in modifying the aerosol properties was examined. In this chapter, the role of meso-scale process (land/sea breeze circulations) as well as planetary scale atmospheric waves/ circulations in
modifying the aerosol properties over tropical stations based on three different case studies has been discussed.

5.2 Role of meso-scale circulations in modifying aerosol concentrations – An interesting case study

Meso-scale variability in both column and surface aerosol characteristics have been reported widely over distinct geographical environments and are mostly associated with the dynamics of the local ABL. At the coastal regions, these meso-scale variabilities are associated with sea/land breeze circulations with higher aerosol concentrations during land breeze time. The sea breeze is having a cleaning effect, in depleting aerosols in the lower atmosphere during the daytime by the re-distribution of aerosols [Babu and Moorthy, 2001; Pillai and Moorthy, 2001; Parameswaran et al., 2004; Niranjan et al., 2005]. Dumka et al, [2008] have reported the presence of strong meso-scale variability (within a daytime) in spectral AODs at a high altitude station Nainital, in central Himalayas and this is attributed to the dynamics of ABL as the aerosols are being transported from the polluted valley region to higher altitudes during afternoon hours when the convection is strong. Similar signatures of meso-scale processes associated with the ABL dynamics has been reported in the vertical distribution of aerosols also [De Wekker et al., 2004].

Here the role of sea/land breeze circulations modified by prevailing winds and orographically down katabatic flow in the dispersal of aerosols and trace gas species has been presented, using aerosol Black Carbon (BC) and Carbon Monoxide (CO) as tracers. The study is based on a mountain grassland fire, occurred at a high altitude hilly location, ~ 20 km upwind of the measurement sites. The duration of the event was ~ 19 hours and the impact of this fire occurred on both BC and CO concentrations at the coastal measurement locations and has been disappeared within the TIBL by the next day.
5.2.1 Measurement sites

Continuous measurements of BC mass concentrations have been carried out from Thumba Equatorial Rocket Launching Station (TERLS, 8.55° N, 76.9° E, 3 m msl), ~ 500 m inland off the Arabian Sea coast, and the concentration of CO from the campus of the Centre for Earth Science Studies (CESS, 8.517° N, 76.91° E, 15 m msl), ~ 6 km southeast to the TERLS site. A site map indicating the local geography and the two sampling sites, are shown in Fig.5.1.

![Site Map](image)

Fig.5.1: The location of the grass land fire (1), TERLS (2) and (3) CESS (3) marked in the Trivandrum district map. The locations of TERLS and grass land fire are separated by a linear distance of 20 km. The location of the measurements on the peninsular India is shown in the inset.

One of the sampling sites, TERLS is a remote, plain, coastal area, not in proximity to any major industrial and/or urban activities, and is located ~ 500 m due east of the Arabian Sea coast and ~ 10 km north, northwest of the urban area. The human activity is highly subdued over an extent of ± ~ 3 km parallel to the coast and about 1 km across it about the measurement site, as the area is almost uninhabited. The other measurement site, CESS is located beside a lake and the laboratory where the instrument is installed is under a canopy with ample vegetation. The urban centre of Trivandrum lies ~ 10 km due southeast of the
measurement sites. The supplementary meteorological data on wind speeds and wind directions at surface level are also measured from TERLS. Being coastal region, the measurement sites experience regular land/sea breeze activity [Narayanan., 1967, Prakash et al., 1992].

5.2.2 Description of the event

The event was an extensive, wild grassland fire, which occurred during 15-16, December 2006, at the high altitude hill-station, Ponmudi (8.75° N, 77.1° E, and 915 m msl) (marked by 1 in Fig.5.1) in the Western Ghats at ~ 20 km radially east/ northeast to TERLS. The fire outbreak occurred at around 16:00 local time on 15 December, 2006 and was completely extinguished by the forest officials by 11:00 on the next day. During this in-between period, thick wild grass growth over 54 hectares of hill-land (bound between 8.754° N - 8.764° N and 77.106° E - 77.114° E) was burnt. The strong and changing winds at the site had made the fire fighting rather tough. The speciality of the event was that, though the source region was hardly 20 km far, it was quite elevated (~ 1 km) with respect to the measurement site, and was upwind of the strong sea breeze, which normally persists strong during the evening hours (around the time of fire outbreak).

Near-real-time measurements of BC mass concentrations, carried out using an Aethalometer at 5 minutes interval from TERLS and CO concentrations made using a CO Analyzer (Monitor Europe Model 9830B) at 5 minutes interval round the clock from CESS (Fig.5.1) for the period December 14 to 17, 2006, formed the database for this investigation.

5.2.3 Prevailing winds

The prevailing winds at Trivandrum are generally modulated by daily land/sea breeze activity during the months of November to March/April, when the synoptic winds [above the Atmospheric Boundary Layer (ABL)] are weak and northeasterly, directed from land. The sea breeze usually sets-in between 09:00 and 11:00 local time and becomes
stronger in the afternoon period. It weakens by sunset and the land-breeze sets-in during 20:00 – 21:00 hrs and continues till next morning [Kunhikrishnan et al., 1993]. Seasonally, the land/sea breeze activity is conspicuous during November to April, when the prevailing winds are off-shore [Narayanan, 1967; Srinivas et al., 2006]. From May onwards, the prevailing winds become strong, change to northwesterlies and this sea-wind almost masks the sea/land breezes until October, when the winds reverse its direction again.

The diurnal pattern of the surface winds (speed and direction) for the period of 13th to 17th December 2006 is shown in Fig.5.2 with the bottom panel (b) showing the direction and the top panel (a) depicting the speed. The local coastline of Trivandrum being along 145- 325° azimuth, measured clockwise from north, the urban centre lies upwind of the station during land breeze regime and down-wind during sea breeze regime.

![Wind Speed and Direction](image)

**Fig.5.2:** (a) The wind speed and (b) the wind direction at surface levels at TERLS from 14th to 17th December 2006

When the sea breeze activity prevails over land, the resulting changes in temperature, humidity and roughness that occur near the coast leads to the formation of
Thermal Internal Boundary Layer (TIBL) [Stull, 1988]. This TIBL is quite shallow at the coast but deepens inland and finally merges with the inland boundary layer which is ~ 1 km deep [Kunhikrishnan et al., 1993]. The breeze is being confined within the ABL, propagates horizontally and the circulation is completed by a return flow in the opposite direction at the top of the ABL. This land/sea breeze circulation and TIBL have important implications for coastal meteorology, air quality and pollutant dispersal in coastal regions [Simpson et al., 1998; Panchal, 1993; Miller et al., 2003; Moorthy et al., 2003b]. Both columnar aerosol optical depth and surface concentrations at TERLS site are observed to be influenced by the breeze reversal [eg., Moorthy et al., 1993; Babu and Moorthy, 2002].

5.2.4 Impacts of the event on the concentrations of \( \text{Me} \) and CO

The temporal variations of BC mass concentration (\( \text{Me} \)) for the period of one week from 13 to 17 December, 2006 are shown as a time series in Fig.5.3. In which, each points represents the \( \text{Me} \) value at 5 minute interval as measured by the Aethalometer. The time of outbreak of the forest fire on the hills is marked by an upward pointing arrow on the X-axis and its duration by the horizontal arrow.

