CHAPTER-2
Ocean Colour Remote Sensing: Basics and Review

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Ocean Colour Remote Sensing: Basics and Review

This chapter deals with the basics of remote sensing, a brief discussion of the physical basis of ocean colour remote sensing and describes the present status of ocean colour sensor.

2.1 Fundamentals of Remote Sensing and Electromagnetic Spectrum

Remote Sensing is defined as the science and technology by which the characteristics of objects of interest can be identified, measured or analyzed without direct contact. Electro-magnetic radiation, which is reflected or emitted from an object, is the usual source of remote sensing data. A device to detect the electro-magnetic radiation reflected or emitted from an object is called a "remote sensor" or "sensor". Cameras or scanners are examples of remote sensors. A vehicle to carry the sensor is called a "platform". Aircraft or satellites are used as platforms. The technical term "remote sensing" was first used in 1960's, and encompassed photogrammetry, photo-interpretation, photogeology etc. Since the launch of Landsat-1, the first earth observation satellite in 1972 (http://www.profc.udec.cl/~gabriel/tutoriales/rsnote/cp1/cp1-1.htm); remote sensing has become a very important tool for all earth science studies.

The characteristics of an object can be determined, using reflected or emitted electro-magnetic radiation, from the object. Each object has unique characteristics of reflection or emission, which undergo certain changes during different types of environmental conditions. Remote sensing is a technology to identify and understand the object through the uniqueness of its reflective or emissive properties.

Remote sensing is classified into three types with respect to the wavelength regions (see fig. 2.1):
(1) Visible and Near Infrared Remote Sensing
(2) Thermal Infrared Remote Sensing and
(3) Microwave Remote Sensing

Figure 2.1 shows the radiation sources, objects and spectral radiances in these bands with their approximate bandwidths and the sensors used.

Figure 2.1 Basic classification of remote sensing into three types with respect to the wavelength regions

(Adopted from: http://www.profc.udec.cl/~gabriel/tutoriales/rsnote/ep1/ep1-5.htm)
Regions of the Electromagnetic Spectrum

The electromagnetic spectrum is the distribution of electromagnetic radiation as a function of frequency or wavelength. The table 2.1 gives approximate wavelengths, and frequencies for different regions of the electromagnetic spectrum.

<table>
<thead>
<tr>
<th>class</th>
<th>wavelength</th>
<th>frequency</th>
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<tbody>
<tr>
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<td>100A – 0.4 µm</td>
<td>750 – 3,000 THz</td>
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<tr>
<td>visible</td>
<td>0.4 – 0.7 µm</td>
<td>430 – 750 THz</td>
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<tr>
<td>infrared</td>
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<td></td>
</tr>
<tr>
<td>near infrared</td>
<td>0.7 – 1.3 µm</td>
<td>230 – 430 THz</td>
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<tr>
<td>short wave infrared</td>
<td>1.3 – 3 µm</td>
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<tr>
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<td>3 – 8 µm</td>
<td>38 – 100 THz</td>
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<tr>
<td>thermal infrared</td>
<td>8 – 14 µm</td>
<td>22 – 38 THz</td>
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<tr>
<td>far infrared</td>
<td>14 µm – 1 mm</td>
<td>0.3 – 22 THz</td>
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<td>3 – 3 THz</td>
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<tr>
<td>millimeter</td>
<td>(EHF) 1 – 10 mm</td>
<td>30 – 300 GHz</td>
</tr>
<tr>
<td>centimeter</td>
<td>(SHF) 1 – 10 cm</td>
<td>3 – 30 GHz</td>
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<tr>
<td>decimeter</td>
<td>(UHF) 0.1 – 1 m</td>
<td>0.3 – 3 GHz</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>very short wave</td>
<td>(VHF) 1 – 10 m</td>
<td>30 – 300 MHz</td>
</tr>
<tr>
<td>short wave</td>
<td>(HF) 10 – 100 m</td>
<td>3 – 30 MHz</td>
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<tr>
<td>medium wave</td>
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<td>0.3 – 3 MHz</td>
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<tr>
<td>long wave</td>
<td>(LF) 1 – 10 km</td>
<td>30 – 300 KHz</td>
</tr>
<tr>
<td>very long wave</td>
<td>(VLF) 10 – 100 km</td>
<td>3 – 30 KHz</td>
</tr>
</tbody>
</table>

Table 2.1 Frequency and wavelength of electromagnetic radiation

(Adopted from: http://www.profc.udec.cl/~gabriel/tutoriales/rsnote/cpl/tl-4-1.gif)

The visible part of the spectrum may be further subdivided according to the colour, with red at the long wavelength end and violet at the short wavelength end.

