Chapter 2

Data and Methodology

2.1. Introduction

A study on the time evolution of synoptic scale distribution of aerosols essentially requires detailed information on aerosols over a large geographical region on a routine basis. Space borne remote sensing is the most effective means for this purpose. The Moderate Resolution Imaging Spectroradiometer (MODIS) onboard the Earth Observing Satellite (EOS) Terra and Aqua is a unique sensor with capabilities to attain nearly a full global coverage every day both over the land and the ocean. The specifications of MODIS for the aerosol measurement are unmatched by any other orbiting sensor at present.

The next most important parameter required for studying the aerosol transport is the atmospheric wind field. Reanalyzed wind field generated using the global circulation model by the National Centres for Environmental Prediction (NCEP) is highly reliable and well suited for this purpose. Wind over the ocean surface, measured by QuikSCAT (Quick Scatterometer) is sufficiently accurate to estimate the in-situ production of sea-salt aerosols. An appropriate combination of these data is effective in studying the role of dynamics that governs the aerosol distribution. Incorporating these into the aerosol flux continuity equation, proves to be an ideal technique to identify the major sources of aerosols and assess their strengths. This chapter provides a brief outline of the data sets used for the
present study and the methodologies adopted for addressing the aerosol transport dynamics and for the estimation of the strength of aerosol sources.

2.2 Aerosol properties from MODIS

2.2.1 Sensor details

MODIS sensor is the first of its kind which can characterize the spatial variation of aerosol properties over the land and ocean globally on a daily basis. The features of this sensor onboard the polar sun-synchronous satellites Terra (descending node, 10:30 am, equatorial crossing local time) launched in December, 1999 and Aqua (ascending node, 1:30 pm, equatorial crossing local time) launched in May, 2002, are far superior to those of any other space-borne aerosol sensor till date. From an altitude of 705 km, at an inclination of 98.2° and field of view ±55°, the sensor’s swath is about 2330 km. This makes it possible to cover almost the entire globe once in a day. Exact revisit over a particular region occurs once in every 16 days. Onboard calibrators are used to perform regular radiometric, spatial and spectral calibrations. Figure 2.1 shows the visible image (composite of red, green and blue wavelengths) of MODIS from Terra satellite on a typical day, April 19, 2000. This shows the capability of MODIS in acquiring almost full global coverage in a single day except for a few gaps between 30°N and 30°S.

Fig.2.1. True colour RGB image from MODIS sensor onboard Terra on April 19, 2000 showing its near daily global coverage capability.
Chapter 2

MODIS operates at 36 spectral channels ranging from 0.42 to 14.24 μm, out of which seven (0.47, 0.55, 0.66, 0.86, 1.2, 1.6, 2.1 μm) are dedicated for aerosol retrieval. The primary resolutions of these seven bands range between 250 - 500m. The spectral stability of MODIS sensor is better than 0.002 μm. Further details of the spectral channels are summarized in the Table 2.1.

**Table 2.1. Characteristics of MODIS channels used in aerosol retrieval.** \( \rho \) is the spectral reflectance and \( \Delta \rho \) is the 'Noise equivalent differential spectral reflectance' and SNR is the 'Signal-to-Noise Ratio' of the sensor.

<table>
<thead>
<tr>
<th>Band /Channel</th>
<th>Wavelength band (μm)</th>
<th>Spatial Resolution (m)</th>
<th>( \Delta \rho ) (x ( 10^{-4} ))</th>
<th>SNR</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.620 – 0.670</td>
<td>250</td>
<td>3.39</td>
<td>128</td>
</tr>
<tr>
<td>2</td>
<td>0.841 – 0.876</td>
<td>250</td>
<td>3.99</td>
<td>201</td>
</tr>
<tr>
<td>3</td>
<td>0.459 – 0.479</td>
<td>500</td>
<td>2.35</td>
<td>243</td>
</tr>
<tr>
<td>4</td>
<td>0.545 – 0.565</td>
<td>500</td>
<td>2.11</td>
<td>228</td>
</tr>
<tr>
<td>5</td>
<td>1.230 – 1.250</td>
<td>500</td>
<td>3.12</td>
<td>74</td>
</tr>
<tr>
<td>6</td>
<td>1.628 – 1.652</td>
<td>500</td>
<td>3.63</td>
<td>275</td>
</tr>
<tr>
<td>7</td>
<td>2.105 – 2.155</td>
<td>500</td>
<td>3.06</td>
<td>110</td>
</tr>
</tbody>
</table>

In agreement with common references in aerosol literature the central wavelengths of the bands (or channels) 1 – 7 are identified respectively as 0.66, 0.86, 0.47, 0.55, 1.2, 1.6 and 2.1 μm. To retrieve aerosol parameters from MODIS data, independent algorithms are used over the ocean and over the land. The principal aerosol parameter retrieved by MODIS is the spectral AOD at six bands corresponding to 0.55, 0.66, 0.86, 1.2, 1.6, 2.1 μm over the ocean and at two bands corresponding to 0.47 and 0.66 μm over the land. The wide spectral coverage of the sensor provides an opportunity to determine different aerosol parameters such as the spectral optical depth, the fine mode fraction (FMF), the angstrom exponent, the effective radius and the column integrated mass concentration. The parameter FMF is specified for 0.55 μm which is the fraction of AOD contributed by the fine (small) mode particles to the total AOD, where fine mode and coarse mode comprise of particles with effective radii in the ranges, 0.1 – 0.25 μm and 1 – 2.5 μm respectively. Over the land, Angstrom exponent is derived from the AODs at 0.47 and 0.66 μm wavelengths while over
the ocean, two Angstrom exponents are derived; one using the wave lengths 0.55 and 0.86 μm and the other using 0.86 and 2.1 μm.

2.2.2 Inversion algorithms

Details of the aerosol retrieval algorithm over the land and the ocean are documented by Kaufman et al. [1997b] and Tanre et al. [1997]. The basic formulation of the retrieval method has not changed since its initial inception except that it has been upgraded [Levy et al., 2003; Remer et al., 2005; 2008] a few times by incorporating appropriate changes in the aerosol models, surface reflectance characteristics and radiative transfer computations. These changes could refine the algorithm to a great extent. The MODIS data are classified into ‘collections’, where the collection number corresponds to data produced by a definite version of the algorithm. The evaluations of Collection 002 (C002) and Collection 003 (C003) products [Chu et al., 2002; Remer et al., 2002] led to the generation of Collection 004 (C004) with a few improvements in this version of the algorithm. A comprehensive evaluation of regional and global C004 aerosol products [Remer et al., 2005] resulted in a complete overhaul of the retrieval algorithm over the land and a few upgradings for the retrieval over the ocean incorporated into the Version 5.2 that forms the C005 products.

