2

Data and Methodology

2.1 Introduction

The characteristic features of an upwelling region are low SST, high CHLA, lower SSH and preferably, presence of along shore wind stress, if on the eastern boundaries of the ocean. The vertical motion rotates water in cyclonic (anti clock wise) or anti-cyclonic (clock wise) circulation (in northern hemisphere). Cyclonic circulation leads to upwelling, while anti-cyclonic circulation results in downwelling. To study the relation between above stated parameters and their impacts on upwelling, data has been amassed from various sources. The following are the satellite platforms and the sources from where the data has been accessed / downloaded.

SST data was obtained from the Advanced Very High Resolution Radiometer (AVHRR) on board a series of NOAA satellites (http://poet.jpl.nasa.gov) and the Tropical Rainfall Measuring Machine (TRMM) Microwave Imager (TMI) (http://las.incois.gov.in). Sea Surface Wind data was obtained from QuikScat scatterometer onboard Quikbird satellite (www.ifremer.fr). The SLA data has been obtained from Archiving, Validation and Interpretation of Satellite Oceanographic data (AVISO) that distributes satellite altimetry data of the available altimeters (ftp://ftp.aviso.oceanobs.com). The ocean colour data from Sea viewing Wide Field of view Sensor (SeaWiFS) and Moderate resolution Imaging Spectroradiometer (MODIS) on board Aqua satellite was obtained from ocean colour group at Goddard Space Flight Centre of NASA (http://oceancolor.gsfc.nasa.gov).
Apart from these satellite data products, other auxiliary data products like temperature and salinity profiles to compute mixed layer depth were obtained from CORIOLIS (http://www.coriolis.eu.org), and world ocean atlases 2001 and 2009 were obtained from National Climate Data Centre of NOAA (www.nodc.noaa.gov).

2.2 Principle of Measurement and Processing

2.2.1 Sea Surface Winds

Sea Surface wind is a vector quantity and space-borne microwave scatterometers are the only proven instruments that can measure both wind speed and direction over the ocean under all weather conditions [Wentz et al., 2001]. Scatterometer is one of the active remote sensors used in satellite oceanography that works on the principles of radar. It is an oblique viewing radar pointing towards the sea surface from aircraft or satellites at incidence angles normally between 20° and 70°. Backscattered energy received by the receiver from the field of view of the sensor determines the sea surface roughness and thereby the wind speeds over the sea surface. The backscatter is governed by the in-phase reflections from surface waves where, for a smooth surface the radar receives no return when viewing at an angle [Wentz et al., 2001]. As the surface roughness increases, backscatter occurs as constructive interference of scattering from periodic structures in the surface roughness. The backscatter does not only depend on the magnitude of the wind stress but also the wind direction relative to the direction of azimuth angle of the radar beam. The retrieval of wind speed and direction from the scatterometer measurements requires knowledge on the backscatter variation with wind speed and direction relative to the radar azimuth [Robinson, 2004]. The empirical formula used to derive the winds for a particular frequency of the radar:

$$\sigma_o = \sigma_o(U, \chi, \theta, p)$$

(2.1)

where $\sigma_o$ is the normalized radar back scatter function, $(U, \chi)$ are the wind speed and direction relative to the radar azimuth, $\theta$ is the radar incidence angle and 'p' is the polarization. Figure 2.1 illustrates graphically the wind speed and
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Figure 2.1: Principle of measurement of QuikScat Scatterometer, taken from http://nsidc.org

direction relative to the radar azimuth and the schematic of measurement principle used in QuikScat Scatterometer. QuikScat operates in ku band (∼14 GHz) frequency in a sun synchronous near polar orbit with 98.6° inclination angle and an altitude of 803km (Wentz et al., 2000, Wentz et al., 2001). Its revisit time is 4 days. The radar radiates microwave pulses at 13.4 GHz using twin pencil beams at angles 46° (horizontal polarization) and 54° (vertical polarization) with an Instantaneous Field of View (IFOV) for each pencil beam as 30 x 40 km. Thus, each point on the ground is viewed from multiple directions while maintaining constant incidence angle for each of the beams. The overall spatial resolution is 25 x 25 km with an accuracy of 2 ms\(^{-1}\) and 20° for wind speed and direction respectively. Scatterometer data has become a source of real time information regarding global wind pattern for both meteorological and oceanographic purposes. In upwelling studies, the wind stress and curl computed from the scatterometer measurements provide an indirect signal on intensity of upwelling. The signatures of upwelling thus arrived at, from the stresses are the negative (positive) wind stress along the eastern boundaries in the northern hemisphere (southern hemisphere). The curl of wind stress should be positive (negative) in the northern hemisphere (southern hemisphere) to boast of divergence pattern over the ocean surface [Pickard and
2.2 Principle of Measurement and Processing

