### Chapter 3
### Ocean Remote Sensing

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Chapter 3
Ocean Remote Sensing

This Chapter deals with two aspects: first, the principles involved in extraction of information from remotely sensed data and second, the data and methods used in the present study. Principles of remote sensing, role of remote sensing in ocean studies, different types of data used in the present study, the corrections used in data processing and basic approaches towards information extraction are discussed.

3.1 Principles of ocean remote sensing

The possibility to observe various ocean parameters and processes using existing satellite sensors, such as optical instruments, infrared radiometers, passive microwave radiometers, active microwave systems, altimeter, scatterometer and SAR is discussed. The basic ocean parameters are: sea surface temperature observed by infrared radiometers, ocean colour by spectrometers, sea surface elevation by altimeters and surface roughness by active and passive microwave systems for deriving surface winds and waves. A number of ocean processes can be derived from synoptic mapping of the basic parameters over larger sea areas, such as current patterns, fronts, eddies, water mass distribution and various water quality parameters including chlorophyll, surface slicks and suspended sediments. The suitability of existing satellite data to fulfil the operational requirements for temporal and spatial coverage, data delivery in near-real-time and long-term access to data are discussed in light of the fact that optical/infrared data are severely hampered by frequent cloud cover, while microwave techniques can provide useful data independent of weather and light conditions.

The growth in space technology increased the opportunity for better understanding of environmental issues especially the processes and dynamics of oceans. SEASAT - the first satellite totally dedicated to oceanography opened a new era in ocean sampling. Alongside the carrying out of a wide range of meteorological studies aimed at the creation of earth Observing Systems (EOS, including oceanic component), the 1980’s saw the appearance of a quite new phenomenon - the transition to practical implementations of satellite data in oceanography without waiting for a properly balanced development of all constituent parts of satellite oceanography. The next decade was the most exciting age of oceanic remote sensing which provided information that facilitated optimal study of the ocean-atmosphere coupled system. Not only has the remote sensing community demonstrated the capability for
observing the oceans from space, but also they, together with the oceanographic community, were well along the way in developing a capability to physically interpret the resulting observations. In three decades of ocean observation from space, remote sensing techniques have gone through phenomenal advancements, which at present provide more consistent and significant information about the seas.

Remote sensing is the observation of a target by a device separated from it by some distance. In the widest sense, it is concerned with detecting and recording electromagnetic radiation from the target areas in the field-of-view of the sensor instrument. This radiation may have originated directly from separate components or the target area; it may be solar energy reflected from them; or it may be reflections of energy transmitted to the target area from the sensor itself (White, 1977). For example, eyes and ears are excellent remote sensors. Vision is a form of optical remote sensing, whereas listening is a form of acoustic remote sensing. Remote sensing makes use of a wide variety of media and technologies like radio energy remote sensing (radar) and high-energy remote sensing (X-ray photography). Eyes and ears are passive detectors relying upon other phenomena to supply the energy (room light or a car horn). RADAR and SONAR, in contrast, broadcast their own energy source and derive information from its reflection and scattering. They are active sensors. Satellite remote sensing is used to obtain information about, and to take measurements of, a place or phenomenon without direct physical sampling. The desired end product of remote sensing is scientifically valid when quantitative analyses are performed with these data. Few environmental products derived from satellite remote sensing include description of current weather conditions, status of wetland habitats, coastal erosion processes, location of oil spills, the extent of algal blooms, and so on.

Satellite images are valuable tools for observing large and inaccessible expanses of the earth. Ocean features such as large-scale circulation, currents, river-outflow and water quality can be examined by analysing variations in colour, turbidity and temperature of the water. These observations are essential in activities like ship routing, environmental monitoring of sensitive coastal zones, hazard assessments, management of fishing fleets, etc. High-resolution coastal images can be used to analyse and map sediment transport, bathymetry, erosion, and for aquaculture applications. However, many of these are possible only in cloud-free conditions.

Applications of remote sensing to coastal management activities include monitoring
changes in vegetative habitat using infrared imagery, understanding fish and invertebrate
distributions using water temperature and colour data and, real-time weather forecasting with
the aid of atmospheric data. Remote sensing techniques are becoming increasingly cost-
effective, given the rapid pace of innovation in computer technology, information networks,
and improvements in sensing systems for satellites.

3.1.1 Remote sensing of sea

Many of the oceanic features that are influenced by various dynamic processes are
characterised by horizontal gradients in bio-geophysical properties of ocean water. However,
the biggest problem in oceanography is the spread of time- and space-sampling that is
inherent in conventional ship-based (in situ) measurement systems. Because of speed
limitations, it is practically impossible to observe large areas synoptically and oceanographers
had to satisfy themselves with long-term averages of data collected over a period of time. In
coastal regime, the property distributions have time scales shorter than the time taken to make
the measurements from a vessel. Hence, synoptic coverage of ocean data using traditional
methods from ships and buoys is inadequate and permits only a limited assessment of the
basin-wide ocean dynamics on climatic time scales. For some stationary and long-period
phenomena such averages may be adequate, but the need for real-time sampling is increasing
with increase in studies on time-dependant oceanic processes. The advent of remote sensing
satellites provided the opportunity for such assessments because of their ability to observe the
earth over large areas and at frequent time intervals. For example, the elevation and
depression of eddies and their movements can be estimated using scatterometry, and ocean
colour annotations and temperature gradients can be obtained from visible/infrared
observations. Effective monitoring of regional and localised circulation patterns and
dispersal of pollutant discharges from fluvial systems are also promising areas of remote
sensing applications of ocean colour sensors. In view of the fast temporal and spatial
developments of coastal features, ship surveys are insufficient tools. Satellite visible data and
infrared temperature data are very useful in these regions because of the large horizontal
gradients. Only satellites can realistically provide the capability to measure surface
temperature and colour over regional, hemispheric or global areas on a repetitive and timely
basis.

However, conventional oceanographers are apprehensive about the capability of
satellite data on processes within the volume of the oceans. This is only partially true as
many of the changes in marine environment that affect human activity – such as waves, tides, storm surges, ice fields, pollution and weather patterns – occur at the sea surface. The processes that occur in the first few meters of ocean depth are of importance in oceanography and meteorology. Remote sensing provides two dimensional synoptic view, very high spatial resolution and low frequency time-series over longer periods.

The methodology of remote sensing of sea differs significantly from the approaches developed for land applications (Barrett and Curtiss, 1976; Lillesand and Kiefer, 1979; Reeves, 1975; Sabins, 1978). However, the problem of sensors for all parts of the electromagnetic spectrum can be scrutinised in four different steps: (a) Sensor calibration (or Radiometric correction), (b) Atmospheric correction, (c) Geometric correction and (d) Geophysical calibration. Details of these calibration techniques are not presented in this study. Problems associated with remote sensing of oceanic parameters and the related instruments are discussed briefly in the following sections.

3.1.2 Electromagnetic spectrum for remote sensing of sea

All remote sensing sensors use electromagnetic radiation (EMR) to view the sea. Generally, active and passive detectors are sensitive to the visual (0.4–0.7 μm), near infrared (0.7–0.9 μm), infrared (0.9–12 μm) and microwave (0.3–30 cm) portions of the electromagnetic spectrum (Figure 3.1). Data from the sensors are used to detect four basic properties of the ocean: colour, temperature, height and roughness. Many applications have been derived from the quantitative detection of these properties. Microwave instruments, such as Synthetic Aperture Radar (SAR), can be used to map oceanographic features including ice fields, internal waves, fronts, eddies and coastal habitats in all weather conditions. High-resolution SAR instruments have been used to detect oil spills, locate ships, monitor topography in the ocean surface to identify changes in coastline and to map bottom topography in shallow waters. When data from multiple sensors are integrated, the product provides additional environmental details. For example, sea surface heights (from altimetry), colour (from optical radiometry) and temperature (from infrared detectors) can be used effectively to study dynamics ocean circulation.