Each MODIS pass is split into individual scenes called ‘granules’, containing measurements for an interval of 5 minutes. Before processing, the calibrated and geolocated radiances in each granule (termed the Level 1B products) are first organized into nominal 10 km × 10 km boxes, each corresponding to 20 by 20 pixels in the case of 500 m resolution channels. Reflectances in channels with 250 m resolution are degraded to 500 m resolution and are also organized into 10 km × 10 km boxes. Each of the 20 by 20 (= 400) pixels in the box is first checked for cloudiness, snow/ice or water, using MODIS cloud mask (MOD/MYD35) which contains all the masking information. Ocean algorithm is implemented only if all the pixels in the box are identified as water otherwise the land algorithm is followed. Note that the quality of aerosol retrieval along the coast will be rather poor as the domain box considered in this case will contain pixels over water as well as over the land.
Chapter 2

2.2.1 Algorithm over the land

The reflectance measured at the top of the atmosphere (TOA) is a function of all the successive orders of interaction within the coupled surface-atmosphere system. According to Kaufman et al. [1997b], this can be approximated as

\[ \rho_{\lambda}^*(\theta_s, \theta_v, \phi_d) = \rho_{\lambda}^a(\theta_s, \theta_v, \phi_d) + \frac{F_{\lambda}(\theta_s, \theta_v) T_{\lambda}(\theta_s, \theta_v) \rho_{\lambda}^s(\theta_s, \theta_v, \phi_d)}{1 - s_{\lambda} \rho_{\lambda}^s(\theta_s, \theta_v, \phi_d)} \] (2.1)

where \( \rho_{\lambda}^a \) is the 'atmospheric path reflectance', \( F_{\lambda} \) is the 'normalized downward flux for zero surface reflectance', \( T_{\lambda} \) is the 'upward total transmission' into the satellite, \( s_{\lambda} \) is the 'atmospheric backscattering ratio' and \( \rho_{\lambda}^s \) is the 'angular surface reflectance', all corresponding to the wavelength, \( \lambda \). These are functions of solar zenith angle, satellite zenith angle and relative azimuth angle (azimuth difference between solar incident and satellite viewing directions) represented here as \( \theta_s, \theta_v \) and \( \phi_d \) respectively. All the terms in Eq.(2.1), except surface reflectance, are functions of aerosol type and its concentration. A 'look-up-table' (LUT) is generated through simulation of the satellite detected radiances for a range of aerosol and surface properties. The TOA reflectance is computed for seven aerosol loadings (\( \tau = 0, 0.25, 0.5, 1.0, 2.0, 3.0, 5.0 \)), nine solar zenith angles (between 0.0° and 66.0°), 16 sensor zenith angles (between 0.0° and 66.0°) and 16 relative azimuth angles (between 0.0° and 180.0°). For aerosols, the algorithm assumes a fine (spherical) particle dominated model and a coarse (spheroid) particle dominated model, each comprising of multiple log normal size distribution combined with proper weightage, to represent the ambient aerosols over the target. The algorithm works by comparing the satellite observed radiances with the computed radiances in the LUT to retrieve the aerosol conditions which mimics best, the observed radiation field.

2.2.2.1 Recent modifications in the land algorithm

Evaluation of C004 algorithm with measurements from 132 Aerosol Robotic Network (AERONET) stations located at different parts of the globe showed very good agreement in most of the cases. However, the retrieved AOD showed a systematic bias which varied from place to place over the land [Remer et al., 2005]. At many places the bias was positive when the AOD values were relatively low and negative when the AOD
values were high [Ichoku et al., 2003; 2005; Levy et al., 2005]. Moreover, the FMF retrieved over the land showed significant deviations in many cases. These problems indicated the need of incorporating region and season specific refinements in surface reflectance and aerosol optical properties to improve the aerosol retrieval. For certain viewing and solar illumination geometries, neglecting the effect of polarization in the radiative transfer calculation was also found to be a source of error in the estimated TOA reflectance. This can lead to discrepancies greater than 10% in the retrieved values of AOD [Levy et al., 2004].

In the MODIS C005 algorithm, a basic change is incorporated in the parameterization of land surface reflectance. Besides, a vector radiative transfer code, RT3 [Evans and Stephens, 1991] is implemented in the generation of the LUT instead of the scalar code used in C004 and the aerosol scattering phase functions are calculated using MIEV Mie code [Wiscombe et al., 1980] or the T-matrix kernel code [Dubovik et al., 2002b; 2006] depending on the spherical and spheroidal model for the aerosols. Further, the Rayleigh optical depth which was assumed same everywhere as at the sea level, is corrected for the elevation of the surface [Fraser et al., 1989]. In addition to these modifications in the algorithm, the sensor calibration coefficients are also updated in C005.

2.2.2.1.2 Parameterization of the land surface reflectance

Up to the C004 algorithm, the retrieval philosophy of MODIS over the land, had assumed aerosols to be transparent at 2.1 μm channel over vegetated surfaces which appear darker in the mid-IR wavelengths. Based on observations over vegetated and dark soil surfaces, Kaufman and colleagues [Kaufman et al., 1997b; Kaufman et al., 1997c] had arrived at the following relations between the surface reflectances in different wavelengths:

\[
\rho_{647}^s = 0.25 \rho_{2.1}^s \quad \text{and} \quad \rho_{0.66}^s = 0.5 \rho_{2.1}^s
\]

(2.2)

where \( \rho_{\lambda}^s \) denotes the surface reflectance at the wavelength \( \lambda \).

Later, after extensive validations of MODIS derived AOD [Chu et al., 2002; Remer et al., 2002; Levy et al., 2005] and surface reflectance measurements in the visible (VIS) and short wave IR (SWIR) [Gatebe et al., 2001], it was observed that the ratio VIS/SWIR vary according to the surface type, scattering geometry and vegetation of the surface.
expressed in terms of NDVI_s (Normalised Difference Vegetation Index) [Gatebe et al., 2001; Remer et al., 2001; Levy et al., 2007]. NDVI_s is defined as

$$\text{NDVI}_s = \frac{\rho^{m}_{1.2} - \rho^{m}_{2.1}}{\rho^{m}_{1.2} + \rho^{m}_{2.1}}$$  \hspace{1cm} (2.3)$$

where $\rho^{m}_{1.2}$ and $\rho^{m}_{2.1}$ are MODIS measured reflectances in 1.2 and 2.1 $\mu$m channels respectively which are not very much influenced by the aerosols.