![Fig. 5.3. The diurnal variations of \( \text{Me} \) from 13\textsuperscript{th} to 17\textsuperscript{th} December 2006 (From noon to noon). The caps identify nocturnal peaks of each day. The nocturnal peak on the day of fire is identified by a downward pointing arrow on the top abscissa. The event duration, the time of fire outbreak and the time of fire deletion are also shown in figure.](image-url)
The nocturnal peak of the event day is marked by the downward pointing arrow, whereas the nocturnal peaks of the other days are marked by the horizontal caps. It is observed that the concentrations (including the daytime low values) remained substantially high during the event duration. The nocturnal peak of the event day and fumigation peak of the next day also remained high compared to the other days. To examine the effects in detail, the diurnal variations in MB from 14th to 17th December 2006 are plotted in Fig. 5.4(a) for MB and in Fig. 5.4(b) for CO concentrations (CO). The times of local sunrise, sunset and fire outbreak on 15th December are marked and identified in the figures. It is seen that within half an hour of the fire outbreak, (elevated and downwind of the prevailing sea breeze) at the hills, MB and CO started increasing and the increase has become quite substantial by 18:00 hrs. This is in sharp contrast to the normal day diurnal pattern seen on the previous days or on 18th and 19th, when the evening rise in MB starts appearing only after 19:00 local time. Moreover, the nocturnal peak of 15th night is ~ 50% higher than the peak on any other day of that week (Fig.5.4 and Fig.5.5), even though the effect is less pronounced in CO.

Fig.5.4 Diurnal variation of (a) MB and (b) CO from 14th to 17th December 2006. The vertical arrow marks the out break of fire event and the two vertical lines on the x axis represents the sunrise and sunset. The double sided arrow represents the width of the fumigation peak of 16th morning. The fumigation peak for both species of 16th is taller and wider than that of other days.
In the typical diurnal pattern, which is quite similar for both MB and CO, there is a gradual build up in the species (BC and CO) concentration in the morning and a sharp fumigation peak occurs between 7:00 and 9:00 LT almost an hour after the local sunrise and this arises from the combined effects of (i) the well-known fumigation effect in the boundary layer, which brings-in aerosols from the nocturnal residual layer shortly after the sunrise [Stull, 1988] and (ii) the morning build up of local anthropogenic activities in the urban area from where the wind is still directed.

As the day advances, the increased solar heating leads to deeper and more turbulent boundary layer leading to a faster dispersion and hence a dilution of pollutants near the surface. The cleaner sea-breeze adds further to this dilution. This continues until sunset. Since the nocturnal boundary layer is shallower than its daytime counter part by a factor of ~ 3 [Kunhikrishnan et al., 1993] and as the wind speeds are lower during night, there is a rapid reduction in the ventilation coefficient. This results in the confinement of aerosols in a much smaller atmospheric volume and a consequent increase in their concentration during early night period, added with the advection and transport during the evening hours.
The sea breeze stops as night advances, the progressive reduction in anthropogenic activities in the urban area results in a reduction in the generation of BC (as well as CO), while the particles closer to the surface are lost by sedimentation. Similar diurnal variations are also reported for BC and CO at other continental sites [Allen et al., 1999; Chen et al., 2001].

However, on 15th December 2006, the evening enhancement in $M_B$ started much earlier, from around 17:00 LT, about half an hour after the outbreak of the fire at Ponmudi. This is ~ 2 hrs before the normal time of starting of the nocturnal enhancement during other days. A peak (comparable to that of the normal-day nocturnal peak) occurred shortly after 18:00 LT, which is still about 1 hour ahead of the normal time nocturnal peak. From there, $M_B$ (and CO) continued to increase to reach the respective nocturnal peaks, which were as high as 9000 ng m$^{-3}$ for BC (0.8 ppmv for CO), almost double the value of the corresponding peaks of 14th and 17th for BC and CO. These peaks are identified by the downward pointing arrows on the abscissa of Fig.5.3. Comparing with the nocturnal peaks of other days of the week (identified by short horizontal caps in Fig.5.3) on the day of the fire, there is an enhancement in the nighttime peak in BC by ~ 3000 ng m$^{-3}$. Consequently $M_B$ (and CO) remained at a considerably higher levels during the early morning of 16th [Fig.5.4 (a) and (b)], though the decreasing trend was present. The fumigation peak after the sunrise on 16th, is much taller and broader than on the other days (Fig.5.4). As the fumigation effect is due to particles (species) in the residual layer of the previous night, these observations show presence of large abundance (well above the normal day abundance) of BC and CO (associated with the grassland fire on the hills) above the nocturnal layer. The fumigation peaks of 16th (next day of the event) were also broader (by ~ 2 hrs) than on other days as the effluents reaching at night stay in the entrainment zone of 15th night and would reach the surface by the next day fumigation time.
To quantify the impact, the mean variations of \( M_B \) and CO representing normal conditions are estimated using the data for 12, 13, 20 and 21 December 2006, which represented typical normal day conditions without any specific source impact. The day 14\(^{th}\) was not considered because of a dawn-to-dusk strike called by the automobile operators, which might have resulted in some underestimation of \( M_B \) and CO on 14\(^{th}\). In order to quantify the day-to-day variability produced by the effluents of the pollutants from the forest fire, the normal day mean is subtracted from the diurnal variations of 15\(^{th}\) and 16\(^{th}\). The deviations are calculated as \( \Delta_{BC} = [M_B - <M_B>_N] \) in BC and, \( \Delta_{CO} = [CO - <CO>_N] \) in CO concentrations respectively. These mean deviations are therefore attributed to the grassland fire. The variations of \( \Delta_{BC} \) and \( \Delta_{CO} \) are shown in Fig.5.5 [(a) and (b)] from noon (12:00) of 15\(^{th}\) to midnight of 16\(^{th}\) (24:00) December 2006. The zero lines correspond to the normal-day means \( <M_B>_N \) of \( M_B \) in Fig.5(a) and \( <CO>_N \) of CO in Fig.5(b), and the vertical bars in these two figures are their corresponding standard deviations (at 5 minutes interval). Values of \( \Delta_{BC} \) and \( \Delta_{CO} \) above the standard deviations are considered significant, and if they are positive, show an enhancement from normal. In Fig.5.5, \( \Delta_{BC} \) remained significantly positive from 15\(^{th}\) afternoon to 16\(^{th}\) night (~ 20:00 hrs), whereas for \( \Delta_{CO} \) (Fig. 5.5b) the significant positive values are observed from 15\(^{th}\) afternoon to 16\(^{th}\) noon (~ 12:00 hrs) and the deviations became indistinguishable thereafter (probably due to the better miscibility of CO and dispersion by the strong winds and other (ABL) processes). The fumigation peak and the long afternoon low remained higher than the normal. No rainfall was observed in the region during the week of the event and strong northeasterlies prevailed at upper levels. Wind speeds at 850 hPa (~ 1.5 km above msl, 500 m above the source region) were exceeding 10 m s\(^{-1}\) during event time. The NCEP (National Centre for Environmental Prediction) derived wind vectors at 850 hPa on 15\(^{th}\) December 2006 at 17:30 local time (LT), 23:30 LT and 5:30 LT (on 16\(^{th}\)) are shown in Fig.5.6. Near the source region, highest wind speed (~ 10 ms\(^{-1}\)) is observed at 17:30 LT [Fig.5.6 (extreme
left panel), the wind speed decreasing gradually and reaching up to 5 to 7 m s\(^{-1}\) by 16\(^{th}\) early morning [Fig. 5.6 (extreme right panel)].