Emery, 1982]. Wind stress is computed from QuikScat measured winds using the bulk aerodynamic formula:

\[ \tau = \rho_a C_d U^2 \]  

where \( \tau \) is wind stress over the ocean, \( \rho_a \) is the density of air (1.25 kg m\(^{-3}\)), \( C_d \) is the wind dependent drag coefficient and \( U \) is the wind speed following Smith [1988]. The zonal (u) meridional (v) component of wind speeds in network Common Data Format (netCDF) were obtained from the Asia Pacific Data Research Centre (APDRC). The validation statistics of QuikScat measured winds over the Indian Ocean region were reported by Goswami and Rajagopal [2003] and Satheesan et al. [2007].

2.2.2 Sea Surface height

Sea Surface Height (SSH) is precisely measured using satellite altimeters. These altimeters are radars that transmit sharp pulses toward the Earths surface and receive the return pulse. Height of the satellite above the sea surface is obtained by measuring the time required by the pulse to travel from the altimeter to the surface and back [Robinson, 2004]. Amplitude and shape of the reflected pulse provide additional information about the surface such as sea surface roughness. Basic understanding of altimetry is derived from the knowledge of potential gravity due to Earths atmosphere and the potential gravity due to the solid earth and water along with the centrifugal acceleration due to Earths rotation. Assuming no atmosphere, still water results in an equi-potential surface, and this equi-potential surface is called Geoid. Geoid is a property of gravitational field and responds to global distributions of mass. The displacement of the sea surface from the geoid is known as the sea surface topography. This difference is primarily due to the currents and tides [Stewart, 1985]. In order to attain the sea level deviation from the geoid, one should have detailed knowledge of the global geoid, but this is not available at present, in such cases the long term altimeter measurements of the available altimeter records provide sea surface topography. Essentially geoid is time invariant and thus the long term altimeter records even without the geoid information provide better data of the time varying ocean dynamic topography. At least two altimeters are required to monitor the ocean precisely at very high
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Figure 2.2: Principle of altimetry: Radar altimeters measure the distance between the satellite and the sea surface (E). The distance between the satellite and the reference ellipsoid (S) is derived by using the Doppler Effect associated with signals emitted from marker points on the Earth's surface as the satellite orbits overhead. Variations in sea surface height (SS, i.e., S-E), are caused by the combined effect of the geoid (G) and ocean circulation (dynamic topography, DT), from (www.eohandbook.com/eohb05)

resolutions to understand the mesoscale variability over the ocean [Robinson, 2004]. Figure 2.2 shows the schematic of the principle of altimetry. Of all sensors carried on satellites, the altimeter is most dependent upon its orbit to be capable of successful calibration and interpretation. An altitude over 1300 km is advised for altimeter missions because:

- of the atmospheric drag, there is an order of magnitude less than at 800 km,
- ground stations can much better track the satellite,
- the satellite orbit error resulting from irregularities of the Earth gravitation field at a high orbit is less than at a lower one and
- Air, water vapor, clouds, and rain slow down the return of the microwave signal. A second instrument called a radiometer is used to correct for the influence of water in the atmosphere.
The dual-frequency NASA radar altimeter (TOPEX / Poseidon) works by sending radio pulses at 13.6 GHz and 5.3 GHz toward the earth and measuring the characteristics of the echo [Fu et al., 1992]. By combining this measurement with data from the microwave radiometer and with other information from the spacecraft and the ground, scientists can calculate the height / level of the sea surface. Data from the SLR (Satellite Laser Ranging) and DORIS (Doppler Orbitography and Radio positioning Integrated by Satellite) systems are used to determine the orbit of TOPEX/Poseidon. Together these systems provide all-weather, global tracking of the satellite. There are however, some limitations in land-based systems. Sea level anomaly (SLA) is derived by subtracting the real time observations from the altimeters from the long term mean. The data from different sensors are merged to arrive at better coverage and accuracy of the sea level to less than 4cm www.aviso.oceanobs.com. The sea level anomaly data used for the present study is a merged product of different altimeter missions and is obtained from AVISO data extraction service [LeTraon and Dibarboure, 1999]. The spatial resolution of the data is 0.25° x 0.25°. The temporal resolution ranging from weekly to monthly was selected based on the process to be studied. The geostrophic currents were computed from the sea level anomalies using the geostrophic relation:

\[ 2\Omega \sin(\phi).V = g \tan(i) \]  