The functioning of all passive sensors (spectro-radiometers) relies on naturally occurring EM radiation (i.e., reflection of the solar radiation or natural emission from earth due to thermal excitation). The blackbody emittance maxima for Sun (at 6000°K) and earth (at 300°K) fall around 0.5μm and 10μm respectively. However, not all parts of the EMR are
useful, since the atmosphere permits only a selective portion of the EMR to transmit through it (Figure 3.2). There are two nearly-transparent ranges, the optical window and the radio window, and several narrow, partial infrared windows. The optical window allows visible light, from red (as far as the A-band of molecular oxygen \((\text{O}_2)\) at 7600 Å) to violet and a little beyond (as far as the ozone \((\text{O}_3)\) cut-off at 2950 Å). The radio window spans a wavelength range from about 1 mm to about 30 m. Lower wavelengths are reflected by the ionosphere while shorter wavelengths suffer increasing amounts of molecular absorption. Water vapour is one of the main absorbers of infrared. Several narrow infrared windows exist at micrometer wavelengths (viz., 1.25 μm, 1.6 μm, 2.2 μm, 3.6 μm, 5.0 μm, 10.2 μm, and 21 μm). There are also small but useable windows at 350μm and 460μm. The radio window, beyond 0.5 cm (~6 GHz) is least affected by the absorbing gases.

Thus, different families of sensors (active / passive) are made in relation to the frequency (or wavelength) range of EMR and the atmospheric windows. The basic passive sensor is a radiometer which responds to fluxes of EM-energy reaching from a given solid angle within a desired wavelength / frequency band. Radiometers are employed to measure the visible (VIS), thermal infrared (TIR) and microwave (MW) parts of the EM-spectrum. Energy received by earth from the Sun is in the form of radiation, nearly all the energy being in wavelengths between 0.2-4 μm. About 40% of this is in visible part of the spectrum (0.4–0.7 μm). The average flux (i.e., solar irradiance from photons of all wavelengths) from Sun at mean radius of earth, commonly known as Solar Constant, \(E_s\), is nearly about 1367 ± 3 W/m² (Hickey et al., 1980; Frohlich, 1983). However, its value varies by a fraction of a percent in time scales of minutes to decades (Frohlich, 1983). Depending on the position of earth in its orbit about Sun, the solar irradiance received by earth varies by about ±50 W/m² in the course of a year or by ±3.5 seasonally (Kondratyev 1969).
3.2 Ocean parameter retrieval from satellite measurements

Almost all physical and bio-geochemical parameters of marine environment namely, ocean colour, sea surface temperature, sea-surface topography and wind stress can be successfully estimated with reasonable accuracies from space-borne sensors. Conventional methods for standard retrievals involve solving an inverse problem using a retrieval algorithm, which relates a parameter to the satellite measurement (e.g., radiance / reflectance, brightness, temperature, etc.).

The principles and physical basis of satellite measurements of four ocean parameters viz., colour, temperature, surface wind, and surface height and their applications are briefly discussed in the following sections.
Figure 3.2: (a) The electromagnetic spectrum in ultraviolet to microwave range with atmospheric transmission windows and radiation absorbing gases which block transmission at specific wavelengths (after Green et al., 2000). (b) Absorption of Sun’s incident radiation in 0.1-30µm. (c) Combined effect of atmospheric absorption, scattering and reflectance reducing the amount of solar irradiance reaching earth’s surface at sea level (after Slater, 1980).

3.2.1 Ocean colour measurement

Observations of changes in ocean colour through measurements of optical properties have been a matter of curiosity and of great practical value as they provide a window into the complex physical, biological and chemical processes of the ocean. So marine monitoring strategy utilises earth observation satellite data in order to provide a representative spatial picture of ocean surface with improved temporal resolution. The colours of ocean and
coastal waters provide information on their contents and their recent history and possible present productivity. For example, clear waters do not contain much suspended material, such as algae or silt; opaque, muddy waters indicate high concentrations of suspended sediment; and bright green waters normally indicate dense concentrations of algae, typically phytoplankton. These microscopic plants are important because they constitute the lowest trophic level of marine food chain, and are involved in many geo-chemical processes including fixation of carbon and nitrogen. Coastal applications of ocean colour monitoring also include quantitative estimation of riverine input into estuaries, coastal erosion (the magnitude and direction of sediment transport in nearshore regions) and location and extent of human impacts on the marine environment. However, the geographic scale of coastal events is often so small that the spatial resolution and / or radiometric sensitivity of available sensors are of minimal utility in coastal resource management. Nevertheless, unlike other detectors (infrared or microwave), the ocean colour sensors can provide depth-integrated information, because visible radiation can penetrate to maximum depth in clear water under a cloudless sky.

The narrow, visible portion of the electromagnetic spectrum (around 400 – 700 nm) is used to record ocean colour, which can be measured only during daylight hours under cloud-free conditions. However, the atmosphere between water and sensor also affects the quality and quantity of light detected at the sensor. To ensure accurate calibration of the numbers from the remote sensor, it is necessary to have frequent in situ measurements of the waters. The ocean colour sensors (passive radiometers) measure the radiant flux, $\Phi_\lambda$ (rate change of radiant energy, $Q$) in a restricted wavelength range, $\lambda$ passing through the aperture of the sensor from a narrow solid angle, $\Omega$

$$\Omega = \frac{dQ}{dt} \text{ Watts}$$

In other words, these sensors measure the reflected spectral radiance, $L_\lambda$ of the ocean surface, which is a measure of the radiant energy, $dQ$ per unit time, $dt$ in a unit spectral bandwidth, $d\lambda$. per unit area of its surface, $dA$ into a solid angle, $d\Omega$ where the element of the radiating area has a projection in direction $\theta$ of $dA \cos \theta$ at depth ($z$). The zenith and azimuth angles ($\theta$, $\phi$) represent the directional terms and are important in satellite remote sensing. The total flux measured at the sensor level is given by
Radiance calculated from Equation 3(2) carries bulk information (including the atmosphere). In fact, the optical pathways to the sensor contain a major part from the atmospheric components. Figure 3.3 illustrates the possible pathways for light rays, which eventually reach the sensor. The labels in Figure 3.3 are explained in the following way:

Left panel: (a) Light scattered by atmosphere. Multiple scattering and reflection at sea surface before or after atmospheric scattering; (b) Specular reflection of direct sunlight at the sea surface; (c) Upwelling light leaving the water surface (water-leaving radiance) and travelling in the direction of the sensor, \( L \). This carries useful information about the water body.

Right panel: (a) Inorganic suspended matter, (b) water molecules, (c) Dissolved organic matters, (d) Sea bottom and (e) phytoplankton pigments. Adapted from Sathyendranath (1986)

Right panel: Before looking at the factors influencing the light upwelled from sea surface, it is worth mentioning that light from Sun gets scattered by atmospheric constituents before it reaches the sea surface and after it leaves the sea surface. In this way, the water-leaving radiance may be scattered away from, or towards the remote sensor by atmosphere. The factors are: (a) Upward scattering by inorganic suspended matters, (b) Upward scattering by water molecules, (c) Absorption by the yellow substances, (d) Reflection off the bottom, (e) Upward scattering by the phytoplankton components.
Thus, the spectral radiance at the top of the atmosphere $L_{TOA}(\lambda)$ (i.e., the radiance received by the sensor) can be expressed as

$$L_{TOA}(\lambda) = L_{atm}(\lambda) + t_d(\lambda)*L_w(\lambda) \quad \text{W/m}^2/\text{sr/\text{nm}}$$

Now Equation 3(3) can be decoupled as

$$L_{TOA}(\lambda) = L_a(\lambda) + L_r(\lambda) + t_d(\lambda)*L_w(\lambda)$$

where $L_{TOA}$ = Sensor detected radiance; $L_{atm}$ = Atmospheric radiance; $L_a$ = Aerosol path radiance; $L_r$ = Rayleigh path radiance; $L_w$ = Water-leaving radiance; $t_d$ = Atmospheric diffuse transmittance.

Since $L_w$ carries information about the water constituents, our basic goal is to obtain this with sufficient accuracy. Equation 3(4) is used directly for explicit calibration and is easily inverted for obtaining $L_w(\lambda)$. For this, one needs accurate measurements of $L_r(\lambda)$ and $L_a(\lambda)$ also because the latter is likely to be spatially variable. Details of procedures of the atmospheric correction for ocean colour sensors and retrieval of oceanic bio-optical constituents (viz., Chlorophyll-a and suspended particulate matter) from $L(\lambda)$ will be discussed in forthcoming sections.

