Characteristics of aerosols over a remote island, Minicoy in the Arabian Sea: Optical properties and retrieved size characteristics

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SUMMARY

Changes in the characteristics of atmospheric aerosol over a marine environment are investigated by making regular spectral extinction measurements in the visible and near-infrared region from a tiny island location, Minicoy (8.3°N, 73.04°E), situated in the Arabian Sea about 400 km due west of the southern tip of the Indian peninsula. The role of seasonally changing air-mass type in causing a regular annual variation in the spectral optical depths is delineated. The association between aerosol optical depths, surface wind speed and rainfall is examined. An increase in wind speed causes an increase in optical depths, the effect is predominant when a marine air mass prevails. The impact of changes in wind speed on optical depths (due to sea-spray production over the sea) is parametrized in the case that the island is influenced by a marine air mass.

Columnar size distributions, retrieved from the spectral optical depths, in general, show a bimodal log-normal distribution in the optically active size range. The accumulation mode is more sensitive to continental air-mass types, while the coarse mode is influenced by the marine conditions. The coarse mode is sharper but its position is variable. Increase in wind speed leads to a remarkable enhancement in the concentration and relative abundance of coarse particles, particularly during the monsoon season. The mass loading and effective radius are well associated and depend on wind speed histories. The findings are discussed.

KEYWORDS: Aerosol characterization Aerosol optical depth Aerosol size distribution Marine aerosols

1. INTRODUCTION

Marine aerosols constitute one of the major components of the global aerosol system. This is important because of its direct and indirect interaction with the incoming solar radiation and consequent effects on climate (Charlson et al. 1987; Fitzgerald 1991; Quinn et al. 1993; Russell et al. 1994; Pandis et al. 1994; Ramanathan et al. 1996). At locations near the coast, marine aerosol characteristics are modified by the continental aerosols advected by offshore winds (Hoppel et al. 1986, 1989; Kim et al. 1995; Satheesh and Moorthy 1997). Similarly, strong onshore winds associated with synoptic weather phenomena advect marine aerosols to the coastal continental locations, producing significant modifications in the coastal aerosol characteristics (Suzuki and Tsunogai 1988; Moorthy et al. 1991, 1993). Tropical oceanic regions adjacent to the Indian subcontinent are under the influence of seasonally varying air-mass types associated with the seasonal movement of the Inter Tropical Convergence Zone. Furthermore, transport of mineral dust from arid regions of Asia would influence the aerosol properties over the coastal and offshore regions of the Arabian Sea. The cyclically changing land–sea breeze circulations mix the boundary-layer aerosols at coastal locations (over the land and the sea). However, these mesoscale processes become unimportant over tiny islands located far off the mainland, and only synoptic systems and local processes would contribute to the aerosol properties there. Thus the marine environment adjacent to the Indian peninsula provides an ideal scenario for studying the aerosol characteristics influenced by all the above processes. Characterization of marine aerosols using long-term observational data has been one of the major objectives of the Aerosol Climatology and Effects (ACE) project, pursued under the Geosphere Biosphere Programme of ISRO/DOS (Indian Space Research Organization of the Department of Space). In this paper the details of a study on aerosol characteristics made from continuous ground-based observations from a tiny island, Minicoy, off the western coast of India, are presented, and the results are analysed from the above perspective.

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Figure 1. Sketch of Minicoy Island (8.3°N; 73.04°E) with the location of the multi-wavelength solar radiometer (MWR). The dashed curve represents the coral reef with the open ocean beyond it (west) and a lagoon to the east of it up to the island.

2. MINICOY ISLAND: PHYSICAL FEATURES

Minicoy (8.3°N, 73.04°E) is a tiny (crescent-shaped) island lying at the southern tip of the Lakshadweep (also known as Laccadives) archipelago, which forms with the Maldives a long, narrow belt of a series of isolated islands and submerged banks (all very small in size) spread over the Arabian Sea and extending north–south from 14°N to 0.6°S in latitude. The island is located about 400 km due west of Trivandrum (8.55°N, 77°E) on the mainland (Indian sub-continent) and has a very small area of $\sim 4.4 \text{ km}^2$. It is $\sim 10.4$ km long, roughly orientated north–south and its width at the widest point is only $\sim 800$ m. A schematic representation of the island is given in Fig. 1. The island is thickly vegetated and its soil is coral lime sand. With a population of over 9000, the island is totally devoid of any industrial or urban-like activities and thus provides an ideal large marine platform from which long-term studies on aerosol properties can be made.

(a) General meteorological conditions

Being a tiny island with negligible land mass, the diurnal changes in solar heating do not produce sufficient land–sea thermal gradients to drive mesoscale circulations. Consequently the prevailing winds are synoptic in nature and generally are those associated with the Indian monsoons. These are highly seasonal, being north-westerlies/westerlies during the period April to October, changing to north-easterlies/easterlies for the rest of the year. During the former period, winds are generally strong, reaching peak values up to 15 to 20 m s$^{-1}$ (during June/July) while during the latter period they are weak or calm ($<3$ m s$^{-1}$). The daily meteorological conditions during the study period are inferred from the regular measurements of the surface meteorological parameters (wind speed $U$, m s$^{-1}$), wind direction ($\theta^\circ$), relative humidity (RH, %), pressure ($P$, mb) and temperature ($T$, °C) at the surface (3 m above ground level) made at three-hourly intervals starting from 0230 IST to 2330 IST for each day. (IST represents Indian
TABLE 1. CLIMATOLOGICAL MEAN VALUES OF SURFACE METEOROLOGICAL PARAMETERS AT MINICOLY

<table>
<thead>
<tr>
<th>Month</th>
<th>$U$ (m s$^{-1}$)</th>
<th>$\theta$ (°)</th>
<th>$\theta'$ (°)</th>
<th>Total rainfall (mm)</th>
<th>$T_{\text{min}}$ (°C)</th>
<th>$T_{\text{max}}$ (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>1.78</td>
<td>83.3</td>
<td>15.5</td>
<td>35.0</td>
<td>22.7</td>
<td>29.6</td>
</tr>
<tr>
<td>February</td>
<td>1.86</td>
<td>70.5</td>
<td>27.6</td>
<td>25.4</td>
<td>23.5</td>
<td>29.8</td>
</tr>
<tr>
<td>March</td>
<td>1.92</td>
<td>59.9</td>
<td>38.4</td>
<td>17.0</td>
<td>24.7</td>
<td>30.5</td>
</tr>
<tr>
<td>April</td>
<td>2.08</td>
<td>33.0</td>
<td>66.4</td>
<td>53.5</td>
<td>26.2</td>
<td>31.1</td>
</tr>
<tr>
<td>May</td>
<td>3.14</td>
<td>13.0</td>
<td>87.0</td>
<td>199.8</td>
<td>26.3</td>
<td>31.3</td>
</tr>
<tr>
<td>June</td>
<td>4.75</td>
<td>1.5</td>
<td>98.5</td>
<td>293.6</td>
<td>25.3</td>
<td>30.0</td>
</tr>
<tr>
<td>July</td>
<td>4.53</td>
<td>1.0</td>
<td>99.0</td>
<td>217.6</td>
<td>25.7</td>
<td>29.5</td>
</tr>
<tr>
<td>August</td>
<td>4.14</td>
<td>5.0</td>
<td>95.0</td>
<td>199.8</td>
<td>25.1</td>
<td>29.4</td>
</tr>
<tr>
<td>September</td>
<td>3.61</td>
<td>11.0</td>
<td>89.0</td>
<td>144.1</td>
<td>25.1</td>
<td>29.5</td>
</tr>
<tr>
<td>October</td>
<td>2.83</td>
<td>28.5</td>
<td>70.8</td>
<td>185.1</td>
<td>24.6</td>
<td>29.6</td>
</tr>
<tr>
<td>November</td>
<td>1.94</td>
<td>32.0</td>
<td>77.5</td>
<td>141.4</td>
<td>23.6</td>
<td>29.2</td>
</tr>
<tr>
<td>December</td>
<td>1.72</td>
<td>57.5</td>
<td>42.0</td>
<td>75.7</td>
<td>23.3</td>
<td>29.7</td>
</tr>
</tbody>
</table>