![Fig. 5.6: 850 hPa wind vectors and total winds for 15\(^{th}\) December 2006 at 12 UT (left panel), and at 18 UT for the same day (middle panel) and at 16\(^{th}\) December 2006 at 00 UT (right panel). The measurement is located on the map by a black dot. Strong northeasterly winds from source region to measurement site. The wind speed is found to be decreasing from (~ 9 m s\(^{-1}\)) from 15\(^{th}\) December 2006 at 12 UT to 16\(^{th}\) December 2006 at 00 UT (> 7 m s\(^{-1}\)) at the measurement site. It appears that the extremely high winds (speed ~ 10 m s\(^{-1}\)) at 850 hPa at 12:00 UTC (17:30 LT) along with the ascending limb of the sea breeze circulation at Ponmudi did the major role in transporting the BC and CO effluent from the source location. This high northeasterly wind would intensify the return circulation of the local sea breeze, which itself has an inherent enhancement in the late afternoon periods. Observations have shown that the sea breeze cell has a vertical extend of ≥ 1 km and has an inland fetch of 20 to 40 km [Kunhikrishnan et al., 1993; Mckendry and Roulet, 1994; Finkele et al., 1995]. Measurements of the offshore extent of land/sea breeze are far less numerous [Simpson, 1994; Atkinson, 1981]. Measurements off Trivandrum during March 1999 have shown that land/sea breeze circulation cell can have an offshore extent of ~ 80 km [Subrahmanyan et al., 2000]. Since the terrain altitude increases from ~ 3 m at the measurement site to ~ 1 km, at the Western Ghats (Ponmudi area, ~ 20 km away) the inland fetch of sea breeze will be limited to that as the mountain peaks, provide an orographic barrier. Since the fire occurred at this high altitude region, the emitted BC particles would circulate along with
the ascending limb of the sea breeze circulation, back towards sea by the return flow shown schematically in Fig. 5.7 (left panel).

Fig. 5.7: Schematic diagram of sea breeze circulation (left panel). The return flow of the sea breeze circulation carrying the BC aerosols and gases in the smoke from the source region to the remote coastal location. The measurement site (2) and the source region (1) are marked in the figure. Schematic diagram of land breeze circulation (right panel). The Katabatic flow carrying the BC and gases down the valley, and then to coastal location along with the land breeze. The locations of measurement site (2) and event location (1) are also marked in figure.

As the flow completes over the ocean, a sharp pollution gradient is encountered when this descending airmass over sea with high concentration of BC particles meets the cleaner marine airmass. This results in recycling of the species towards the coastal region by the low level on-shore flow. Since the impact was seen within ~ 30 minutes of the fire outbreak, it is possible that the off-shore extent of sea breeze circulation was short. However, we did not have any measurements towards this. The fire outbreak occurred ~ 16:00 and the sharp primary peak in BC occurred after 18:00 hrs (eventhough a gradual increase started within half an hour). But this primary peak is less conspicuous in CO measured at another location which was 6 km away from the coast. The primary peak in \( M_B \) occurred in sea breeze regime of the event day [~ 18 hrs, Fig. 5.4 (a)]. Half an hour later, the circulation switched to land breeze (~ 19 hrs) and the corresponding surface wind speed near the measurement site (denoted by 2 in Fig. 5.1) decreased to ~ 0.5 m s\(^{-1}\) [Fig. 5.2 (a)].
Fig. 5.6 reveals the prevalence of strong winds at the elevated source region throughout the period of event duration. By evening, the nocturnal boundary layer forms over the land and the low level inversion associated with this (at ~ 50 m above the ground level at the coastal plains) shields the layer below from BC and CO enriched air at the higher levels. Consequently, the pollutants enhanced region remains within the residual layer. The breeze, which now reversed to land breeze, contributes to this enhancement through the katabatic flow down the hilly region, near the source, carrying pollutants to the valley and from there by the low level land breeze, which carries it towards the coast as shown schematically in Fig. 5.7 (right panel). However, the local activities, in the urban area, which cause the normal nocturnal enhancement in $M_B$ and CO, are still relevant and would add to the already elevated pollutants level at near-surface and with the nocturnal layer capping it, $M_B$ and CO shoots up. This explains the large nocturnal peak (second peak) on the night of 15 December 2006.

Thus the combined effect of land breeze and the katabatic flow down, loaded with BC and CO levels is responsible for the very high values of $M_B$ and CO throughout the night [Fig. 5.5 (a) and (b)]. After sunrise of the next day, the warming of the surface leads to gradual lifting of the nocturnal layer and the capping inversion, leading to bringing-in the pollutants from the residual layer during the fumigation period. The fumigation effect, though present every day, was more significant on the morning of 16th, because of the enhanced concentrations of BC and CO residing in the previous night's residual layer. This results in a very large fumigation peak for both species on 16th morning, the excess amplitude of it above the normal day is almost as large as the nocturnal enhancement ($\Delta_{BC}$ and $\Delta_{CO}$) [Fig. 5.5 (a) and (b)]. It also results in the broader fumigation peak, as the daytime convective mixing needs to become stronger (than the normal days) to flush out the enhanced pollutants concentration. This requires the convective daytime boundary layer to be elevated and the breeze to become stronger and this takes more time resulting in
broadening of the peak. The fire was completely extinguished by 11:00 on 16th and the strong upper air winds and the deeper ABL, both are conducive for efficient dispersion of the particles and gases so that the effect becomes insignificant by the subsequent night.

The case-study, provides an interesting experimental evidence of the importance of the meso-scale processes (land/sea breeze and ABL dynamics) in producing strong modulations to aerosol characteristics (concentration and composition) at a given location, even when the source region is located at apparent downwind and is considerably elevated. Such impacts, though are of considerable short duration than those caused by seasonal changes in airmass (detailed description are given in Chapter 1) would produce quite strong modulations as seen here.

As discussed above, the meso-scale variabilities in aerosol characteristics are more prominent in coastal or high-land locations. For other stations, much longer variabilities of the order of days (planetary waves) are contributing towards the observed features in aerosol properties.

5.3 Planetary scale modulations in Spectral Aerosol Optical depths

As have already been discussed in Chapter 1, planetary scale wave propagation in different regions of the atmosphere produce strong modulations to atmospheric parameters such as pressure, winds, humidity and clouds. As the aerosols are well-mixed within the ambient air in the troposphere, these modulations would reflect in the properties of aerosols also. Several of these planetary scale oscillations are significant only to the tropics (such as MJO) and especially strong over the Indian and west Pacific longitudes [Sperber, 2003]. The Indian Peninsula, lying well within the tropics and surrounded by tropical oceans provide an excellent opportunity to examine the effects of such waves on AOD. Nevertheless, such attempts are virtually non-existent over this region. Here the results of an investigation on the modulations in aerosol properties by the planetary scale (below the
annual scale) atmospheric waves, using continuous measurements of spectral Aerosol Optimal Depth (AOD) during boreal winter (January to March) spanning over a period of six years (2001-2006) from a tropical semi-arid station Anantapur (14.7° N, 77.6° E, 331 m msl, ATP), in the peninsular India are presented.