3.2.2 Sea surface temperature measurements

Unlike many physical parameters of oceans, Sea Surface Temperature (SST) relates more directly with ocean-atmosphere interactions. The circulation in the upper layers of ocean is, to some extent, reflected in the distribution of water properties at sea surface. Hence, SST can be an indicator of the circulation features at surface. SST also plays the role of an indicator of the Indian monsoon and El Niño Southern Oscillation (ENSO) events (Bigg 1995), which have devastating effects on regional climate and agricultural economy (WMO 1995). Surface temperature patterns in the tropical oceans can provide reliable forecasting of climate anomalies such as El Niño. SST can also be treated as a quasi-conservative tracer of fronts, gyres, eddies and upwelling regions. Winter bloom associated with winter cooling in Arabian Sea is well known. SST is also useful in identifying the western boundary currents (Legeckis, 1987; Ratna Reddy et al., 1995). The surface temperature of ocean and coastal waters may also provide information on the waters’ origin and recent history. Waters upwelled from great depths are cold, nutrient-rich and clearer than the surrounding water. Many of the world’s major surface currents are warmer than adjacent water masses. In coastal areas, SST measurements help to locate upwelling, fronts, river outflows and
intrusions of water masses. Regional SST measurements are useful in identifying location and areal extent of major currents and their associated eddies and meanders.

3.2.3 SST measurements from satellites

Monitoring SST from earth-orbiting infrared radiometers has had the widest impact on marine remote sensing. Satellite measurements of SST began in the 1970s, using infrared radiometers flying aboard National Oceanic and Atmospheric Administration’s (NOAA) geostationary and polar orbiting platforms (McClain et al. 1985). Very narrow infrared portion of the electromagnetic spectrum (3–14 μm) is typically used for high-resolution temperature observations, which can be made any time of the day but only under cloud-free conditions. Thermal infrared energy from Sun reflected off the water surface can lead to daytime interpretation problems. SST measurements can also be made by satellite microwave radiometry. Microwaves can penetrate clouds with little attenuation, giving a clear view of the sea surface under all weather conditions except rains. At frequencies below about 12 GHz, the surface radiance is proportional to SST; therefore passive microwave sensors offer an uninterrupted view of the ocean surface temperature through clouds, though with a significant decrease in thermal accuracy and spatial resolution. This is a distinct advantage over infrared measurements of SST, which are obstructed by clouds. Thus microwave SST measurements provide us with insight into a number of areas, including tropical instability waves, marine boundary layer dynamics and prediction of hurricane intensity. Sea surface temperature is measured using infrared sensors. We must remember that remote sensing systems operating away from the visible spectrum can view only the top few millimetres to centimetres of the water and thus, cannot provide information on subsurface temperatures. To ensure accurate calibration of the temperature numbers from remote sensors, frequent in situ measurements are required.

The incoming solar radiation which reaches ocean surface is composed of visible and infrared components. Experiments show that clear ocean water absorbs 74% of this flux in the top 1.5m and the remaining portion is absorbed exponentially over 15m (Ivanoff, 1977; Paulson and Simpson, 1977). Sun emits radiation at shorter wavelengths (peak around 500 nm) compared to thermal emission of earth (peak around 10 μm). So SST can be measured from the radiation emitted in the 10–12 μm waveband using passive IR radiometers. The spectral exitance, $M_\lambda$ (often called ‘thermal emittance’), of a body (whose surface is a non-perfect emitter, often called a ‘grey body’) at temperature $T$ °K can be derived from Plank’s
radiation law:

\[ M_A(\theta, \phi) = \frac{2hc}{\lambda^2} \left( \frac{\varepsilon_A(\theta, \phi)}{\exp(hc/\lambda kT) - 1} \right) \text{ W/m}^2\mu\text{m} \]  

where \( h \) - Plank's constant = 6.626 \times 10^{-34} \text{ Js}; \( c \) - speed of light = 2.9979 \times 10^8 \text{ m/s}; \( k \) - Boltzmann's constant = 1.38 \times 10^{-23} \text{ J/K}; \( \varepsilon_A(\theta, \phi) \) - surface emissivity in the direction \((\theta, \phi)\) as a function of wavelength \((\lambda)\).

The spectral emissivity, \( \varepsilon(\lambda) \), at temperature \( T \) describes emitting properties of a real surface as:

\[ \varepsilon(\lambda) = \frac{M_A(\text{real surface})}{M_A(\text{perfect emitter})} \]  

For sea surface, \( \varepsilon \approx 98.0 \); however, it is worth noting that it may vary little with wavelength, temperature, or surface roughness in 3 – 14 \( \mu \text{m} \) wavelength range.

Assuming sea surface to be Lambertian, the radiance measured by satellite sensor is given by:

\[ L_\lambda = \frac{M_A}{\pi} \]  

Inverting Equation 3(5) we can obtain the apparent temperature, which is known as brightness temperature, \( T_B \), that can be expressed as:

\[ T_B(\theta, \phi) = \varepsilon(\theta, \phi) T(\theta, \phi) \]  

Hence, for a known emissivity, \( \varepsilon \) and measured brightness temperature, \( T_B \) the real surface temperature can be estimated using Equation 3(6).

The principles of passive microwave radiometry are essentially similar to the ones described for infrared radiometers except for the operating wavelength range \((i.e., 1.5–300 \text{ nm equivalent to 1–200 GHz})\). Thus, passive microwave radiometers observe thermal radiation in the microwave part of electromagnetic spectrum though thermal emission is weaker at these longer wavelengths. Thermal emission from sea surface is measured as the radiation received by the microwave radiometer’s antenna with some approximation to Planck’s radiation formula. It is often customary to write Equation 3(7) as a function of frequency for microwave radiometry (using the relation \( v = c/\lambda \))
But \((\hbar v/\kappa T) \ll 1\) for micro wave frequencies; hence \(\exp(\hbar v/\kappa T)\) is approximately equal to 1. Thus, Equation 3(8) takes the form

\[
B_s(\theta, \phi) = \frac{2v^3}{c^2} \frac{c(\theta, \phi)}{\exp(hv/kT) - 1}
\]

Equation 3(9) is known as Rayleigh-Jeans approximation and is widely used in radiometry.

Methods employed in microwave antenna response correction, atmospheric correction for SST retrieval and SST calibration correction are beyond the scope of this thesis. However, helpful readings can be found in Reeves (1975), Singh and Warren (1983), Wentz (1983), Stewart (1985), Ulaby et al. (1981, 1982 & 1986). The algorithm involved in retrieval of SST from brightness temperatures will be discussed in subsequent sections.

Currently AVHRR on TIROS, ATSR on ERS and MODIS on EOS provide data for regional and ocean-basin SST determinations. ATSR provides more accurate measurements, while the wider viewing swath of AVHRR (2,580 km) provides more coverage (with spatial resolution of 1.1°×1.1°). Sensors such as OCTS on ADEOS (12 channels; 700 m resolution) and MODIS (36 channels; 250 m resolution) improve available spatial and spectral resolution significantly. Among microwave radiometers, SMMR, SSMI, TMI and MSMR have been providing unique, all weather surface thermal structure of the world ocean since 1970.

3.3 Sea surface wind measurements

Wind stress, \(\tau\) on the sea surface drives the dynamics of the boundary layer and is responsible for generation of surface waves, production of wind-driven surface currents and the stirring processes which keep the upper ocean well-mixed down to the thermocline. Climate research requires precise information on surface fluxes of momentum, heat, moisture, and chemical constituents across the air-sea interface. High resolution, gridded fields of surface wind vectors are necessary to determine exchanges of momentum across the interface. The remaining thermodynamic and tracer fluxes require ancillary fields (e.g., temperature, moisture, chemical concentrations) that are temporally and spatially consistent with the surface wind. Surface wind stress provides the forcing for ocean circulation and hence the global surface wind fields obtained from scatterometers are key elements in large-scale ocean circulation studies. Information on velocity of coastal and
Oceans and wind are very important in resource management. This is especially true during response efforts to hazardous material (pollutant) releases, since disasters seldom happen in ideal weather. It is also useful in weather forecasting, ship routing, and air-sea flux studies. Winds transfer some energy to the surface layer of sea, generating ripples. The ripples develop into wavelets and waves in proportion to the direction and magnitude of wind. Wind stress also induces wind-driven circulation of the oceans. Although the direct effects of wind stress are felt close to ocean surface (the oceanic Ekman layer, extending down to about 100 m), it is the deeper components of the circulation that bring into play a significant volume of seawater. Wind field causing surface water drift in the presence of topographic constraints favors upwelling/sinking (because of divergence/convergence of surface currents) that eventually influence biological activity in the sea.