$U$ is the monthly mean wind speed, $\theta$ is wind direction and $T_{\text{min}}$ and $T_{\text{max}}$ are the monthly mean minimum and maximum temperatures.

Standard Time which corresponds to 82.5°E longitude.) In addition to this, a hand-held cup anemometer was also used to measure the wind speed at periods not covered by the regular observations. The daily mean values of wind speed and direction for the period of study (January 1995 to December 1996) are shown separately for each month of the year in the polar diagrams in Figs. 2(a) and 2(b), where the individual points represent the daily mean wind speed. The wind speeds are represented radially with values indicated at discrete intervals increasing in steps of 5 m s$^{-1}$ by the three concentric circles. High wind speeds ($U > 5$ m s$^{-1}$) occur more frequently (>50% of days) during May to October compared with other months. During these months, the winds are generally confined to the fourth quadrant (clockwise with respect to north) and are coming from west of north. During December, January and February, winds are weak (generally <5 m s$^{-1}$), and show very little variation from day to day. Also, they come mostly from east of north and are confined to the first quadrant. Transitions from one pattern to the other occur generally during March–April and November. These features, in general, conform to the climatological patterns observed at this station and are given in Table 1, which is based on long-term data (for the last 20 years) of surface meteorological parameters, provided by the Meteorological Centre, Trivandrum. The relevance of the seasonally changing wind patterns is that they represent basically distinct air-mass types. From the geography (Fig. 1) it is evident that surface winds blowing from east of north would represent winds from the Indian sub-continent having a shorter sea track, and thus with a continental influence (as far as aerosol characteristics are concerned) whereas those blowing from west of north would represent a marine air mass with the air having a much longer trajectory over the ocean.

The monthly total rainfall experienced at Minicoy during the 1995 and 1996 period is shown in the histogram in Fig. 3, which shows that most of the rainfall experienced is during the summer monsoon period and is, in general, in agreement with climatological features given in Table 1, according to which ~88% of the annual rainfall occurs in this season. Very little rain occurs during the December–January–February months. The ambient temperature does not show significant variations from month to month and is typical of the marine environment. In view of the close similarities of the behaviour during the study period with the climatological patterns, it is reasonable to consider that
Figure 2. Polar diagrams showing the distribution of daily mean surface wind combined for 1995 and 1996: (a) for January to June and (b) for July to December. The angles are measured clockwise with respect to north.
Figure 2. Continued.
the aerosol characteristics inferred from this two-year period are representative of the general features prevailing in this region.

3. INSTRUMENTAL SET-UP, DATABASE AND ANALYSIS

Aerosol studies are made using a multi-wavelength solar radiometer (MWR) capable of making automatic and continuous spectral extinction measurements at ten narrow wavelength bands centred at 380, 400, 450, 500, 600, 650, 750, 850, 935 and 1025 nm, with full width half maximum band width in the range 6 to 10 nm at different wavelengths. This MWR is identical to the one used for aerosol optical depth studies and described in detail elsewhere (e.g. Moorthy et al. 1997). The field-of-view is limited to \( \sim 2^\circ \) using lens-pinhole-detector optics, so that the effect of diffuse radiation entering in to the system field-of-view on the retrieved optical depths is insignificant (Shaw 1976; Russell et al. 1993; Satheesh and Moorthy 1997). The MWR is installed at the premises of the meteorological observatory, Minicoy, almost at the centre of the island (the approximate location is marked in Fig. 1), and operated in association with observers of the observatory.

Total columnar atmospheric optical depths (\( \tau_\lambda \)) are estimated from the MWR data following the Langley technique (e.g. Shaw et al. 1973). \( \tau_\lambda \) is the sum of contributions from molecular scattering (\( \tau_{RL\lambda} \)), scattering and absorption due to atmospheric aerosols (\( \tau_{p\lambda} \)), absorption due to ozone (\( \tau_{O_3\lambda} \)), absorption due to NO\(_2\) (\( \tau_{NO_2\lambda} \)), and absorption due to water vapour (\( \tau_{w\lambda} \)); each of these being separate functions of wavelength. The aerosol optical depth (\( \tau_{p\lambda} \)) is determined by subtracting the other components from the total optical depth (\( \tau_\lambda \)), so that,

\[
\tau_{p\lambda} = \tau_\lambda - \tau_{RL\lambda} - \tau_{O_3\lambda} - \tau_{w\lambda} - \tau_{NO_2\lambda}.
\]  

(1)

Details of estimating the individual optical depth components on the right-hand side (r.h.s.) of Eq. (1) and errors in the estimated \( \tau_p \) are described elsewhere (Satheesh and Moorthy 1997).

The MWR system was installed at Minicoy in January 1995 and since then was operated regularly on all clear days and during periods when clouds are far away from the solar disc (when viewed from the observation location). The data obtained for the period from January 1995 to December 1996 are analysed to deduce \( \tau_p \) at all ten wavelengths. (The filter at 380 nm was included in November 1995 and, as such, the data at this wavelength are only available since then.) Generally, the raw data obtained
Table 2. The MWR Database at Minicoy

<table>
<thead>
<tr>
<th>Month</th>
<th>Number of datasets</th>
<th>Month</th>
<th>Number of datasets</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>20</td>
<td>July</td>
<td>5</td>
</tr>
<tr>
<td>February</td>
<td>45</td>
<td>August</td>
<td>11</td>
</tr>
<tr>
<td>March</td>
<td>57</td>
<td>September</td>
<td>12</td>
</tr>
<tr>
<td>April</td>
<td>34</td>
<td>October</td>
<td>14</td>
</tr>
<tr>
<td>May</td>
<td>16</td>
<td>November</td>
<td>33</td>
</tr>
<tr>
<td>June</td>
<td>5</td>
<td>December</td>
<td>20</td>
</tr>
</tbody>
</table>

during one day is analysed as a single set to deduce one set of \( \tau_{p,\lambda} \) values for that day, which is valid only if \( \tau_p \) remained statistically stable during the observation period for that day. This has been checked and confirmed by performing a Student's \( t \)-test on the data and accepting only those sets where the mean variance of the regression slope was small enough for a confidence level (of acceptance) of better than 0.95 for the appropriate number of degrees of freedom (Kleinbaum and Kupper 1978). During some occasions the Langley plot showed two distinct slopes; one for the forenoon and the other for the afternoon; each one being separately significant in accordance with the \( t \)-test. On such occasions, the data are analysed separately, considering the forenoon and afternoon data as two independent sets. A total of 272 sets of \( \tau_{p,\lambda} \) values thus obtained formed the database for the present study. A detailed error analysis, following Russell et al. (1993), considering the contributions due to the variation in the Langley intercept (which was typically less than 5%), uncertainties in the estimates of the various terms on the r.h.s. of Eq. (1), and other instrumental errors, yields an overall maximum error in the estimated value of \( \tau_{p,\lambda} \) as \( \sim 0.01 \) to 0.015 (Satheesh and Moorthy 1997). The monthly distribution of this database, considering both years together, is given in Table 2. As can be seen from it, for the months from June to October the number of observations are far less numerous compared with other months. This is because of cloudy sky conditions prevailing during these months, due to the monsoon activity.