Values of NDVI_s > 0.6 correspond to dense vegetation while NDVI_s < 0.2 indicates sparse vegetation. An increase in NDVI_s indicates increase in the density of vegetation near the surface. Thus the values of $\rho^{s}_{0.66}$ and $\rho^{s}_{0.47}$ used in C005 algorithm, are modified as functions of $\rho^{s}_{2.1}$, NDVI_s and scattering angle $\gamma$, defined as

$$\rho^{s}_{0.66} = f_1(\rho^{s}_{2.1}, \text{NDVI}_s, \gamma) = \rho^{s}_{2.1} \times \text{slope}_{0.66/2.1} + \text{int}_{0.66/2.1}$$

$$\rho^{s}_{0.47} = f_2(\rho^{s}_{0.66}) = \rho^{s}_{0.66} \times \text{slope}_{0.47/0.66} + \text{int}_{0.47/0.66}$$  \hspace{1cm} (2.4)$$

where

$$\text{slope}_{0.66/2.1} = \text{slope}_{0.66/2.1} + 0.002\gamma - 0.27,$$

$$\text{int}_{0.66/2.1} = 0.00025\gamma + 0.033,$$  \hspace{1cm} (2.5)$$

$$\text{slope}_{0.47/0.66} = 0.49,$$

$$\text{int}_{0.47/0.66} = 0.005$$

where in turn

$$\text{slope}_{0.66/2.1} = 0.48; \text{ for NDVI}_s < 0.25,$$

$$\text{slope}_{0.66/2.1} = 0.58; \text{ for NDVI}_s > 0.75,$$  \hspace{1cm} (2.6)$$

$$\text{slope}_{0.66/2.1} = 0.48 + 0.2(\text{NDVI}_s - 0.25),$$

for $0.25 \leq \text{NDVI}_s \leq 0.75$

### 2.2.2.1.3 Inversion procedure

The C005 land algorithm combines fine and coarse particle dominated models to match with the sensor detected spectral reflectance. By simultaneously inverting aerosol
and surface information in the three channels 0.47, 0.66 and 2.12 μm, three output parameters; τ and FMF (η) at 0.55 μm and the surface reflectance \( \rho_{2.12}^s \) are derived. It has been shown by Remer et al. [2005] that η can be used to express the TOA spectral reflectance \( \rho_λ \) as a weighted sum of the spectral reflectances from a combination of fine (\( \rho_λ^{*f} \)) and coarse (\( \rho_λ^{*c} \)) dominated aerosol models in the form

\[
\rho_λ = \eta \rho_λ^{*f} + (1 - \eta) \rho_λ^{*c}
\]

(2.7)

where \( \rho_λ^{*f} \) and \( \rho_λ^{*c} \) are composed of surface reflectance and atmospheric path reflectance [as described by Eq. (2.1)] of separate aerosol models. These parameters are defined as

\[
\rho_λ^{*f} = \rho_λ^{af} + \frac{F_λ^f T_λ^f \rho_λ^s}{1 - s_λ^f \rho_λ^s}
\]

(2.8)

\[
\rho_λ^{*c} = \rho_λ^{ac} + \frac{F_λ^c T_λ^c \rho_λ^s}{1 - s_λ^c \rho_λ^s}
\]

where \( \rho_λ^{af} \) and \( \rho_λ^{ac} \) are the atmospheric path reflectances for the fine mode and coarse mode aerosols, \( F_λ^f \) and \( F_λ^c \) are normalized downward fluxes for zero surface reflectance, \( T_λ^f \) and \( T_λ^c \) represent upward total transmission in the satellite field of view, \( s_λ^f \) and \( s_λ^c \) are atmospheric backscattering ratios. In the retrieval procedure the value of η is varied in discrete steps of 0.1 from -0.1 to 1.1. The details of the procedure are described by Levy et al [2007].

2.2.2.1.4 Correction for Rayleigh optical depth

In the C005 algorithm the Rayleigh optical depth is computed considering the elevation of the surface [Dutton et al., 1994; Bodhaine et al., 1999] as,

\[
\tau_{R,λ} = 0.00877 \lambda_p^{-0.05}
\]

(2.9)
In the computation of $\tau_{R,\lambda}$, the surface elevation $z$ is incorporated in Eq. (2.9) through $\lambda_p$ using the expression [Fraser et al., 1989],

$$\lambda_p(z) = \lambda \exp \left( \frac{z}{34} \right)$$  \hspace{1cm} (2.10)

where $\lambda$ is the true wavelength of the desired channel.

Combining Eq.(2.9) and (2.10), the Rayleigh optical depth is estimated in the algorithm as

$$\tau_{R,\lambda} = 0.00877\left[ \lambda \exp \left( \frac{z}{34} \right) \right]^{-4.05}$$  \hspace{1cm} (2.11)

2.2.2.1.5 Inversion procedure for dark pixels

The reflectance data in the 400 pixels encompassed over a domain of 10 km x 10 km [described in Sect. (2.2.2)] is screened based on the value of surface reflectance for each pixel. Dark pixels are those with dense vegetation for which the value of $\rho_{2,1}$ will be small while for the pixels with unvegetated and partly vegetated surface the value of $\rho_{2,1}$ will be large. With this assumption only those pixels which satisfy $0.01 < \rho_{2,1} < 0.25$ are considered for deriving the aerosol parameters over the land. In order to reduce the possible cloud and surface contaminations, the darkest 20% and brightest 50% of the pixels in the 0.66 $\mu$m band are also discarded. After this, if the number of pixels remaining is 12 or more, the average of the reflectances at 0.47, 0.66, 2.1 and 1.2 $\mu$m channels and the NDVI, are computed. Depending on the number of the screened in pixels, a Quality Assurance Confidence (QAC) value between 0 and 3 is assigned for each box. In the algorithm, the fine dominated aerosol model is selected based on the geography and season at the respective region while the coarse dominated model (dust) is always kept fixed. The fine mode (spherical) model include either of the three models viz ‘absorbing/heavy smoke, ‘neutral/generic’ and ‘non-absorbing/urban-industrial’ depending on the case applicable with appropriate single scattering albedos, refractive indices and scattering phase functions [Remer et al., 2006; Levy et al., 2007].

The values of the parameters $\rho^s$, $F$, $T$ and $s$ for fine as well as coarse models are interpolated from the LUT for the solar illumination and sensor viewing angles corresponding to seven values of AOD (0, 0.25, 0.5, 1.0, 2.0, 3.0 and 5.0) indexed at 0.55 $\mu$m. The algorithm then solves for $\tau_{0.55}$, and $\rho^s_{2.12}$ for each of the 13 values of $\eta$ ranging
from -0.1 to 1.1 in interval of 0.1, to match exactly the MODIS measured reflectances at 0.47 μm ($\rho_{0.47}^m$) and 2.1 μm ($\rho_{2.1}^m$) with those obtained from the LUT ($\rho_{0.47}^*, \rho_{2.1}^*$) [Remer et al., 2005; Levy et al., 2007]

$$\left| \frac{\rho_{0.47}^* - \rho_{0.47}^m}{\rho_{0.47}^m} \right| = 0$$ (2.12)

$$\left| \frac{\rho_{2.12}^* - \rho_{2.12}^m}{\rho_{2.12}^m} \right| = 0$$

The final solution for $\tau_{0.55}$ and $\rho_{2.12}^*$ is arrived at iteratively using different values of $\eta$ such that the error $\varepsilon$ defined by,

$$\left| \frac{\rho_{0.66}^* - \rho_{0.66}^m}{\rho_{0.66}^m} \right| - \varepsilon$$ (2.13)

is reduced to a minimum value.