Ocean circulation models require information on wind stress and wind stress curl. Until recently, oceanographers used wind datasets such as Hellermann and Rosenstein (1983), Comprehensive Ocean-Atmosphere Data Set (COADS; Slutz et al., 1985) and ship observations available in atlas (Levitus, 1982). Conventional observations are often of poor accuracy covering only limited regions of oceans, and are sampled at irregular intervals in time and space. Moreover, wind observations from ships need many corrections before use. Buoy observations have higher accuracy but extremely sparse coverage. Scatterometers aboard earth observing satellites provide a means of improving accuracy as well as temporal and spatial coverage of winds over oceans.

3.3.1 Scatterometer as an anemometer

Scatterometer is essentially a microwave non-nadir (nadir means direction pointing directly below) looking monostatic real-aperture radar that transmits microwave pulses (in wavelengths that are Bragg-scattered by centimetre long ocean waves) to the wind roughened ocean surface and measures the backscattered power received at the instrument. In a radar system, the relation between the received power \( P_r \) and the transmitted power \( P_t \) is given by (Ulaby et al., 1982)

\[
P_r = \frac{\lambda^2}{64\pi^3} \int \frac{PG^2 \sigma^0}{R^2} dA, \quad 3(10)
\]

\[
P_r = \frac{\lambda^2}{64\pi^3} \int \frac{PG \sigma^0}{R^2} dA, \quad 3(11)
\]
where \( \lambda \) - radar beam wavelength, \( G \) - antenna gain, \( R \) - antenna-target distance, \( A \) - effective area (radar footprint), and \( \sigma \) - normalized radar cross-section. The subscripts \( t \) and \( r \) stand for transmitter and receiver, respectively. Equation 3(10) represents the most generic formulation of radar equation and corresponds to bistatic radar (i.e., transmitter and receiver use different antennae and can therefore be in separate locations). This is simplified in Equation 3(11) for monostatic radar. With the assumption that \( \sigma^0 \) does not vary over \( A \), the expression for the averaged \( \sigma^0 \) in \( A \) becomes

\[
\sigma^0 = \frac{64\pi^3 R^4 P_r}{\lambda^2 G^2 AP_r}  
\]

3(12)

The amplitude of the return signal is interpreted empirically as a measure of sea surface roughness. The roughness elements on ocean surface depend largely on the local wind condition, which exhibits large variability. Since atmospheric motions themselves do not substantially affect the radiation emitted and received by radar, scatterometers use an indirect technique to measure wind velocity over ocean. Depending on frequency of radar, and hence on wavelength of Bragg resonant surface roughness, the magnitude of the return signal can be related to the surface wind field. This relationship is known as the geophysical model function (GMF). Wind stress over ocean generates ripples and small waves which roughen the sea surface. These waves modify the radar cross-section (\( \sigma^0 \)) of ocean surface and hence the magnitude of backscattered power. In other words, strength of the return signal is related to amplitude and density of wind-driven waves. Their amplitude and density are proportional to wind speed and shape is dependent on wind direction, so that two looks at different angles from the radar give information about wind direction. The empirical relation (forward GMF) between wind fields and normalized radar cross section (NRCS) can be generalized as

\[
\sigma^0(\theta, \phi, U_w) = aU_w^{\alpha(\theta)}(1 + b\cos\phi + c\cos2\phi) 
\]

3(13)

where \( \theta \) is incident angle of view, \( \phi \) is azimuth angle between upwind direction and radar look angle, \( U_w \) is wind speed and \( a, b, c \) are constants value of which depend upon wind speed, wind direction and incident angle.

In order to extract wind velocity from these measurements, one must understand relationship between \( \sigma^0 \) and near-surface winds. Since GMF contains two unknowns (speed and direction), if only one backscatter measurement from one view is available, then the
inversion problem is underestimated, because there are infinite wind speed and direction solutions, which satisfy Equation 3(13). Most scatterometers operate at Ku band frequencies of ~13.9 GHz with radar wavelengths of ~2.2 cm; nevertheless, these are not the only bands suitable for wind retrieval. Scatterometers extrapolate wind speed by comparing microwave pulse transmitted from satellite with the waveform of the reflected pulse. Since spatial resolution is insufficient, the results are of limited direct use in nearshore processes. However, they can provide warnings of surface wind conditions heading toward shores. Examples of recent Scatterometers are NSCAT (NASA-SCATterometer) and QuickSCAT (followed by NSCAT)

3.4 Sea surface measurements

Sea levels contain information about surface geostrophic currents, tides, waves and bottom topography. Sea level drives ocean dynamics and distribution of properties in one and many ways. Precise knowledge of sea level contributes to our understanding of global ocean dynamics. In coastal waters, knowledge of speed and direction of parcels of water known to contain toxic red tide algal blooms or hazardous materials (e.g., spilled oil, industrial waste, etc.) is essential to plan appropriate remedial measures. Data on ocean circulation is a significant component of global climate programs. During the last decades, rapid developments in satellite altimetry as a remote sensing method have provided a wealth of information on sea levels. Satellite altimeter provides data on height of sea surface for studying dynamics of circulation of oceans, and for other applications, including study of ocean tides, geodesy and geodynamics, wave heights and wind speeds. Since they are active microwave instruments that calculate the round trip time of a pulse transmitted from a satellite in space, altimeters are usable in all weather conditions. With more recent achievement of higher spatial resolution (~25 km), remote sensing using altimeters helps us to study coastal processes such as beach erosion, salt-marsh subsidence, and barrier island development.

3.4.1 Principles of satellite altimetry

Altimetry is a major component of Satellite Oceanography. The physics of altimetric measurements is nothing new, but its application led to phenomenal advancements in Oceanography. Measurement of absolute slope of the sea-surface from SEASAT altimeter in 1978, with an accuracy of around 10 cm, was a dream achievement for oceanographic community. Contemporary altimeters in orbit provide high-precision (~3 cm) information on
sea level and earth’s geoid. An altimeter utilises a microwave radar pulse sent from the orbiting satellite (space station), which bounces off the earth’s surface and returns to the orbiting spacecraft. The radar altimeters transmit short, narrow pulses towards the target from the vertical top and determine:

(a) The time delay, $T_d$, between transmission of pulse and return of the backscattered energy with an accuracy of 1 ns;
(b) The shape of the returned pulse (esp. its leading edge) and
(c) The absolute value of the backscattered coefficient, $\sigma^o$.

Concept of satellite radar altimetry is schematically illustrated in Figure 3.4. The time delay together with velocity of propagation through ionosphere and wet troposphere can be converted to altitude estimation with a precision of the order of a few centimetres. Knowing the electromagnetic refractive index for ionosphere (through ionosondes) and the moist atmosphere (via., surface pressure and integrated water vapour measurements), if $c(z)$ is the resultant profile of the speed of light, then the altitude to the surface, $h(x, y)$, can be determined from the round trip time delay, $T_d$

$$T_d = 2 \int_0^h \frac{dz}{c(z)} \quad \text{(3.14)}$$

Departure of the actual sea surface topography from the equipotential surface (the geoid) can be determined if the geoidal heights are known independently. This departure is known as the sea surface elevation or setup (Figure 3.4). However, errors might arise in calculating the distance from measured time because of variability of wave speed in ionosphere and in atmosphere, as well as due to rains, clouds, ocean waves and tide influences and inverse barometer effects.