4. RESULTS AND DISCUSSION

Aerosol spectral optical depth values showed day-to-day variations, which were generally of random nature. However, these random variations are found to be superposed on a more regular variation, which has a longer periodicity. Generally, low values of optical depth are encountered frequently during January and February. Higher values occur more frequently in April to July. Thus, the optical depth values are averaged for each month and for both years of the study period to smooth out the random components. These monthly averages are examined for temporal variations.

(a) Temporal variations

The annual variations (month to month) of aerosol optical depths are shown in Fig. 4 (at all the ten wavelengths) where the solid points represent the mean value of \( \tau_p \) for each month, estimated using the individual values in that month combined for both the years. The vertical bars through the points are the standard errors (equal to \( 1/\sqrt{N} \) times the ensemble standard deviations, \( N \) being the number of individual \( \tau_p \) values in the ensemble as given in Table 1). The variations are generally consistent over the two-year period. It is seen that:
Figure 4. Annual variations at Minicoy of monthly mean aerosol optical depth at ten wavelengths. The vertical bars through the points are the standard errors. The gap in the 380 nm record is due to absence of data owing to the late introduction of this filter.
(i) Low values of aerosol optical depths occur (annually) during the months of January–February at the longer wavelengths (λ ≥ 650 nm). At the shorter wavelengths low values occur during the post-monsoon month (September).

(ii) Variations in \( \tau_p \) within a month are generally higher during June to October, while quite small during January to March, as indicated by the length of the error bars (notwithstanding the fact that the number of days of data per month is lower in the former case).

(iii) The general nature of the variations of \( \tau_p \) resembles that of wind speed shown in Fig. 2.

(b) Spectral characteristics

The spectral variation of \( \tau_p \) is important in as much as it is indicative of the changes in aerosol size characteristics. In Fig. 5, the variation of monthly mean aerosol optical depth is shown as a function of wavelength (λ), where the solid line represents spectral variations during 1995 and the dotted line those for 1996. In general, the spectral dependence of \( \tau_p \) remains nearly similar in 1995 and 1996, particularly during January to May and November. During the period November to March, when the prevailing winds are predominantly from the north-east, \( \tau_p \) shows a general decrease as λ increases, which is typical of continental aerosols (Junge 1963). The spectral variations become shallower during April and May due to the relative increase in \( \tau_p \) at the near-infrared (NIR) wavelengths. During the period August to October \( \tau_p \) is almost wavelength independent (typical of a marine environment (Moorthy et al. 1997)) due to a large reduction at the shorter wavelengths. There is also a large increase in \( \tau_p \) in June–July followed by a sharp decrease. As the Mie theory for scattering (McCartney 1976) shows that maximum extinction at a given λ is caused by aerosols having a size in the range from λ/2 to λ, the above feature is indicative of an increase in the relative concentration of larger particles during the monsoon months. These aspects are examined in detail in section 6 of this paper that deals with the size distributions and related aerosol size parameters.

(c) Dependence of \( \tau_p \) on wind speed

The dependence of aerosol abundance, size distribution and optical depth in the marine (and coastal) boundary layer (MBL) on wind speed over the sea has been a subject of intense investigation over the years (Woodcock 1953; Blanchard 1963; Lovett 1978; Exton et al. 1985; Hoppel et al. 1990; O'Dowd and Smith 1993; Moorthy et al. 1997). All these observations have indicated that an increase in wind speed results in an increase in the abundance of aerosols, particularly of the super micron-sized aerosols in the MBL with consequent effects on \( \tau_p \) and size distributions. As Minicozzi is a tiny land mass situated in open ocean, the aerosol optical depths over it are likely to be influenced by changes in wind speed, analogous to open-sea environments. Besides, the temporal and spectral variations of \( \tau_p \) (in Figs. 4 and 5) reveal a similarity in the nature of variation with those of wind speed (Fig. 2).

With a view to examining the correlation of \( \tau_p \) with wind and rainfall, and also keeping in mind the facts that higher wind speeds occur mostly during the monsoon period when the rainfall also is intense, and that rainfall will lead to a wash out of aerosols from the MBL, a partial correlation analysis was carried out between (1) the monthly mean \( \tau_p \) values, (2) the monthly mean wind speed, \( U \), and (3) monthly total
Figure 5. Spectral variation of aerosol optical depth at Minicoy showing distinctive changes from month to month. Continuous lines are for 1995 and dashed lines for 1996.
TABLE 3. CORRELATION ANALYSIS BETWEEN $\tau_p$, $U$ AND $R_f$ FOR 21 MONTHLY MEANS

<table>
<thead>
<tr>
<th>$\lambda$</th>
<th>$\rho_{12}$</th>
<th>$\rho_{13}$</th>
<th>$\rho_{23}$</th>
<th>$\rho_{12.3}$</th>
<th>$\rho_{13.2}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>400</td>
<td>0.11</td>
<td>0.001</td>
<td>0.14</td>
<td>-0.01</td>
<td></td>
</tr>
<tr>
<td>450</td>
<td>0.39</td>
<td>0.006</td>
<td>0.51</td>
<td>-0.36</td>
<td></td>
</tr>
<tr>
<td>500</td>
<td>0.36</td>
<td>0.003</td>
<td>0.48</td>
<td>-0.33</td>
<td></td>
</tr>
<tr>
<td>600</td>
<td>0.28</td>
<td>0.003</td>
<td>0.66</td>
<td>0.37</td>
<td>-0.24</td>
</tr>
<tr>
<td>650</td>
<td>0.35</td>
<td>0.008</td>
<td>0.46</td>
<td>-0.32</td>
<td></td>
</tr>
<tr>
<td>750</td>
<td>0.50</td>
<td>0.013</td>
<td>0.65</td>
<td>-0.48</td>
<td></td>
</tr>
<tr>
<td>850</td>
<td>0.44</td>
<td>0.011</td>
<td>0.57</td>
<td>-0.41</td>
<td></td>
</tr>
<tr>
<td>935</td>
<td>0.61</td>
<td>0.014</td>
<td>0.79</td>
<td>-0.63</td>
<td></td>
</tr>
<tr>
<td>1025</td>
<td>0.44</td>
<td>0.012</td>
<td>0.57</td>
<td>-0.41</td>
<td></td>
</tr>
</tbody>
</table>

Note that in Table 3 the coefficient $\rho_{23}$ between wind speed and rainfall is wavelength independent. See text for explanation of symbols.

rainfall, $R_f$. The partial correlations are estimated following the equation (Fisher 1970)

$$\rho_{12.3} = \frac{\rho_{12} - \rho_{13} \cdot \rho_{23}}{\sqrt{(1 - \rho_{13}^2)(1 - \rho_{23}^2)}}$$  \hspace{1cm} (2)

where $\rho_{12.3}$ is the partial correlation between parameter 1 and 2 (mean $\tau_p$, and wind speed in this study) after removing the contribution due to parameter 3 (rainfall), and $\rho_{12}$, $\rho_{13}$ and $\rho_{23}$ are the direct correlation coefficients between 1 and 2, 1 and 3 and 2 and 3 respectively. The results are summarized in Table 3. The analysis revealed the following.