(2.3)

where \( \Omega \) is the earth’s angular velocity, \( \phi \) is the latitude, \( V \) is the velocity and \( \tan(i) \) is the slope of the sea surface [Pond and Pickard, 1983]. The geostrophic currents are made use of to understand the circulation pattern in the region of interest during different seasons.

2.2.3 Sea Surface Temperature

Infrared Radiometers  Present study makes use of SST data obtained from Advanced Very High Resolution Radiometer (AVHRR) onboard NOAA series of satellites and Tropical Rainfall Measuring Machine - Microwave Imager (TMI). The data products were chosen based on their availability. The AVHRR functions based on infrared radiometry, where the fundamental basis is that all surfaces emit radiation whose strength is in-turn dependent on the surface temperature.
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![Figure 2.3: Black body emittance spectrum at different temperatures, from http://cimss.wisc.edu](image)

The higher the temperature, the greater is the radiant energy. By measuring the emitted radiation, the temperature can be calculated, provided the physics of the process is well defined Robinson [2004]. The spectral characteristics of thermal emission from body at temperature \( T \) in K are determined by Planks radiation law:

\[
M(\lambda, T) = \frac{C_1}{\lambda^4} \left[ \exp\left(\frac{C_2}{\lambda T}\right) - 1 \right]
\]  

(2.4)

where \( \lambda \) is the wavelength in metres, \( M \) is the spectral exitance (also called mittance: The radiant flux density of radiation per unit band width centred at \( \lambda \), leaving unit area of surface, irrespective of direction), \( C_1 \) and \( C_2 \) are constants. By integrating over all the wavelengths gives the total exitance of a black body:

\[
M = \sigma T^4
\]

(2.5)

where \( \sigma = 5.669 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4} \) (Stefan’s constant). The spectral peak of each temperature is found at wavelength:

\[
\lambda_{max} T = C_3
\]

(2.6)