Shape of the pulse helps to determine significant wave height $H_{1/3}$ over a range of 0.5-20 m. When an altimeter pulse is incident on a rough sea surface, the backscattered pulse will be broadened out by distribution of reflecting heights presented by surface. When the backscatter is measured over several hundred pulses, a smooth envelope is obtained, which slowly changes with geographical location as the sea state evolves. This returned pulse shape is a convolution of the outgoing pulse, the delta-function response of sea surface and the distribution of surface elevations. If sea surface is Gaussian, the apparent rise time of leading edge of the pulse, $t_r$, can be determined from the radar pulse width ($t_o$ or the compressed
pulse length) and \( H_{1/3} \)

\[
l_p^2 = l_0^2 + \frac{e^{2/3}}{e^{2}} \ln 2
\]

Figure 3.4: Satellite altimetry and its measurement system (Courtesy: Robinson, 1985)

A compressed pulse (when a sharp pulse is dispersed by a filter into a very much longer swept frequency pulse) carries more energy but conveys exactly the same information. Degradation of pulse shape on reflection from a rough sea surface may lead to possible errors, which can be eliminated by applying corrections. Peak value of \( \sigma^0 \) can also be used to determine wind speed knowing its dependence on surface roughness. The joint NASA-CNES (Centre National d'Etudes Spatials) satellite on ocean topography mission, TOPEX, is a dual frequency space borne radar altimeter that is capable of retrieving ionospheric delay of radar signals. The mission also carries two experimental instruments: a single-frequency solid-state altimeter (POSEIDON) for low-power, low-weight altimetry and a Global Positioning System (GPS) receiver for continuous, precise satellite tracking. Skylab, Seasat, GEOSAT, T/P, Jason-1 are few among recent successful satellite altimeter missions.

Satellite observations are of great advantage in understanding ocean dynamics. Moreover, synergistic approach with a blend of in situ measurements can provide complete and lucid picture of complex oceanic processes. Data from different satellite sources used in
this work will be discussed in the coming sections followed by their analyses and results in subsequent Chapters.

3.5 Data and methods

The present study utilises a large quantity of remotely sensed data to elucidate sediment dynamics of the coastal sea along Mangalore coast. The data requirements are:

1. Daily gridded wind field NASA-QuickSCAT for wave hind-casting
2. Sea surface height anomaly from TOPEX/POSEIDON for flow model boundary conditions
3. Time-dependent total suspended matter (TSM) from IRS-P4 OCM data
4. In situ CTD data and Temperature profiles from radiometer and data from wave rider buoy

Brief discussions on data types, sources and related instruments explaining their technical characteristics, measured parameters, accuracies, associated retrieval algorithms and corrections applied to each product are presented in following sections.

3.5.1 QuickSCAT daily winds

The wind datasets used in this study are daily mean products generated from SeaWinds (on NASA-QuickSCAT) scatterometer data. The 1,800-kilometer swath during each orbit provides approximately 90% coverage of earth's oceans every day. The SeaWinds instrument uses a rotating dish antenna with two spot beams that sweep in a circular pattern. The antenna radiates microwave pulses at a frequency of 13.4 GHz across broad regions on earth's surface. SeaWinds scatterometer design used for QuickSCAT is significantly different from the fan-beam scatterometers flown on previous missions (SeaSat SASS and NSCAT). QuickSCAT employs a single 1-metre parabolic antenna dish with twin-offset feeds for vertical and horizontal polarizations. The antenna spins at a rate of 18 rpm, scanning two pencil-beam footprint paths at incident angles of 46° (H-pol.) and 54° (V-pol.). The system measures winds between 3 and 30 m/s with accuracies better than 2 m/s (or 10%) in speed and 20° in direction with a spatial resolution of ~25 km. Performance of QuickSCAT wind data has been found satisfactory in comparison with deep sea moored buoy data over Indian Ocean region (Goswami and Rajagopal, 2003).

3.5.2 Principles of operation

Space-borne scatterometers transmit microwave pulses to ocean surface and measure the backscattered power received at the instrument. Since atmospheric motions themselves
do not substantially affect radiation emitted and received by the radar, scatterometers use an indirect technique to measure wind velocity over ocean. Wind stress over ocean generates ripples and small waves which roughen the ocean surface. These waves modify the radar cross section ($\sigma^0$) of the ocean surface and hence the magnitude of backscattered power. The transmitted radar pulse is modulated or 'chirped', and the received pulse (after Doppler compensation) is passed through an FFT stage to provide sub-footprint range resolution. The range resolution is modifiable between 2 km and 10 km, with a nominal value set at about 6 km. The nominal pulse repetition frequency is 187.5 Hz. Each telemetry frame contains data for 100 pulses. Signal and noise measurements are returned in the telemetry for each of the 12 sub-footprint 'slices'. Ground processing locates the pulse 'egg and slice' centroids on earth's surface. $\sigma^0$ value is then computed for both 'egg' and the best 8 of the 12 'slices' (based on location within antenna gain pattern). SeaWinds antenna footprint is an ellipse approximately 25 km in azimuth by 37 km in the look (or range) direction. Signal processing provides commendable variable range resolution of approximately 2 to 10 km. QuickSCAT generates an internal calibration pulse and an associated load pulse every half-scan of the antenna. In ground processing, the load pulses are averaged over a 20 minute window, and the calibration pulses over a 10 pulse (approximately 18 second) window, to provide current instrument gain calibration needed to convert telemetry data numbers into power measurements for $\sigma^0$ calculation. Since the receiver gain is very stable, the noise power measurements can, with careful calibration, be converted into apparent brightness temperatures measurements. The observed apparent brightness temperatures are functions of instrument noise (continuously monitored), surface emissivity, temperature, and attenuation/emission of the intervening atmosphere. Over the ocean, brightness temperature measurements are used to locate rain cells and to flag wind vector values which can be contaminated by rain effects. In the polar regions, they are mainly useful in discriminating sea ice covered areas from ice-free 'open ocean' since surface emissivity of sea ice is almost twice that of seawater at the operating frequency of QuickSCAT.
3.5.3 Retrieving wind vectors from scatterometer measurements

Scatterometers on board satellites can routinely provide surface wind vector with high spatial and temporal resolution over all ocean basins. Although the exact mechanisms responsible for measured backscatter power under realistic oceanic conditions are not fully understood, theoretical analysis, controlled laboratory and field experiments, and measurements from space-borne radars confirm that backscatter power at moderate incident angles is substantially dependent on near-surface wind characteristics (speed and direction with respect to radar viewing geometry). Presently, microwave scatterometer is the only satellite sensor that observes wind in terms of speed and direction. Scatterometer measurements rely on empirically derived algorithms. An empirical relationship is typically given by the following harmonic formula:

$$\sigma^0 = \sum_{j=0}^{k} A_j (\lambda, P, \theta, U) \cos(j \chi)$$  \hspace{1cm} (3.16)

where $k$ is degree of $\sigma^0$ representation that uses cosines as orthogonal basis (number of harmonics), $\lambda$ is scatterometer wavelength, $P$ is polarization, $\theta$ is radar incidence angle, $U$ is wind speed for neutral stability and $\chi$ is the angle between wind direction and radar azimuth. $A_j (\lambda, P, \theta, U)$ are model coefficients to be determined through regression analysis.

Surface wind speed and direction at a given height are retrieved through minimization in $U$ and $\chi$ space. The Maximum Likelihood Estimator (MLE) function is defined by

$$F = \frac{\sum_{j=1}^{N} (\sigma_j^o - \sigma_j^e)^2}{\text{var}(\sigma^o_m)}$$  \hspace{1cm} (3.17)

where $\sigma_j^o$ and $\sigma_j^e$ are measured and estimated backscatter coefficients respectively. $\text{var}(\sigma^o_m)$ stands for $\sigma^o$ variance estimation. $N$ is number of measured $\sigma^o$ used in wind vector estimation. This approach yields up to four solutions and an ambiguity removal procedure is needed to estimate the most probable wind vector (NASA, 1997).

