Chapter 4

Modeling Photosynthetically Available Radiation (PAR) from satellite data over Indian Ocean

This chapter describes estimation of PAR from OCEANSAT-1 & 2 Ocean Colour Monitor (OCM) using two different methods under both clear and cloudy sky conditions. In the first method, the atmosphere is treated as a single layer in clear sky conditions, and as a double layer in cloudy conditions. Surface reflectance has been neglected in the first approach. Aerosol optical depth and cloud optical depth has been estimated at 865 nm and 443 nm spectral bands of OCM. PAR has been estimated from OCM using second method which assumes that the effects of clouds and clear atmosphere can be decoupled with cloud system and ocean surface albedo. Ocean surface albedo and cloud albedo have been estimated from TOA (top of the atmosphere) radiance data of OCM on a pixel by pixel basis.
4.1 Introduction

PAR estimation based on satellite observation has become increasingly important as PAR estimated from satellite provide information at desired temporal and spatial resolution required by different ecosystem models and radiation budget models. Ocean colour remote sensing is a useful tool and it provides quantitative information of seawater constituents. OCEANSAT-1 OCM and OCEANSAT-2 OCM are two Indian Ocean observation satellites which are extensively used for various societal and scientific applications like Potential Fishing Zone (PFZ) identification, estimation of primary productivity, algal bloom detection and studying the coastal processes. Ocean primary production models use PAR at noon (Platt and Sathyendranath 1993) and daily averaged PAR (Behrenfeld and Falkowski 1997a) as a one of the main input. Two PAR products such as PAR at noon and daily averaged PAR have been estimated from OCM using two different methods.

4.2 OCEANSAT-1 & 2 OCM characteristics

Indian Space Research Organisation launched Ocean Colour Monitor (OCM) sensor on-board OCEANSAT-1 satellite in May 1999 to realize the importance of ocean colour measurements from the space. OCEANSAT-1 OCM was the first Indian ocean colour satellite and provides ocean colour measurements around the sea adjoining the Indian subcontinent. OCEANSAT-1 OCM had eight bands. The first six bands centered at 412 nm, 443 nm, 490 nm, 510 nm, 555 nm and 670 nm are used for analyzing ocean colour components in the water column. The last two bands of central wavelength at 765 nm and 865 nm are used for atmospheric correction. The spectral resolution of the first six bands and last two bands are 20 nm and 40 nm, respectively.
Table 4.1: Major specifications and features of OCEANSAT-1 & 2 OCM

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Specifications</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>OCEANSAT-1 OCM</td>
</tr>
<tr>
<td>Spectral Range (nm)</td>
<td>404-882</td>
</tr>
<tr>
<td>No. of Channels</td>
<td>8</td>
</tr>
<tr>
<td>Wavelength (nm)</td>
<td></td>
</tr>
<tr>
<td>Band 1: 404-423</td>
<td>Band 1 : 404-424</td>
</tr>
<tr>
<td>Band 3: 475-495</td>
<td>Band 3: 476-496</td>
</tr>
<tr>
<td>Band 4: 501-520</td>
<td>Band 4: 500-520</td>
</tr>
<tr>
<td>Band 5: 547-565</td>
<td>Band 5: 546-566</td>
</tr>
<tr>
<td>Band 6: 660-677</td>
<td>Band 6: 610-630</td>
</tr>
<tr>
<td>Band 7: 745-785</td>
<td>Band 7: 725-755</td>
</tr>
<tr>
<td>Band 8: 845-885</td>
<td>Band 8: 845-885</td>
</tr>
<tr>
<td>Satellite altitude (km)</td>
<td>720</td>
</tr>
<tr>
<td>Spatial Resolution (m)</td>
<td>360×236</td>
</tr>
<tr>
<td>Swath (km)</td>
<td>1420</td>
</tr>
<tr>
<td>Repetitive (days)</td>
<td>2</td>
</tr>
<tr>
<td>Quantisation</td>
<td>12 bits</td>
</tr>
<tr>
<td>Equatorial crossing time</td>
<td>12 noon</td>
</tr>
<tr>
<td>Along Track steering</td>
<td>±20°</td>
</tr>
<tr>
<td>(to avoid sunglint)</td>
<td></td>
</tr>
<tr>
<td>Data acquisition modes</td>
<td>Local Area Coverage (LAC)</td>
</tr>
</tbody>
</table>

OCEANSAT-2 spacecraft of Indian Space Research Organization (ISRO) is the second satellite in ocean series, which was successfully launched on September 23, 2009. OCEANSAT-2 OCM is the continuity satellite of OCEANSAT-1 OCM. OCEANSAT-2 OCM satellite carried three main instruments namely i) Ku band pencil beam scatterometer, ii) modified Ocean Colour
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Monitor (OCM) and iii) Radio Occultation Sounder of Atmosphere (ROSA) instrument of Italian Space Agency (ASI). Spectral bands of OCEANSAT-2 OCM are almost identical to OCEANSAT-1 OCM. However, central wavelengths of two spectral bands i.e band 6 and 7 have been shifted. The spectral band of central wavelength 670 nm in OCEANSAT-1 OCM has been shifted to 620 nm in OCEANSAT-2 OCM to improve suspended sediments quantification. Another spectral band 7 centered at 765 nm in OCEANSAT-1 OCM has been shifted to 740 nm to avoid oxygen absorption in OCEANSAT-2 OCM. Table 4.1 provides the technical details of the OCEANSAT-1 & 2 OCM instrument.

Over the Indian Ocean region covering Bay of Bengal and Arabian Sea OCM (LAC) data was acquired by NRSC ground station. The OCEANSAT-1 OCM (1999-2010) and OCEANSAT-2 OCM (2009 to present) data are archived at NRSC (National Remote Sensing Centre, Hyderabad) for use in various ocean colour applications. Data products are available as standard RAD (radiometrically corrected) and GEO (radiometrically and geometrically corrected) products.

The raw data of OCEANSAT-1 OCM are provided in two-byte generic binary integer format. The header file of the OCEANSAT-1 OCM data contains information about the geographic grid and solar and sensor viewing geometry and calibration coefficients. The geographic control point (GCP) information for projecting the OCEANSAT-1 OCM image data onto a geographic map projection as well as the information on the solar and sensor viewing geometry such as sun zenith angle ($\theta_s$), sun azimuth angle ($\phi_s$), sensor zenith angle ($\theta_v$), and sensor azimuth angle ($\phi_v$) from OCEANSAT-1 OCM data CD are obtained by using an in-house developed program.

OCEANSAT-2 OCM data products are available in the HDF 4.0 format providing all the required ancillary information such as sun zenith angle ($\theta_s$), sun azimuth angle ($\phi_s$), sensor zenith angle ($\theta_v$), and sensor azimuth angle ($\phi_v$). The data products from OCEANSAT-2 OCM are available at 360 meter spatial resolution for regional studies, which are also called local area...
coverage (LAC) products. The global area coverage (GAC) products are available at 1 km spatial resolution for global studies. The Level 1B top of the atmosphere (TOA) radiance data from all the eight bands of OCM sensor is used along with the ancillary information to generate various bio-geophysical data products.

Standard products of OCM are provided as Path/Row products based on a referencing scheme, which is a method for convenient geographic location of areas on Earth. This scheme is designated by Path and Rows based on the nominal orbital characteristics. The ground trace of a satellite’s orbit in space is called a ‘PATH’. Along a path, the continuous stream of data is segmented into a number of scenes. The lines joining the corresponding scene centers of different paths are parallel to the equator and are called ‘ROWS’. The region of Arabian Sea (55°E-78°E and 6°N - 28°N) is covered by OCM orbital paths 8 and 9 and rows 13 and 14. In the present study, OCM data of eastern Arabian Sea (path 9; rows 13 & 14; 6°N-25°N and 68°E-78°E) were obtained in band-separated band-sequential (BSQ) binary format on CD-ROM media.