Figure 2.3 shows the emittance at different wavelengths for different temperatures. The solar emitted energy has a peak in the visible part of the spectrum and typical SSTs emission peak lies between 9\( \mu \)m and 11\( \mu \)m. Thus, the thermal IR is an optimal region for monitoring SST. An IR radiometer can detect the brightness temperature of the radiation at the sea surface. The measured brightness
temperature differs from the actual temperature of the observed surface because of its non-unit emissivity and also because of the intervening atmosphere. The presence of atmosphere restricts the IR radiometry of the sea surface to two spectral windows $3.5 - 4.1 \mu m$ and $10.0 - 12.5 \mu m$. The emissivity of the sea surface in the IR region is approximately 0.98 to 0.99 [Masuda et al., 1988]. The sensor, when viewing the sea surface with an angle zero, the maximum emissivity is 0.992 at $11 \mu m$ wavelength. However, the sensor monitors the surface from different incident angles during different sea states and thus the emissivity varies with wind speed [Watts et al., 1996]. It could be said that the IR radiance above the sea surface depends not only on the temperature of the surface skin, but also on the surface emissivity, the incident radiation after removing the direct sun glitter. Since the emissivity is $\sim 1$ in the IR region, the measured brightness temperature is only slightly less than the true SST and thus can be easily incorporated in the atmospheric corrections. The main difficulty in the IR radiometry is the correction for reflection of the incoming solar radiation. Even though the emittance from the sun is greater than the sea surface, it fills only small part of the sky and thus the solar irradiance reaching the top of the atmosphere is about $10^{-5}$ of its value near the solar surface which is about $1/300$ of the radiation emitted by the sea surface (Maull, 1985, Robinson, 2004). At $10 \mu m$, the solar reflection is small and thus this spectral window is effective both during the day and night. At $3.7 \mu m$, the incoming solar irradiance is of same order as the surface emittance and the diffusion reflection of sunlight contributes an unacceptable error which cannot be corrected. Therefore $3.7 \mu m$ cannot be used during day time. Thus one can deduce that the spectral windows used for SST measurements are the bands between $3.5 - 4.1 \mu m$ during night time and between $10.0 - 12.5 \mu m$ during both day and night. These signals are derived after necessary atmospheric corrections. For IR sensor calibration, a target of known temperature is used. This temperature is measured and transmitted to ground receiving station along with the signal measured by the IR sensor. Atmospheric correction is based on multispectral approach, when the differences between brightness temperatures measured at different wavelengths are used to estimate the contribution of the atmosphere to the signal. For cloud detection, the thermal and near-infrared waveband thresholds are used, as well as different spatial coherency tests. The
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Accuracy of the IR sensors is high but they cannot see the sea surface through the clouds under overcast conditions. AVHRR is the most successful and widely used IR sensor onboard NOAA series of satellites. It has 5 spectral bands as shown in the following table: The scanner has an IFOV of approximately 1.3m rads and a cross-track scan of ±55.4°. With a nominal height of 833 km the ground FOV at nadir are 1.1 km and the swath width about 2500 km. The orbit period is about 102 min and 14 orbits are completed per day. The swath of adjacent orbits overlap, ensuring that the whole Earth surface is viewed at least twice a day, once from the ascending (daylight) passes and once from the descending (night) overpasses. The pathfinder SST that is employed in the present study is systematically analyzed AVHRR data from 1985. The data is provided on the public domain on daily, 8-day and monthly averaged scales with a spatial resolution of ~ 9 km. The processing methodology and validation statistics of the Pathfinder project are explained explicitly in Podesta et al. [1995] and Kilpatrick et al. [2001].

**Table 2.1:** AVHRR spectral channels and their characteristics, taken from http://noaasis.noaa.gov/NOAASIS/ml/avhrr

<table>
<thead>
<tr>
<th>Channel No.</th>
<th>Resolution</th>
<th>Wavelength (µm)</th>
<th>Typical Use</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1.09km</td>
<td>0.58 - 0.68</td>
<td>Daytime cloud and surface mapping</td>
</tr>
<tr>
<td>2</td>
<td>1.09km</td>
<td>0.725 - 1.00</td>
<td>Land and water boundaries</td>
</tr>
<tr>
<td>3</td>
<td>1.09km</td>
<td>1.58 - 1.64</td>
<td>Snow and Ice detection</td>
</tr>
<tr>
<td>4</td>
<td>1.09km</td>
<td>3.55 - 3.93</td>
<td>Night cloud mapping, SST</td>
</tr>
<tr>
<td>5</td>
<td>1.09km</td>
<td>10.30 - 11.30</td>
<td>Night cloud mapping, SST</td>
</tr>
<tr>
<td>6</td>
<td>1.09km</td>
<td>11.50 - 12.50</td>
<td>SST</td>
</tr>
</tbody>
</table>

**Microwave Radiometers** To overcome the lack of AVHRR data during the overcast conditions, the data products derived from the microwave (MW) radiometers were made use of. Microwave radiometers are capable of measuring SST independently of the cloud cover. These radiometers are normally operated at wavelengths between 1.5mm and 300mm (200GHz to 1GHz) depending on the parameter to be measured. They observe the thermal radiation emitted by the sea surface in the microwave part of the spectrum under all weather conditions.
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At longer wavelengths, there is little or no absorption or scattering by the intervening atmosphere, aerosols, haze, dust or small water droplets in clouds. Only problem is the liquid water in the form of precipitation. But, at longer wavelengths, the thermal emission is weak and thus the signal received is relatively weak. Therefore to maintain the signal strength and overcome the noise levels, the field of view (FOV) must be wider than for IR radiometers. This makes the spatial resolution of MW radiometers coarser than the IR radiometers. The physical principle behind the MW radiometer is the Planks radiation law for non-perfect emitter:

\[ L_\lambda(\theta, \phi) = 2\frac{hc}{\lambda^5} \left[ e^{\frac{hc}{\lambda KT}} - 1 \right] \]  