A main task for a scatterometer investigator is calibration of the sensor data. The calibration involves both determination of the empirical model 3.16 and the development of the surface wind retrieval algorithm. A second task consists of validating accuracy of backscatter coefficients and wind estimates and their comparison with other sources of data. Since July 1999, two scatterometers are available providing surface wind estimates with different instrumental configurations. The first one is on board the European Remote Sensing
satellite 2 (ERS-2) and the second is the NASA scatterometer SeaWinds on board QuickSCAT. The use of both wind estimates should potentially lead to a more refined wind field analysis calculated from satellite data. Details of QuickSCAT have been presented in the previous Chapter.

3.5.4 Estimation of gridded wind fields

QuickSCAT wind is estimated using a statistical technique for objective analysis of remote sensor wind data. This statistical interpolation is a minimum variance method related to the widely employed kriging technique. The analysis scheme is based on determining the estimator of surface parameters derived from scatterometer measurements. The computational details of constructing regular wind fields from polar orbiting satellite data are given by Bentamy et al. (1996).

Let \(v(x)\) be an observation at point \(X = (x,y,t)\) where \(x\) and \(y\) are spatial locations and \(t\) indicates time. We suppose that \(v(x)\) is a realization of the variable \(<U>(X)\). We assume that each measurement consists of the true value plus a random error.

\[
v(x) = <U>(x) + \epsilon(x)
\]

The analysis scheme is based on the determination of estimator \(\hat{U}\) of \(<U>\), at a grid point \(X_0\), of the surface variables using \(N\) observations of \(v\) at the point \(X_i\). Here \(X_i\) stands for spatial and temporal coordinates. The weights \(\lambda\) are determined as the minimum of a linear system using kriging method. Kriging method provides an expression for variance error, named kriging variance, which indicates accuracy of the estimated wind variable at each grid point. The solution of kriging system is used to calculate variance of the difference between the estimated value \(\hat{U}\) and the true value \(<U>\) of the surface parameter.

In order to resolve kriging system, it is necessary to acquire the best possible knowledge of variogram (\(\Gamma\)). Variogram (\(\Gamma\)) is resolved using an exponential formulation. A very sensitive step in this process is, for each grid cell, determination of a neighbourhood containing scatterometer data used to estimate wind vector (wind speeds, zonal and meridional components). Indeed, due to highly irregular spatial and temporal arrangements and density of scatterometer wind observations, the determination of a local neighbourhood is not straightforward; a compromise has to be found between an adequate spatial and temporal sampling number and computing duration. This is done by taking (for daily mean fields),
every hour in the averaged period, the 4 closest (to the grid cell centre) observations available (if exists) within a 600 km radius depending on variogram parameters.

3.6 **TOPEX/POSEIDON monthly sea surface height anomaly**

Radar altimeters provide continuous samples that are unimpeded by cloud cover. Warm, freshwaters rise above surrounding waters providing an excellent signal for radar altimetry. Sea level data from joint NASA-CNEN TOPEX/POSEIDON altimeter (T/P; launched on August 10, 1992 on Ariane 42P launch vehicle), which covers almost 95% of the ice-free oceans every 10 days, were used to compute the quasi-geostrophic flow fields and to observe the meso-scale eddy features.

Originally, T/P sea surface height (SSH) along the ground tracks were pre-processed to remove estimates of tide model errors by removing fitted harmonics at the tidal alias frequencies and to remove long wavelength high frequency signals that were interpreted as orbit errors or errors in environmental corrections (e.g., ionospheric or water vapour corrections to the travel time). Standard processing of altimeter data takes into account all significant aspects of sea surface variability, including tides and eddies. Data used in almost all literatures have been adjusted to remove fluctuations due to changes in local atmospheric pressure (i.e., inverse barometer effect). Ocean Surface Topography Experiment (OSTE) mission claims unprecedented accuracy of 4.2 cm in sea-level measurements. The T/P data are based on T/P MGDR-B data released in 1997 by NASA and CNES (Benada, 1997). NASA/JPL and CNES/AVISO versions have small differences that are usually rounded-off very near land. T/P monthly gridded sea level anomalies (SLA) are obtained from the Physical Oceanography Distributed Active Archive Center (PO.DAAC) of the Jet Propulsion Laboratory (JPL), Pasadena. The SLA fields are differences from the 4-year mean from 1993 to 1996.

Sea surface height anomalies are not used directly in the present study. The flow model is simulated with extracted boundary conditions from a global model that is validated with the help of satellite derived sea surface heights.

3.7 **Ocean colour products**

The ocean colour refers to accurate measurements of light intensity at visible wavelengths. This is largely determined by spectral changes imposed by water on the incoming radiation, as well as by various suspended and dissolved organic and inorganic
substances in the sea. It is, therefore, reasonable to estimate the absolute concentrations of
materials at ocean surface by inverting radiance spectra received by the remote sensor.
Ocean colour observations made from space allow an oceanographic perspective, i.e., a
global picture of physical as well as biological activity in the surface waters of world oceans.

Data from Ocean Colour Monitor (OCM) onboard IRS-P4 (known as OCEANSAT-1)
are used for estimating Total Suspended Matter (TSM). OCEANSAT-1, launched on May
26, 1999, is a polar orbiting, Sun-synchronous satellite, crossing the equator at around 12-
noon Local Standard Time once in two-days. The high spatial resolution of OCM provides
better opportunity to observe local and, small to medium scale changes on the sea surface,
whereas the wide view of SeaWiFS is unique for basin to global scale studies. Technical
specifications of OCM are given in the following Table.

The specific bands of the sensor are selected based on laboratory experiments and
past experience Coastal Zone Colour Scanner (CZCS). Absorption and reflection by various
ocean constituents show different spectral characteristics that are also dependent on satellite
viewing properties and Sun illumination. When visible light from Sun illuminates the ocean
surface, it is subjected to several optical effects. Foremost among these effects are light
reflection and absorption. Reflection beneath the water surface is generally inefficient,
returning only a small percentage of the light intensity falling on the ocean surface.
Absorption selectively removes some wavelengths of light while allowing transmission of
other wavelengths. In the ocean, light reflection is mainly due to suspended particulate
matter, and absorption occurs primarily due to photosynthetic pigments (e.g., chlorophyll-a)
present in phytoplankton. The net result of these optical interactions is light radiating from
the ocean surface – the water-leaving radiance ($L_w$). However, in the context of remote
sensing, it is common to deal with spectral remote-sensing reflectance ($R_{rs}(\lambda)$), which is
closely related to sea surface irradiance reflectance ($R(\lambda)$) but makes use of upwelling
radiance (more precisely, water-leaving radiance) instead of irradiance. The remote-sensing
reflectance is defined as

$$R_{rs}(\theta, \phi, \lambda) = \frac{L_w(\theta, \phi, \lambda)}{E_d(\lambda)}$$

where $L_w(\theta, \phi, \lambda)$ is water-leaving radiance and $E_d(\lambda)$ is downwelling irradiance which has a
directional relationship with radiance as:
$$E_d(\lambda) = \int_{\phi=0}^{2\pi} \int_{\theta=0}^{\pi/2} L(\theta, \phi, \lambda) \cos \theta \sin \theta \, d\theta \, d\phi$$

where \(z\) is depth, \(\theta\) is zenith angle and \(\phi\) is azimuth angle. In other words, remote-sensing reflectance decomposes the reflectance into its component radiances as a function of the viewing angles, \(\theta\) and \(\phi\).

### Spectral Range

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<td>Channel 1: 404-423 (340.5)*</td>
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<tr>
<td>Channel 2: 431-451 (440.7)*</td>
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<tr>
<td>Channel 3: 475-495 (427.6)*</td>
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<tr>
<td>Channel 4: 501-520 (408.8)*</td>
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<tr>
<td>Channel 5: 547-565 (412.2)*</td>
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<tr>
<td>Channel 6: 660-677 (345.6)*</td>
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<tr>
<td>Channel 7: 745-785 (393.7)*</td>
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<tr>
<td>Channel 8: 845-885 (253.6)*</td>
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<table>
<thead>
<tr>
<th>Data rate</th>
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</thead>
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<tr>
<td>20.8 Kbps</td>
</tr>
</tbody>
</table>

*(SNR at typical ocean radiance)

Table 3.1 Technical specifications of IRS-4 OCM

The measured radiance/reflectance is then quantitatively related to various constituents in the water column that interact with visible light, such as chlorophyll-\(a\), inorganic suspended sediments/minerals and degraded organic matters (CDOM) based on their wavelength-dependent natures. Some well-calibrated empirical methods for the retrieval of ocean constituents (total suspended matter) are discussed in subsequent sections.