4.3 Sun glitter estimation in OCEANSAT-1 & 2 OCM images

In any satellite images, sunglint regions are observed when there is direct reflectance or specular reflectance of the incoming solar radiation from the ocean surface to the sensor. Specular reflection is the mirror like reflection of light from a surface, in which light from a single incoming direction is reflected into a single outgoing direction. The sunglint occurs at one point when zenith angle of Sun and satellite are same and their azimuth angles are opposite in absolutely flat ocean surface. However, ocean surface is never flat. Surface roughness generated because of wind enlarges the sunglint area (Mohan and Chauhan, 2001). The shape of the sunglint area is changing with the change of orientation of the sensor. For a given tilt angle of the sensor, sunglint area moves from the north to south or south to north with the change of solar declination angle. If sunglint region has been not masked, it will be treated as cloudy region and misleading PAR will be estimated under sunglint condition.
The OCEANSAT-1 & 2 OCM was designed to tilt the sensor operationally ± 20° (Mohan and Chauhan, 2001) away from the nadir to minimize sun glint affect. However, it has been observed that even with the scheme proposed by Mohan and Chauhan (2001) some OCEANSAT-1 & 2 OCM data shows sun-glint in the months of April and August. Sun glint area has been masked by a method proposed by McClain and Yeh (1994). The probability of a pixel affected by sunglint is function of sea surface wind speed $W$, solar azimuth ($\phi_s$), solar zenith angle ($\theta_s$), satellite azimuth ($\phi_v$) and satellite zenith angle ($\theta_v$). A probability parameter $P_\sigma$ (McClain and Yeh, 1994) is defined by

$$P_\sigma = \frac{1}{\pi \sigma^2} \exp \left( -\frac{\tan^2 \theta_n}{\sigma^2} \right)$$

(4.1)

$\sigma^2$ is the mean square surface slope distribution. $\sigma^2$ increases linearly with wind speed.

$$\sigma^2 = 0.003 + 0.00512W$$

(4.2)

$\theta_n$ is the vector normal to the surface vector for which sunglint will be observed. $\theta_n$ can be derived from the surface reflection angle, $\omega$.

$$\theta_n = \cos^{-1} \left[ \frac{\cos \theta_s \cos \theta_v}{2 \cos \omega} \right]$$

(4.3)

$$\cos 2\omega = \cos \theta_s \cos \theta_v + \sin \theta_s \sin \theta_v \sin (\phi_v - \phi_s)$$

(4.4)

Probability parameter greater than 1.5 (McClain and Yeh, 1994) has been chosen as sunglint affected region and it has been masked. Figure 4.1 shows false colour composite (FCC) image of OCEANSAT-1 OCM in the Arabian Sea dated on 18th April, 2006. Sunglint region has been estimated using equation (4.1). Masked elongated region oriented towards north-south direction in the Figure 4.1 shows sunglint affected area in the OCEANSAT-1 OCM and it has been masked.
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Figure 4.1: Sunglint region showed in false colour composite (FCC) image of OCEANSAT-1 OCM data over the Arabian Sea (18th April, 2006).

4.4 Distinction between clear and cloudy sky in OCEANSAT-1 & 2 OCM

Almost seventy percent of the Earth’s surface is covered by clouds (Rossow and Schiffer, 1999). Clouds are generally characterized by higher reflectance and lower temperature (Ackerman, et al., 1998) compared to the underlying surface. Based on the properties, cloudy sky is distinguished from the clear sky from space using threshold value of reflectance estimated at visible band and brightness temperature estimated at infrared band. However, the type of the underlying surface of the clouds modifies the strength of the signal detected at the sensor.
Different types of the clouds have different radiative properties and show various contrast with the underlying surface. Specially, thin cirrus clouds, low stratus at night and small cumulus are difficult to detect from space observation as they show insufficient contrast with the surface radiance (Ackerman, et al., 1998). Further field of view of sensor will not always be completely cloudy or clear at the cloud edge and it creates difficulty to detect cloudy sky at the edge of the clouds. There had been several algorithms developed to mask different types of cloud in MODIS, NOAA AVHRR (advanced very high resolution radiometer) and International Satellite Cloud Climatology Project (ISCCP) (Saunders and Kriebel, 1988; Gesell, 1989; Rossow, 1989; Seze and Rossow, 1991; Rossow and Schiffer, 1991; Rossow and Garder, 1993a; Rossow and Garder, 1993b; Ackerman, et al., 1998; Ackerman, et al., 2006) for different underlying surface such as water, snow, land vegetation and bare soil. Based on the previous work on masking of cloudy pixels from clear pixel, Table 4.2 lists spectral bands and algorithms used to separate cloudy sky from clear sky over water surface.

Table 4.2: Spectral bands and different cloud detection test for clouds over water from space observation (Ackerman, et al., 1998; Ackerman, et al., 2006)

<table>
<thead>
<tr>
<th>Type of cloud</th>
<th>Bands</th>
<th>Cloud detection test</th>
</tr>
</thead>
<tbody>
<tr>
<td>Low cloud over water</td>
<td>i. 870 nm</td>
<td>i. R&lt;sub&gt;870&lt;/sub&gt;</td>
</tr>
<tr>
<td></td>
<td>ii. 670 nm</td>
<td>ii. R&lt;sub&gt;870/670&lt;/sub&gt;</td>
</tr>
<tr>
<td></td>
<td>iii. 11000 nm</td>
<td>iii. BT&lt;sub&gt;11000-3700&lt;/sub&gt;</td>
</tr>
<tr>
<td></td>
<td>iv. 3700nm</td>
<td></td>
</tr>
</tbody>
</table>

| High Thick cloud over water | i. 1380nm | i. R<sub>1380</sub> |
|                            | ii. 870nm | ii. R<sub>870</sub> |
|                            | iii. 670nm | iii. R<sub>870/670</sub> |

| High Thin cloud over water | i. 1380nm | i. R<sub>1380</sub> |

BT= Brightness temperature
R = Reflectance
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OCEANSAT-1 & 2 OCM does not have infrared spectral bands such as 1380 nm, 3700 nm and 11000 nm. Reflection at 865 nm and 670 nm for OCEANSAT-1 OCM and 620 nm for OCEANSAT-2 OCM has been estimated. Figure 4.2 shows false colour composite, reflectance at 865 nm and reflectance ratio between 865 nm and 670 nm spectral bands for OCEANSAT-1 OCM.

Figure 4.2: False colour composite, reflectance at 865 nm and reflectance ratio between 865 nm and 670 nm band for OCEANSAT OCM-1 dated on 16th November 2001.

Figure 4.2 shows that reflectance at 865 nm band is below 0.06 under clear sky. A threshold value of 0.055 reflectance at 865 nm has been taken to screen out cloud mask for MODIS image (Ackerman et al., 2006). Chauhan et al. (2002) have used 1.1% albedo at 865 nm band to screen out clouds, land and sunglint area in OCEANSAT-1 OCM. For thick clouds reflectance at 865 nm is very high and it is greater than 0.2 (Figure 4.2). However, the reflectance at 865 nm for thin clouds is highly variable and the ranges are between 0.06 to 0.2 (Figure 4.2).

The reflectance ratio between 865 nm and 670 nm is observed below 0.6 for clear sky (Figure 4.2) condition. For thick clouds, the ratio is very high and it is greater than 1.05 (Figure
4.2). For thin clouds, the ratio is greater than 0.78 (Figure 4.2). In this study, a threshold value of reflectance 0.06 at 870 nm and ratio of reflectance at 870 nm and 670 nm greater than 0.9 has been used to distinguish between cloudy sky from clear sky of OCEANSAT-1 OCM images. In MODIS cloud masking algorithm, ratio between the reflectance at 865 nm and 660 nm has been found to be 0.9 to 1.1 in cloudy region (Ackerman et al, 1998). In NOAA- AVHRR, if the reflectance ratio between spectral channel 2 (720-1100 nm) to 1 (580-680 nm) is found in between 0.7 and 1.1, and then the pixel was declared as cloudy pixels (Saunders and Kriebel 1988). For cloud-free ocean, the ratio is expected to be less than 0.75 (Saunders and Kriebel 1988). So the application of cloud masking algorithm is of importance.

4.5 PAR estimation from OCEANSAT-1 & 2 OCM using method I

PAR has been estimated from OCM using method I which is based on a combination of clear sky irradiance equations given in Iqbal (1983), Bird (1984) and the broadband cloud reflection algorithm given in Stephens et al. (1984). The pixels have been separated first for clear sky or cloudy sky in OCM data as mentioned in section 4.4. For clear pixels, PAR has been estimated by the equations developed for clear sky condition and for cloudy pixels PAR has been estimated by the equations developed for cloudy sky conditions (Van Laake and Azofeifa, 2004). In this radiative transfer model, the atmosphere is treated as a single layer in clear sky conditions, or as a double layer in cloudy conditions i.e., a layer above the cloud top and a layer from the cloud top downwards. Surface reflectance is not included in this first method.

4.5.1 Modeling PAR under clear sky condition

Total PAR is the summation of direct PAR \( I_{dir} \) and diffuse PAR \( I_{diff} \). Direct PAR is modeled with extraterrestrial solar irradiance and transmittance of the atmosphere. Direct PAR \( I_{dir} \) is
expressed with the equation (4.5) (Bird, 1984; Bird et al., 1986; Carder et al., 2003; Van Laake and Azofeifa, 2004)

\[ I_{dir\lambda} = I_0 T \] \hspace{1cm} (4.5)

\( I_0 \) is the extraterrestrial solar irradiance (W m\(^{-2}\) \(\mu\)m\(^{-1}\)) at the top of the atmosphere; \( T \) is transmittance of the atmosphere. \( I_0 \) varies slightly throughout the year because of the eccentric path of the Earth around the Sun.

\[ I_0 = E_0 \left(1 + 0.0344\cos \left(\frac{360N}{365}\right)\right) \] \hspace{1cm} (4.6)

\( E_0 \) is the extraterrestrial solar irradiance (W m\(^{-2}\) \(\mu\)m\(^{-1}\)) without correction for the Earth Sun distance. \( N \) is the day number.