### 2.2.2.1.6 Inversion procedure for brighter pixels

If the number of screened in pixels is less than 12 in the above procedure, the domain is considered to be less vegetated and the pixels within the box are too bright to support the standard retrieval procedure for darker pixels. In this case, the upper limit of $\rho_{2.1}^m$ is relaxed to increase as a function of slant path as

$$0.25 < \rho_{2.1}^m < 0.25G < 0.40$$ (2.14)

where $G = 0.5 ((1/M_\nu) + (1/\sqrt{M_0}))$, $M_\nu = \cos(\theta_\nu)$ and $M_0 = \cos(\theta_0)$ represents the slant path of radiation.

According to this criterion, as the path traversed by the radiation increases, more energy is scattered from the atmosphere, and the contribution from the surface reflectance becomes less significant. This is especially true at the 0.47 μm channel where atmospheric signal is maximum and the surface is very dark. For this reason this alternative path of inversion retrieves aerosol only in the 0.47 μm channel [Remer et al., 2005]. If a minimum number of 12 pixels satisfying the above condition is available in the domain, this inversion procedure for the brighter pixels is followed. Otherwise the procedure ends with no retrieval. In this case, the Quality Assurance Confidence (QAC) is set 0, indicating that data quality is ‘poor’.
Because of the greater retrieval uncertainty over these brighter surfaces as well as due to the fact that the retrieval is made only in one wavelength (unlike the case over the darker pixels), a simple ‘continental model’ is used for the retrieval purpose. The AOD and flux are derived from the LUT for 0.47 µm and are then extrapolated to 0.55 and 0.66 µm using the spectral dependence of the continental model [Remer et al., 2005]. The same VIS to SWIR reflectance relation [given in Eq.(2.4) to (2.6)] used in the earlier method is assumed for this case also and the aerosol properties are indexed to 0.55 µm. Over and above, since a single spheroid aerosol model is used for retrieval, the first term of Eq.(2.7) alone (i.e., \( \eta = 1.0 \)) is used in this case. The primary products derived are \( \tau_{0.55} \) and \( \rho^s_{2.1} \) [Levy et al., 2007].

2.2.2.2 Algorithm over the ocean

Aerosol retrieval algorithm over the ocean employs six MODIS spectral channels (i.e. all except channel 3 in Table.2.1) to derive aerosol properties only under cloud free [Ackerman et al., 1998; Martins et al., 2002], glint free and sediment free [Li et al., 2003] conditions. The 400 pixels, after being screened for clouds, glint and sediment, are sorted in ascending order to avoid the darkest and brightest 25% pixels to eliminate the possible residual clouds, glint and cloud shadow effects. The retrieval algorithm will proceed only if there are at least 10 good pixels at 0.86 µm channel within the 10 km x 10 km domain. A LUT generated using the radiative transfer code by Ahmad and Fraser [1982] consists of TOA reflectances at six wavelengths computed by incorporating the sun glint effect [Cox and Munk, 1954], reflection by white caps [Koepke, 1984] for a surface wind speed 6 ms\(^{-1}\) and a Lambertian reflectance from under water scattering. The water leaving radiance is assumed to be negligible in all the wavelengths except 0.55 µm for which a fixed value of 0.005 is assumed. The LUT further adopts a bimodal lognormal distribution function for aerosol size distribution with nine basic modes including ‘four’ fine modes and ‘five’ coarse modes combined with a weighting parameter (\( \eta \)) which is the FMF such that,

\[
\rho^\text{LUT}_{\lambda} = \eta \rho_f^\lambda + (1 - \eta) \rho_c^\lambda
\]  

(2.15)

The LUT is computed for the 20 combinations of fine mode and coarse mode for 6 AOD values (0, 0.2, 0.5, 1.0, 2.0 and 3.0), 9 solar zenith angles (between 6.0° and 72.0°),
16 satellite zenith angles (between 0.0° and 72.0°) and 16 relative sun/satellite azimuth angles (between 0.0° and 180.0°) for a total of 2304 angular combinations. For each of the 20 combinations the inversion finds the pair of τ_{0.55} and η that minimizes the deviation defined as,

\[
\Delta = \sqrt{\sum_{\lambda=1}^{6} \left( \frac{N_{\lambda}}{\rho_{\lambda}^{m} - \rho_{\lambda}^{LUT}} - \left( \frac{\rho_{\lambda}^{m} - \rho_{\lambda}^{ray} + 0.01}{\rho_{\lambda}^{m} - \rho_{\lambda}^{ray} + 0.01} \right) \right)^2}
\]

(2.16)

where \( N_{\lambda} \) is the sum of good pixels at wavelength \( \lambda \), \( \rho_{\lambda}^{m} \) is the MODIS measured reflectance at \( \lambda \), \( \rho_{\lambda}^{ray} \) is the reflectance contributed by Rayleigh scattering and \( \rho_{\lambda}^{LUT} \) is that provided in the LUT. A fixed value of 0.01 is included in this equation to prevent division by zero for longer wavelength under clean conditions [Tanre et al., 1997]. The \( \rho_{LUT}^{0.87} \) is made to fit exactly with the measured reflectance at this wavelength, as this channel is least affected by variability in water leaving radiance but exhibits strong influence by aerosols. Best fits to other five wavelengths are arrived at using the relation Eq.(2.16). The combination of the modes with accompanying τ_{0.55} and η that minimizes \( \Delta \), gives the best solution. The ‘average’ solution is the average of the solutions with \( \Delta < 3\% \). If no solution has \( \Delta < 3\% \), the average of three best solutions is taken.

To identify dust events over the glint, an additional check is performed in the inversion procedure. Heavy dust activity over glint shows a distinct spectral signature because of increased absorption at blue wavelength. Hence if glint angle is < 40°, and \( \rho_{0.47}/\rho_{0.66} \) is < 0.95, it is identified as heavy dust and retrieval is done, instead of masking the (glint) pixel.

The theory and strategy of C005 aerosol retrieval over the ocean remains the same as that for C004 described above except for a change in the sea salt refractive index in the coarse aerosol model. The primary retrieved products of ocean algorithm are AOD and FMF. From these primary data, other products like effective radius, mass concentration etc. are derived [Remer et al., 2005].
2.2.3 Cloud screening in MODIS

Ackerman et al. [1998] and Gao et al. [2002] have developed an excellent cloud masking scheme for MODIS by making use of a combination of 14 wavelength bands involving about 40 different tests. These different tests confirm the elimination of all types of clouds including thick high clouds, thin clouds, low clouds, upper tropospheric thin clouds and cirrus. But as these tests are clear sky conservative (error towards more clouds), thick aerosol features could be misidentified as clouds. Later Martins et al. [2002] additionally implemented spatial variability test which further produce satisfactory difference between aerosols and clouds. Because of this specialized cloud masking scheme and high spatial resolution, MODIS retrieves aerosol properties near the clouds better than many other sensors.