### 3.7.1 Correction for ocean colour monitor (OCM) data

Solar radiation is absorbed or scattered by atmosphere during transmission to the ground surface, while reflected or emitted radiation from the target is also absorbed or scattered by atmosphere before it reaches a sensor. The ground surface receives not only direct solar radiation but also sky light, or scattered radiation from atmosphere. A sensor will
receive not only direct reflected or emitted radiation from a target, but also scattered radiations from target and atmosphere, which is called path radiance. These effects are to be removed by applying atmospheric correction. The atmospheric correction method is classified into method using radiative transfer equation, method using ground truth data and other methods. In the present study, correction based on radiative transfer model is used in the OCM data. An approximate solution is usually determined for the radiative transfer equation. For atmospheric correction, aerosol density in the visible and near infrared regions and water vapour density in the thermal infrared region are estimated. Because these values cannot be determined from image data, a rigorous solution cannot, however, be used. There are three steps incorporated in the atmospheric correction algorithm developed by the Space Application Centre (SAC), Ahmedabad: (i) cloud screening (ii) removal of Sun glitter and (iii) atmospheric correction based on the radiative transfer equations. This has been used for the removal of atmospheric contamination in Ocean Colour Monitor data in this study.

3.7.2 Methodology for atmospheric correction of OCM data

The radiance detected by a space borne sensor at the top of atmosphere (TOA) in the wavelength \( \lambda \) can be split into four terms (Dorffer, 1992)

\[
L_{\text{TOA}}(\lambda) = L_a(\lambda) + L_r(\lambda) + \tau_d(\lambda)I_w(\lambda)
\]

Rayleigh path radiance, aerosol path radiance and the diffuse transmittance terms can be derived from the following equations:

Aerosol path radiance can be calculated from

\[
L_a = \frac{F_0 \sigma_{oa} \tau_a P_a}{4\pi \cos \theta_v}
\]

Similarly, Raleigh path radiance can be calculated from

\[
L_r = \frac{F_0 \sigma_{ra} \tau_r P_r}{4\pi \cos \theta_v}
\]

where \( \sigma_{oa} \) is aerosol (Rayleigh) single scattering albedo, \( \tau_a \) is aerosol optical depth, \( \tau_r \) is Rayleigh optical depth, \( P_a \) is a function related to aerosol (Rayleigh) scattering phase function, \( \theta_v \) is satellite viewing zenith angle and \( F_0 \) is extra-terrestrial incoming solar irradiance at the top of atmosphere.
Rayleigh single scattering albedo $\sigma_{0r}$ can be safely approximated to ~1.0, whereas the function $P$ related to the scattering phase functions is given by

$$P_m(\gamma) = P_m(\gamma) + [R(\theta_s) + R(\theta_a)]*P_m(\gamma^-)$$

where $R$ is the Fresnel reflectance of water surface and $\gamma^+$ is the forward/backward scattering angle. The phase function for Rayleigh scattering is given by

$$P_r(\gamma^+) = \frac{3}{4}(1 + \cos^2(\gamma))$$

The phase function for aerosol scattering $P(\gamma^+)$ can be modelled by a two-term Heyney Greenstein phase function as

$$P_a(\gamma^+) = A f(\gamma^+, g_1) + (1 - A) f(\gamma^+, g_2)$$

where

$$f(\gamma^+, g) = (1 - g^2)/(1 + g^2 - 2g \cos \gamma^+)$$

with $A = 0.985$, $g_1 = 0.8$ and $g_2 = 0.5$.

Atmospheric diffuse transmittance associated with the water-leaving radiance term in Equation 3(4) is a function of solar and satellite zenith angles and depends on Sun zenith angle and ozone absorption optical thickness.

$$t_d \equiv \exp \left[ \frac{1}{\cos \theta_s} + \frac{1}{\cos \theta_a} \right] \left[ \frac{r_s}{2} + r_o \right]$$

$\theta_s$ and $\theta_o$ are Sun zenith angle and ozone optical depth respectively.

In Equation 3(4), Rayleigh and aerosol path radiances together constitute the atmospheric path radiance. In order to obtain the water-leaving radiance from the detected radiance, $L_r$ and $L_a$ have to be removed from $L_t$ and divided by $t_d$. The Rayleigh path radiance can be computed easily since the spectral dependence of Rayleigh optical depth and Rayleigh phase function are known. The diffuse transmittance is also computed easily since it is dependent only on Rayleigh optical depth and ozone absorption optical depth. For ozone optical depth, we use climatological data since its value is very small and does not have much variability in the spectral range of an ocean colour sensor. However, the aerosol path radiance is difficult to determine since aerosol parameters are highly variable and there is no
a priori information on aerosol optical depths and their spatial distributions. However, by making use of radiances in the wavelengths above 700 nm, we determine the aerosol path radiance indirectly using the following procedure.

We assume that in the spectrum of water-leaving radiance, there will be no radiance for wavelengths $\lambda > 700$ nm due to strong infrared absorption by water. Therefore, the detected top of atmosphere radiance is just the sum of Rayleigh and aerosol path radiances. After removing the Rayleigh path radiance from the detected radiance what remains is the aerosol path radiance. By finding the aerosol path radiance in this way for a few wavelengths above 700 nm and using their variations with respect to wavelength, the aerosol path radiances in the ocean colour regions of the spectrum (i.e., $\lambda < 700$ nm) are computed through extrapolation. Two methods of extrapolation suggested by Gordon (1997) are described below.

### 3.7.2.1 Ångstrom exponent method

Since $L_w \sim 0$ for $\lambda > 700$ nm, Equation 3 (4) can be written as

$$L_t = L_a + L_r$$

That is,

$$L_a = L_t - L_r$$

Observations carried out on different types of aerosols indicate that the wavelength dependence of aerosol optical depth can be modeled by a power law to a good degree of approximation by

$$\tau_a \propto (\lambda)^{-\alpha}$$

This is called the Ångstrom relation where $\alpha$ is known as the Ångstrom exponent. Applying Equation 3(12) in the expression for aerosol path radiance and assuming the phase function to be constant over the range of wavelengths considered,

$$\frac{L_a}{F_0} = K(\lambda)^{-\alpha}$$

where $K$ is a constant. Taking log on both sides,
\[
\log\left[ \frac{L_a}{F_0} \right] = -aK \log \lambda
\]

\((33)\)

\(\alpha\) is assumed to be the negative of slope of the best fit line of \(\log (L_a/F_0) - \log (\lambda)\) for two or more wavelengths greater than 700 nm. In the case of IRS P4-OCM, the wavelengths used are 765 nm (band 7) and 865 nm (band 8). Using these two bands, we can determine the Ångstrom exponent for each pixel of the image as:

\[
\frac{\log(L_{\lambda_1}/F_{\lambda_1}) - \log(L_{\lambda_2}/F_{\lambda_2})}{\log(\lambda_2) - \log(\lambda_1)} = \alpha
\]

\((34)\)

where the suffixes 1 and 2 correspond to the two atmospheric correction bands. Thus, aerosol path radiances in ocean colour wavelengths (corresponding to OCM bands 1-5) may be obtained as:

\[
L_a(\lambda < 700 \text{ nm}) = L_{a_0} (F_0 / F_{\lambda_0}) (\lambda / \lambda_0)
\]

\((35)\)

This can be used to determine the water-leaving radiance

\[
L_w = L_a - L_t - L_r
\]

\((36)\)

### 3.7.2.2 Method of exponential relation

In this approach, instead of power law, the aerosol optical depth is assumed to have an exponential relationship with wavelength. Again, considering the phase function to remain constant over the ocean colour wavelength range, Equation \((310)\) is modified as:

\[
\frac{L_a}{F_0} = K \exp(-C\lambda)
\]

\((37)\)

where \(C\) is a constant. Taking the log,

\[
\log \frac{L_a}{F_0} = -KC\lambda
\]

\((37)\)