1. Direct correlations show only a weak positive association between wind speed and $\tau_p$ which is, however, below the 0.02 significance level; an insignificant association between $\tau_p$ and rainfall and a highly positive association ($\rho \approx 0.66$) between wind speed and rainfall, which is significant at the $P = 0.01$ significance level (Fisher 1970). (The significance level $P$ is the probability that such a correlation would arise by random sampling from an actually uncorrelated population, and thus the lower its value, the more significant is the correlation.)

2. The partial correlations between wind speed and $\tau_p$ are high, positive and significant at the 0.02 level at all wavelengths above 400 nm.

3. The partial correlations between $\tau_p$ and rainfall become negative once the effects of wind speed are removed. They are generally high (0.35 to -0.5) and significant at the 0.05 or above level at all wavelengths above 400 nm. As the wet-removal processes are known to be more efficient for larger aerosols (Flossmann et al. 1985) and the larger aerosols are present more in the MBL, it is possible that their number concentration will be negatively associated with the amount of rainfall. As larger aerosols contribute more to the aerosol optical depth at the longer wavelengths, $\rho_{13}$ and $\rho_{13.2}$ will exhibit wavelength dependence (as seen in Table 3).

The wavelength 380 nm was not considered in this study since it was introduced after about 10 months and did not have sufficient samples. The partial correlation analysis shows clearly that an increase in wind speed would result in an increase in $\tau_p$, and that the rainfall would cause a decrease in it.

5. PARAMETRIZATION OF WIND SPEED DEPENDENCE OF $\tau_p$

In order to parametrize the effect of wind speed changes on aerosol optical depth over Minicoy, the $\tau_p$ values are grouped into ensembles of different wind speeds.
TABLE 4. REGRESSION COEFFICIENTS

<table>
<thead>
<tr>
<th>Wavelength (nm)</th>
<th>380</th>
<th>400</th>
<th>450</th>
<th>500</th>
<th>600</th>
<th>650</th>
<th>750</th>
<th>850</th>
<th>935</th>
<th>1025</th>
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<tbody>
<tr>
<td>$\tau_0$</td>
<td>0.32</td>
<td>0.31</td>
<td>0.28</td>
<td>0.25</td>
<td>0.18</td>
<td>0.17</td>
<td>0.18</td>
<td>0.17</td>
<td>0.16</td>
<td>0.16</td>
</tr>
<tr>
<td>$b$ (s m$^{-1}$)</td>
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<td>0.06</td>
<td>0.06</td>
<td>0.07</td>
<td>0.10</td>
<td>0.08</td>
<td>0.11</td>
<td>0.10</td>
<td>0.11</td>
<td>0.12</td>
</tr>
<tr>
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<td>0.91</td>
<td>0.93</td>
<td>0.89</td>
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<td>0.87</td>
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<td>0.95</td>
<td>0.95</td>
</tr>
<tr>
<td>$\rho^*$</td>
<td>0.54</td>
<td>0.61</td>
<td>0.56</td>
<td>0.74</td>
<td>0.61</td>
<td>0.69</td>
<td>0.64</td>
<td>0.64</td>
<td>0.75</td>
<td>0.77</td>
</tr>
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</table>

See text for explanation of symbols.

More clearly, the observations of $\tau_p$ corresponding to wind speeds in the range 0–1 m s$^{-1}$ constitute one ensemble which is considered to represent a mean wind speed of 0.5 m s$^{-1}$ and those lying in the wind speed range 1–2 m s$^{-1}$ forms the next ensemble which represent a mean wind speed of 1.5 m s$^{-1}$ and so on. This way all the $\tau_p$ data are grouped at 1 m s$^{-1}$ intervals of wind speed. For this purpose only simultaneous data of wind speeds and $\tau_p$ are used. As, generally, individual $\tau_p$ is the mean for each observational day, individual wind speeds also are taken as the mean of the measurements made on the same day from 0530 IST to the end of the MWR observations. The $\tau_p$ data lying in each group are averaged to obtain the mean $\tau_p$ corresponding to the mean wind speed of each group. This way of grouping and averaging randomizes all those variations in $\tau_p$ which are not associated with changes in wind speed. These averaged $\tau_p$ values are plotted on a log–linear scale, with the logarithm of mean $\tau_p$ against the ensemble mean wind speed ($U^*$) for all the wavelengths. A fairly good linear dependence was discernible between $\ln(\tau_p)$ and $U^*$.

A linear regression analysis (Kleinbaum and Kupper 1978) gave values for the correlation coefficients ($\rho^*$) in the range from 0.54 to 0.77 at different wavelengths. However, this analysis did not distinguish the wind in terms of the air-mass types. It is seen from Fig. 3 that the months December to February are mostly under the influence of a continental air mass with winds coming from the first quadrant, i.e. from east of north. So, in order to separate the marine air mass, which is more likely to be influenced by the over sea wind speeds, the observations corresponding to wind directions east of north (which are also influenced by the continental aerosol flux) are removed from the database and the analysis is repeated. A substantial increase in the correlation coefficient is observed with a mean value of 0.91, which is significant at the $P = 0.01$ level (Fisher 1970). A plot of $\ln(\tau_p)$ against $U^*$ corresponding to this dataset is shown in Fig. 6 for all the wavelengths, where the points are the ensemble averaged $\tau_p$ values, and the vertical bars represent the standard errors. The figure reveals a strong dependence of $\tau_p$ on $U^*$, suggesting an exponential increase in $\tau_p$ with $U^*$ in accordance with a relation of the form,

$$\tau_{p\lambda} = \tau_{0\lambda} \exp(bU^*)$$  \hspace{1cm} (3)