5.3.1 Data base

Regular measurements of columnar spectral AODs, made at the rural semi-arid tropical station Anantapur [ATP, a station in the ARFI network (Chapter 1)] using a Multi-Wavelength solar Radiometer (MWR) for six years (from 2001-2006), formed the base data for this study. The location of the station in the peninsular India is shown in the Fig.5.8. In addition to the above experimental data, the daily mean wind-field (zonal and meridional) from the NCEP/NCAR reanalysis database has also been used for the above period (up to the pressure level of 500 hPa).

![Fig.5.8: Measurement location, Anantapur (ATP) in the peninsular India](image)

Continuous data (without any gap) is required for the periodogram analysis and hence the station ATP has been selected for the study. Owing its arid nature, especially during winter season, the days are very clear and hence continuous data is available during the season, whereas as the season progresses towards the summer season, the sky
conditions become adverse associated with summer monsoon rainfall. In addition, the wave activities are found to more active during the winter season. As such, winter season has been selected for the study of wave modulations in the AOD at a tropical inland station, where continuous data is available.

5.3.2 Intraseasonal variations

The time series of the spectral AOD obtained for the period 01 January to 31 March has been examined (winter season, where the wave activity was strong over the tropics) for the station ATP for all the six years from 2001 to 2006 and the resulting time series are given in the Fig.5.9 (in panels a to f). The figure clearly depicts presence of quasi-periodic fluctuations, in spectral AOD throughout the season for all the years superposed over a weak increasing trend from winter to pre-monsoon in AOD. Even in a particular year the fluctuations showed large temporal variability from January to March and also strong interannual variability.

Another interesting point to be noted that, during the year 2005 the fluctuations observed in the spectral AODs were very strong, especially during January-February months, so that it could even mask the seasonal trend (Fig.5.9(e), top panel), whereas, during the years 2001 and 2002, fluctuations appeared weaker compared to other years [Fig.5.9 (a) and (b)]. However a quantitative comparison of the amplitudes is difficult from this figure as the mean AOD itself would differ from year to year. The vertical structure of the corresponding zonal winds given in the bottom most panels in Fig.5.9 also revealed similar quasi-periodic fluctuations with intermittent easterlies and westerlies at all the height levels from 850 hPa to 100 hPa with increasing magnitudes at higher altitudes.

Careful examination of both the times series (AOD and zonal wind) unveiled that variations in both the parameter are more or less similar with the high AODs coinciding with the westerlies approximately. To delineate the dominant periodicities in AODs (as
well as in winds), these data were subjected to wavelet analyses [Torrence and Compo, 1998].

Fig. 5.9: The time series of the spectral AOD from January 01 to March 31 (top panels) and the corresponding time series of zonal winds at 850 hPa (bottom panel)
5.3.2.1 Wavelet analysis

The wavelet transform is useful to analyze time series that contain non stationary power (amplitude) at different discrete frequencies [Daubechies, 1990]. Assuming that one has a time series, \(x_n\), with equal time spacing \(\delta t\) and \(n = 0 \ldots N - 1\); a wavelet function, \(\psi(t)\), that depends on a non-dimensional ‘time’ parameter \(t\), to be “admissible” as a wavelet, this function must have zero mean and be localized in both time and frequency space [Farge, 1992].

In the present analysis, the Morlet wavelet has been used, consisting of a plane wave modulated by a Gaussian of the form,

\[
\psi_0(t) = C \exp\left(-\frac{t^2}{2}\right) \exp(i\omega_0 t)
\]

(5.1)

where \(\omega_0\) is the non dimensional frequency, here taken to be 6 to satisfy the admissibility condition [Farge, 1992]. This wavelet is shown in Fig.5.10, where \(\omega\) is the angular frequency.

![Fig.5.10: The wavelet base for Morlet wavelet. The left panel shows the real part (solid) and imaginary part (dashed) for the wavelet in the time domain. The plots on the right give the corresponding wavelet in the frequency domain (from Torrence and Compo [1998])](image)

The continuous wavelet transform of a discrete sequence \(x_n\) is defined as the convolution of \(x_n\) with a scaled and translated version of \(\psi_0(t)\).

\[
W_n(s) = \sum_{n'=0}^{N-1} x_n \psi^* \left[ \frac{(n' - n)\delta t}{s} \right]
\]

(5.2)

where \(\psi^*\) indicates the complex conjugate, \(s\) is the wavelet scale.
To approximate the continuous wavelet transform, the convolution of equation (5.2) should be done \( N \) times for each of the scale, where \( N \) is the number of points in the time series [Kaiser, 1994]. By choosing \( N \) points, the convolution theorem allows to do all \( N \) convolutions simultaneously in Fourier space using a discrete Fourier transform (DFT). The DFT of \( x_n \) is

\[
x_k = \frac{1}{N} \sum_{n=0}^{N-1} x_n \exp(-2\pi i kn / N)
\]

(5.3)

where \( k = 0,\ldots, N-1 \) is the frequency index.

Fourier transform of a function \( \psi(t/s) \) is given by \( \psi(s\omega) \) in continuous limit. More details of the Fourier Transform are given in Chapter 4. By the convolution theorem, the wavelet transform is the inverse Fourier transform of the product

\[
W_n(s) = \sum_{k=0}^{N-1} x_k \psi^* (s\omega_k) \exp(-i\omega_k n\delta t)
\]

(5.4)

where the angular frequency \( \omega \) is defined as

\[
\omega_k = \begin{cases} 
\frac{2\pi k}{N\delta t} & k \leq \frac{N}{2} \\
\frac{-2\pi k}{N\delta t} & k \geq \frac{N}{2}
\end{cases}
\]

(5.5)

Because the wavelet function \( \psi(t) \) is in general complex, the wavelet transform \( W_n(s) \) is also complex. The transform can then be divided into the real part \( (R[W_n(s)]) \) and imaginary part \( (E[W_n(s)]) \), with amplitude of the wave as \( |W_n(s)| \) and phase, \( \tan^{-1}(E[W_n(s)]/R[W_n(s)]) \).

The resulting wavelet spectra for AOD are shown in Fig.5.11 at three wavelengths; 380 nm, 500 nm and 1025 nm (respectively from top to bottom) for all the years (2001 to 2006). The x-axes of the figure represent the day number (January 1 to March 31) and y-axes represent the periodicities in days. The color indicates the absolute magnitudes of the amplitudes of the corresponding periodicities. It is interesting to observe that in all the
years, the dominant periodicities (present with significant amplitude) observed at all the wavelengths are the 30-60 day periodicity and the 13-22 day (quasi 16 day) periodicity. In addition, much shorter periodicities are also present, probably associated with the day-to-day variations in AOD. However, the focus is only on the 30 to 50 days and the quasi-16 day periodicities, the amplitudes of which (>0.08 at 500 nm in any year) are significantly higher than the largest uncertainty (~ 0.03) in the daily mean AOD, mainly because these periodicities are capable of producing intra-seasonal variations (Fig. 5.11).