The energy source used in the visible and reflective infrared remote sensing is the Sun. The Sun radiates electro-magnetic energy with a peak wavelength of around ~0.5 µm. The source of radiant energy used in thermal infrared remote sensing is the object itself, e.g., an object at 300K will emit electro-magnetic radiation with a peak at about ~10 µm. In passive microwave remote sensing, the microwave radiation emitted from an object is detected. However, in active remote sensing instruments like radar, a signal is sent from
the sensor itself, which is scattered by the target, and the part that reaches back to the sensor is detected and measured by the sensor. This provides the back scattering coefficient of the target.

An object radiates unique spectral radiant flux depending on the temperature and the emissivity of the object. This radiation is called the thermal radiation because it mainly depends on temperature. Thermal radiation can be expressed in terms of black body theory. A black body is an object that absorbs all electro-magnetic energy incident upon it and does not reflect nor transmit any energy. The black body radiation can be given by Planck’s law as a function of temperature $T$ and wavelength $\lambda$.

**Planck’s law of radiation:**

Spectral radiance of a black body is given by

$$B(\lambda) = \left(\frac{2hc^2}{\lambda^5}\right) \frac{1}{\exp\left(\frac{hc}{k\lambda T}\right) - 1}$$

Where

- $B(\lambda)$ = black body spectral radiance ($Wm^{-2}sr^{-1}\mu m^{-1}$)
- $T$ = absolute temperature of body (°K)
- $\lambda$ = Wavelength (μm) & $k$ = Boltzman’s constant = $1.3806x10^{-23} JK^{-1}$
- $c$ = velocity of light & $h$ = Planck’s constant = $6.6160x10^{-34} JS$

The blackbody radiation curve is shown in fig. 2.2. It gives the spectral radiant flux as a function of wavelength for a given temperature of the source.
Normally observed objects are not black bodies. These are called gray bodies. The radiation from the gray bodies is less than that from a black body at the same physical temperature. The ratio of the radiation from a gray body to that from a black body is termed as emissivity. It can be defined by the following formula:

\[
\text{Emissivity} = \frac{\text{Radiant energy of an object}}{\text{Radiant energy of a black body at the same temperature}}
\]

Emissivity ranges between 0 and 1 depending on the dielectric constant of the object, surface roughness, temperature, wavelength, look angle etc. The temperature of the black body, which radiates the same radiant energy as an observed object is called the brightness temperature of the object.

**Stefan-Boltzman's law** describes emissivity of a black body at a given temperature and is obtained by integrating the spectral radiance given by Planck's law, and shows that the radiant emittance is proportional to the fourth power of absolute temperature \((T^4)\). It is
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given as

$$E = sT^4$$

(2.2)

Where

- $E =$ total amount of radiation emitted by an object per unit area (Watts m$^{-2}$)
- $s =$ A constant called the Stefan-Boltzman constant = $5.67 \times 10^{-8}$ Watts m$^{-2}$ K$^{-4}$
- $T =$ Temperature of the object in K

**Wien's displacement law** gives the wavelength at which a body emits maximum radiation and is obtained by differentiating the spectral radiance, which shows that the product of wavelength corresponding to the peak of spectral radiance and the temperature of the body is approximately 3,000 (µm°K).

**Atmospheric Transmission and Absorption**

The sunlight's transmission through the atmosphere is affected by absorption and scattering of atmospheric molecules and aerosols. The reduction of sunlight intensity is called extinction. The rate of extinction per unit distance traversed in the atmosphere is called as extinction coefficient. The optical thickness of the atmosphere corresponds to the integrated value of the extinction coefficient. The following elements will influence the transmittance of the atmosphere.

Scattering by atmospheric molecules with a smaller size than the wavelength of the sunlight is called **Rayleigh scattering**. Raleigh scattering is inversely proportional to the fourth power of the wavelength. Scattering by aerosols with larger size than the wavelength of the sunlight is called **Mie scattering**. The source of aerosols will be suspended particles such as sea water or dust in the atmosphere blown from the sea or the ground, urban garbage, industrial smoke, volcanic ashes etc. Scattering, absorption and transmittance of the atmosphere are different for different wavelengths. Fig. 2.3 shows the spectral transmittance of the atmosphere.
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Figure 2.3 Characteristic of atmospheric transmittance
(Adopted from: http://www.profe.udec.cl/~gabriel/tutoriales/rsnote/cpl/l-1-1.gif)

The low parts of the curve show the effect of absorption by the molecules. Fig. 2.3 shows the spectral transmittance, or conversely absorption, with respect to various atmospheric molecules. The open region with higher transmittance is called "atmospheric window". Since the present study is dealing with optical ocean colour measurements, the fundamental aspects of remote sensing have been discussed here in brief.

2.2 Inherent Optical Properties of Seawater

The Inherent Optical Properties (IOP) of a medium, describe absorption and scattering interactions that modify a vector light field propagating through it. The IOP definitions and relationships from theoretical and mathematical points of view for practical instrument concepts and methods for measuring IOP in seawater are defined.