Transmittance of the atmosphere is a dimensionless quantity. It depends on the wavelength. The transmittance of the atmosphere \( T \) is decomposed into transmittance of different atmospheric constituents such as gas molecules, ozone, aerosol etc. The transmittance of the atmosphere is expresses by the equation (4.7) (Bird, 1984; Bird et al., 1986; Carder et al., 2003; Van Laake and Azofeifa, 2004)

\[ T = T_R T_{OZ} T_A T_u \] \hspace{1cm} (4.7)

In the equation (4.7) \( T_R \) is transmittance of Rayleigh scattering (R), \( T_{OZ} \) is transmittance of ozone (oz) absorption, \( T_A \) is transmittance of aerosol scattering and \( T_u \) is transmittance of uniformly mixed gas (u) absorption. The absorption by water vapour in the PAR range is negligible (Eck and Dye, 1991). Transmittances depend on the concentration of the attenuating element in the atmosphere and pressure corrected air mass.
When considering absorption and scattering within the atmosphere it is always necessary to know the total mass of the absorbing or scattering substance. Absolute optical air mass $m_a$ is defined as (Paltridge and Platt, 1976)

$$m_a(h) = \int \rho \, ds$$

Where $\rho$ is the density of absorbing or scattering substance and $ds$ is the geometrical path element along the solar beam. The integration is taken at a perpendicular height $h$ above the ground surface to the top of the atmosphere along the beam. Further, the mathematics of scattering and absorption involve the approximation of a plane-parallel atmosphere where the optical path between two levels along some zenith angle is related to the vertical optical path between the same levels by the multiplying factor sec of solar zenith angle. Again when referring to the direct solar beam, relative optical air mass $m_r$ is defined formally as

$$m_r(h) = \int_{h_0}^{\infty} \frac{\rho \, ds}{\int_{h_0}^{\infty} \rho \, dh}$$

If $m_p$ is the relative optical mass corrected for local pressure and $m_0$ is the relative optical mass at mean sea level, then $m_p$ relates with $m_0$ by the equation (4.10) (Bird, 1984; Gregg and Carder, 1990; Van Laake and Azofeifa, 2004).

$$m_p = m_0 \frac{P_z}{1013.25}$$

If the local pressure $P_z$ in hPa equals to the pressure at mean sea level 1013.25 hPa then $m_p = m_0$. Relative optical mass $m_0$ varies with the solar zenith angle and has been estimated with the equation (4.11) and equation (4.12) (Gregg and Carder, 1990; Van Laake and Azofeifa, 2004).
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\[ m_0 = \frac{1}{\cos \theta_s}, \quad \theta_s \leq 60^\circ \]  
\[ m_0 = \frac{1}{\cos \theta_s + 0.15(93.885 - \theta_s)^{-1.253}}, \quad \theta_s > 60^\circ \]

(4.11)

(4.12)

\( \theta_s \) is the solar zenith angle.

The fraction of the direct beam irradiance that is scattered by the atmosphere is available as diffuse irradiance. It is much more difficult to model diffuse irradiance with any confidence like direct irradiance (Bird 1984). Brine and Iqbal (1982) first estimated diffuse irradiance based on the broadband method of Davis and Hays (1980). Bird (1984), Bird et al. (1986) estimated diffuse irradiance using Brine and Iqbal (1982) equations. In clear sky condition, diffuse PAR mainly originates from Rayleigh scattering and aerosol scattering if surface reflectance is neglected. Diffuse PAR of Rayleigh scattering origin is modeled with the assumption that 50\% of the total Rayleigh scattering is scattered towards the surface (Bird 1984; Van Laake and Azofeifa, 2004). The aerosol scattered diffuse PAR is modeled with single scattering albedo of aerosol and forward scatterance (Van Laake and Azofeifa, 2004). The forward scatterance depends on the solar zenith angle. Van Laake and Azofeifa (2004) developed equation for forward scatterance based on the data given in Iqbal (1983). Forward scatterance \( F \) can be expressed by the equation (4.13) (Van Laake and Azofeifa, 2004).

\[ F = 0.9302 \cos \theta_s^{0.2556} \]

(4.13)

Under the assumption of single scattering albedo and neglecting surface-reflected radiation, diffuse PAR \( I_{\text{diff}} \) (W m\(^{-2}\) μm\(^{-1}\)) has been estimated by the equation (4.14) (Van Laake and Azofeifa, 2004).

\[ I_{\text{diff}} = I_0 \cos \theta_s T_{oZ}([0.5T_A(1 - T_R)] + [F \omega_0 T_R(1 - T_A)]) \]

(4.14)
The first term of the equation (4.14) in square brackets, refers to the diffuse radiation originating from Rayleigh direct beam scattering and the second term to aerosol direct beam scattering. The total PAR at each wavelength $I_{s\lambda}$ is now the sum of the direct PAR ($I_{dir\lambda}$) and diffuse PAR ($I_{diff\lambda}$)

$$I_{s\lambda} = I_{dir\lambda} \cos \theta_s + I_{diff\lambda}$$ \hspace{1cm} (4.15)

Extraterrestrial solar irradiance from 400 nm to 700 nm has been divided into small wavelength intervals. Under clear sky condition, instantaneous PAR ($I_{par\_clearsky}$) has been estimated by summing $I_{s\lambda}$ at each wavelength interval.

$$I_{par\_clearsky} = \sum_{400 \text{ nm}}^{700 \text{ nm}} I_{s\lambda} \Delta \lambda$$ \hspace{1cm} (4.16)

4.5.2 Inputs for PAR model under clear sky condition

4.5.2.1 Extraterrestrial solar irradiance

Extraterrestrial solar irradiance data at the top of the atmosphere from 400 nm to 700 nm at 1 nm resolution is obtained from Thullier et al. (1998). Figure 4.3 shows the variation of extraterrestrial solar irradiance at the top of the atmosphere with wavelength. The mean absolute uncertainty of the measured solar irradiance data of Thullier et al. (1998) is from 2% to 3%. Subsequently, this data has been subdivided into 15 wavelength band at 20 nm spectral intervals with central wavelength starting from 410 nm to 690 nm. $I_0$ at each wavelength band has been obtained by using trapezoidal method of integration.
4.5.2.2 Rayleigh scattering

Transmittance of Rayleigh scattering is computed by using equation (4.17) (Iqbal 1983, Van Laake and Azofeifa, 2004).

\[ T_R = \exp[-0.008735\lambda^{-4.08}m_0] \] \hspace{1cm} (4.17)

4.5.2.3 Ozone absorption

Transmittance by ozone absorption is estimated using equation (4.18) (Iqbal 1983, Van Laake and Azofeifa, 2004).
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\[ T_{oz} = \exp[-k_{oz}lm_0] \] \hspace{1cm} (4.18)

In the equation (4.18), ozone thickness is \( l \) and \( k_{oz} \) is ozone absorption coefficient. The value of ozone absorption coefficient is obtained from Nicolet (1981). Nicolet (1981) estimated ozone absorption coefficient at 1 nm wavelength resolution from 380 nm to 1110 nm. At each wavelength band, ozone absorption coefficient has been obtained by averaging ozone absorption coefficient at each wavelength band from 400 nm to 700 nm region.

Ozone thickness has been obtained from Ozone monthly data measurements made by the Earth Probe Total Ozone Mapping Spectrometer (EP/TOMS). The EP/TOMS experiment provides measurements of Earth’s total column ozone by measuring the backscattered Earth radiance in the six 1-nm bands (308.60 nm, 313.50 nm, 317.50 nm, 322.30 nm, 331.20 nm, 360.40 nm) (McPeters et al., 1998). Retrieval of total ozone is done by using radiative transfer calculations to generate a table of backscattered radiance as a function of total ozone, viewing geometry, surface pressure, surface reflectivity and latitude. Level-3 ozone monthly data is available from the Goddard Space Flight Center (GSFC) in the form of Hierarchical Data Format (HDF) (ftp://jwocky.gsfc.nasa.gov/pub/eptoms). Level–3 product contains global total ozone on a fixed 1-degree latitude by 1.25-degree longitude grid.

4.5.2.4 Uniformly mixed gas absorption

The uniformly mixed gases has been computed by uniformly mixed gas absorption (Leckner 1978)

\[ T_u = \exp\left[-1.41 \frac{a_u m_0}{(1 + 118.93 a_u m_0)^{0.45}}\right] \] \hspace{1cm} (4.19)

\( a_u \) is the absorption coefficient and obtained from Bird (1984). In this PAR model, wavelength interval, integrated extraterrestrial solar irradiance at the top of the atmosphere, ozone absorption coefficient and absorption coefficient for mixed gases are listed in the Table 4.3.
Table 4.3: Extraterrestrial solar irradiance at the top of the atmosphere, absorption coefficient of ozone ($k_{oz}$), absorption coefficient for mixed gas ($a_u$) at each wavelength band used in the PAR model.