\(\log (L_a/F_0)\) is plotted against \(\lambda\). \(C\) is determined as the negative of the slope of the straight line as:
Once $C$ is known, the aerosol path radiance for the wavelengths below 700 nm can be determined as

$$La(\lambda < 700 \text{ nm}) = L_{a1} * (F_0 / F_{01}) * \exp(-C \lambda / \lambda_1)$$

Then $L_w$ is computed using Equation 3(4)

The algorithm for atmospheric correction and estimation of bio-optical parameter assumes cloud-free skies. It is essential, therefore, that cloud-contaminated pixels be masked out before ocean colour images are processed. Seawater and its constituents alter spectral composition of visible (400-700 nm) radiation reaching the satellite sensor. However, near infrared radiation (NIR) is strongly absorbed by open ocean water, resulting in a uniformly low albedo. Clouds, in contrast, have a wide range of reflectivity at the wavelength measured by OCM sensor. Therefore, band-8 of OCM centred at 865 nm provides the best contrast between clear open ocean waters and clouds, and is used in cloud-masking algorithm. An albedo-based cloud masking approach is used for OCM processing, which uses NIR band of 865 nm. This method for calculation of albedo is adopted from Eckstein and Simpson (1991). It is found that OCM image pixels of 865 nm band having albedo greater than 1.1% could be successfully masked for clouds, thin clouds and haze. This threshold value is also found to be effective for masking land and Sun glint affected pixels.

In ocean colour remote sensing, water constituents are detected from the spectrum of solar radiation backscattered from the top layers of ocean water. For incident angles less than $\sim 50^0$, most of the radiant energy refracts into seawater and the remaining part ($\sim 2-4\%$) reflects specularly at air-water interface. The specularly reflected radiation called Sun glitter is a noise in the detection of water constituents since it carries no information on water constituents. The spread of Sun glitter is governed by the angular distribution of facet normals of the rough sea surface (A facet is a portion of sea surface of the order of the roughness scale size). The greater the wind speed, the wider is the spread. Depending on Sun illumination and sensor viewing directions, Sun glitter can mask a considerable portion of an image. To shift the Sun glitter away from the field of view, IRS-P4 OCM sensor has the capability to tilt its viewing angle within $\pm 20^0$ about the nadir. Here, positive value
corresponds to the sensor pointing in the forward direction with respect to the nadir. Sun glitter radiance has been computed in terms of percentage of the total detected radiance in the third band (490 nm) of OCM for an atmospheric aerosol optical depth of 0.32. As water-leaving radiance forms only 10-15% of detected radiance, the tolerance limit for Sun glitter (for the specified OCM estimation accuracy of ocean parameters) was kept at ~2-3%. The most appropriate tilt angle for the Indian coastal waters is found to be -20° during the period from September to April and +20° during rest of the year.

3.8 Estimation of total suspended matter (TSM) from OCM data

Total suspended inorganic matters (TSM) are highly light-scattering in nature and so can be related to the spectral reflectance of solar radiation in green wave length region. The reflectance-based empirical relation used to derive concentration of TSM from IRS-P4 OCM radiance/reflectance is based on the algorithm proposed by Tassan (1994). This algorithm is obtained by a least-square fit between suspended sediment concentration \( S \) and an optimised variable as explained in the following:

\[
S = 25 \exp [2.166 + 0.991 \times \log X_s] \text{ mg/litre} \tag{3(40)}
\]

where \( X_s \) is a factor relating remote sensing reflectance at 490, 555, and 670 nm and is given by

\[
X_s = \left[ \frac{R_{rs}(555) + R_{rs}(670)}{R_{rs}(490)} \right] \tag{3(41)}
\]

The first factor in above equation, with wavelengths in zones of low chlorophyll and yellow substance absorption, is the sensitive term (because of high sediment scattering) and the second factor, with wavelengths in the slope zone of absorption spectra, is known as the compensating term. The retrieval accuracy of the present algorithm is ± 15% (Chauhan, 2002, personal communication). Although this relation provides an indispensable measure of inorganic matters and their pathways, it has limitations in very high turbid coastal waters (with surface suspended sediment concentrations ~100 mg/l and beyond) due to several obvious reasons, for instance, the uncertainty in atmospheric correction procedure near the coast due to land contamination and the coexistence of non-linearly behaving multiple biogeophysical constituents.

Pradhan et al. (2002) attempted to calibrate ocean colour algorithm for TSM in terms of the diffuse attenuation coefficient \( K \) at selected OCM wavelengths in the Bay of Bengal.
coastal turbid waters (sediment concentrations ∼40-200 mg/l). They found a linear relationship between very high concentration of TSM (more than 40 mg/l) and diffuse attenuation coefficient at 555-nm in the coastal waters of Bay of Bengal. It was also found that attenuation coefficient at 555nm may be correlated with the ratio of normalised water-leaving radiance \((L_{wn})\) at 443 and 670nm.

The relation is

\[
K(555) = 0.07 + \left[ \frac{L_{wn}(443)}{L_{wn}(670)} \right]^{-0.87} \text{ per metre.}
\]

The factor of 0.07 on the RHS represents the attenuation coefficient for pure seawater i.e., the minimum possible value for \(K\) at 555-nm. The relationship between turbid water TSM and \(K\) (555) was then established by regression analysis

\[
TSM = 2.93 K(555) + 13.24 \text{ \{mg/l\}}
\]

with a standard error of estimates 15 mg/l (for 25 ≤ TSM ≤ 200 mg/l). This algorithm fails to measure TSM at lower concentrations.

All OCM CD-ROMs containing Level-1 data (un-calibrated digital values) are obtained from the National Remote Sensing Agency (NRSA) Data Centre (NDC), Hyderabad and the level-3 mapped products are generated using the indigenous software developed at the Space Applications Centre (SAC), Ahmadabad. All ocean colour products were generated on UNIX workstations (SGI-O2 and Escala PowerPC) using C and FORTRAN programming languages. The mapped products are colour coded using SeaDAS-4.4 (SeaWiFS Data Analysis Software) package provided by NASA.

### 3.8.1 In situ temperature/salinity/depth profiles

Temperature- salinity-depth data collected from three cruises during February – March and December, 2001 are used in this study. These cruises were planned for studying the influence of mixing of water masses on primary productivity. Locations of ship observations were pre-determined using the concept of partitioning of water masses in biologically homogeneous domains and were fine tuned according to the current information obtained from Space Application Centre (SAC), Ahmadabad through e-mail during the cruises in near-real time. The data were analysed to find the influence of northern Arabian Sea high salinity water mass on coastal dynamics. A sample profile of salinity is shown Figure 3.5.
The data were collected using the Sea-Bird 911 plus CTD system consisting of a SBE-9 plus underwater unit and a SBE-11 plus Deck unit (for real-time readout using conductive wire). The underwater hardware is housed in a main pressure housing containing power supplies, acquisition electronics, telemetry instruments and a suite of modular sensors. The system provides high resolution sampling (24 Hz) up to 6800 m depth, auxiliary sensor flexibility and modem channel for water sample control. The temperature sensor (model SBE 3 plus) is a compact module consisting of a pressure-protected high-speed thermistor and ‘Wein bridge oscillator’ interface electronics that can be operated at -5°C and +35°C. The conductivity sensor (model SBE 4C) is similar to the temperature sensor in operation and configuration, except that the Wein bridge element is a cell resistance. This sensor measures in the range 0 – 70 mho/cm. The pressure sensor also provides a variable frequency output. The sensor frequencies were measured using high-speed parallel counters and the resulting digital data in the form of count totals were converted to numeric representations of the original frequencies. The data set was processed on board using Seasoft software.

Figure 3.5: Salinity profiles along the cruise track of SS-222. Cruise started from off Cochin towards northern Arabian Sea. Note the influence of Bay of Bengal low salinity water at surface at southern station while subsurface levels of adjacent stations are influenced by Arabian Sea High salinity waters.
3.8.2 Data from buoys

The moored data buoys were equipped with sensors to collect time series data on meteorological and oceanographic parameters. In addition to satellite-derived oceanic parameters, time series wave, tide, wind and current measurements from wave rider buoys located at off Mangalore and off Pipavav were also used in this study for wave parameter validation as well as to check validity of satellite-derived wind vectors.