2.2.4 Calibration of MODIS

The MODIS sensor has an efficient in-built system for performing the radiometric, spectral and spatial calibration of the instrument onboard. This system includes

1. Black Body (BB): Primary source for the calibration of thermal bands located between 3.5 μm and 14.4 μm

2. Solar Diffuser (SD): The source for the calibration of visible, near IR and short wave IR bands.

3. Solar Diffuser Stability Monitor (SDSM): A device which tracks changes in the reflectance of SD in reference to the sun, so that the changes in calibration source will not be incorrectly attributed to the instrument error.

4. Spectro-Radiometric Calibration Assembly (SRCA): This provides in-flight spectral, radiometric and spatial calibration.

Two additional calibration techniques used by MODIS are the viewing of the moon and the deep space. The advantage of "looking" at the moon is that it enables MODIS to view an object that is roughly the same brightness as that of the Earth. Like the on-board Solar Diffuser, the moon is illuminated by the sun. Besides, unlike the Solar Diffuser or the Earth, the moon reflectance is not expected to change over the lifetime of the MODIS mission. "Looking" at the moon provides a second method for tracking degradation of the Solar Diffuser. "Looking" at deep space provides a near zero signal condition, which will
It used as an additional point of reference for calibration (stars are too dim for MODIS to sense).

22.5 Levels of MODIS products

The aerosol products retrieved by processing Level 1B data (discussed in Section 2.2.2) at 10 km × 10 km resolution (at nadir) are called the Level-2 products. Global composites of Level 2 granule products for a day, geometrically corrected and gridded to 1° × 1° resolution for an earth based coordinate system are termed Level-3 daily products [Masouka et al., 1998]. During the re-gridding, the data quality is ascertained by appropriately weighting the Level 2 data based on its quality flags. It should be noted that each grid square in a Level 3 product can contain multiple orbits and hence are not synchronous in time, but is a better representation of daily average. From Level 2, weekly and monthly Level 3 global products are also generated.

22.6 Uncertainties in MODIS retrieval

Extensive exercises for the validation of MODIS have been carried out using data from AERONET stations distributed over the globe. The expected error bars were within ±0.03 ± 0.05τ over the ocean and ±0.05 ± 0.15τ over the land [Remer et al., 2005]. These errors, designated for 0.55 µm are assumed to be applicable to all other channels also. Retrieval results for C005 products over the ocean at 0.55 µm fall within the expected uncertainty 60% of the time. These results are similar to those reported for C004. While over the land, more than 72% of the observations are within the error (against the 68% in the case of C004). High values of bias encountered over the land reduced as much as ~ 0% for Terra and ~ 7% for Aqua (keeping in mind that the expected errors are higher than that over ocean). Over the ocean, even though the implementation of C005 did not show significant effect on the derived AOD, it reduced the positive bias associated with the FMF [Remer et al., 2008].

In the vicinity of clouds (for cloud fraction > 80%), there is a drastic increase in the measured AOD (almost double) both over the land and the ocean, most probably caused by the cloud contamination [Zhang et al., 2005]. However, this condition occurs only 2%
times over the ocean and less than 1% times over the land. It could also be due to factors like 3-D effects [Wen et al., 2006; 2007] and the increased humidity around clouds [Koren et al., 2007; Remer et al., 2008].

In the case of FMF, it does not have a fixed error attached to it. But it shows good correlations with AERONET measurements of fine aerosol weighting defined by O’Neill et al. [2003]. Large bias in FMF over the ocean reported by Kleidman et al. [2005] has now been reduced by improving the aerosol models used for the inversion in the new version of C005. Hence at present, this is in closer agreement with AERONET observations [Remer et al., 2008]. As the FMF under low AOD conditions (AOD<0.2) is highly uncertain, it is not reported under such conditions. In general on an average, an uncertainty up to 30% can exist for MODIS derived FMF over the ocean and is higher in the case of naturally occurring aerosols like sea salt and mineral dust [Kaufman et al., 2005; Kleidman et al., 2005; Bellouin et al., 2005]. The uncertainty of FMF over the land is still larger and hence it is suggested that the values should be used only for a qualitative analysis and that too on a global mean basis. Even though during some occasions the FMF values are closer to the expected ones, they cannot properly represent the transformation of one aerosol type to the other [Jethva et al., 2007]. Hence FMF over the land could be used with caution after a thorough evaluation and validation. A better option for representing sub-micron sized aerosol contribution is fine AOD which is the product of FMF and AOD. The uncertainties associated with this are almost comparable to that of the AOD. This quantity can be used as an indicator for the anthropogenic contribution to the measured AOD [Kaufman et al., 2005].

2.2.7 MODIS data used for the present study

As the present study involves an investigation on the day-to-day variations in the spatial distribution of AOD over a large domain comprising the Indian subcontinent and adjoining oceans, the Level 3 daily data at 1° × 1° grid resolution from both Terra and Aqua satellites are used optimally. Gaps between the satellite passes and presence of clouds reduce the coverage of a satellite sensor. But in the case of MODIS, this can be improved considerably by combining data from Terra and Aqua. The inclinations of the tracks of these satellites on the day side are different and there is a time gap of around 3 hours.
between their passes over a given location. Combining data from both the satellites, the difference in their inclination helps to reduce the gaps between the passes of each satellite. In addition to this, the movement of clouds during the time gap between the two satellites would reduce data loss due to clouds. Therefore, in order to achieve maximal spatial coverage and better representation of the daily aerosol properties, the data from Terra and Aqua are combined optimally. For doing this, the pixels are divided into three types in terms of data content: (i) those having data from both Terra and Aqua (ii) those having data from either Terra or Aqua (iii) those with data from neither. In the case of pixels having data from both the satellites, the mean is taken and for those having data from one satellite, that data is retained. The pixels with no data from either of the satellites are left blank. Figure 2.2 shows a typical example of MODIS derived AOD on March 20, 2006, over the Indian subcontinent and the surrounding oceanic areas (between 0° - 30°N and 40° - 100°E), from Terra and Aqua satellites separately and that obtained through optimal combination as described above. One can notice the considerable improvement in the spatial coverage achieved.

![Figure 2.2. Spatial distribution of Level 3 AOD data from MODIS on board Terra (a) Aqua (b) and that obtained by the optimal combination of data from both Terra and Aqua (c) for March 20, 2006](image)
2.3 Aerosol optical depth from surface measurements

Microtops II Sunphotometer (Solar Light Company, Inc., USA) is a multi-channel hand held instrument, that measures direct solar radiation the attenuation of which in different wavelengths is used to derive the AOD in those wavelengths. This instrument, being portable, is widely used for the measurement of AOD from ground.