<table>
<thead>
<tr>
<th>Wavelength interval (nm)</th>
<th>Central wavelength (nm)</th>
<th>Integrated extraterrestrial solar irradiance at the top of the atmosphere (W m$^{-2}$)</th>
<th>Ozone absorption coefficient $k_{oz}$ (cm$^{-1}$)</th>
<th>Absorption coefficient for mixed gas $a_u$</th>
</tr>
</thead>
<tbody>
<tr>
<td>400-420</td>
<td>410</td>
<td>35.216</td>
<td>0.000769754</td>
<td>0</td>
</tr>
<tr>
<td>420-440</td>
<td>430</td>
<td>32.335</td>
<td>0.002042187</td>
<td>0</td>
</tr>
<tr>
<td>440-460</td>
<td>450</td>
<td>38.987</td>
<td>0.005253237</td>
<td>0</td>
</tr>
<tr>
<td>460-480</td>
<td>470</td>
<td>39.754</td>
<td>0.012178648</td>
<td>0</td>
</tr>
<tr>
<td>480-500</td>
<td>490</td>
<td>37.927</td>
<td>0.023915919</td>
<td>0</td>
</tr>
<tr>
<td>500-520</td>
<td>510</td>
<td>37.983</td>
<td>0.042220376</td>
<td>0</td>
</tr>
<tr>
<td>520-540</td>
<td>530</td>
<td>37.654</td>
<td>0.064968176</td>
<td>0</td>
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<td>540-560</td>
<td>550</td>
<td>35.646</td>
<td>0.087357776</td>
<td>0</td>
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<td>560-580</td>
<td>570</td>
<td>36.406</td>
<td>0.120187333</td>
<td>0</td>
</tr>
<tr>
<td>580-600</td>
<td>590</td>
<td>35.832</td>
<td>0.121824857</td>
<td>0</td>
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<tr>
<td>600-620</td>
<td>610</td>
<td>36.019</td>
<td>0.120865286</td>
<td>0</td>
</tr>
<tr>
<td>620-640</td>
<td>630</td>
<td>33.255</td>
<td>0.090607429</td>
<td>0</td>
</tr>
<tr>
<td>640-660</td>
<td>650</td>
<td>33.255</td>
<td>0.087357776</td>
<td>0</td>
</tr>
<tr>
<td>660-680</td>
<td>670</td>
<td>32.182</td>
<td>0.045572414</td>
<td>0</td>
</tr>
<tr>
<td>680-700</td>
<td>690</td>
<td>31.091</td>
<td>0.030667357</td>
<td>0.03</td>
</tr>
</tbody>
</table>
4.5.2.5 Aerosol transmission from OCM

Aerosol transmission is modeled with the spectrally varied aerosol optical depth (AOD) and relative optical mass at mean sea level. Aerosol transmission is computed by using equation (4.20) (Carder et al., 2003).

\[ T_a = \exp(-m_0 \tau_a) \]  

(4.20)

\( \tau_a \) is aerosol optical depth. According to Angstrom (1964), the spectral variation of aerosol optical depth is expressed by the equation (4.21).

\[ \tau_a \propto (\lambda)^{-\alpha} \]  

(4.21)

\( \alpha \) is known as angstrom coefficient. \( \alpha \) depends on the size distribution of the aerosol.

AOD has been estimated from OCM at 865 nm wavelength. Same methodology developed by Chauhan et al. (2009) to estimate aerosol optical depth from OCEANSAT-1 OCM has been followed. AOD estimation from satellite observation is based on the principle that for case I water for the wavelength greater than 700 nm, open ocean water absorbs strongly and water leaving radiance detected at the satellite is the contribution of Rayleigh and aerosol scattering. Thus, at the top of the atmosphere in the wavelength greater than 700 nm, the radiance detected by a satellite sensor \( L_{\lambda} \) is summation of Rayleigh \( L_{r\lambda} \) and aerosol path radiance \( L_{a\lambda} \) (Doerffer, 1992).

\[ L_{\lambda}(\lambda) = L_{a\lambda}(\lambda) + L_{r\lambda}(\lambda) \]  

(4.22)

Where

\[ L_{a\lambda} = \frac{F_{0\lambda} \omega_{0a} \tau_{a\lambda} P_a}{4\pi \cos \theta_v} = \text{aerosol path radiance} \]  

(4.23)
\( L_{r,\lambda} = \frac{F_{0,\lambda} \omega_0 \tau_0 \omega_\lambda}{4\pi \cos \theta_v} \) = Rayleigh path radiance .......................................................... (4.24)

\( \omega_{0a/0r} = \frac{\text{aerosol/Rayleigh single scattering albedo.}}{} \)

\( \tau_{a/\lambda} = \frac{\text{aerosol/Rayleigh optical depth at spectral band } \lambda.}{\text{ }} \)

\( p_{a/\lambda} = \text{Function related to aerosol/Rayleigh scattering phase function.} \)

\( F_{0,\lambda} \) is the extraterrestrial solar flux at the spectral band \( \lambda \) of OCEANSAT-1 & 2 OCM satellite. \( \theta_v \) is the satellite view angle. Assuming \( \omega_{0a} \approx 1 \) and \( \omega_{0r} \approx 1 \) for marine aerosols (Chauhan et al., 2009), the aerosol optical depth is estimated using equation (4.25).

\[ \tau_a = \left[ \frac{(L_t - L_r) 4\pi \cos \theta_v}{F_0 \omega_{0a} p_{a/\lambda}} \right] \] .......................................................... (4.25)

Rayleigh phase function is estimated using equation (4.26) (Doerffer, 1992).

\[ p_r(\gamma^\pm) = \frac{3}{4} [1 + \cos^2(\gamma_a^\pm)] \] .......................................................... (4.26)

The forward/backward scattering angle of aerosol is \( \gamma_a^\pm \) and it is computed using equation (4.27)

\[ \cos \gamma_a^\pm = \pm (\cos \theta_v \cos \theta_s - \sin \theta_v \sin \theta_s \cos \phi) \] .......................................................... (4.27)

\( \phi \) is relative azimuth angle. Das et al. (2002) have shown that the aerosol phase function can be approximated by the two term Henyey-Greenstein phase function of the following form

\[ p_a(\gamma^\pm) = A f(\gamma^\pm, g_1) + (1 - A) f(\gamma^\pm, g_2) \] .......................................................... (4.28)
Modeling PAR from satellite data

\[ f(\gamma^\pm, g) = \frac{1 - g^2}{(1 + g^2 - 2g \cos \gamma^\pm)^2} \]  \hspace{1cm} (4.29)

With \( A=0.985, g_1=0.8, g_2=0.5 \) for marine aerosols (Doerffer, 1992). Figure 4.4 shows the spatial variation of aerosol optical depth at 865 nm band and angstrom exponent estimated from OCEANSAT-1 OCM in the Arabian Sea.

![AOD at 865 nm](image1)

![Angstrom coefficient](image2)

**Figure 4.4: Aerosol optical depth estimated at 865 nm band and angstrom coefficient of OCEANSAT-1 OCM (Date: 8 November, 2001).**

In the open ocean variation of AOD at 865 nm band was from 0.05 to 0.3. The retrieval range of AOD from ocean colour remote sensing in the open ocean is from 0.01 to 0.3 (IOCCG, 2012). The high aerosol loading around the cloud patches is because of high reflectance at the edge of the cloud. Correction at the edge of the cloud has not done at the present model. The
variation of alpha is mainly 0.8 to 1.3 in the open ocean. OCEANSAT-1 OCM estimated AOD has been compared with limited *in-situ* measured AOD during November 2001 and January 2003. *In-situ* measurement of AOD by Sunphotometer has been discussed in Chapter 3 (Page No. 56). Figure 4.5 shows the comparison between OCEANSAT-1 OCM estimated AOD at 865 nm and *in-situ* measured AOD (865 nm). Figure 4.5 shows a good correlation between OCEANSAT-1 OCM estimated AOD (865 nm) and *in-situ* measured AOD (865 nm).

Figure 4.5: Comparison between OCEANSAT-1 OCM estimated AOD and *in-situ* measured AOD during November 2001 and January 2003 at oceanic locations.