where $\tau_{0\lambda}$ represents the wind independent quiescent background level of aerosol optical depth and $b$ (s m$^{-1}$) the index for wind speed dependence. The values of $\tau_0$ and $b$ are estimated for each of the ten wavelengths and are given in Table 4 along with the correlation coefficients, $\rho$, for the database, excluding the north-east wind conditions. In the last row of the same table the correlation coefficients $\rho^*$ are given when the entire data (both continental and marine air-mass types) are considered. Comparison of $\rho$ and $\rho^*$ leads to the inference that sea-spray aerosols produced by action of surface winds significantly influence the aerosol optical depth characteristics over Minicoy when the station is under the influence of a marine air mass.
During continental air-mass conditions the effect of wind appears to be less significant. It is to be recalled in this context that based on aerosol optical depth measurements over the equatorial Indian Ocean on board a scientific cruise, Moorthy et al. (1997) observed that influence of changes in wind speed on aerosol spectral depth is most important over remote ocean locations and when the winds do not have a significant continental history. Moreover, the value of the index obtained from Minicoy data is lower than that obtained for the tropical Indian Ocean region. Similarly, τ₀λ over Minicoy exhibits strong wavelength dependence, decreasing rapidly with an increase in wavelength similar to that observed over rural locations (Pruppacher and Klett 1978). This is unlike the behaviour over a remote oceanic environment, where τ₀λ values are significantly lower.
than those observed over Minicoy and also less wavelength dependent (Moorthy et al. 1997). It appears that the high values of the steady-state optical depths over Minicoy would be the prevailing values, resulting from the aerosols transported from different regions of the Indian sub-continent and the adjoining areas by the synoptic wind systems and mixed with the marine aerosols over the Arabian sea. The large increase in $\tau_{p\lambda}$ at shorter wavelengths is suggestive of the presence of a substantial amount of small particles, possibly of continental origin. Strong surface winds modify this environment by additional aerosol loading (due to sea spray), leading to an increase in $\tau_p$ in accordance with Eq. (3). This high value of $\tau_{p\lambda}$ might be responsible for the lower value of $b$ (compared with open ocean) at Minicoy. The index $b$ shows a gradual increase with $\lambda$, which means that the NIR optical depths are affected more by the surface winds. Since NIR optical depths are attributable to relatively larger aerosols, the above feature suggests that larger aerosols are affected more by the prevailing surface wind speeds. This aspect is examined in detail by retrieving the column size distributions representative of each month/season and examining their characteristics.

6. RETRIEVED SIZE DISTRIBUTIONS

Size distribution of atmospheric aerosols is very important in estimating their radiative effects (e.g. Charlson et al. 1987; Russell et al. 1994). The size distribution of aerosols at a given location is determined by the relative strengths of different production and removal processes (Jaenicke 1993; Raes 1995). The size distribution of aerosols over marine locations depends on a number of additional processes, including the transport of aerosols from the continents (e.g. Prospero 1979; Tyson et al. 1996), local aerosol production from bursting of bubbles associated with whitecaps produced by the action of winds on sea surface (Woodcock 1953; Blanchard and Woodcock 1980; Monahan et al. 1986; Fitzgerald 1991; O’Dowd and Smith 1993) and the processes like precipitating and non-precipitating cloud cycling (Hoppel et al. 1990). The changes in the spectral variation of $\tau_p$ with changes in the air-mass types as well as with wind speed observed over Minicoy (in the earlier sections) are indicative of changes in size distributions. This is because the spectral variation of aerosol optical depth depends on the nature of the size distribution through the scattering equation.

$$\tau_{p\lambda} = \int_{r_a}^{r_b} \pi r^2 Q_{\text{ext}}(m, \lambda, r)n_c(r) \, dr$$

(4)

where $Q_{\text{ext}}$ is the Mie extinction efficiency parameter which is a function of the particle radius $r$, wavelength $\lambda$ and the complex refractive index $m$, and $n_c(r)$ is the columnar size distribution (CSD) function giving the number density of aerosols in a vertical column of unit cross-section, within a small radius range $dr$ centred at $r$. $r_a$ and $r_b$ are, respectively, the lower and upper limits of the integration such that the particles with sizes within $r_a$ and $r_b$ contribute significantly to the integrand of Eq. (4) and their values depend on the extreme values of $\lambda$ used in estimating $\tau_{p\lambda}$.

Several numerical methods are available to retrieve $n_c(r)$ from $\tau_{p\lambda}$ estimated over a number of wavelengths, of which the constrained linear inversion technique (King et al. 1978; King 1982) is quite extensively used. We have employed this technique to retrieve the size distributions from $\tau_{p\lambda}$ estimates, and the details of application of this method are given elsewhere (Moorthy et al. 1991, 1997) and hence are not repeated. The values of $r_a$ and $r_b$ are selected by evaluating the kernels (integrand of Eq. (4)) for the extreme wavelengths used in the MWR (i.e. 380 and 1025 nm) for different types of aerosol size distribution models. We found that $r_a = 0.05$ and $r_b = 3.0 \mu m$ is an optimal choice. For
the estimation of $Q_{\text{ext}}$, refractive index values as a function of wavelength as given by Shettle and Fenn (1979) for maritime aerosols at RH = 70% are used.

(a) Annual variations of CSDs

Representative CSD has been obtained for each month by inverting the monthly mean $\tau_{p\lambda}$ values (obtained as described earlier). These are examined for general features and their temporal variation. The retrieved CSDs generally showed a secondary large-particle mode, most of the time preceded by a primary small-particle mode, which sometimes is seen explicitly and sometimes is only indicated by the slanting nature of CSD towards smaller values of $r$. The latter case arises mainly when the primary mode occurs at a value of $r$ which is close to, but less than, $r_n$ (e.g. Moorthy et al. 1997). All these CSDs are designated as bimodal distributions. A typical example is shown in Fig. 7, which is the CSD obtained for March 1996. In Fig. 7(b), $n_c(r)$ is plotted against $r$ on a log–log scale, while Fig. 7(a) shows the monthly mean $\tau_{p\lambda}$ values, estimated using the MWR, as points with the error bars, and the $\tau_{p\lambda}$ values, re-estimated using the retrieved size distribution in Eq. (4), by the continuous line. A total of 22 such CSDs are obtained for the years 1995 and 1996 together. Composite plots of all these CSDs are shown in Figs. 8(c) and (d). Figures 8(a) and 8(b) show the monthly mean $\tau_{p\lambda}$ values, again as composite plots, from which the CSDs are retrieved. In Fig. 8, the panels on the left are for 1995 and on the right are for 1996. From the figure, the bimodal feature is seen almost always. In fact, of the 22 CSDs, all except those obtained for July and September 1995 showed a bimodal nature. The CSDs are generally steeper during December to May, compared with the monsoon months (June to November) when they are almost flat except near the lower and upper radii limits. It is also apparent from Fig. 8 that the position of the mode, particularly the coarse-particle mode, is variable from month to month, more so in 1995 than in 1996. Nevertheless, the secondary mode never occurs below $r = 0.5$ $\mu$m. However, in order to examine the features more quantitatively and to understand their dependence on other processes, the CSDs are parametrized in terms of physical properties of aerosols using analytical functions.