2.3.1 Principle of Operation of Microtops II

Microtops-II measures solar irradiance in five user defined wavelength channels with a set of five accurately aligned optical collimators having a full field of view of 2.5°. A narrow-band interference filter (with a typical full width at half maximum band pass of 0.01 μm) and a photodiode optimized for the respective wavelength are fitted to each of these channels.

A sun target and pointing assembly is permanently attached to the laser aligned optical block of the instrument to ensure precise alignment with each of the optical channels. All the optical channels are oriented accurately towards the solar disk by focusing the sun exactly on the cross hairs screen provided on the instrument. Radiation entering the collimators is detected by a photodiode after passing through the interference filters. The anode current of the photodiode is proportional to the radiant power received. These signals are further amplified and converted to digital signals using a high resolution A/D converter. Microtops has built-in pressure and temperature sensors. A GPS (Global Positioning System) associated with this system provides the exact geographical coordinates as well as the time information for each of the observations. Using all these parameters recorded by the instrument, a built-in microprocessor computes AOD along with the column integrated concentrations of ozone or water vapor in real time and displays on an LCD screen [Morys et al., 2001; Porter et al., 2001].

2.3.2 Estimation of AOD using Microtops II

The computation of AOD by the instrument from the ground reaching solar radiation in the visible to near IR wavelengths is based on the Lambert – Beer’s law

\[ V_i = V_{0i} \exp(-\tau_i M_a) \]  

(2.17)
where, \( V_\lambda \) is the signal measured at the wavelength, \( \lambda \) 

\( V_0 \) is the extraterrestrial signal at wavelength, \( \lambda \) corrected for mean sun-earth distance 

\( \tau_\lambda \) is the total optical depth \( = \tau_\lambda + \tau_{\text{R}_\lambda} + \tau_{\text{O}_\lambda} \) where 

\( \tau_\lambda \), \( \tau_{\text{R}_\lambda} \) and \( \tau_{\text{O}_\lambda} \) are optical depths for aerosols, molecules and ozone respectively

\( M_a \) is the optical air mass

Optical air mass \( (M_a) \) is defined as the ratio of the actual path length traversed by the radiation as it reaches the detector to the corresponding vertical path length. This is expressed in terms of the solar zenith angle \( (\theta_z) \) as

\[
M_a = \sec \theta_z - 0.0019167 \left( \sec \theta_z - 1 \right) - 0.002875 \left( \sec \theta_z - 1 \right)^2 - 0.0008083 \left( \sec \theta_z - 1 \right)^3
\] (2.18)

For the estimation of AOD, Rayleigh and Ozone optical depths, \( \tau_{\text{R}_\lambda} \) and \( \tau_{\text{O}_\lambda} \) defined (based on atmospheric models) through the following relation are to be subtracted from the total measured AOD by the instrument.

\[
\tau_{\text{R}_\lambda} = R_4 \exp \left[ -z / (273/29.3) \right] \quad \text{(2.19)}
\]

\[
\tau_{\text{O}_\lambda} = \text{Ozabs} \times \text{DOBS} / 1000 \quad \text{(2.20)}
\]

where,

\( z \) is the altitude of the place of observation in meters

\[
R_4 = 28773.6 \times \left\{ R_2 \times (2 + R_2) \times \lambda^{-2} \right\}^2
\]

\[
R_2 = 10^{-8} \times \left\{ 8342.13 + 2406030 / (130 - \lambda^{-2}) + 15997 / (38.9 - \lambda^{-2}) \right\}
\]

Here, \( \lambda \) is the wavelength of radiation in microns \([\text{Edlén}, 1966; \text{Teillet}, 1990; \text{Bodhaine et al.} 1999]\)

Ozabs is the Ozone absorption cross section, obtained from a LUT based on wavelength \([\text{e.g., Vigroux, 1953; Molina and Molina, 1986}]\),

DOBS is the columnar Ozone amount in Dobson unit, extracted from a LUT based on latitude and date of observation \([\text{e.g., London et al., 1976}]\)
From Eq.(2.17), AOD ($\tau_{\lambda}$) can be derived as

$$\tau_{\lambda} = \left[ \ln(V_{0\lambda}) - \ln(V_{2\lambda}) \right] / M_a - (\tau_{2\lambda} + \tau_{O3\lambda})$$  \hspace{1cm} (2.21)

$V_{0\lambda}$ is obtained by calibrating the instrument through the extrapolation of Langley plot ($\ln V_{\lambda}$ against $M_a$) [Shaw et al., 1973] for zero air mass on clear days at a high altitude location.

### 2.3.3 Errors and precautions

1. Sufficient care has to be taken during the sunphotometer measurements to avoid cloud contamination. Any cloud patch close to the line of sight of the sun can lead to unreasonably high values in AOD. In fact, measurements should be restricted to maintain at least an angular distance of 30° between sun and the nearest cloud patch [Ichoku et al., 2002].

2. The instrument should be pointed accurately towards the Sun such that the centre of solar disk exactly coincides with the centre of cross hairs. Capturing the Sun’s image away from the optical axis increases AOD exponentially [Ichoku et al., 2002].

3. The measurements must be made closer to the local noon to avoid optical distortions arising at larger solar zenith angles.

4. The quartz window of the instrument in the front side has to be cleaned periodically to avoid false measurements (unreally high values of AOD).

5. Being a hand held instrument, pointing accuracy is very critical, especially on moving platforms such as ships, which has been discussed extensively in Porter et al. [2001]. In such cases, it is advised to make 4-5 measurements continuously in quick succession and select the least value among them as the real optical depth. This is because pointing inaccuracies will always lead to decrease in the collected radiance thereby erroneously leading to an apparent increase in the measured AOD.

6. Sunphotometers have to be calibrated at regular intervals (once a year) and well cleaned before measurements to get accurate values of AOD.

With all these precautions, this instrument is expected to provide AOD with an accuracy of ±0.03.
2.4 NCEP reanalysis of meteorological fields

The National Centers of Environmental Prediction (NCEP) / National Center for Atmospheric Research (NCAR) are collaboratively working for a project called ‘reanalysis’ which produces a global record of atmospheric fields for the use of research and climate prediction communities. The basic aim of the project is to use a frozen state-of-art forecast system and execute assimilation of a large collection of data from 1957 onwards. The collection includes data from land surface measurements, rawinsonde, aircraft, satellite and all other possible measurements [Kalnay et al., 1996].