The beam attenuation coefficient $c(z,\lambda)$, the volume absorption coefficient $a(z,\lambda)$, and the volume scattering function (VSF) $\beta(z,\lambda,\psi)$ at one or more scattering angles $\psi$, are the fundamental quantities to describe IOP of the medium (at depth $z$ and wavelength $\lambda$).

The fluorescence, or Raman scattering are also IOP of seawater, which contains suspended particles and dissolved materials also. However, only elastic scattering processes are considered in this thesis. Figure 2.4 is an illustration of scattering and
absorption properties of a medium.

Fig. 2.4 describes beam transmission and scattering geometry. A collimated beam of radiometric flux \( \Phi_o(\lambda, 0, 0, \bullet) \) is emitted from a source at the origin \( \hat{x}_o \) and traverses towards Detector 1. It is reduced by scattering and absorption and a reduced flux \( \Phi_r(\lambda, r, 0, \bullet) \) is measured by Detector 1 at position \( \hat{x}_r \). At the intermediate location \( \hat{x}_s \), some fraction of the flux \( \Phi_t(\lambda, r_s, 0, \bullet) \) is scattered into the direction \((\psi, \varphi)\). The directionally scattered flux \( \Phi_s(\lambda, 0, \psi, \varphi) \) is then transmitted to a distance \( r_D \) with further losses caused by scattering and absorption, and the reduced flux \( \Phi_t(\lambda, r_D, 0, \bullet) \) is measured by Detector 2 at position \( \hat{x}_o \).

The shaded portion on the extended \( z_m \)-axis in fig. 2.4 illustrates a cylinder (of cross section \( \Delta \varphi \)) representing the radiative flux transmitted from the source to Detector 1. During transmission over a path interval from \( r \) to \( r + \Delta r \), the fraction of radiant flux...
absorbed in the volume \( \Delta s \Delta r \) is termed as the spectral absorptance, \( A(\lambda) \) and the fraction scattered out of the beam in the volume \( \Delta s \Delta r \) is spectral scatterance \( B(\lambda) \).

The beam attenuation of radiant flux over an optical path length \( r_T \) in seawater is given by the *Beer-Lambert-Bouguer Law*

\[
\Phi_t(\lambda, r_T, 0, \star) = \Phi_o(\lambda, 0, 0, \star) e^{-c(\lambda) r_T}.
\]  

(2.3)

The beam attenuation coefficient \( c(\lambda) \) is given by

\[
c(\lambda) = a(\lambda) + b(\lambda), \text{ m}^{-1},
\]  

(2.4)

Here, the volume absorption and scattering coefficients \( a(\lambda) \) and \( b(\lambda) \) are defined in terms of absorptance \( A(\lambda) \) and scatterance \( B(\lambda) \) as the optical path length \( \Delta r \) approaches to zero as

\[
a(\lambda) = \lim_{\Delta r \to 0} \frac{A(\lambda)}{\Delta r}, \text{ and } b(\lambda) = \lim_{\Delta r \to 0} \frac{B(\lambda)}{\Delta r}, \text{ m}^{-1}
\]  

(2.5)

and the beam transmittance (the fraction of radiant flux transmitted over the path distance \( r_T \)) is given by

\[
T(\lambda, r_T) = \frac{\Phi_t(\lambda, r_T, 0, \star)}{\Phi_o(\lambda, 0, 0, \star)}
\]  

(2.6)

The flux \( \Phi \) can be expressed as

\[
\frac{d\Phi(\lambda)}{\Phi(\lambda, r)} = -c(\lambda) dr
\]  

(2.7)

On integration over pathlength \( r_T \), we get

\[
c(\lambda) \int_0^{r_T} dr = -\int_0^{r_T} \frac{d\Phi(\lambda)}{\Phi(\lambda, r)}
\]  

(2.8)
This gives us,

$$c(\lambda) = \frac{\ln \Phi(\lambda, r, r, r) - \ln \Phi(\lambda, r, r, r)}{r} = -\frac{\ln T(\lambda, r)}{r} \text{ m}^{-1} \quad (2.9)$$

Above equation can be used to determine beam attenuation coefficient from the transmissometer measurements.

**Transmissometers**

Transmissometers are of basically three types direct path transmissometer, folded path transmissometer and variable path length transmissometer.

In a direct-path transmissometer, either a white light, or a light emitting diode, source is combined with a pinhole to provide a point source. A lens is inserted into the path to collimate the light beam, an interference filter is inserted to select the waveband of the measurement, and the light is passed into the water through a window. At the other end of the optical path, the light enters the detector assembly through another window and is focused by a lens. An aperture at the focal point removes off-axis scattered light, and the transmitted light falls on the detector. Several commercial transmissometers including, some laboratory spectrophotometers, and the (former) SeaTech and WET Labs field instruments are being used.