(b) Parametrization of the CSDs and physical properties of aerosols

The retrieved size distributions are parametrized by least squares, fitting them to a bimodal log–normal distribution function of the form,

$$n_c(r) = \sum_{i=1}^{2} \frac{N_{0i}}{\sqrt{2\pi} \sigma_i r} \exp \left( -\frac{(\ln r - \ln r_{mi})^2}{2\sigma_i^2} \right)$$  \hspace{1cm} (5)

following the procedure described by Moorthy et al. (1998). In Eq. (5) $r_{mi}$ and $\sigma_i$ represent, respectively, the mode (geometric mean) radius and its standard deviation, with $i = 1$ representing the primary (small particle) mode and $i = 2$ the secondary (coarse particle) mode. $N_{0i}$ are scaling constants which depend on total aerosol concentration. Occasionally when the CSDs indicated the presence of only a single mode, the same function, Eq. (5), is used with $i = 1$, and for CSDs with a secondary large-particle mode preceded by an inverse power-law type of behaviour with no indication of a primary mode, Eq. (5) is replaced by,

$$n_c(r) = N_{01} r^{-v} + \frac{N_{02}}{\sqrt{2\pi} \sigma_2 r} \exp \left( -\frac{(\ln r - \ln r_{m2})^2}{2\sigma_2^2} \right)$$  \hspace{1cm} (6)
Figure 7. (a) The input $\tau_0$ values (see text) estimated from the multi-wavelength solar radiometer measurements as points with error bars. The continuous line represents $\tau_0$ values re-estimated using the retrieved CSD (shown in (b)) in Eq. (4). (b) Typical example of columnar size distribution (CSD) retrieved from monthly mean spectral optical depth for the month of March 1996 at Minicoy.

where $\nu$ is the power-law index. All the CSDs shown in Fig. 8 are parametrized by bimodal distribution functions (Eq. 5), except for July and September 1995. The distribution for July 1995 is parametrized using Eq. (6) as a combination of the power law and a unimodal distribution function, and that for September 1995 with a unimodal function with $i = 2$ in Eq. (5). By evolving a least-squares fit between the retrieved CSDs of Fig. 9 and the Eqs. (5) and (6) as the case is, $r_{m1}$, $r_{m2}$, $\sigma_1$ and $\sigma_2$ ($\nu$, $r_{m2}$, $\sigma_2$ when Eq. (6) is used) are evaluated for each of the 22 CSDs. From the CSDs other physically important parameters describing the aerosols, such as the effective radius ($R_{\text{eff}}$; a measure of the total volume to area of aerosols and the columnar mass loading, $m_L$), are estimated using the following equations.

$$R_{\text{eff}} = \frac{\int_{r_a}^{r_b} \pi r^2 n_c(r) \, dr}{\int_{r_a}^{r_b} r^2 n_c(r) \, dr}$$  \hspace{1cm} (7)$$

$$m_L = \frac{4}{3} \pi \rho_a \int_{r_a}^{r_b} r^3 n_c(r) \, dr$$ \hspace{1cm} (8)$$

where $\rho_a$ is the density of aerosols, taken as 2.2 g cm$^{-3}$.  

Figure 8. Composite plots of (a) and (b) monthly mean $\tau_{\text{ol}}$ (see text) estimated using the multi-wavelength solar radiometer and (c) and (d) the retrieved columnar size distributions. (a) and (c) are for 1995 and (b) and (d) are for 1996.
The month-to-month variations of the mode radii \( r_{m1} \) and the standard deviations \( \sigma_m \) of the mode, estimated from the corresponding CSDs, are shown in Fig. 9. No systematic monthly variations are seen except for random fluctuations about the mean. The ensemble mean value and ensemble standard deviations of these parameters, considering all the CSDs as a single ensemble, are given in Table 5. It can be seen that the primary mode is broader and is generally consistent, whereas the secondary mode is sharper and its position is variable, as shown by the larger standard deviation for \( r_{m2} \). These features can also be seen in Fig. 8.
TABLE 5. PARAMETERS OF AEROSOL SIZE DISTRIBUTION

<table>
<thead>
<tr>
<th>Parameter</th>
<th>( r_{m1} (\mu m) )</th>
<th>( \sigma_1 )</th>
<th>( r_{m2} (\mu m) )</th>
<th>( \sigma_1 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>0.133</td>
<td>0.335</td>
<td>0.857</td>
<td>0.218</td>
</tr>
<tr>
<td>Standard deviation</td>
<td>0.075</td>
<td>0.079</td>
<td>0.219</td>
<td>0.034</td>
</tr>
</tbody>
</table>

See text for explanation of symbols.

The above observations suggest that the two modes, observed with the CSDs, are attributable mainly to two different production mechanisms. The primary small-particle (accumulation) mode arises mainly from secondary production processes such as gas-to-particle conversion reactions of continental gaseous effluents and their subsequent growth by condensation, coagulation and cloud cycling. As most of these aerosols might have originated over the continent (as Minicoy is totally unindustrialized) they would be sufficiently aged by the time they reach Minicoy which would take about 48 hours (for the average wind speeds encountered during winter season) when the air mass is predominantly of continental nature. As a result, the accumulation mode is aged, stable, and quite broad. On the other hand, the coarse-particle mode appears to be more associated with local production of sea-spray aerosols over the ocean. Consequently their characteristics, particularly the mode radius and abundance, will vary depending on the wind characteristics. This leads to the mode being sharp (as wind is the major source), but its position being largely variable (as shown by the CSDs of Fig. 8) due to the variable nature of wind speed in the monsoon season. This is also borne out from Fig. 9, which shows that irrespective of the variability in the positions of the mode, \( r_{m1} \) is always less than 0.3 \( \mu m \), and \( r_{m2} \) is mostly greater than 0.6 \( \mu m \), again suggesting these to be associated mainly with different causative processes. Besides the above, submicron non sea-salt aerosols also would contribute to the accumulation mode. These non sea-salt aerosols are produced mainly by the photolysis of low volatile gaseous emissions (such as dimethyl sulphide) from the ocean, and their sizes generally are less than 0.2 \( \mu m \) (e.g. Clark et al. 1987; Fitzgerald 1991).

7. DISCUSSIONS

The studies of aerosol optical depths described in section 5 have shown clear monthly and seasonal patterns, which suggest seasonal changes in the physical characteristics of aerosols. But since the basic characteristics of the aerosol size distributions (mode radii and standard deviations, or in other words the shape) do not show any clear patterns, the other derived parameters characterizing the aerosol environment, like \( m_L \), \( R_{eff} \), and columnar number concentration of aerosols, \( N_t \), are examined. The monthly variation of \( m_L \) and \( R_{eff} \) are shown in Fig. 10. Both these parameters show low values in January and high values during June to August. From January to February \( m_L \) increases, though \( R_{eff} \) does not. A similar feature is noticed in October. March to May show varying patterns. It may be noted (also evident from Eqs. (7) and (8)) that \( m_L \) depends both on \( N_t \) as well as on the concentration of larger aerosols, while \( R_{eff} \) is dependent more on the relative dominance of larger to smaller particles. Thus the observations in Fig. 10 are indicative of seasonal changes in total aerosol concentration coupled with seasonal changes in the relative dominance of different sized aerosol particles. This aspect is examined by dividing \( N_t \) into two regimes, the accumulation regime (small aerosols) and the coarse-particle regime (large aerosols) keeping a radius \( r' \approx r_{m2} - \sigma_2 \approx 0.5 \mu m \) as a cut-off that divides the CSD into the two distinct regimes. This cut-off point is also
selected based on the examination of CSDs (Fig. 8), where a clear inflexion is seen in the CSDs at $r \approx 0.5 \, \mu m$ and the accumulative mode never goes above $0.3 \, \mu m$. The total concentration of accumulation mode particles ($N_a$) and coarse particles ($N_c$) are thus estimated as the area under the CSD curve below and above $r = 0.5 \, \mu m$ respectively, so that,