(2.7)

where \( L_\lambda \) = spectral radiance at wavelength \( \lambda \), \( h \) = Plank’s constant = 6.626 x 10^{-34} \text{Js}, \( c \) = speed of light, \( k \) = Boltzmann’s constant = 1.38 x 10^{-23} \text{J K}^{-1}, \( T \) = temperature of the emitter in degrees K, \( \epsilon(\theta, \phi) \), \( \theta \) and \( \phi \) are the zenith and azimuth angles. Following the Rayleigh-Jeans approximation the spectral radiance \( B_f \) per unit frequency bandwidth is:

\[ B_f = kT \frac{\epsilon(\theta, \phi)}{\lambda^2} \]  

(2.8)

For a black body at temperature 300k typical of the ocean, the Rayleigh-Jean approximation deviates by less than1% as long as \( f \) is less than 117 GHz, corresponding to \( \lambda > 2.57 \text{mm} \). Thus the emitted radiance is directly proportional to the temperature of the emitting surface. Therefore, the emitted radiance is called Brightness temperature while the radiance measured by the microwave antenna is apparent temperature at a certain frequency. But it is typically much lower than the actual temperature of the emitting ocean surface because for the sea water, the emissivity is less than 0.5 at microwave frequencies. The Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) is a passive microwave radiometer that measures SST [Wentz et al., 2000]. Apart from SST, the other parameters that TRMM measures are the precipitation, Earth’s radiant energy and lightening. The TMI has five wavebands at 10.7, 19.4, 21.3, 37.0 and 85.5 GHz. The channel at 10.7 GHz is used to measure SST. Its orbit is not sun synchronous and has complete ground coverage between ±40° latitude and is
2.2 Principle of Measurement and Processing

Figure 2.4: Schematic depicting the temperature structure near the sea surface (a) at night and (b) during the day in conditions suitable for diurnal warming. The figure shows where the skin, sub-skin and depth measurements of SST are defined. $SST_f$ represents the foundation temperature at the base of any diurnal thermocline that may be present (after Donlon et al. [2002]).

accomplished in 3 days at an altitude of 402 km and an angle of inclination 35°. The spatial resolution at which TMI provides SST is $0.25^\circ \times 0.25^\circ$. The accuracy of TMI is measured at $0.6^\circ$C with the drifter buoys [Wentz et al., 2000]. Figure 2.4 presents the schematic of different SST measurements near the sea surface. Every SST observation depends on the measurement technique, the sensor that is used, the vertical position of the measurement within the water column, the local history of all the component heat flux conditions and the time of the day the measurement was obtained. The vertical structure of SST can be defined [Donlon et al., 2002] as follows:

1. The interface SST, $SST_{int}$ is the temperature of an infinitely thin layer at the exact air-sea interface. It represents the temperature at the top of the SSTskin layer and this cannot be measured using current technology.

2. The SSTskin is the temperature measured by a radiometer at depth within a thin layer ($\sim 500 \mu m$) at the water side of the air-sea interface. This
is the basis for measuring the skin temperature using IR radiometers at wavelengths shorter than 5µm [McKeown and Asher, 1997].

3 The SSTsub-skin is the representative of the SST at the bottom of the SSTskin temperature gradient. It could be measured by MW radiometers operating at low frequencies (6 - 10GHz). In this region of electromagnetic spectrum, the penetration depth in seawater is much greater and the resultant measurements depths are greater than 1mm.

4 The SSTdepth (also known as SSTbulk) is the temperature beneath the SSTskin, therefore SSTdepth should always be quoted with specific depth in the water column. SSTdepth is measured using traditional temperature sensors mounted on buoys, profilers and ships at any depth beneath SSTsubskin.

The AVHRR SST data products were obtained from Jet Propulsion Laboratory of NASA and the TMI data was obtained from SSMI. The data products were on different time scales ranging from daily, 8-day and monthly. SST (skin) was used to compute upwelling indices and also in synergy with other data products to ascertain the upwelling region. SST obtained from TMI is used for understanding the temporal relationship between the forcing factors and the upwelling response on SST as the AVHRR data is available on 8-day basis.