The folded pathlength beam transmissometer design uses one or more reflectors to create a longer pathlength. An aperture at the focal point removes off-axis scattered light, and the transmitted light falls on the detector. The basic idea for this design can be attributed to Petterson (1934), and later refinements were made by several researchers, e.g., Wattenberg, 1938; Timofeeva, 1960, Nikolayev and Zhiltsov 1968.

A variable pathlength transmissometer is a complex concept. One important factor of such an instrument is its ability to adjust the pathlength to make it optimal for the
measuring conditions. Moreover, it is self-calibrating. It assumes that the beam attenuation coefficient is constant for the measurements made. Barth et al. (1997) described the design and application of a variable pathlength instrument for use in coastal waters. Small changes in the alignment of the reference detector will cause large errors in the beam attenuation coefficient. Moreover, the beam collimation errors will introduce errors in the measurement of beam attenuation coefficient.

**Spectral absorption coefficient**

The spectral absorption coefficient is one of the inherent optical properties of the aquatic systems. The absorption coefficient $a(\lambda)$, in $\text{m}^{-1}$, at any point within a natural water body can be described in terms of the additive contribution of its components as

$$a(\lambda) = a_w(\lambda) + a_p(\lambda) + a_s(\lambda) \text{ m}^{-1} \quad (2.10)$$

where $a_w(\lambda)$, $a_p(\lambda)$, and $a_s(\lambda)$ are the spectral absorption coefficients of water, particles, and soluble components, respectively. The spectral absorption coefficients of pure water can be referenced from the Kou et al. (1993), Pope and Fry (1997), Sogandares and Fry (1997) and Fry (2000).

The particle absorption coefficient may be further decomposed as

$$a_p(\lambda) = a_{p}(\lambda) + a_{d}(\lambda) \text{ m}^{-1} \quad (2.11)$$

where $a_{p}(\lambda)$ and $a_{d}(\lambda)$ are the spectral absorption coefficients of phytoplankton, and de-pigmented particles, respectively. It is conceptually possible to further separate $a_{d}(\lambda)$ into absorption fractions due to de-pigmented organic and inorganic particles, but at present, there are no well established relations for separately determining the absorption coefficient for inorganic particles. To interpret aquatic spectral reflectance and better understand photochemical and photobiological processes in natural waters, it is essential to quantify the contributions of the individual constituents to the total absorption coefficients in the ultraviolet (UV) and visible regions of the spectrum. Kalle (1938) and Yentsch
(1962), presented methods for analyzing the absorption by soluble and particulate material in natural waters. The most widely used approach for estimating absorption by particulate matter in water samples is given by Yentsch (1957).

Mitchell and Kiefer (1984, 1988a) made direct estimates of volume absorption coefficients for phytoplankton suspensions. This procedure is the basis of most laboratory methods for determining particle absorption in water samples. Field applications of these quantitative estimates of \( a_p(\lambda) \) were reported by Mitchell and Kiefer (1984) and Bricaud and Stramski (1990).

Separation of the particle fraction into phytoplanktons and other components is of considerable ecological and biogeochemical interest. Early efforts to separate absorbing components for natural particles included treatment with organic solvents, UV radiation, and potassium permanganate (Shifrin, 1988; Bricaud and Stramski, 1990). The most widely used chemical method is based on methanol extraction (Kishino et al. 1985, 1986). A recent method consists of bleaching the phytoplankton pigments by sodium hypochlorite (Tassan and Ferrari, 1995a; Ferrari, 1999). Soluble absorption observations were described by Bricaud et al. (1981) for diverse ocean environments, including oligotrophic and eutrophic regions. Several other investigators have carried out research in this field, (Carder et al., 1989a; Blough et al., 1993; Vodacek et al., 1996; Hoge et al., 1993; Nelson et al., 1998; D'Sa et al., 1999). Spectrophotometric measurement of absorption by dissolved materials is straightforward, but has limits due to the very small signal for short pathlengths routinely employed, and also due to difficulties in maintaining quality control of purified water used as a reference.

**Volume scattering**

There is no practical way to directly measure the volume scattering coefficient \( b(\lambda) \). Given the measurements of beam attenuation and absorption coefficients, however, the volume scattering coefficient may be computed from as \( b(\lambda) = c(\lambda) - a(\lambda) \), m\(^3\). In practice it's not quite that simple, and several interrelated scattering and absorption
corrections must be applied. Alternatively, attempts have been made to calculate $b(\lambda)$ by integrating measurements of the volume scattering function (VSF), e.g., by Petzold (1972). He designed some general angle scattering meters (GASM) with a well-collimated beam and very small detector acceptance angle and reported VSF's measured for selected natural waters using these instruments. Mobley et al. (2002) found that nadir-viewing radiance reflectance is less sensitive to the detailed shape of the forward scattering lobe of $b(\lambda)$.