NCEP/NCAR provides several meteorological fields at 2.5° × 2.5° resolution for 17 atmospheric levels from 1000 hPa to 10 hPa. In order to investigate the effects of wind field on aerosols which are mainly confined within the lower troposphere, the data of the lowest seven levels (1000 – 400 hPa) are sufficient. Hence NCEP data of the lowest seven levels are used in the present analysis. The meteorological parameters from NCEP reanalysis used for the present study are the zonal, meridional and vertical components of vector wind and relative humidity.

2.5 Ocean surface wind from QuikSCAT

‘SeaWinds’ is a scatterometer housed in the NASA’s Quick scatterometer (termed QuikSCAT). The scientific objective of this sensor is to carry out high resolution, continuous, all weather measurement of near-surface winds over the ice-free global oceans. The name ‘quik’ is added to this sensor because this satellite was launched with a ‘quick’ planning in 1999 to fill the data gap of the NASA scatterometer (NSCAT), which lost its power in June 1997 while in orbit.

2.5.1 QuikSCAT sensor and measurement technique

QuikSCAT orbits at an altitude 803 km in a Sun-synchronous orbit inclined at 98.6° inclination and has a rotating dish antenna operating in the Ku-band (13.4 GHz) with two pencil beams sweeping the ground in two circles. The inner beam oriented at 40° (angle of incidence) and outer beam at 46° polarized horizontally and vertically respectively produce a wide swath of around 1800 km covering almost 90% of the global oceans in a single day. The centimeter scale or the capillary waves generated by the winds on the
Chapter 2

Ocean surface scatter back the radar power primarily through Bragg resonance modulated by the roughness of the surface. This scattered radiation received by the satellite is used for estimating the vector wind. The measurement geometry of the QuikSCAT is shown in Fig. 2.3

![Measurement geometry of QuikSCAT scatterometer](image)

**Fig. 2.3. Measurement geometry of QuikSCAT scatterometer**

The concept of pencil beam Scatterometry was introduced by Kirimoto and Moore [1985]. Because of the helical scan pattern swept out by the two pencil beams [as shown in Fig. 2.4(a)] of the scatterometer, each measurement cell within the radius of the inner beam will be sampled in four directions: first by the outer beam (T1) followed by the inner beam, looking ahead of the spacecraft (T2) and later when the two beams look backwards, the inner beam scans first (T3) and the outer beam (T4) follows. The typical geometry of these four scans are presented in Fig 2.4(b). These multiple collated measurements from different directions can be used to determine wind speed and direction simultaneously [Kramer, 1994]. Most of the time, four wind solutions are found to be consistent with the observed backscatter at each measurement cell. From these, the absolute wind speed and direction at 10m height from the surface are derived using the 'Geophysical Model Function (GMF)'. These are empirical relations derived based on number of experiments for wide range of incidence angles, frequencies, polarization and azimuth angles. The GMF was first developed for NSCAT [Wentz and Smith, 1999] and later modified to suit QuikSCAT [Lungu, 2001].
Fig. 2.4 The helical scan pattern swept out by the QuikSCAT scatterometer (a) and two beam system which shows a single point being looked at from four different directions (T₁, T₂, T₃ and T₄) (b)

2.5.2 Retrieval uncertainties & difficulties

Maximum uncertainty in the QuikSCAT retrieval is when the wind speed is very low (< 3 m/s), and the spatial variation in wind direction is high. At low wind speed, the backscattered signal will be too low to separate it from noise. For wind speeds in the range 3 – 20 m/s⁻¹, the accuracy is within 2 m/s⁻¹ and in the range 20 – 30 m/s⁻¹, it is within 10%. The wind direction can be measured within an uncertainty of 20° [Freilich, SeaWinds Algorithm Basis Theoretical Document]. The GMF tuned with buoy observations and weather forecast models is found to underestimate the wind speed when it is very large. Hence the ambiguity for wind speed > 25 m/s⁻¹ is large compared to that for the low winds [Hoffman and Leidner, 2005].

Rain also introduces contamination in the retrieved wind data as it affects the ocean surface roughness, and attenuates and scatters the radiation. Hence QuikSCAT cannot retrieve accurate ocean winds during moderate to heavy rain conditions. Hoffman and Leidner [2005] have shown that this condition arises when vertically integrated rain rate (estimated from SSM/I) exceeds 2 km mm hr⁻¹.

The data retrieved in the mid swath lying in the range 200 to 700 km on either sides of nadir will be sufficiently accurate since this region is viewed for maximum different viewing geometry. But the data from regions very near to nadir as well as far from the swath region are expected to be inferior in quality.
2.5.3 Details of wind data from QuikSCAT

For the present study, daily data of meridional and zonal components of the ocean surface winds gridded at 0.25° × 0.25° resolution from QuikSCAT are used. Each data file contains individual data sets for both ascending and descending nodes of the satellite pass. Data from both these passes at every 0.25° × 0.25° grid are optimally combined to represent the daily average value of wind at that location. This has been done similar to the case of combining data optimally from MODIS Terra and Aqua satellites, discussed in Sect. 2.2.7.

2.6 Meteorological parameters influencing the aerosol properties

Atmospheric aerosols are highly heterogeneous in their spatial distribution and also show significant temporal variation. The wide range of aerosol generation mechanism, their varying strengths and their long range transport through atmospheric circulation are the influencing factors responsible for these spatio-temporal variabilities. The day-to-day variations in the spatial distribution of AOD through atmospheric circulation cannot be attributed entirely to wind speed and/or wind direction. It is also influenced by other features of wind such as wind convergence, vorticity and up/down drafts. In addition to these dynamical features of transport, the prevailing atmospheric relative humidity also plays an important role in determining the AOD at a particular location at a given time [Nair and Moorthy, 1998]. The influence of each of these parameters on AOD is detailed below.

2.6.1 Horizontal wind convergence

Wind field in lower atmosphere is not a well organized stream line flow. It is highly turbulent with speed and direction significantly varying from place to place depending on many atmospheric parameters as well as topography. This can lead to convergences and divergences in the wind field which varies with time. Being buoyant, aerosols are carried over long distances, up to thousands of kilometers from their sources depending on the nature of wind. During the transport, dynamical features of wind such as divergence and convergence can influence the spatial distribution of these aerosols. When there is a convergence in horizontal wind, the aerosols accumulate leading to an increase in the concentration even in the absence of sources. Similarly when there is a divergence aerosols
get dispersed very fast even in the presence of an intense aerosol source. The convergence of horizontal wind can be expressed as

\[
\text{Wind convergence} = - \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right)
\]  

(2.22)

where \( u \) and \( v \) are zonal and meridional components of the horizontal wind field. The positive and negative values indicate convergence and divergence respectively.