Alpha has been estimated using a ratio at wavelength 865 nm and 765 nm. AOD at other wavelength has been estimated by angstrom relationship of AOD with wavelengths with the equation (4.30).

$$\tau_{a} = \tau_{865} \left( \frac{\lambda}{865} \right)^{-\alpha} \tag{4.30}$$

Aerosol transmission $T_a$ at each of fifteen wavelength band has been computed using equation (4.20).
4.5.3 Modeling PAR under cloudy sky condition

Clouds regulate the flow of radiant energy in the atmosphere through the process of scattering and absorption of shortwave and longwave radiation. They reflect incoming shortwave radiation to the space. Reflectance and transmittance of incoming solar radiation are both sensitive to optical thickness of the cloud (McBride et al., 2011). The optical thickness of a cloud involves the integration over cloud droplet size distribution \( n(r) \) and cloud depth \( z \) varying from zero to \( \Delta z \). Cloud optical depth is formally defined as (Stephens, 1978)

\[
\tau_c = \int_0^{\Delta z} \int_0^\infty n(r)Q_{\text{ext}}(x)\pi r^2 dr dz \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad (4.31)
\]

Where \( x = \frac{2\pi r}{\lambda} \), \( Q_{\text{ext}} \) is efficiency factor for extinction, \( r \) is the radius of the cloud droplets. Effective radius is an area weighted mean radius of the cloud droplets. Traditionally, effective radius is defined by the equation (4.32) (Stephens, 1978).

\[
r_e = \frac{\int_0^\infty n(r)r^3 dr}{\int_0^\infty n(r)r^2 dr} \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad \cdots \quad (4.32)
\]

Figure 4.6 shows the variation of transmittance through the water cloud for different effective radius ranging from 5 \( \mu \text{m} \) to 25 \( \mu \text{m} \). Larger droplet size is associated with stronger forward scattering and thus larger transmittance. However, because droplet absorption increases with wavelength, the opposite effect dominates at wavelengths greater than about 1400 nm and larger droplet size leads to decrease transmittance. From the Figure 4.6 it is evident that transmittance through the water cloud for a particular droplet size is constant in the wavelength region from 400 nm to 700 nm (McBride et al., 2011).
In the wavelength region from 400 nm to 700 nm, Stephens et al. (1984) estimated cloud top reflectance $R_{θ}$ for unit single scattering albedo based on the solution of parameterization model of radiation profile in the extended water clouds.

$$R_{θ} = \frac{\beta_{θ} \tau_c / \cosθ}{(1 + \beta_{θ} \tau_c / \cosθ)}$$  \hspace{1cm} (4.33)$$

$τ_c$ is the optical thickness of the cloud, $β_{θ}$ is the backscattered fraction of incident radiation as a function of the solar zenith angle. $T_{θ}$ is the transmittance through the cloud and has been obtained from reflectance by using equation (4.34) (Stephens et al., 1984).

$$T_{θ} = 1 - R_{θ}$$  \hspace{1cm} (4.34)$$
Modeling PAR from satellite data

In the presence of clouds, direct ($I_{dir, top\_cloud}$) and diffuse ($I_{diff, top\_cloud}$) PAR at the top of the cloud is estimated using equation (4.5) and (4.14). Cloud top pressure has been used to compute transmittance of aerosol absorption, uniformly mixed gas and Rayleigh scattering under cloudy conditions instead of mean sea level pressure since in cloudy condition. In cloudy condition, $m_p$ is not equal to $m_0$. To compute PAR under cloudy condition following assumptions have been made.

i. All ozone absorption is assumed to occur above clouds.

ii. Cloud cover is homogeneous, plane parallel and spatially stationary.

iii. Satellite observations suggest that complex interactions occur between coexisting cloud and aerosol layers (Kaufman et al., 2005). Chauhan et al. (2009) showed that the variation of aerosol optical depth at 870 nm in the Arabian Sea was from 0.12 to 0.14 with average value 0.13. Fixed aerosol optical depth 0.13 at 865 nm wavelength was used to estimate PAR under cloudy condition.

The scattered radiation above the cloud top originates from Rayleigh and aerosol scattering have been estimated using equation (4.35).

$$I_{s\_top\_cloud} = I_{dir\_top\_cloud} \cos \theta_s + I_{diff\_top\_cloud}$$ ………………………… (4.35)

The radiation transmitted through the clouds is then used to compute the attenuation in the lower part of the atmosphere. Total PAR under cloudy condition have been computed using equation (4.36).

$$I_{par\_cloudy} = \sum_{400nm}^{700nm} I_{s\_top\_cloud} T_{\theta_s \Delta \lambda}$$ ………………………… (4.36)
4.5.4 Inputs for PAR model under cloudy sky condition

4.5.4.1 Cloud top pressure

Cloud top pressure has a moderate influence on the estimation (Van Laake and Azofeifa, 2004) of PAR. ISCCP (International Satellite cloud climatology Project) global average cloud top pressure has been estimated 574 hPa (Kokhanovsky et al., 2011). A fixed cloud top pressure 574 hPa was used to compute PAR under cloudy condition.

4.5.4.2 Backscattered fraction of incident radiation ($\beta_{\theta s}$)

The backscattering fraction is a function of solar zenith angle and cloud optical depth. Stephan et al. (1978) employed a detailed multiple scattering model to calculate a series of values of reflection, absorption and transmission for a number of model cloud types. These calculations were used to tune the value of $\beta_{\theta s}$. Values of $\beta_{\theta s}$ have been obtained from Stephens et al. (1984) for wavelength less than 750 nm.

4.5.4.3 Cloud optical depth (COD)

Cloud optical depth is estimated using a suitable model which use transmitted radiance data measured by ground-based observation or reflected radiance data measured by space-based observation. However, ground based observation data are not available in systematic way globally. Transmission-based algorithm use spectral irradiance (Min and Harrison 1996) or broadband irradiances (Leontyeva and Stamnes 1994; Dong et al., 1997; Boers 1997; Barker et al., 1998).
Cloud optical depth has been estimated from transmitted radiance data by ground-based observation at 415 nm, 440 nm, 675 nm, 870 nm wavelengths (Min and Harrison, 1996; Barnard and Long, 2004; Chiu et al., 2010). Barnard and Long (2004) use a simple empirical equation to calculate cloud optical thickness as a function of cosine of solar zenith angle, surface albedo, broadband diffuse irradiance and broadband clear sky total irradiance. The Aerosol Robotic Network (AERONET) is a ground-based network that is designed to measure microphysical and optical properties of aerosol at wavelengths of 440 nm, 675 nm, 870 nm and 1020 nm. In cloudy condition, radiance detected by AERONET radiometers has been used to estimate cloud optical depth (Chiu et al., 2010) at each wavelength.

To estimate cloud optical depth from space based observation, Nakajima and King (1990) showed that reflectance at 750 nm and 2160 nm can be used to estimate cloud optical thickness and mean effective radius. The measurements at 750 nm are preferred as it is free of considerable influence of atmospheric absorption due to gases or liquid water.

International Satellite Cloud Climatology Project provides cloud optical depth using radiance data from five geostationary such as GMS, METEOSAT, GOES-WEST, GOES-EAST, INSAT and two polar orbiting satellites such as NOAA-AVHRR-Afternoon, NOAA-AVHRR-morning (Rossow et al., 1996) for 280-km grid cell over the globe. Cloud optical thickness at 600 nm values retrieved using two different cloud microphysical models (Schiffer and Rossow, 1983; Schiffer and Rossow, 1985; Rossow and Schiffer, 1991; Rossow et al., 1996; Rossow and Schiffer, 1999). A liquid water droplet model with a water sphere size distribution described by a gamma distribution with effective mean radius 10 μm and effective variance 0.15 has been used. Second model is an ice crystal model with a random fractal crystal shape and power law size distribution from 20 μm to 50 μm, giving an effective radius of 30 μm and an effective variance of 0.10.

Cloud optical depth has been estimated from MODIS data based on asymptotic theory (King 1987; Nakajima and King, 1990; King et al., 1992; King et al., 1997). The wavelength are used to estimate cloud optical thickness are 645 nm over land surface, 858 nm over ocean surface.
and 1240 nm over snow surface. A combination of 645 nm, 1640 nm and 2130 nm bands was used for cloud thermodynamic phase function.

A polynomial approach is used to estimate cloud optical depth at 753 nm where cloud optical depth are related to a polynomial function of the MERIS radiance (Fischer et al., 2000). However, using ADEOS-POLDER radiance measurement, cloud optical depth has been estimated at 443 nm, 670 nm, 865 nm band (Parol et al., 2000).

Kokhanovsky et al. (2003) developed a simple semi-analytical model to estimate cloud optical depth. However, this algorithm is valid for optically thick clouds or clouds having cloud optical thickness greater than 5 m. The theory is based on the premise that reflectance estimated at visible (non-absorbing) wavelength is used to estimate cloud optical thickness and reflectance in near infrared and mid infrared is used to estimate effective radius of cloud droplet.

Using the equation (4.33), cloud optical depth (COD) from OCM for thick water clouds (COD > 4m) has been obtained from a semi-analytical model developed by Kokhanovsky et al. (2003) with input of OCM reflectance at 443 nm band. For thin clouds (COD < 4m), a quadratic relationship between COD and TOA radiance at 443 nm band of OCM was used as described in the section 4.5.4.3 (B).

A) Cloud optical depth estimation for thick clouds

Cloud optical depth has been estimated at 443 nm spectral band of OCM using a semi-analytical model developed by Kokhanovsky et al. (2003). The reflectance $R_{443}(\tau_c; \mu, \mu_0, \phi)$ at 443 nm band of OCM has been calculated using equation (4.37).

$$R_{443}(\tau_c; \mu, \mu_0, \phi) = \frac{\pi L_{443}(\tau_c = 0; \mu, \phi)}{\mu_0 F_{443}} \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots \ldots (4.37)$$
Here $L_{443}$ denotes TOA radiance at 443 nm band of OCM, $\mu = |\cos \theta_v|$ and $\mu_0 = |\cos \theta_s|$. $\theta_v$ is the satellite view angle, $\theta_s$ is the solar zenith angle, $\phi$ is the relative azimuth angle between solar and satellite directions. $F_{443}$ is the extra-terrestrial solar irradiance corresponding to the 443 nm spectral band of OCM sensor derived using band specific Relative Spectral Response (RSR) function of OCM band (Chauhan et al., 2002). Figure 4.7 shows the variation of reflectance estimated at 443 nm of OCM in the Arabian Sea under cloudy sky condition.