2.3 Apparent Optical Properties of Seawater

Apparent optical properties are those that modify light field while passing through it. The spectral hemispherical reflectance and spectral remote sensing reflectance are considered as apparent optical properties.

The *spectral hemispherical reflectance* is defined as the ratio of spectral upward to downward irradiance:

$$R(\lambda, z) = \frac{E_u(\lambda, z)}{E_d(\lambda, z)}$$  \hspace{1cm} (2.12)

The *spectral remote sensing reflectance* $R_{rs}$ is defined as the ratio of the water leaving radiance to the downward irradiance just above the sea surface ($z=0^+$):

$$R_{\text{rs}}(\theta, \phi, \lambda) = \frac{L_w(\theta, \phi, \lambda)}{E_d(\lambda)}$$ \hspace{1cm} (2.13)

The spectral remote sensing reflectance specifies that portion of downward light incident onto the water surface that is returned through the surface into direction ($\theta$, $\phi$).

2.4 Physical Basis of Remote Sensing of Ocean Colour

Sunlight is the information carrier in all passive ocean colour remote sensing techniques. Water is only transparent for the visible parts of the electromagnetic spectrum; therefore we are restricted to this small spectral window including the near infrared (NIR), used for atmospheric correction procedures. The following basic steps are involved in
remote sensing of ocean colour

1. Sunlight penetrates the atmosphere, whereby atmospheric gases, such as ozone, oxygen and water vapor, selectively absorb part of the light. Another part of the radiation is scattered by air molecules. Both processes, absorption and scattering in the atmosphere, are wavelength dependent. They determine the transmission of light through the atmosphere and the light backscattered from the atmosphere into the sensor.

2. At the sea surface, part of direct sunlight and the scattered light are specularly reflected according to the Fresnel’s law. The reflected sunlight has to be avoided when measuring the radiance of the ocean surface by tilting the instrument away from the nadir into the opposite direction of the sun’s azimuth angle. At high sun elevations and a rough sea it is difficult or even impossible to fully avoid the sun glitter. In this case, the image has to be masked using the wind climatology at the glitter locations. The specularly reflected diffuse sky light also reaches the sensor and has to be taken into account in the data evaluation process. According to the Fresnel’s law it increases with increasing viewing angles and is modified by the sea surface roughness. Furthermore, substances floating on the water surface, such as mineralic or natural oil films, floating plankton (blue-green algae), foam and white caps, may mask the water leaving radiance.

3. Within the water column, sunlight is partly absorbed and partly scattered by water molecules and suspended constituents. Light energy absorbed by phytoplankton and gelbstoff is also partly re-emitted at longer wavelengths in the form of fluorescence. Substances may be in distributed within the water column with maxima occurring at certain depths.

Figure 2.5 shows the schematic of the interaction of the solar radiation with the atmosphere and the ocean with emphasis on scattering and attenuation of the radiation within the atmosphere.
Figure 2.5 Interaction of solar radiation with atmosphere and ocean

(Concept from: Robinson, I.S., 1983: Philo. Trans. Royal Soc. of Lo London, Series A, Volume 309, 338-347, Figure 6.10)

The various symbols used in figure 2.5 are explained below:

- **a** - The light path of the water-leaving radiance.
- **b** - Shows the attenuation of the water-leaving radiance.
- **c** - Scattering of the water-leaving radiance out of the sensor's Field Of View (FOV).
- **d** - Sun glint (reflection from the water surface).
- **e** - Sky glint (scattered light reflecting from the surface).
- **f** - Scattering of reflected light out of the sensor's FOV.
- **g** - Reflected light is also attenuated towards the sensor.
- **h** - Scattered light from the sun that is directed toward the sensor.
- **i** - Light already scattered by the atmosphere that is then scattered toward the sensor.
- **j** - Water-leaving radiance originating out of the sensor FOV but scattered into the sensor.
\[ k \] - Surface reflection out of the sensor FOV that is then scattered toward the sensor.
\[ L_w \] - Total water-leaving radianc.
\[ L_r \] - Radiance above the sea surface due to all surface reflection effects within the IFOV.
\[ L_p \] - Atmospheric path radiance.

In shallow and clear water, the bottom of the sea may also contribute to the reflected sunlight and change the ocean colour. All of these basic processes have to be taken into account for the measuring strategy including ground truth program and data evaluation.