2.6.2 Vorticity

Another important parameter relevant to aerosol transport is the vertical component of wind vorticity. Locations of positive and negative vorticities and their strengths are indicators of pressure distributions in the atmosphere, which block and guide the flow of aerosols by the winds. Winds flow faster around localized strong pressure gradients. Thus spatial distribution of vorticity becomes very helpful in identifying the origin and transport pathways of aerosols. Vorticity is defined as

\[
\text{Vorticity} = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}
\]  

(2.23)

2.6.3 Vertical wind

Vertical component of wind influences the vertical distribution of aerosols. The uplift of aerosols to higher altitudes and their subsidence to lower altitudes are also governed by this component of vector wind. In several cases, the elevated aerosol layers in the atmosphere are formed as a consequence of change in the direction of the vertical wind.

In the NCEP reanalysis, the vertical wind is expressed in Pascal per second (Pas\(^{-1}\)) with positive and negative signs representing downdraft and updraft respectively. Note that if the vertical velocity is expressed in terms of ms\(^{-1}\), the opposite will be the case.

2.6.4 Relative Humidity

The effect of RH on the aerosol optical properties are detailed in Chapter 1, Sect. 1.6.2. As the RH in the atmosphere increases water vapour condenses on the aerosol particles leading to a change in their size, density and refractive index thereby modifying
their optical properties. All these effects, lead to an increase in the aerosol extinction coefficient and hence, the AOD. Oceanic aerosols which consist mostly of salt particles, generally show high affinity towards water [Parameswaran and Vijayakumar, 1994]. Hence RH will have a substantial influence on the AOD over the oceanic environment. Hygroscopic growth of aerosols however, is found significant only when RH exceeds around 70% [Hanel, 1976; Parameswaran, 1996].

2.7 Investigation of aerosol source strength and continuity equation

The spatial distribution of aerosols is governed by source activities and features of prevailing circulation. Thus, in order to identify the actual role of sources, it is essential to eliminate the influence of wind field. For this, an analysis using an aerosol flux continuity equation is very effective. Using this the spatial distribution of sources can be derived and their strengths can be estimated.

2.7.1 Aerosol flux continuity equation

The dynamics of aerosol transport and generation /loss processes can be described using the aerosol flux continuity equation

$$\frac{\partial \beta}{\partial t} + \nabla \cdot [\beta \mathbf{V}] = s_0$$  (2.24)

where $\beta$ is the aerosol extinction coefficient, $\mathbf{V}$ is the three dimensional vector wind and $s_0$ the strength of aerosol source which are in general, functions of the space co-ordinates and time $(x, y, z$ and $t)$ . The aerosol parameter provided by the satellite is the columnar AOD which is related to $\beta$ as

$$\tau(x, y) = \int_0^z \beta(x, y, z) \, dz$$  (2.25)

Integrating Eq.(2.24) over $z$,

$$\frac{\partial \tau}{\partial t} + \frac{\partial}{\partial x} \left[ \int_0^z \beta u \, dz \right] + \frac{\partial}{\partial y} \left[ \int_0^z \beta v \, dz \right] = S(x, y, t)$$  (2.26)
here \( u \) and \( v \) are the zonal and meridional components of the wind and \( S \) is the column
integrated net source strength. By incorporating the daily values of AOD (from MODIS)
and the daily wind field at different atmospheric altitudes (from NCEP reanalysis) into this
equation, the strength of aerosol source/sink can be estimated.

As the spatial resolution of NCEP wind field is 2.5° × 2.5°, MODIS derived AOD at
1° × 1° resolution is first degraded to match the resolution of the winds. This degrading is
performed by imposing the condition that the AOD values must be available over 60% of
the area covered by the NCEP wind field grid. Otherwise, the grid is kept blank.

A time interval of one day is considered for the first term on the LHS of Eq.(2.26)
which represents the change in AOD in each pixel from one day to the next. The second
and the third terms together constitute aerosol flux divergence. The aerosol flux at different
pressure levels are expressed through the product of aerosol extinction coefficient at those
altitudes with the corresponding NCEP winds. Appropriate models for the altitude
distribution of aerosol extinction are adopted (depending on the region and period of
interest) based on available information.

In the present study, as the time resolution of AOD data is one day, the mean of the
flux divergences for the two successive days are considered along with the day-to-day
changes in AOD for solving Eq.(2.26). It should be noted that to compute the LHS of this
equation, the AOD for a given pixel has to be available on the two successive days along
with the condition that data from at least one adjacent pixel in the zonal and meridional
directions should also be available (for estimating the flux divergence terms). In the
divergence terms, the latitude and longitude grid sizes are converted to corresponding
distances taking into account the earth’s curvature. The vertical integration of the flux term
is carried out over seven levels (from 1000 hPa to 400 hPa) following the standard
Lagrangian interpolation scheme [Jain et al., 1993] by dividing the entire altitude range
into three sections with three sub-pressure levels in each.

On the RHS of the Eq.(2.26), \( S \) is the net aerosol generation expressed in terms of
AOD change per day. This will include aerosol generation from all sources present (natural
and anthropogenic) like wind blown dust, sea-salt particles, industrial exhaust, fuel/biomass
burning etc., and loss due to removal processes such as dry/wet deposition, impaction,
coagulation, diffusion, etc. The sign of the net source will determine the net change in
aerosol loading at a particular location. If the value of net source is zero it suggests that the
aerosol generation is fully compensated by the prevailing loss processes, while positive value of $S$ suggests that the aerosol generation is dominating the loss processes and negative value indicates that the loss process is dominating.

In the present study, the flux continuity equation is used to identify major sources of aerosol generation (with positive values of $S$) and to quantify their strength at different locations over the Indian subcontinent as well as over the adjacent oceanic regions. The use of the flux continuity equation in different study domains are explained in the following chapters.

2.8 Summary

This chapter mainly deals with different sources of data used for the present study (the transport dynamics of aerosols and its influence on the spatio-temporal distribution of aerosols over the Indian subcontinent and the adjoining oceans). The major database required for this study is the spatial distribution of AOD, aerosol size distribution, wind field and atmospheric relative humidity on a synoptic scale. The primary source of aerosol data is MODIS onboard Aqua and Terra satellites from which AOD and FMF are optimally combined to study the spatio-temporal variations. The details of MODIS data and algorithm used for inversion (from the observed upwelling radiance) are briefly described along with the assumptions involved as well as the sources and magnitude of possible errors. Before using the MODIS data it is to be validated (in each case) with ground based observations, for which Microtops II is used. The principle used in this measurement and its possible errors are also described. Synoptic wind field (on a daily basis) from NCEP reanalysis is used for studying the aerosol transport features. The details of NCEP data are briefly discussed. The ocean surface wind from QuikSCAT is used to examine the marine aerosol production over the oceanic regions. The measurement technique used in this space-borne sensor and the probable errors associated with the data are also briefly described. Finally, the wind variables used for studying the transport dynamics and the method adopted for evaluating the source strength using the flux continuity equation are described. These are implemented effectively for studying the spatio-temporal variation in aerosol properties, the results of which are detailed in subsequent chapters.