4.7: Reflectance estimated at 443 nm spectral band of OCEANSAT-1 OCM for cloudy pixels (16$^{th}$ November, 2001).

The variation of reflectance at 443 nm spectral band of OCEANSAT-1 OCM is 0.17 to 0.25 for very thin cloud. For thick cloud patches, reflectance value is less at the edge of the cloud and it is higher at the middle of the cloud patches.
For ideally white reflector, the value of reflectance should be one. However, the value of reflectance of clouds estimated from OCM is not equal to one. Reflectance can be smaller and larger than one depending on the incidence angle (Karlgard, 2008). This implies that, for a particular viewing geometry a cloud is even more reflective than ideally white reflector. This is mostly due to peculiarities of the scattering phase function of the cloud (Karlgard, 2008). Phase function $P(\zeta)$ of a cloud describes the probability of a photon being scattered in the direction of scattering angle ($\zeta$). For a given particle size and wavelength, the phase function $P(\zeta)$ can be derived using Mie solutions. Kokhanovsky et al. (2003) derived an approximate fifth order polynomial to estimate phase function using equation (4.38).

\[
P(\zeta) = Q e^{-\sigma} + \sum_{i=1}^{5} b_i e^{\beta_i (\zeta - \zeta_i)^2} \quad \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots (4.38)
\]

Scattering angle in radians is estimated using equation (4.39).

\[
\zeta = \arccos(-\cos\theta_s \cos\theta_s + \sin\theta_s \sin\theta_s \cos\phi) \quad \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots \cdots (4.39)
\]

Constants such as $Q, \sigma, b_i, \beta_i, \zeta_i$ in the equation (4.39) have been obtained from Kokhanovsky et al. (2003)

Table 4.4: Values of constants used to estimate phase function of clouds.

<table>
<thead>
<tr>
<th>$i$</th>
<th>$b_i$</th>
<th>$\beta_i$</th>
<th>$\zeta_i$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1744.0</td>
<td>1200.0</td>
<td>0.0</td>
</tr>
<tr>
<td>2</td>
<td>0.17</td>
<td>75.0</td>
<td>2.5</td>
</tr>
<tr>
<td>3</td>
<td>0.30</td>
<td>4826.0</td>
<td>$\pi$</td>
</tr>
<tr>
<td>4</td>
<td>0.20</td>
<td>50.0</td>
<td>$\pi$</td>
</tr>
<tr>
<td>5</td>
<td>0.15</td>
<td>1.0</td>
<td>$\pi$</td>
</tr>
</tbody>
</table>
In the case of conservative scattering i.e, single scattering albedo of cloud =1, reflectance can be expressed by the equation (4.40) (King, 1987; Nakajima and King, 1990; Kokhanovsky et al., 2003; Kokhanovsky and Nauss, 2006)

\[ R(\tau_c; \mu, \mu_0, \phi) = R_\infty^0(\mu, \mu_0, \phi) - \frac{t_c(r_e, w)(1 - A_s)K_0(\mu)K_0(\mu_0)}{1 - A_s(1 - t_c(r_e, w))} \] \hspace{1cm} (4.40)

\( R_\infty(\mu, \mu_0, \phi) \) is the reflectance of an idealized semi-infinite non-absorbing water cloud. \( R_\infty(\mu, \mu_0, \phi) \) can be represented by the simple approximate equation (4.41) (Kokhanovsky et al., 2003).

\[ R_\infty^0(\mu, \mu_0, \phi) = \frac{b_1 + b_2 \cos \theta_s \cos \theta_\nu + P(\zeta)}{4(\cos \theta_s + \cos \theta_\nu)} \] \hspace{1cm} (4.41)

\( b_1 \) and \( b_2 \) are constants. \( b_1=1.48, b_2=7.76 \) (Kokhanovsky et al., 2003). In the equation (4.41), \( K_0(\mu) \), \( K_0(\mu_0) \) are escape functions. Escape functions describes the angular distribution of light escaping a semi-infinite cloud from sources located deep inside the medium. Because multiple scattering dominates in the cloud, typical features of single scattering becomes less pronounced and escape functions can be well represented for isotropic scattering by the functions represented in the equation (4.42) and (4.43) (Kokhanovsky et al., 2003).

\[ K_0(\mu) = \frac{\sqrt{3}}{4} (1 + 2\mu) \] \hspace{1cm} (4.42)

\[ K_0(\mu_0) = \frac{\sqrt{3}}{4} (1 + 2\mu_0) \] \hspace{1cm} (4.43)

In the equation (4.40), \( t_c(r_e, w) \) is the diffused transmittance of a cloud. \( t_c(r_e, w) \) is governed by the effective radius \( r_e \) and the liquid water path \( w \). However, it can be expressed in terms of asymmetry parameter \( g(r_e) \) and optical thickness \( \tau_c \) (Kokhanovsky et al., 2003).
\[ t_c = \frac{1}{\alpha_c + \frac{3}{4} \tau_c (1 - g(r_e))} \]  \hspace{2cm} (4.44)

The value of \( \alpha_c \) is 1.072 for water clouds (Kokhanovsky et al., 2003). The optical thickness is expressed by the equation (4.45).

\[ \tau_c = \frac{4(t_c^{-1} - 1.072)}{3(1 - g(r_e))} \]  \hspace{2cm} (4.45)

Where \( t_c = \frac{1}{(R_{t_o} - R(t_c, \mu_0, \varphi) - A_t - A_s)} \)  \hspace{2cm} (4.46)

In the above equation, ocean surface albedo \( \frac{A_t}{1 - A_s} \) has been neglected. Asymmetry function \( g(r_e) \) is taken as 0.85 for water clouds (King, 1987).

**B) Cloud optical depth estimation for thin clouds**

Semi-analytical model to estimate PAR is valid for optically thick clouds or clouds, which are having cloud optical depth greater than 4 m. For very thin ice clouds of 34.3 µm effective radius, a quadratic relationship between cloud optical depth and TOA radiance at 443 nm band at OCM viewing geometry has been obtained from COART (Coupled Ocean Atmosphere Radiative Transfer) model (http://www-cave.larc.nasa.gov/cave/). Figure 4.8 shows the relationship between TOA radiance at 443 nm band of OCM and cloud optical depth obtained from COART model. Figure 4.9 shows the variation of cloud optical depth both for thick and thin cloud. Based on the analysis, the minimum value of COD is fixed at 2 m. Clouds are having particle sizes much larger than the wavelength of light and show extremely weak wavelength dependence (Hu et al., 1993), in the wavelength region from 400 nm to 700 nm. Cloud optical depth estimated at any non-absorbing wavelength, has been used at other wavelengths in the PAR wavelength range.
Modeling PAR from satellite data

\[ y = 0.01x^2 + 0.02x - 0.56 \]
\[ r^2 = 0.99 \]

Figure 4.8: Variation of TOA radiance at 443 nm of OCM with cloud optical depth obtained from COART model.

Figure 4.9: Cloud optical depth estimated at 443 nm band of OCEANSAT-1 OCM (16th November, 2001) for thick and thin clouds.
Figure 4.10 shows Flow chart to estimate PAR from OCEANSAT-1 & 2 OCM using method I.

Figure 4.10: Flow chart to estimate PAR from OCEANSAT-1 & 2 OCM.
4.5.5 Output of PAR model under clear and cloudy sky condition

Instantaneous total PAR $I_{par}$ in W m$^{-2}$ is calculated using equation (4.47) under both clear and cloudy condition.

$$I_{par} = \sum_{400nm}^{700nm} I_{s\lambda} \Delta \lambda \text{ for clear sky}$$

$$= \sum_{400nm}^{700nm} I_{s\lambda, top\_cloud} T_{\theta_2} \Delta \lambda \text{ for cloudy sky} \quad \text{...(4.47)}$$

Output of the PAR model is PAR estimated at noon in Watt m$^{-2}$ and daily (24 hour) averaged PAR in Einstein m$^{-2}$ day$^{-1}$. As equatorial crossing time of OCM is 12 noon, PAR at noon $I_{om}$ is calculated based on the state of the atmosphere when satellite data was acquired. Daily total PAR has been calculated using equation (4.48) based on a sinusoidal approximation of light and assuming the state of the atmosphere is not going to change for full day. This is an assumption as observations at other times are not available.

$$I_T = \int_0^D I_{om} \cdot \sin \left( t \cdot \frac{\pi i}{D} \right) \, dt$$

$$I_T = -\frac{I_{om}D}{pi} \left[ \cos \left( t \cdot \frac{\pi i}{D} \right) \right]_0^D$$

$$I_T = -\frac{I_{om}D}{pi} \left[ \cos pi - \cos 0 \right] = -\frac{I_{om}D}{pi} \left[ -1 - 1 \right] = \frac{2I_{om}D}{\pi} \quad \text{...} \quad \text{...}(4.48)$$