Fig. 5.11: The wavelet spectra of the spectral AOD at three wavelengths of 380 nm, 500 nm and 1025 nm (from top to bottom) for all the years from 2001 to 2006.
These intra-seasonal variations are very common in almost all meteorological parameters such as wind vectors, pressure, temperature etc [Sperber, 2003]. With a view to identifying these quasi periodic fluctuations, the zonal and meridional wind spectra are examined at the different height levels. The Fig.5.12 [(a) and (b)] shows the resulting wavelet spectra for zonal and meridional winds at 850 hPa for a representative year of 2004. The top panel of the figure reveals the presence of the two dominant periodicities in the zonal wind; one having a period of 30 to 50 days and the other with 13 to 22 days (called the quasi-16 day periodicity).

![Wavelet Spectra](image)

Fig.5.12: The wavelet spectra of (a) zonal wind and (b) meridional winds at 850 hPa for a representative year of 2004.

Examining the meridional wind spectrum in Fig.5.12 (b), it is seen that the quasi-16 day periodicity persisted in the meridional wind (significantly) while the amplitude of the
30-50 days periodicity was insignificant. This gives an indication that, the 30 to 50 days oscillations are the Madden Julian Oscillations (MJO) and shorter periodicities are associated with the quasi-16 day planetary scale Rossby waves.

Examination of the phase propagation helps to ascertain the type of the wave phenomena observed in the wind field, which result in corresponding modulations in the AOD. As such, the phase propagation of these waves has been deduced by analyzing the NCEP zonal wind data as a function of longitude. The results are shown in Fig.5.13 again for the year 2004. Fig.5.13(a) confirms the eastward propagation characteristics of the MJO, while Fig.5.13(b) signifies the westward propagating characteristics of the quasi-16 day Rossby wave.

![Fig.5.13: The phase propagation (top) for MJO and (bottom) for quasi-16 day waves](image-url)

With a view to quantifying the contributions of these waves to the seasonal mean AOD, their amplitudes and phases were estimated; and from the amplitudes, the Percentage
contribution of each periodicity to the Seasonal Mean AOD (termed as PSM) at three wavelengths of 380 nm, 500 nm and 1025 nm are estimated for each year and these are shown in the bottom panels of the Fig. 5.14 [(a) and (b)].

![Graph showing amplitude of periodicities and PSM contributions](image)

**Fig. 5.14:** The amplitude of the (a) MJO and (b) Quasi-16 day periodicity in zonal wind and the percentage contributions of these waves to the seasonal mean AODs.

The top panel of the Figure [Fig. 5.14 (a) and (b)] shows the corresponding zonal wind amplitudes at 850 hPa. It is observed that (i) the PSMs of both MJO and Quasi-16 day oscillations to AOD at 500 nm varied between 10% to as high as ~ 24% during the six years and (ii) These waves exhibited distinct phase relations with those in the winds. While the MJO contribution had the least amplitude in the year 2003, for the quasi-16 day wave this occurred in 2002. Similarly, MJO contributed significantly to the mean AOD in the
years 2001 and 2003; while quasi-16 day contribution peaked in 2004 and 2006. Both these modulations together contributed > 45% to the seasonal mean AOD in 2004, while in other years the combined contributions ranged from 25% to 40%. The variations were quite similar at the short wavelength of 380 nm and at the near infrared wavelength of 1025 nm also, though the absolute magnitudes differed. In all the cases, the values of PSM were closely associated with the zonal wind amplitudes.

Despite the differences being marginal, the PSM of MJO with the AOD at 1025 nm showed a higher correlation coefficient (0.82) and the linear least square fit between the PSM and zonal wind amplitudes yielded a higher value for the slope (0.16 ± 0.04), than with AOD at the shorter wavelength of 500 nm where the correlation coefficient was 0.77 and slope was 0.11 ± 0.05 [Fig. 5.15(a)], the difference in the correlation coefficients being significant. In contrast to these, the PSM of quasi-16 day periodicity to AOD at 500 nm showed a higher correlation coefficient (0.93) and stronger association with AOD than at 1025 nm (0.76) [Fig. 5.15(b)].

These indicated that these waves apparently had a preferred effect on different particle size regimes, with MJO influencing the coarser particle regime more in the Q-16 wave, while the latter inferred the accumulation and fine size regime.

Earlier experiments over the Arabian Sea and equatorial Indian Ocean during winter of 1999 (during INDOEX) using a series of constant altitude balloons have shown strong persistence of MJO activity in the wind field at 900 hPa over the oceans around peninsular India [Appu et al., 2001]. A recent work by Tian et al. [2008] has shown strong intraseasonal oscillations in the MODIS and AVHRR derived AODs, (with magnitudes ± 0.1 in AOD) comparable to those of the annual variations over the MJO-active equatorial Indian and west Pacific Oceans, as well as over Tropical Africa and Atlantic Oceans where high AODs persisted. They also reported the AOD anomalies to be related to rainfall anomalies. There are only a few recent studies on the impact on large scale atmospheric
motions on atmospheric composition and AODs [e.g. Tian et al., 2007; 2008; Wong and Dessler, 2007].

![Graph showing scatter plot between periodicities in zonal wind amplitudes and PSMs for two wavelengths.](image)

**Fig. 5.15**: The scatter plot between the periodicities in zonal wind amplitudes and the corresponding PSMs for two wavelengths of 380 nm and 1025 nm for (a) MJO and (b) Q16.

Based on measurements from several AERONET sites, Khutorova and Teptin [2006] have reported the presence of fluctuations in AODs having periods of 8 to 15 days as well as 30 to 45 days. Our study, however, was confined to clear and cloud-free winter season, with no rainfall. Since the MJO circumnavigates the globe, and the strongest MJO activity is observed over tropical Indian Ocean and western Pacific during boreal winter, within a latitudinal spread of 30°, this wave, with its eastward propagating characteristics, could transport airmass from the mostly arid regions of west Asia and Africa to the eastern locations, across the Arabian Sea. The consequent modulation in the AOD (caused by advected mineral dust and sea-salt aerosols, which are coarse in nature) would be stronger.
at the longer wavelengths as has been seen earlier [Fig. 5.15(a)]. In addition, with a typical propagation speed of \( \sim 5 \) m s\(^{-1}\), these waves would carry particles very efficiently from the vast desert regions lying even from the western part of the Sahara within a week. This is also supported by several recent observations of the impact of advected coarse mode particles over the Arabian Sea during winter/spring season [Li and Ramanathan 2002; Moorthy et al. 2005a; Satheesh et al., 2006b]. On the other hand as the quasi-16 day planetary wave being easterly, the regions of traverse of this wave prior to arrival at Anantapur are mostly far East Asia, central Bay of Bengal and the eastern part of Peninsular India. As the speed of this wave is less than that of MJO (\( \sim 1 - 2 \) m s\(^{-1}\)), the fine particles from these regions take a longer time (\( \sim 10 - 14 \) days) to reach the measurement site than that of MJO. Since the removal mechanisms of aerosol from the atmosphere is weak during the winter season, the life time of these particles are more, so that these modulations (MJO and quasi 16 day) are clearly seen in coarse and fine mode aerosols respectively. Several investigators have shown the dominance of accumulation mode aerosols in the airmass advected from the east [eg., Moorthy et al., 2003a; Huebert et al., 2003]. Spatially resolved aerosol measurements over the land during the winter of 2004 have shown significantly high accumulation mode fraction in the aerosol concentration over the eastern part of peninsular India than in the western part [Moorthy et al., 2005c]. This might be leading to the better association of the PSM of quasi-16 day periodicity with AOD at 380 nm.