2.5 Status of Ocean Colour Sensors

Fargion (2003) discusses the ocean colour sensors flown onboard various satellites and also outlines their operational characteristics (see table no 2.2). Many of these sensors have additional bands, not listed here, addressing data requirements in terrestrial or atmospheric sciences.
### Table 2.2 Details of various ocean colour missions and sensors

The details of the Moderate Resolution Imaging Spectroradiometer (MODIS) are given in a separate table (table no 2.3).
<table>
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<th>Primary Use</th>
<th>Band</th>
<th>Bandwidth (nm)</th>
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<th>CZCS</th>
<th>AVHRR</th>
<th>SeaWiFS</th>
<th>MVISR</th>
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<td>3.929 - 3.989</td>
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<td>22</td>
<td>3.929 - 3.989</td>
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<td>1.360 - 1.390</td>
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<td>8</td>
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</table>

(Contd.)
Table 2.3 Details of Moderate Resolution Imaging Spectroradiometer (MODIS)

(Adopted from: http://modis.gsfc.nasa.gov/about/specs.html)

Remark:

- **Resolution**: Bands 1 and 2 are 250m resolution, bands 3 to 7 are 500m; the rest are 1km.
- **Bands 13 and 14** both have a high detector and a low detector, giving increased sensitivity.
- **Bands 21 and 22** cover the same part of the spectrum but have different saturation points. Channel 21 saturates at about 500 K, and channel 22 saturates at about 335 K. Channel 22 is also less noisy.

Since in this thesis, work has been mainly carried out for the IRS-P4 OCM sensor, a separate table has been provided to describe the characteristics of the orbital and sensor related parameters of OCM (table no 2.4).
2.6 Remote Sensing of Water

2.6.1 Bio-optical Properties of Water

A fundamental requirement for interpreting remotely sensed radiance spectra with inverse modeling technique, is the knowledge of the optical properties of all water constituents and of the optical processes that modify the ocean water leaving radiance.
The absorption of water cannot be measured directly but derived from measurements of the total beam attenuation coefficient. The absorption shows a strong increase with increasing wavelengths. As a result of scattering and absorption, the transparency of pure water and pure seawater has its maximum around $\lambda = 0.47 \, \mu m$ and the remote observations of water constituents are only possible within the wavelength range $\lambda \sim 0.4 - 0.7 \, \mu m$. Another consequence for imaging spectroscopy is that we are looking deeper into clear water at shorter wavelengths than at longer wavelengths. In case of an inhomogeneous vertical distribution of a substance, this may imply difficulties in data interpretation. Particularly the use of the fluorescence (with its maximum around $\lambda = 0.685 \, \mu m$) is limited only to the upper few meters of the sea due to the absorption by pure water (Doerffer, 1992). The increasing absorption of pure water towards the red can be used to estimate the vertical distribution of chlorophyll within the penetration depth of case I water (i.e. where phytoplankton chlorophyll is dominant) by combining measurements at different wavelengths using the absorption effect in the blue/green (absorption band) and the fluorescence in the red.

The Inherent Optical Properties (IOP) of a medium, include absorption and scattering interactions that modify a vector light field propagating through it. These properties describe the loss in radiant intensity (or energy) per meter of a parallel beam of the light and are wavelength dependent. Therefore, all processes of the propagation of light through the water column and the atmosphere depend on the wavelength.

It is difficult to experimentally determine the absorption of pure water in the laboratory, principally due to the difficulty of making and maintaining pure water during the course of an experiment. However, there have been several successful experiments over the past few decades, and the spectral absorption and scattering coefficients of pure water are reasonably well known. The optical properties of pure water include polarization also but this study is mainly concerned only with absorption and scattering. The contrast in angular distribution characteristics of scattering by water and by particles is an important element in instrument design.
Absorption by Pure Water

The absorption by pure water shows a strong increase with increasing wavelength. The spectral values recommended for the volume absorption coefficients of pure water, \(a_w(\lambda) \text{ m}^{-1}\), are those of Sogandares and Fry (1997) for wavelengths between 340 nm and 380 nm, Pope and Fry (1997) for wavelengths between 380 nm and 700 nm, and Smith and Baker (1981) for wavelengths between 700 nm and 800 nm. Further work was carried out for wavelengths > 700 nm by Kou et al. (1993) and Van Zee et al. (2002). The composite \(a_w(\lambda)\) spectrum derived from these sources together with the linear temperature dependency \(\frac{\partial a_w(\lambda)}{\partial T} \text{ [m}^{-1}\text{°C]}\) is reported by Pegau and Zaneveld (1993) and Pegau et al. (1997).