$I_T$ is daily total PAR, $I_{om}$ is PAR at noon and $D$ is daylength. Daylength has been calculated using equations given by Teets (2003). To estimate daily (24 hour) averaged PAR ($I_D$), $I_T$ is divided by 24 hours as done for SeaWiFS based calculations (Frouin et al., 2003). The unit of daily averaged PAR in W m$^{-2}$ was converted in Einstein m$^{-2}$ day$^{-1}$ (Dye 2004) as shown in the equation (4.49).

$$I_D = \frac{I_T}{24} = \frac{2 \times I_{om}D}{24 \times pi} = 0.0104 \times I_{om} \times D \quad \text{...} \quad \text{...}(4.49)$$
Figure 4.11: Variation of PAR at noon, daily total PAR and daily averaged PAR estimated from OCM over the Arabian Sea (Date 12 November, 2001).
In the equation (4.49) $I_{0m}$ is in W m$^{-2}$ and D is in hour, then daily averaged PAR $I_D$ is in Einstein m$^{-2}$ day$^{-1}$ unit. Figure 4.11 shows variation of PAR at noon (W m$^{-2}$), daily total PAR $I_T$ (W m$^{-2}$) and daily averaged PAR $I_D$ (Einstein m$^{-2}$ day$^{-1}$) estimated from OCM.

### 4.5.6 Sensitivity study of the PAR model (Method I)

A sensitivity analysis has been studied to understand the importance of different input parameters to estimate PAR. The default parameter values used in the sensitivity study are listed in Table 4.5.

**Table 4.5: Input default parameters for sensitivity study.**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Clear Sky</th>
<th>Cloudy Sky</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ozone concentration</td>
<td>250 DU*</td>
<td>250 DU*</td>
</tr>
<tr>
<td>Aerosol optical depth at 865 nm</td>
<td>0.15</td>
<td>0.13</td>
</tr>
<tr>
<td>Angstrom exponent</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>Cloud top pressure</td>
<td></td>
<td>574mb</td>
</tr>
<tr>
<td>Cloud optical depth</td>
<td></td>
<td>9.5</td>
</tr>
<tr>
<td>Backscattered fraction of incident radiation</td>
<td></td>
<td>0.0692</td>
</tr>
<tr>
<td>Latitude, Longitude</td>
<td>19.18°N, 66.73°E</td>
<td></td>
</tr>
<tr>
<td>Date, Time</td>
<td>10Jan2003, 12 p.m</td>
<td></td>
</tr>
</tbody>
</table>

*DU= Dobson unit

Figure 4.12 shows sensitivity study of PAR at noon with changes of month (a), ozone concentrations (b), aerosol optical depth (c) and angstrom coefficient (d) under clear sky condition. PAR has been estimated at every month using default parameters mentioned in the Table 4.5. To understand monthly variation of PAR, PAR estimated at each month has been
normalized to the average value of PAR during 12 months. Minimum value of PAR has been observed in December. In May month, maximum PAR has been observed (Figure 4.12 a). During April to August the variation of PAR is minimum and it is situated at the plateau region of the graph (Figure 4.12 a).

![Graphs showing sensitivity to various factors]

**Figure 4.12:** Sensitivity to a) monthly variation b) ozone concentration c) aerosol optical depth (AOD) d) angstrom coefficient ($\alpha$) for PAR estimation under clear sky.

PAR has been decreased linearly with increase of ozone concentration (Figure 4.12 b). With the change of 500 D.U ozone concentrations PAR has decreased maximum 4% compared to 0 ozone concentration. This indicates that PAR estimation is not so much sensitive to ozone concentration.
Modeling PAR from satellite data

Up to aerosol optical depth 1, PAR has been decreased exponentially with increasing aerosol optical depth (Figure 4.12 c). However, greater than one aerosol optical depth value, decrease of PAR is almost constant with the change of aerosol optical depth (Figure 4.12 c). PAR has decreased 7.8 % and 13.8% for aerosol optical depths 0.3 and 1, respectively compared to no aerosol loading. The variation of PAR is observed only ± 2% when angstrom coefficient varies from 0 to 1.8 compared to angstrom exponent 1 (Figure 4.12 d). Under clear sky condition, aerosol optical depth plays important role to attenuate PAR.

Figure 4.13 shows globally cloud classification in terms of cloud optical thickness and cloud top pressure obtained from International satellite Cloud Climatology Project (Rossow and Schiffer, 1999).

**Figure 4.13: Different cloud types in terms of cloud top pressure and cloud optical thickness (Rossow and Schiffer, 1999).**
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Sensitivity study has also done to understand variation of PAR in terms of cloud optical thickness and cloud top pressure for different types of clouds mentioned in the Figure 4.13. Figure 4.14 shows sensitivity study about the effect of cloud optical depth (a) and cloud top (b) on PAR estimation. PAR estimation is highly sensitive to cloud optical depth, particularly at lower value of cloud optical depth above 10 (Figure 4.14 a). PAR has been decreased by 5%, 18%, 44%, 68% and 85%, respectively for cloud optical depth 1.5, 3.5, 9.5, 23 and 60 compared to clear sky (Figure 4.14 a).

![a) Sensitivity to cloud optical depth](image)

![b) Sensitivity to cloud top pressure](image)

**Figure 4.14: Sensitivity to a) cloud optical depth and b) cloud top pressure for PAR estimation under cloudy sky condition**

A fixed cloud top pressure 574 mb has been taken to estimate PAR under cloudy condition. PAR value obtained at different cloud top pressure is normalized to PAR estimated at 574 mb and the variation of PAR/PAR\text{574} is shown in the Figure 4.14 b). The sensitivity results of the effect of cloud top pressure on PAR estimation shows that if cloud top pressure is higher than 574 mb, maximum 2.3% lower PAR is estimated compared to PAR estimated at 574 mb. Higher value of PAR such as 0.15%, 1.45%, 2.90%, 4.40%, 5.96% has been observed for cloud top 560 mb, 440 mb, 310 mb, 180 mb and 50 mb, respectively compared to 574 mb.
PAR has been estimated using second method according to Frouin et al. (2003) which assumes that the effects of clouds and clear atmosphere can be decoupled with the cloud system and surface albedo. PAR under both the clear and cloudy sky $I_{par}$ reaching at the ocean surface is then given by

$$ I_{par} = < F_0 > \cos(\theta_s) < T_d > < T_g > (1 - < A >)(1 - < S_a > < A >)^{-1} \left(1 - < S > < A >\right)^{-1} \\
$$

$$..........................(4.50)$$

$< >$ symbolizes average value over the PAR range. $\theta_s$ is the solar zenith angle, $T_d$ is the clear sky diffuse transmittance, $T_g$ is the gaseous transmittance, $S_a$ is the spherical albedo for molecular and aerosol scattering. $A_s$ is the ocean surface albedo and $A$ is the cloud albedo. In the clear sky condition, ocean surface albedo $A_s$ reduces to cloud albedo $A$. Absorption due to water vapour has been neglected in the method II, similar to method I.

$F_o$ is the extra-terrestrial solar irradiance corresponding to the six discrete spectral bands in the visible region of OCM sensor derived using band specific Relative Spectral Response (RSR) function of OCM bands. $F_o$ at each band of OCM is calculated with the equation

$$ F_o = \frac{\int_{?}^{\lambda_n} I_0(\lambda) S(\lambda) d\lambda}{\int_{?}^{\lambda_n} S(\lambda) d\lambda} \quad ...........................................(4.51)$$

Where $S(\lambda)$ is the response function of OCM, $\lambda$ is wavelength. It starts from $\lambda_o$ to $\lambda_n$ for different spectral band of OCM from band 1 to band 6 as mentioned in the Table 4.1. $F_o$ varies slightly throughout the year because of the eccentric path of the Earth around the Sun. $F_o$ has been corrected for the Earth Sun distance (Gregg and Carder, 1990).
Gaseous transmittance $T_g$ essentially due to ozone has been estimated from Frouin et al (2003).

$$T_g = \exp\left(-\frac{k_{oz} \times l}{\cos \theta_s}\right)$$

$\frac{k_{oz}}{\cos \theta_s}$ is ozone absorption coefficient and $l$ is ozone concentration. Table 4.6 shows values of $F_0$ and ozone absorption coefficient of each band starting from band 1 to band 6 of OCM used in the method II.