With a view to examining this further, daily mean values of the Angstrom exponents (\( \alpha \)), have been estimated from the corresponding daily mean AOD spectra by least squares fitting the measurements to the Angstrom equation (equation 3.1 in Chapter 3) over the entire wavelength region. The resulting timeseries of \( \alpha \) has been subjected to wavelet analysis and the spectrum for the representative year of 2004 is shown in Fig.5.16 (top panel). It clearly indicates the presence of MJO and quasi-16 day periodicities similar
to those seen in AOD, indicating that these waves not only modulate the column abundance but the size spectrum of aerosols as well. By spectrum analyzing the timeseries data of $\alpha$, the amplitudes and phases of the above periodicities were estimated. The amplitudes ($|\Delta \alpha|$) provide the magnitude of the extent to which $\alpha$ deviated from its seasonal mean value while the phase showed the sign of this deviation. The results, shown in the middle and bottom panels of the Fig.5.16, reveal that the deviation in $\alpha$ from its seasonal mean is linearly related to the amplitudes of the corresponding periodicities in zonal wind. The phases (Table 5.1) showed an almost out-of-phase relation between MJO in $\alpha$ and zonal wind in all the six years; where a nearly in-phase relationship prevailed between the fluctuations in $\alpha$ and zonal wind for the quasi-16 day wave. This means that stronger MJO in winds leads to a larger decrease in $\alpha$ from its seasonal mean value, whereas a stronger quasi-16 day wave leads to an increase. This lends further support that the zonal wind MJO modulated coarse mode aerosols and quasi-16 day periodicity modulated accumulation mode aerosols. This strong association of AOD anomalies with the MJO and quasi-16 day wave activities is quite important in understanding the aerosol indirect effect on climate, as well as in reducing the uncertainties in AOD retrievals from satellite data [Tian et al., 2008].

**Table 5.1:** Phase relationship between the fluctuations in zonal wind and $\alpha$.

<table>
<thead>
<tr>
<th>Year</th>
<th>MJO $\alpha$ (degrees)</th>
<th>Phase diff</th>
<th>Quasi-16 day $\alpha$ (degrees)</th>
<th>Phase diff</th>
</tr>
</thead>
<tbody>
<tr>
<td>2001</td>
<td>45</td>
<td>180</td>
<td>230</td>
<td>241</td>
</tr>
<tr>
<td>2002</td>
<td>1</td>
<td>189</td>
<td>260</td>
<td>233</td>
</tr>
<tr>
<td>2003</td>
<td>5</td>
<td>186</td>
<td>214</td>
<td>215</td>
</tr>
<tr>
<td>2004</td>
<td>6</td>
<td>194</td>
<td>221</td>
<td>241</td>
</tr>
<tr>
<td>2005</td>
<td>3</td>
<td>193</td>
<td>223</td>
<td>255</td>
</tr>
<tr>
<td>2006</td>
<td>2</td>
<td>199</td>
<td>190</td>
<td>201</td>
</tr>
</tbody>
</table>
Having examined the effectiveness of the meso-scale circulations and the intraseasonal oscillations in modulating column AOD significantly, the role of inter-annual scale oscillations has been examined.

5.4 Quasi biennial oscillations in Spectral Aerosol Optical Depths

Examining AOD spectra over several sites over the peninsular India during winter season of 1996 to 2003, Saha et al. [2005] have observed enhancements from the climatological mean, nearly every alternate years, with varying magnitudes and attributed it primarily to the large scale atmospheric dynamics. The studies that have preceded have clearly demonstrated the role of the atmospheric oscillations of different time scales in
modulating the aerosol characteristics at short (within a week), seasonal and intra seasonal time scales. With all these with back drop, the response of aerosols to other longer period oscillations such as QBO is examined. For this, AOD data from a few stations having continuous data over a long time period are considered.

5.4.1 Data base

Regular measurements of spectral AODs made at 4 stations, Trivandrum (8.55° N, 76.9° E, 3 m msl, TVM) and Anantapur (14.7° N, 77.6° E, 331 m msl, ATP) in India (South Asia), and Ouagadougou (12.2 °N, 1.4 °W, 290 m msl, OGU) in Africa and Solar Village (25.0 °N, 46.4 °E, 764 m msl, SV) in Arabia, are considered in this study. Details of the stations, instruments used and database considered are given in Table 5.2 and the spatial distributions of the stations are given in the Fig.5.17.

![Fig.5.17: The spatial distribution of the stations](image)

Of these TVM is a semi-urban, coastal equatorial location on the west coast of southern tip of India; Anantapur is rural, semi-arid location in the central peninsula; Ouagadougou, is semi-arid equatorial region of Western Africa lying south of Sahara; and Solar Village, a remote extra-tropical continental site in Saudi Arabia. Out of these four stations, SV and OGU formed part of the AERONET (AErosol RObotic NETwork), where the
measurements are made using CIMEL Sunphotometer, where as TVM and ATP are the IGBP stations with measurements of AOD using MWR. From the individual days AOD spectra, the monthly mean AOD spectra are estimated at each station for the study period and the time series thus generated has been examined for QBO signatures.

5.4.2 Quasi Biennial Oscillations

The time series of monthly mean AOD at 500 nm, smoothed using a five point running mean filter to smooth off the seasonal and intraseasonal fluctuations, is shown in Fig. 5.18 in separate panels for each of the 4 stations. Superposed with each of these are the corresponding monthly mean zonal wind speed U, at 50 hPa level, obtained from the NCEP data and smoothed using the same 5 point running mean filter.

![Figure 5.18](image_url)

**Fig. 5.18.** Time series of the five point running mean AOD (at 500 nm) along with the zonal wind at 50 hPa.

Despite the differences in the magnitudes of the variables plotted, presence of strong oscillations with periods of 2-3 years (Quasi Biennial Oscillations, QBO) is clearly discernible in AOD as well as in U, though less conspicuously at Solar Village. Another
interesting feature seen in Fig. 5.18 is that while the nature of variations in AOD and U appeared to be nearly out-of-phase at TVM and OGU, the near-equatorial stations, they appeared to be nearly in-phase at the off-equatorial stations of ATP and SV.

With a view to resolve the dominant periodicities, wavelet analysis was performed on the time series data as discussed earlier. The results are shown in the Fig. 5.19 for all the stations, for AOD on the left panels and the corresponding zonal winds (at 50 hPa) on the right.