$$N_a = \int_{r_a}^{0.5} n_c(r) \, dr,$$

and

$$N_c = \int_{0.5}^{r_a} n_c(r) \, dr.$$  

From this a dimensionless ratio $N_c/N_a$ is evaluated, which represents a measure of the relative abundance of large (coarse) particles to the total aerosol concentration $N_t$. (This is so because $N_c$ is generally about 2 to 4 orders smaller than $N_a$, so that $N_a$ is approximately equal to $N_t$.) The month-to-month variations of this ratio and $N_t$ are shown in Fig. 11. Though $N_t$ is relatively featureless, the ratio $N_c/N_a$ shows monthly patterns similar to that seen in Fig. 10 for $m_l$ and $R_{eff}$. In fact, during June–September, the ratio $N_c/N_a$ increases by more than two orders of magnitude from its value during December–January, indicating increased dominance of large particles in the size spectrum during the monsoon season. Since $N_a$ contains about 99% of the total content, the variations in $N_c$ may not reflect in $N_t$, whereas a change in the number of
larger aerosols affect the mass loading. Considering the errors in the retrieved CSDs and also the fact that these are more significant to \( N_c \), only changes in \( N_c / N_a \) exceeding a factor of 5 are considered significant. Similarly changes by a factor of 1.5 or more in \( m_L \) are significant. For a given number of aerosols, a change in the ratio of larger to smaller aerosols affects the \( R_{eff} \). This explains the features in Figs. 10 and 11.

(a) Seasonal variations

The seasonal mean aerosol optical depths are inverted to retrieve the representative size distribution function for each season, and from these the different aerosol parameters are estimated and are given in Table 6. The monsoon season is divided into two in view of the change in the direction of the prevailing wind, from those blowing from over the ocean (south-west) to those blowing from over the land (north-east). Also, the south-west monsoon season exhibits high wind speeds and extensive rainfall. During the north-east monsoon season, though the rainfall is extensive, wind speeds tend to be much lower (as can be seen from Fig. 3).

Table 6 very clearly brings out the dominant role played by the wind over the ocean in changing the aerosol size spectrum. There is a significant enhancement in \( m_L \) and \( R_{eff} \) during the south-west monsoon seasons when the winds are strongest and are directed from the open ocean, irrespective of the fact that this season also experiences extensive
rainfall (Fig. 3). It can also be seen from the last column of Table 6 that associated with this there is a substantial increase in the relative abundance of coarse particles \((N_c)\). As the coarse aerosols are more associated with sea-spray production (Exton et al. 1985; O’Dowd and Smith 1993), which is a strong function of wind speed, the studies conclude that over Minicoy the monsoon winds lead to significant enhancement in aerosol optical depth and mass loading. \(R_{\text{eff}}\) shows only an insignificant increase along with \(m_L\). In fact the variations in \(m_L\) and \(R_{\text{eff}}\) are not very well associated as can be seen from Fig. 12(a), where \(m_L\) is plotted against \(R_{\text{eff}}\). This, when viewed along with the fact that the coarse particles contribute to most of \(m_L\) whereas \(R_{\text{eff}}\) depends on both \(N_c\) and \(N_a\), shows that characteristics are brought about by the changes in relative abundance of coarse aerosols. This is seen from Fig. 12(b), where \(N_c/N_a\) is plotted against \(R_{\text{eff}}\). The accumulation mode becomes finer (i.e. the radius becomes smaller) during the north-east monsoon and winter seasons when the air mass is predominantly continental (i.e. from the north-east), showing the advection of fine submicron aerosols (possibly of anthropogenic origin) from the mainland. During the periods of marine air-mass influence (spring and south-west monsoon seasons, when the wind is mainly from west of north) the accumulation mode radius \(r_{m1}\) becomes larger, showing the influence of locally produced aerosols also (due to film droplets).

### (b) Effect of wind speed on aerosol parameters

The effect of wind speed on aerosol characteristics has been extensively investigated in the past (e.g. Woodcock 1953; Blanchard 1963; Monahan 1968; Prodi et al. 1983; O’Dowd and Smith 1993; Mooorthy et al. 1997). We have already seen (section 5) that aerosol optical depths at Minicoy are closely associated with wind speed when a marine air mass prevails. Spectral behaviour of this dependence shows that NIR optical depths are influenced more by surface winds. In view of this, different aerosol physical parameters derived from the aerosol size distribution are examined to study their association with wind speed. As we have seen in section 5, the effect of wind speed on aerosol optical depth is influenced by prevailing air-mass types. In view of this, data pertaining to the months characterized by north-east winds (January, February and December) are not used in this analysis. The results are presented in Fig. 13 where \(m_L\) and \(R_{\text{eff}}\) (estimated from the CSDs retrieved from monthly mean \(\tau_{\text{pA}}\) values) are plotted on a logarithmic scale against the monthly mean wind \((\bar{U})\) on a linear scale. In the figure the points represent the values estimated from the MWR measurements and

<table>
<thead>
<tr>
<th>Season</th>
<th>(r_{m1}) ((\mu m))</th>
<th>(\sigma_1)</th>
<th>(r_{m2}) ((\mu m))</th>
<th>(\sigma_2)</th>
<th>(m_L) (mg m(^{-2}))</th>
<th>(R_{\text{eff}}) ((\mu m))</th>
<th>(N_c/N_a) ((x 10^{-4}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>W94</td>
<td>0.11</td>
<td>0.33</td>
<td>1.19</td>
<td>0.18</td>
<td>173.39</td>
<td>0.43</td>
<td>37.24</td>
</tr>
<tr>
<td>S95</td>
<td>0.13</td>
<td>0.28</td>
<td>1.27</td>
<td>0.19</td>
<td>312.23</td>
<td>0.76</td>
<td>10.56</td>
</tr>
<tr>
<td>M195</td>
<td>0.27</td>
<td>0.25</td>
<td>1.15</td>
<td>0.29</td>
<td>387.23</td>
<td>1.02</td>
<td>2233.04</td>
</tr>
<tr>
<td>M295</td>
<td>0.24</td>
<td>0.27</td>
<td>0.88</td>
<td>0.20</td>
<td>147.06</td>
<td>0.48</td>
<td>622.32</td>
</tr>
<tr>
<td>W95</td>
<td>0.06</td>
<td>0.39</td>
<td>0.65</td>
<td>0.20</td>
<td>227.93</td>
<td>0.23</td>
<td>300.02</td>
</tr>
<tr>
<td>S96</td>
<td>0.07</td>
<td>0.34</td>
<td>0.95</td>
<td>0.20</td>
<td>268.18</td>
<td>0.28</td>
<td>24.43</td>
</tr>
<tr>
<td>M196</td>
<td>0.32</td>
<td>0.30</td>
<td>0.92</td>
<td>0.34</td>
<td>352.51</td>
<td>1.00</td>
<td>2000.68</td>
</tr>
<tr>
<td>M296</td>
<td>0.10</td>
<td>0.28</td>
<td>0.98</td>
<td>0.17</td>
<td>226.21</td>
<td>0.33</td>
<td>47.59</td>
</tr>
</tbody>
</table>