Scattering by Pure Water

The scattering of pure water into the solid angle \(\gamma\) normalized to the total scattering (into the sphere \(4\pi\)) can be described by the Rayleigh phase function \(P_\gamma(\gamma)\)

\[
P_\gamma(\gamma) = \frac{3}{4} (1 + \cos^2 \gamma)
\]

(2.14)

A better approach for liquids is the fluctuation theory, which includes the depolarization factor \(\delta\) with a value 0.09 for seawater. Then the phase function is described by

\[
P_\gamma(\gamma) = \frac{3}{2} \left(1 + \frac{1 - \delta}{1 + \delta} \cos^2 \gamma\right) \frac{1 + \delta}{2 + \delta}
\]

(2.15)

The scattering coefficient \(b(\lambda) \text{ (m}^{-1}\)) is wavelength dependent. It decreases with increasing wavelength for pure seawater by \(\lambda^{-0.432}\) (Doerffer, 1992). The scattering coefficient for pure seawater at wavelength 0.45 \(\mu\text{m}\) ranges from 0.00349 m\(^{-1}\) to 0.00454 m\(^{-1}\). It can be calculated for other wavelengths from the following formula (for seawater)
The typical angular distribution of the molecular scattering phase function for pure water is illustrated in fig. 2.6. The magnitude of \( \beta_\psi(\psi) \) represents the probability that a photon scattering interaction with a water molecule will redirect the photon path direction by an angle \( \psi \) measured from its original path. The shape of \( \beta_\psi(\psi) \) is sometimes referred, as "isotropic" in the literature, a characterization that is true only in that the function is axially symmetric and the probabilities of forward and backward scattering are equal. (Pegau, 2003)

\[
b(\lambda) = b(\lambda_0) \left( \frac{\lambda}{\lambda_0} \right)^{-4.32}
\]

(2.16)

Fig. 2.6 The dashed curve represent the angular distribution of the scattering phase function for pure water, calculated using the approximate Rayleigh scattering model (Adopted from: Pegau, 2003)

2.6.2 Bio-optical Properties of Phytoplanktons

Although phytoplankton is part of suspended matter in the sea, it has to be treated separately because of its pigment, that selectively absorbs radiance of the visible spectrum. The most dominant pigment is chlorophyll-a that is used by the photo-organisms, to absorb light energy for photosynthesis. Other accessory pigments, such as carotenoids, assist
chlorophyll in harvesting light energy at those wavelengths for which chlorophyll has no or weak absorption bands. Besides the concentration of these pigments, the structure of the chloroplasts also has an important influence on the absorption. This structure is a major cause of the variability of the specific absorption coefficient. The absorption per unit pigment concentration depends on:

- the species or species composition of a plankton population,
- the light conditions under which the plankton population grows (season, climate, turbulence within the water column, etc.) and
- the nutrient conditions.

As a consequence, it is not possible to use standard absorption spectra for all areas, seasons etc. One of the important prerequisites for utilizing the information of an imaging spectrometer is the knowledge of the optical properties of the phytoplankton that is actually present in the area under research. Exceptional plankton blooms, such as ‘Red Tides’, ‘White Water’ coccolithorides, floating blue green algae, etc. cause a strong discolouration of the sea. This can be easily detected by eye or a visual inspection of the images during phases of such blooms.

The scattering properties (scattering function; total, forward and backward scattering coefficients) of phytoplankton are characterized by a narrow size distribution compared to other suspended particles, particularly in case of mono species bloom. They have to be calculated using the Mie Theory with the help of size distribution, refractive index and absorption coefficient.

The absorption spectrum cannot be derived from pigment extraction, because the structure of the chromatophores has to be maintained. Therefore, in situ absorption is determined, which can be performed by vector and scalar irradiance measurements from in situ systems, or by correcting the absorption spectra of a pigment extract. Biogenic detritus, such as fecal pellets of zooplankton or fragments of zoo and phytoplankton, may also contain pigments and their degradation products. They are present particularly in coastal zones near estuaries, especially during phases of decaying phytoplankton blooms.
The most prominent optical property of phytoplankton with respect to selective passive remote sensing is the sun light stimulated fluorescence. The fluorescence effect is studied by using non-imaging spectrometers. The sunlight energy, re-emitted by phytoplankton chlorophyll in the form of fluorescence augment the water leaving radiance around $\lambda = 0.685 \, \mu m$ where it produces a small peak. The height of this peak can be calculated by constructing a baseline from radiances at wavelength before and after the fluorescence peak. The difference between the baseline and the fluorescence peak is called the fluorescence line height (F.L.H).

Many experiments have shown that the FLH can be linearly correlated with the chlorophyll concentration and thus, used to determine the concentration of chlorophyll pigment. However the regression may vary depending on area and season as a function of the phytoplankton population and its physiological state. The physiological state depends on the actual light conditions, the light history and the nutrients availability.

Thus, the relation between fluorescence line height and the chlorophyll concentration has to be determined for each experiment through ground truth. Furthermore, it has to be considered that, with increasing concentration, the shape of the fluorescence peak as well as its maximum may shift to the red side of the spectrum due to the absorption of chlorophyll with its maximum in the range $\lambda = 0.670$ to 0.675 $\mu m$. Suspended matter may attenuate the fluorescence signal or phaeopigment containing detritus and can give misleading high values of chlorophyll concentration. The strong oxygen and water vapor absorption bands at $\lambda > 0.687 \, \mu m$ have to be avoided when selecting appropriate spectral channels for measuring the fluorescence peak and baselines. In cases, where the phytoplankton maximum is below that depth, the fluorescence signal becomes very weak or is not observable.