![Wavelet spectra of AOD and zonal winds](image)

**Fig. 5.19:** The wavelet spectra of AOD (at 500 nm) and the corresponding zonal winds at 50 hPa at all the stations

All the four stations revealed strong annual component [Annual Oscillation (AO), period ~ 12 months] in both AOD and zonal wind and significant, yet less prominent, presence of QBO with periodicity in the range around 20 to 32 months. To quantify the strength of these oscillations, the amplitudes as well as phases of these periodicities have been estimated for all the four stations and from the amplitudes, the percentage contributions of
AO and QBO to the long term annual mean AOD were also computed and are given in Table 5.2. At all the stations, with the exception of ATP, the annual variations were quite strong, accounting for 25% to 41% of the climatological annual mean AOD. At ATP, however, the AOs contributed only ~ 10% to the mean. The annual variations are basically associated with annual variations in synoptic meteorology leading to changes in wind field and precipitation (as given in Chapter 6). The role of synoptic meteorology in bringing about the AO over several Indian stations has been well documented [Moorthy et al., 2005a; 2007a; Gogoi et al., 2009] and discussed in Chapter 3 and 4. The semi-arid nature of ATP with very little precipitation (the annual rainfall < 300 mm, less removal) and its location in the central plateau would probably be the reasons for the reduced amplitude of AO.

At this juncture it is desirable to examine the periodicities present in the wind vectors at lower levels, as the winds in the lower troposphere are much favorable for the modulations in the AOD. As such, the wavelet analysis of the monthly mean time-series of zonal wind vectors at 850 hPa have been done for all the above stations and the resulting spectra are shown in Fig.5.20.

![Wavelet Spectra of Zonal Wind Vectors at 850 hPa](image)

**Fig.5.20:** The wavelet spectra of zonal wind vectors at 850 hPa at all the stations
It is very interesting to observe that there is no signatures of biennial oscillations in any of the stations, rather strong signatures of annual oscillations alone are observed. This rules out the possibility of the modulations by the tropospheric biennial oscillations (TBO) in modulating the aerosol characteristics at these locations. Hence it is asserted that that QBO signatures observed in AOD (quantified in the former part of this section) is exclusively related to the stratospheric quasi biennial oscillations.

However, the more significant observation is that the QBO, which is basically a stratospheric phenomenon, contributed as much as 10% to 19% to the annual mean AOD at 500 nm; with the highest contribution (19%) at OGU and least (10%) at SV. A similar analysis done for the longer wavelength (1025/1020 nm, not shown in the figure) revealed that the QBO amplitudes are quite smaller and are often within the measurement uncertainties in AOD, except at OGU. This suggests that the QBO\textsubscript{AOD} are caused more by particles in the accumulation size regime. It is also interesting to note that at ATP, the contribution of QBO to the climatological mean AOD is as much as that of the AO. The phase estimation of these periodicities revealed that, while the QBO\textsubscript{AOD} was out-of-phase with the QBO\textsubscript{U} at the near equatorial stations of TVM and OGU, they showed an in-phase relation at the off-equatorial stations of ATP and SV. To be specific, the easterly (westerly) phase of QBO\textsubscript{U} was associated with the positive phase of QBO\textsubscript{AOD} at equatorial (off-equatorial) stations. The highest amplitude of QBO\textsubscript{AOD} was noticed at the African station of OGU.

As the amplitude of QBO\textsubscript{U} decreases rapidly with latitudes off the equator [Wallace, 1973], its impact on AOD is expected to decrease at the off-equatorial stations such as ATP and SV. Despite, significant QBO\textsubscript{AOD} amplitude was noticed at SV, lying nearly in the subtropics. Occurrence of an anomalous meridional circulation has been reported associated with the stratospheric QBO\textsubscript{U} as the easterly and westerly wind shear regimes propagate downwards [Plumb and Bell, 1982; Collimore et al., 2003]. During the
easterly (westerly) phase of the QBOU, the contrasting thermal regimes in the lower and upper stratosphere favor equatorial convection (subsidence) and an off-equatorial subsidence (convection) [Gray et al., 1992]; hence a cooler and higher (warmer and lower) tropopause near the equator and a warmer and lower (cooler and higher) tropopause at the off-equatorial regions [Collimore et al., 2003]. Hence, during the westerly phase of the QBOU, the divergence from the equatorial upper troposphere would enhance the aerosol loading at the off-equatorial station SV, where the amplitude of QBOU is normally negligible. This leads to the mixing up of tropospheric air, resulting in the entrainment of significant amount of tropospheric aerosols to the stratosphere, leading to an increase in the abundance of stratospheric aerosols, which could then be modulated by the stratospheric QBOU.

Reports of the signatures of QBO in AOD are sparse in the literature. This could be primarily because the QBO is basically a stratospheric phenomenon and its signatures on the column AOD will be small except when either the stratospheric optical depths is high or troposphere-stratosphere exchange takes-place. The MWR analysis has shown that QBO produces discernible signature in AOD (at visible wavelengths), which contributed significantly to the annual mean AOD, particularly at the equatorial locations over Asia and Africa. Recent airborne measurements of aerosol extinction over the Indian region have shown significant contribution of upper atmospheric aerosols to the column AOD (as much as 50-60%) [Satheesh et al., 2009]. Based on the lidar measurements at Gadanki, Kulkarni et al. [2008] reported an increasing trend in the aerosols in the upper troposphere-lower stratosphere region and have reported that the aerosols in these regions contribute ~ 12% to column AOD deduced from concurrent MODIS data.

It is known that the QBO in zonal wind can affect the height of convective clouds. Gray et al. [1992] have reported that the changes in the vertical extent of convection associated with the QBO are directly related to the extent of cloud amount. Strong zonal
wind shear between lower-stratosphere and upper-troposphere would break the internal structure of convective clouds as the clouds penetrate into the stratosphere, while a weak cross tropopause shear would allow clouds to penetrate to higher altitudes. The state of the upper-tropospheric winds at the time and location of cloud formation decides whether the east or west phase of the QBO will be causing the strong shear. As such, the variation in the amount of tropical deep convections (as revealed by OLR) with the QBO has been examined. As cloud scavenging and rainout are effective removal mechanisms of the atmospheric aerosols (particularly in the optically active size range [Flossmann et al., 1985; Saha and Moorthy, 2004], such oscillations in the OLR (cloudiness) and rainfall would modulate the AOD variations.

![Wavelet spectra of OLR at the stations during the study period](image)

**Fig.5.21:** wavelet spectra of OLR at the stations during the study period

The wavelet spectra of the time-series of the monthly mean, NCEP derived OLR (2.5°x 2.5°) for all the stations considered in this study are shown in Fig.5.21. Presence of oscillations of 2-3 years (QBO) as well as of 3-5 years (QTO) periodicities (in addition to the strong AO) are clearly discernible in the figure, with weaker amplitudes at the off-
equatorial station of SV. This evidences the following. In the tropics, the easterly phase of the QBO causes higher tropopause and hence it allows convection to penetrate deeper-than-normal. The deeper clouds will have larger horizontal as well as vertical extent leading to more convergence of mass, moisture, and energy at lower levels. Hence deeper clouds lead to increased cloud amount. Whereas, at the off-equatorial stations, as the QBO effect on tropopause height is weak, the strong zonal wind shear across the tropopause would reduce the convection. The reverse scenario would occur during the westerly phase of QBOu [Collimore et al., 2003]. Our analysis (Table 5.2) revealed an in-phase relation between QBOu and OLR at the equatorial stations of TVM and OGU and an out-of-phase relation at the off-equatorial stations (ATP and SV). An easterly (westerly) QBOu causes cooler (warmer) and higher (lower) tropopause and thereby enhanced (reduced) convection. An out-of-phase relationship between the oscillations (AO/QBO) in AOD and OLR has been seen at all the stations, implying that the AODs tend to be higher during periods of low OLR or increased cloud cover. Similar AOD-cloud cover relationships have been reported in the recent literature [Kaufman et al., 2005; Loeb and Manalo-Smith, 2005; Matheson et al., 2005]. Humidification, wet deposition and cloud (cirrus) contamination (of the measured AODs) are the possible mechanisms for this. Lidar measurements from a tropical location in India have shown higher frequency of occurrence of thin and sub-visible cirrus clouds around the tropopause level associated with deep convections during summer [Thampi et al., 2009]. Growth of hygroscopic aerosols at higher relative humidity (humidification) would lead to increase in AOD and is usually more effective for scattering aerosols [e.g., Jeong et al., 2007].