See text for explanation of column headings. W, S, M1 and M2 stand for winter (December to February), spring (March to May), south-west monsoon (June to September) and north-east monsoon (October to November) respectively, together with the last two figures of the year. Note: W94 only contains data for January and February 1995.
the lines are linear least squares fitted. Notwithstanding the scatter of points a fairly good linear relation significant at the $P = 0.1$ level (Fisher 1970) is seen with wind speed as indicated by the regression line. The dependence can be represented analytically as

$$m_L = 116 \exp(0.14\bar{U})$$

(11)

$$R_{\text{eff}} = 0.16 \exp(0.20\bar{U})$$

(12)

where 116 mg m$^{-2}$ and 0.16 $\mu$m represent the values of $m_L$ and $R_{\text{eff}}$ respectively at $\bar{U} = 0$. The scatter of the points could be the consequences of other influencing parameters and the averaging of wind speed and $\tau_p$ over the months. Nevertheless the dependence is definite. The increases in $m_L$ and $R_{\text{eff}}$ with wind speed suggest a relative increase of larger aerosols with wind speed because they contribute more to the mass loading. This is evident from Fig. 14, in which the ratio $N_C/N_a$ is plotted against mean wind speed, where a significant increase in $N_C/N_a$ (at the $P = 0.2$ level, $\rho = 0.66$) is seen with wind speed. The addition of data points corresponding to the north-east wind months does not affect the correlation coefficient ($\rho = 0.65$) in this case. This confirms that larger ($r > 0.5 \mu$m) aerosols are produced more locally compared
with small aerosols by the surface winds over oceanic areas. The fact that $\rho$ remains the same irrespective of wind direction shows that the increased generation of (large) aerosols with increase in wind speeds occurs locally over the ocean surface and is not due to increased advection from continental areas. As sea-spray production of aerosols depends on the wind stress on the water surface, which is related to the wind speed, it is not significant whether the winds are from the north-east or north-west, as long as they have a significant sea travel. The size distributions of aerosols retrieved from the aerosol optical-depth measurements at Minicoy have shown consistent bimodal shape in most of the months, irrespective of the season. Also, the mode radii do not show any consistent seasonal variations. Examination of the dependence of mode radii with wind speed have shown the correlation coefficients are insignificant with values of $\rho < 0.20$, as is apparent from the scatter of points shown in Fig. 15. The removal of the data points corresponding to north-east winds does not improve the situation. A similar study carried out for the standard deviations ($\sigma$) revealed no clear dependence of $\sigma_1$ with $U(\rho = -0.18)$ whereas $\sigma_2$ shows a weak increase ($\rho = 0.35$) as is seen from Fig. 15. Increase in $\sigma_2$ with $\bar{U}$ indicates an increase in the spectral width of coarse aerosols with increase in wind speed. The significance of this is discussed subsequently.
Studies on the production of sea-salt aerosols reveal two distinct mechanisms, both associated with the bursting of white-cap bubbles, which produce two types of aerosols with distinct characteristics (Blanchard and Woodcock 1957, 1980; Fitzgerald 1991). These are jet droplets and film droplets. The jet droplets are larger in size compared with film droplets and are about one tenth of the parent bubble size (Keintzler et al. 1954; Blanchard 1989). The investigations on the formation of bubbles and their characteristics at the sea surface show that bubble-size distribution peaks at around
100 μm range which suggests a peak in the jet droplet spectrum at ~10 μm (Blanchard and Woodcock 1957; Wu 1992). After production the droplets are transported to higher altitudes and are reduced in size in order to keep them in equilibrium with the ambient relative humidity. At the sea surface RH is close to 100% and decreases sharply with height. The initial droplet size is reduced to one quarter when it reaches equilibrium with the ambient RH, which means the peak in the aerosol size produced from jet droplets in the MBL occurs at a diameter of ~2.5 μm (Blanchard and Woodcock 1980). The studies on the film droplet spectrum reveal peaks in the submicron size (r < 0.1 μm) range (Resch et al. 1986; Resch and Afeli 1992). Thus it is clear that film droplets contribute to the primary (small particle) mode and jet droplets to secondary (large particle) mode. Also studies on the relative contribution of jet and film droplets have shown that aerosols with radii r > 0.2 μm are contributed to by jet droplets, and those with r < 0.2 μm are contributed to by film droplets (Woolf et al. 1987). The oceanic surface films, present over the sea surface, are also ejected with the film droplets since these are produced from the bubble films (whereas the jet droplets are produced from an unstable jet of water formed as a result of the bubble cavity). As such, jet droplets will have almost similar characteristics to the seawater, whereas film droplets have distinct features. Viewed in the above perspective, the poor association of the mode radii with wind speed observed in Fig. 16 is logical. Since the two modes are associated with different production mechanisms an increase in wind speed causes enhancement in all the sizes (though the magnitude of production varies), the mode radii are not affected significantly. As wind speed increases, in addition to aerosol production by bubble bursting, aerosols are produced directly from the sea spray, the size range of which is high compared with jet droplets (Monahan 1968; Decker and DeLeeuw 1993). Moreover, increases in wind speed result in increased turbulent exchange in the MBL (Smirnov et al. 1995; Moorthy et al. 1997) and thus leads to increased uptake of jet and spray aerosols deeper into the MBL where they get mixed with the prevailing aerosols and undergo size changes in accordance with the RH profile. The deeper mixing carries coarse aerosols to higher altitudes of the MBL, which eventually increases their lifetime. The combined effects of all these would result in a broader size spectrum of coarse aerosols under increased wind speed conditions and this explains the observed increase in σ2 with U in Fig. 15.

8. CONCLUSIONS

Our aerosol characterization studies from Minicoy have revealed the following.

1. The aerosol characteristics over Minicoy resemble in many aspects those of a remote marine environment. However, there is a strong influence of the adjoining continental land mass (Indian mainland) particularly in seasons when continental air mass prevails.


3. Spectral variations are similar from year to year. During the period June to September when the winds are very strong and have a long marine history, the optical depths are nearly spectral independent, typical of a marine environment. During winter, when north-easterly winds prevail, there is a significant influence of continental submicron aerosols advected from the mainland, causing the spectral variations of τp to become steeper.

4. Optical depths are significantly influenced by changes in surface wind speed, with τp increasing exponentially with wind speed. The effect is much more remarkable when the winds are from the north-west with considerable marine history than when the entire
database is considered. However, the index of wind speed dependence of $\tau_p$ is lower than those values reported for open-ocean environments.

5. The wind-independent component of the optical depths ($\tau_{0,0}$) is significantly higher than those observed over the remote tropical Indian Ocean, probably due to the continental impact.

6. Aerosol columnar size distributions retrieved from spectral optical depths depicted, in general, bimodal characteristics. The basic shape of the aerosol size distribution does not change with season. The CSDs are found to be flatter during the monsoon season compared with the winter seasons. These modes appear to be associated with different generation mechanisms, with the coarse-particle mode being produced locally over the sea surface while the accumulation mode has a continental influence.

7. The columnar mass loading and effective radius show maximum values during June–July and minimum during November to February. These changes are attributed to similar changes occurring in the relative concentration of coarse particles in the MBL, which are in turn strongly dependent on wind speed characteristics. The aerosol parameters, derived from the CSDs, are found to be dependent on wind characteristics. An increase in wind speed results in an increase in the relative abundance of coarse particles, which in turn results in increases in columnar mass loading and effective radius. However, the mode radii are not significantly affected.

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