Figure 2.7 shows the sample spectra for pure water, chlorophyll-a and Combined Dissolved Organic Matter (CDOM). The X-axis shows the wavelengths ranging from 350nm-750 nm and the Y-axis represents the arbitrary absorption coefficients, but in arbitrary units. The amplitude of each absorption spectrum is arbitrarily scaled to illustrate the characteristic difference in shapes between the constant water background absorption
and the two varying absorption components associated with chlorophyll and CDOM concentrations. The unique shape and magnitude of the absorption spectrum describes relative values of pure seawater, chlorophyll and CDOM concentrations at different wavelengths.

The pure water has low absorption coefficient up to about 500 nm that primarily lies in the blue green regions of the light spectrum. Thereafter it starts rising and shows steep rise beyond 700 nm (i.e., in Near Infra Red, NIR).

The chlorophyll-a exhibits two strong peaks one around 443 nm and the other around 675 nm. The first peak is the chlorophyll absorption peak and the second one is its fluorescence peak.

The curve for Combined Dissolved Organic Matter (CDOM) shows strong absorption in the ultra violet region and gradually decreases as we go to visible and NIR regions.

Figure 2.7 Qualitative comparisons of the shapes of absorption spectra of pure water, Chlorophyll-a and Combined Dissolved Organic Matter.
(Adopted from: Prieur and Sathyendranath, 1981)
2.6.3 Bio-optical Properties of Suspended Particles and Coloured Dissolved Organic Material (CDOM)

Suspended matter is defined as particulate matter in seawater and includes all the particles that do not pass through a filter with a pore size of 0.47 \( \mu \text{m} \). For remote sensing purposes, all suspended particles in the water without particular pigments are considered as suspended matter. The phytoplankton pigments are treated separately. Thus, suspended matter consists mainly of mineral particles (e.g., grains of quartz) and degradation products of organisms (e.g., broken shells of diatoms, zooplankton) materials transported by rivers or from the atmosphere in particulate form into the sea. The absorption spectrum of this class of constituents is characterized by a smooth and flat decrease with increasing wavelength without any prominent maxima (see figure 2.7). A similar shape exhibits the scattering spectrum, which depends on the size distribution of the particles. For particles which have a size with a radius of the order of a wavelength or larger, scattering is described by the Mie theory. The most prominent feature of the scattering is a very steep increase at small angles. This property makes measurements of the scattering function very difficult. But it is possible to calculate the scattering function by the Mie theory if one knows the particle size distribution, the refractive index of the particles in water and the absorption coefficient.

The Mie theory assumes particles with a spherical form and a homogeneous refractive index, which, in fact is often not the case. However, deviations from these assumptions do not cause substantial problems. By using the Mie theory, one can calculate a phase function (the normalized scattering function, i.e. the integral over all scattering directions), which can be assumed as constant for all wavelengths within the visible part of the spectrum. Also the relation between scattering and absorption can be assumed as constant for this wavelength range. For areas of cases where material with other optical properties is dominant, these normal assumptions have to be modified and use other size distributions; absorption coefficients and another single scattering albedo. This is particularly the case for plankton blooms where the dominating species have a narrow size range.
About 70% of the dissolved carbon in the ocean is formed by a mixture of various molecules that can only be described by a sum parameter and not by a single structure formula. This fraction, which passes a filter with a pore size of 0.47 μm, is called Gelbstoff, since it causes – in case of high concentrations – a yellow discolouration of the water by its high absorption of blue light. Gelbstoff is formed from excretion or degradation products of organisms under the activity of microbes. After microbial degradation and decomposition, new complex molecules are composed by biochemical condensation. Most of Gelbstoff is transported by rivers into the sea, but it has been also found in patches of high phytoplankton concentrations. Gelbstoff molecules are very stable and thus can be used as natural tracers. In coastal zones, where estuaries are transporting large quantities of Gelbstoff into sea, the absorption of Gelbstoff can be used to trace the distribution of fresh water within the coastal area.

Absorption by Suspended Particulates and Coloured Dissolved Organic Material (CDOM)

Variations in the spectral absorption of natural waters result directly from variations in the concentrations and chemical compositions of material substances distributed within the water volume. These absorbing materials may be present in seawater either in suspended particulate form, such as pigment-bearing phytoplanktons, or as solubles (i.e. CDOM).

In Case 1 waters, it is often useful to assume (Gordon and Morel 1983; Morel and Maritorena 2001) that particle absorption is dominated by phytoplankton pigments and may be expressed as a function of chlorophyll concentration [mg m⁻³] and a Chl-specific absorption spectrum (Prieur and Sathyendranath 1981), and that CDOM concentration is correlated with chlorophyll.