Since increased cloudiness would, in general, be closely associated with increased precipitation, the wavelet spectra of the monthly mean rainfall (in mm day\(^{-1}\)) at TVM and ATP (where the data were available) have been examined in Fig.5.22. The results clearly show the presence of AO and QBO in rainfall, at both the stations, with appreciable
amplitudes of $\geq 1$ mm day$^{-1}$; even though the highest contribution came from AO. In addition to these two, the signatures of periodicity of the order of 3-5 years (QTO) were also observed. Comparing the phases of QBO and AO in rainfall with that of the corresponding periodicities in AOD revealed an out-of-phase relationship (Table-5.2).

![Wavelet spectra of rainfall at the Indian stations of TVM and ATP](image)

Fig.5.22: Wavelet spectra of rainfall at the Indian stations of TVM and ATP

As the rainfall is a highly efficient aerosol removal mechanism through wet deposition [e.g. Pruppecher and Klett; 1978; Koch et al., 2003; Saha and Moorthy, 2004], the positive phase of QBO in rainfall would lead to increased wet removal, hence reduced aerosol loading and lower AOD.

### 5.5 Summary

Meso-scale atmospheric circulations as well planetary scales of various periodicities significantly modulate the columnar AOD and its spectral dependencies over the tropics and extra-tropics depending on the phases. The study has brought out for the first time the significant role of these natural processes in causing regional heterogeneity in aerosol properties. The main conclusions are:
<table>
<thead>
<tr>
<th>Station and instruments used</th>
<th>Period of Data</th>
<th>Mean AOD</th>
<th>AOD Amplitude</th>
<th>% Contribution</th>
<th>Phase angle (in degrees) Between AOD and U</th>
<th>Phase angle (in degrees) Between AOD and OLR</th>
<th>Phase angle (in degrees) Between AOD and RF</th>
</tr>
</thead>
<tbody>
<tr>
<td>TVM (8.55° N, 76.9° E, 3 m) MWR</td>
<td>1995-2003</td>
<td>0.35</td>
<td>0.09, 0.04</td>
<td>24.4, 11.9</td>
<td>18, 192</td>
<td>201, 195</td>
<td>193, 202</td>
</tr>
<tr>
<td>OGU (12.2.7° N, 1.4° W, 290 m) CIMEL (AERONET)</td>
<td>2000-2006</td>
<td>0.52</td>
<td>0.17, 0.10</td>
<td>29.8, 19</td>
<td>227, 215</td>
<td>202, 229</td>
<td>-</td>
</tr>
<tr>
<td>ATP (14.7° N, 77.6° E, 331m) MWR</td>
<td>2001-2007</td>
<td>0.42</td>
<td>0.04, 0.05</td>
<td>10.0, 11.6</td>
<td>18, 29</td>
<td>199, 215</td>
<td>224, 199</td>
</tr>
<tr>
<td>SV (46.38° N, 24.98° E, 764 m) CIMEL (AERONET)</td>
<td>1999-2007</td>
<td>0.29</td>
<td>0.12, 0.03</td>
<td>39.6, 10.3</td>
<td>15, 22</td>
<td>300, 217</td>
<td>--, --</td>
</tr>
</tbody>
</table>

Table 5.2: Details of the Annual Oscillations and Quasi Biennial oscillations
• The land/sea breeze circulation is found to disperse recycling of the pollutant species (such as BC and CO), which are produced by a mountain grass land fire, very efficiently based on the measurements from the two coastal stations nearby at Trivandrum and such modulations have been scales of a few hours to days. The prevailing upper level circulation, along with the land/sea breeze activity and the atmospheric boundary layer dynamics resulted in remarkable enhancements in the concentration of these species, eventhough the fire outbreak occurred downwind and at an elevated location. During the sea breeze regime, (when the fire broke out), the effluents are transported offshore through the return flow of the sea breeze circulation via the sea breeze and during the land-breeze regime, the effluents came directly from the source region to the valley via katabatic flow and were then transported to the coast through the land breeze.

• Planetary waves, the eastward propagating periodicities of 30-60 days (Madden Julian Oscillations, MJO) and the westward propagating 13-22 days (quasi 16 day) Rossby waves are found to modulate the aerosol properties over the tropics at intra-seasonal time scales. The MJOs are conducive for advection of coarse mode aerosols (dust and /or sea-salt) from the west while the westward propagating quasi-16 day waves appear to favour advection of accumulation mode aerosols from the east. The most important implication of this result is that large scale natural variabilities in atmospheric circulation systems, even within a season, can result in as much as 45% changes in the mean aerosol optical depths as well as in its spectral dependencies (change in aerosol abundance and type) over the tropics. These changes would be important in characterizing regional radiative impacts of aerosols as well as in AOD retrievals from remote-sensed data.

• Stratospheric Quasi Biennial Oscillations were found to produce strong interannual variations to column AOD not only over the tropics but the extra-tropics too, based
on the long-term time series of monthly mean AOD at four stations over the equatorial and off-equatorial stations. The contribution of QBO in AOD (at 500 nm) varied from 10 to 19% to the climatological mean AOD and these were well associated with the QBO in the stratospheric (50 hPa) zonal wind, U. At the longer wavelength (1025nm), the QBO amplitudes in AOD were weak/ inconspicuous. While the QBO_{AOD} and QBO_U, were out-of-phase at the equatorial stations they were in-phase at the off equatorial and subtropical stations.

- As the QBO_U induces a meridional circulation with equatorial convection (subsidence) and off equatorial subsidence (convection) during its east (west) phase, with a cooler (warmer) and higher (lower) tropopause at the equatorial (off-equatorial) regions, the associated vertical as well as horizontal mixing of mass flux would be modulating the AOD.

- Analysis of OLR showed an in-phase (opposite phase) relationship at the equatorial (off-equatorial) stations between the corresponding periodicity in QBO_U, implying that the easterly phase of QBO_U favors more convection (cloudiness) and hence low OLR. At all the stations, QBO_{OLR} had an opposite phase relationship with the QBO_{AOD}.

- It was observed over the Indian stations QBO_{AOD} was out-of-phase with the QBO in rainfall. It is attributed to the increased wet removal by increased precipitation.

The observations point to a section, specially permit to the tropics but generally ignored or overlooked in developing regional aerosol model.

In this chapter, the modulations in aerosol properties by various time scales atmospheric processes have been examined. The resulting changes in the microphysics and source strengths at distinct geographical environments have been discussed in the next chapter.