**Table 4.6: Extraterrestrial solar irradiance and ozone absorption coefficient at six spectral bands of OCM used in method II.**

<table>
<thead>
<tr>
<th>Spectral Band</th>
<th>(OCEANSAT-1 OCM)</th>
<th>(OCEANSAT-2 OCM)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$F_0$ ( mW cm$^{-2}$ μm$^{-1}$)</td>
<td>$k_{oz}$ (cm$^{-1}$)</td>
</tr>
<tr>
<td>Band 1</td>
<td>170.7943</td>
<td>0.0006014</td>
</tr>
<tr>
<td>Band 2</td>
<td>189.4438</td>
<td>0.002996</td>
</tr>
<tr>
<td>Band 3</td>
<td>193.6842</td>
<td>0.02297</td>
</tr>
<tr>
<td>Band 4</td>
<td>188.3675</td>
<td>0.04291</td>
</tr>
<tr>
<td>Band 5</td>
<td>185.3973</td>
<td>0.1011</td>
</tr>
<tr>
<td>Band 6</td>
<td>153.3877</td>
<td>0.1090</td>
</tr>
</tbody>
</table>

Average gaseous transmittance has been expressed by

$$< T_g > = \frac{\sum_i T_{gi} \times F_{0i}}{\sum_i F_{0i}}$$

$\frac{\sum_i T_{gi} \times F_{0i}}{\sum_i F_{0i}}$ ...............................(4.53)
i is the band numbers of OCM. i varies from 1 to 6.

Direct $T_{dir}$ and Diffuse transmittance $T_d$ has been computed using equation (4.54) and equation (4.55) (Tanre et al., 1979; Frouin et al., 2003).

\[ T_{dir} = \exp\left[-(\tau_{r\lambda} + \tau_{a\lambda})/\cos\theta_s\right] \] ... (4.54)

\[ T_d = \exp\left[-(\tau_{r\lambda} + \tau_{a\lambda})/\cos\theta_s\right] \times \exp\left[-(0.52\tau_{r\lambda} + 0.83\tau_{a\lambda})/\cos\theta_s\right] \] ... (4.55)

$\tau_{r\lambda}$, $\tau_{a\lambda}$ are Rayleigh optical depth and aerosol optical depth at six visible band of OCM. $\tau_{r\lambda}$ has been estimated from Bird (1984). $\tau_{a\lambda}$ has been estimated at each visible band from AOD at 865 nm band of OCM using angstrom relationship. AOD at 865 nm and angstrom coefficient estimated from OCM has been used in this method II as input.

Average direct and diffuse transmittance has been expressed by

\[ < T_{dir} > = \frac{\sum_i T_{diri} \times F_{0i}}{\sum_i F_{0i}} \] ... (4.56)

\[ < T_d > = \frac{\sum_i T_{di} \times F_{0i}}{\sum_i F_{0i}} \] ... (4.57)

Ocean surface albedo $A_s$ and cloud albedo $A$ has been estimated for clear and cloudy pixels for OCM. Albedo is the total reflectance of the surface integrated over all the angles of the upward hemisphere. Taylor et al. (1996) estimated ocean surface albedo based on aircraft measurement and proposed ocean surface albedo is a function of solar zenith angle. Froiun et al. (2003) incorporated the effect of atmospheric optical depth with solar zenith angle to estimate ocean surface albedo. Average ocean surface albedo has been calculated using equation (4.58) (Frouin et al., 2003).
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\[
< A_s > = \frac{< T_{dir} >}{< T_d >} \left[ \frac{0.05}{1.1(\cos \theta_s)^{1.4} + 0.15} \right] + 0.08 \left( \frac{1 - < T_d >}{< T_d >} \right) \tag{4.58}
\]

Spherical albedo represents the ratio of the total energy reflected by aerosol and molecules present in the entire atmosphere to the energy incident on it. Spherical albedo \( S_a \) has been computed using analytical formulas developed by Tanre et al. (1979).

\[
S_a = (0.92 \tau_{r\lambda} + 0.33 \tau_{a\lambda}) \exp[-(\tau_{r\lambda} + \tau_{a\lambda})] \tag{4.59}
\]

Average spherical albedo has been expressed by

\[
< S_a > = \frac{\sum_i S_{a,i} \times F_{0,i}}{\sum_i F_{0,i}} \tag{4.60}
\]

Average cloud albedo has been expressed by the reflectance measured by OCM in the PAR spectral range and angular factor \( F \) (Frouin et al., 2003)

\[
< A > = F \times< R (t^*) > \tag{4.61}
\]

Where \( t^* \) is the OCM observation time. Angular factor \( F \) has been calculated by analytical formulas proposed by Zege (1991).

For each pixel of OCM which is not contaminated by Sun glitter, OCM radiance \( L_i \) in band \( i \) (\( i = 1, 2, \ldots, 6 \)) has been transformed into reflectance \( R_i \).

\[
R_i = \frac{\pi L_i}{\cos \theta_s F_{0,i}} \tag{4.62}
\]

Reflectance estimated at each band of OCM has been corrected for ozone and also for intrinsic atmospheric reflectance as given in Frouin et al. (2003). Intrinsic reflectance is defined.
as reflectance by photons that have not interacted with cloud or ocean surface layers. \( <R(t^*)> \) is average reflectance of six visible OCM bands. Average reflectance has been estimated by

\[
<R(t^*)> = \frac{\sum_i R_i(t^*) \times F_{0i}}{\sum_i F_{0i}} \quad (4.63)
\]

Figure 4.15 shows the variation of ocean surface albedo and cloud albedo estimated from OCM on a pixel by pixel basis in the Arabian Sea. Ocean surface albedo estimated from OCM was varying from 5-6% (Figure 4.15). For thin clouds, cloud albedo was varying 18-20%. Maximum 40-45% cloud albedo was observed for thick clouds (Figure 4.15).

Figure 4.15: OCM derived average ocean surface albedo and cloud albedo in PAR wavelength range (Date: 16 November, 2001).
4.7 Summary and conclusions

Two PAR products such as PAR at noon and daily averaged PAR have been estimated from OCEANSAT-1 & 2 using two different methods for sunglint free region. Sunglint region in OCM has been estimated using wind climatology and it has been masked. Major specifications and features of OCEANSAT-1 & 2 OCM have been discussed. Pixels of OCEANSAT-1 & 2 OCM have been separated under clear and cloudy sky condition based on threshold value. Threshold value has been estimated from reflectance at band 8 and reflectance ratio between band 8 and band 6 of OCM data.

In the method I, the atmosphere is treated as a single layer for clear sky conditions, or as a double layer for cloudy conditions i.e., a layer above the cloud top and a layer from the cloud top downwards. Surface reflectance is not included in this method. Extra-terrestrial solar irradiance at the top of the atmosphere has been subdivided into continuous fifteen wavelength band at 20 nm spectral intervals. Extra-terrestrial solar irradiance at each wavelength band has been obtained by using trapezoidal method of integration. Extra-terrestrial solar irradiance at each band has been corrected for Rayleigh scattering, ozone absorption, uniformly mixed gas absorption, aerosol transmittance under cloud free condition. Aerosol transmittance is expressed in terms of aerosol optical depth and angstrom coefficient property. Methodology to estimate aerosol optical depth and angstrom coefficient from OCEANSAT-1 & 2 OCM have been discussed. Under cloudy sky input parameters are extra-terrestrial solar irradiance, cloud top pressure, backscatter fraction of incident radiation and cloud optical depth. A fixed cloud top pressure of 574 mb and fixed aerosol optical depth of 0.13 have been used to estimate PAR under cloudy sky. Cloud optical depth has been estimated at band 2 of OCM using a semi-analytical model for thick clouds. Reflectance at band 2 of OCM, reflectance of an idealized semi-infinite non absorbing cloud, escape function, solar zenith angle, satellite view angle and relative azimuth angle are used as input parameters to estimate cloud optical depth from OCM for thick clouds. For thin clouds i.e., clouds which are having cloud optical depth less than 4, a quadratic relationship between cloud optical depth and TOA radiance at band 2 of OCM has been used.
In method I, sensitivity of input parameters such as monthly variation, ozone concentrations, aerosol optical depth, angstrom exponent, cloud optical depth and cloud top pressure has been studied. Sensitivity results show that PAR changes with month wise. PAR is maximum during April to August. Ozone is the least sensitive parameter. The variation of PAR is observed only ± 2% when angstrom coefficient varies from 0 to 1.8. The variation of PAR is observed ±3% when cloud top pressure varies from 310 mb to 800 mb compared to fixed cloud top pressure 574 mb. The error to estimate PAR is under cloudy sky 4.5-6 % when cloud top pressure varies from 180 mb to 50 mb compared to fixed cloud top pressure 574 mb. PAR has decreased 7.8% and 13.8% for aerosol optical depths 0.3 and 1 respectively compared to no aerosol loading. PAR estimation is highly sensitive to cloud optical depth, particularly at lower value of cloud optical depth above 10. Under clear sky condition aerosol optical depth and under cloudy sky cloud optical depth plays important role for a particular season compared to the other variables in PAR estimation.

For method II, the effect of clouds and clear atmosphere is assumed to be decoupled with the cloud system and ocean surface albedo. In this method, mean of extra-terrestrial solar irradiance corresponding to six discrete bands in visible wavelength of OCM has been used. AOD at 865 nm, angstrom coefficient and ocean surface albedo estimated from OCM has been used in this method II as input. TOA radiance of OCM at each visible band has been used to estimate cloud albedo. Comparisons of OCM PAR estimated using two methods with in-situ measured PAR and also PAR estimated from other ocean colour sensors have been discussed in the next chapter.