2.2.4 Chlorophyll-a Concentration

The chlorophyll-a concentration (CHLA) is the primary photosynthetic pigment in phytoplankton that primarily absorbs in the blue and red regions than in green of the visible spectrum of electromagnetic radiation. The measurements of ocean color are based on the electromagnetic energy between 400 and 700 nm. The sunlight is not merely reflected from the sea surface. The color of water surface results from sunlight that has entered the ocean, been selectively absorbed, scattered and reflected by phytoplankton and other suspended material in the upper layers, and then backscattered through the surface. The back scattered energy from the ocean surface progressively shifts from deep blue to green as the concentration of phytoplankton increases [Yentsch, 1960]. Ocean color radiances in
the blue-green can be coalesced from depths as great as 50 m. The transparency of open ocean waters is very high; the upper layer of tens of meters depth contributes to ocean color, this contribution decreasing with depth whereas in the turbid coastal waters, the depth of the upper layers and also the transparency of the water column are very less. Satellite ocean color depends on CHLA, which in turn depends on phytoplankton biomass. Thus the ocean color data provides the practical means for monitoring the spatial and seasonal variations of near surface phytoplankton, oceanic primary production, global carbon and biogeochemical cycles and fishery research. The CHLA is derived from the back scattered radiances using semi empirical algorithms based on the regression of radiance versus chlorophyll [O’Reilly et al., 1998]. The following are the drivers that can change the ocean color:

1. Phytoplankton and its pigments

2. Colored Dissolved Organic Material (CDOM, or yellow matter, or gelbstoff) which is derived from decaying vegetable matter (land) and phytoplankton degraded by grazing or photolysis.

3. Suspended particulate matter

   - These particulate matter consist of both Organic and Inorganic substances that can alter the ocean color.

   - The organic particulates (detritus) consist of phytoplankton and zooplankton cell fragments and zooplankton fecal pellets.

   - The inorganic particulates consist of sand and dust created by erosion of land-based rocks and soils. These enter the ocean through river run off, deposition of wind-blown dust and wave or current suspension of bottom sediments.

Depending on the density of both dissolved and suspended matter, Morel and Prieur [1977] divided the ocean into Case 1 and Case 2 waters. In case 1 waters, phytoplankton pigments and their co-varying detritus pigments dominate the seawater optical properties and in case 2 waters, other substances that do
not co-vary with CHLA (such as suspended sediments, organic particles, and CDOM) are dominant. Though the case 2 waters occur in relatively smaller area of the oceans, they are important and optically complex because of their constant interaction with human habitat in the coastal regions, prevalence of large river run off. These optically dominant parameters are retrieved from the ocean color sensor observations depending on their inherent nature to absorb and backscatter the energy in particular wavelengths as shown in the following:

1 Chlorophyll absorption peak is at 443 nm,

2 Measurements must also be made in the 500 - 550 nm range where the chlorophyll absorption is zero and the absorption of other plant pigments (i.e., carotenoids) dominate.

3 CDOM-dominated wavelength is at 410 nm and

4 Suspended particulate matter dominates the red region of the spectrum.

All the ocean color measurements are to be subjected to atmospheric corrections before finally being used. Sunlight backscattered by the atmosphere contributes 80 - 90% of the radiance measured by a satellite sensor at visible wavelengths. Such scattering arises from dust particles and other aerosols, and from molecular (Rayleigh) scattering. Such atmospheric contribution can be calculated and removed by the additional measurements made in the red and near-infrared spectral regions (e.g., 670 and 750 nm). Since blue ocean water reflects very little radiation at these longer wavelengths, the radiance measured is due almost entirely to scattering by the atmosphere. Long-wavelength measurements, combined with the predictions of models of atmospheric properties, can therefore be used to remove the contribution to the signal from aerosol and molecular scattering. The data was obtained by Sea viewing Wide Field of view Sensor (SeaWiFS) radiometer onboard Orbview spacecraft of NASA. It has a sun synchronous orbit with an altitude of 705km with a spatial resolution of 1.1 km at Local Area Coverage (LAC) and 4.5km at Global Area Coverage (GAC). SeaWiFS observes the earth surface using its 8 spectral channels: SeaWiFS Data Analysis System (SeaDAS) is a software tool provided by the Ocean color group
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**Table 2.2:** Spectral channels of SeaWiFS and their purposes, taken from website http://oceancolor.gsfc.nasa.gov

<table>
<thead>
<tr>
<th>Spectral Channel</th>
<th>Central Wavelength (nm)</th>
<th>Primary Use</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>412 (Violet)</td>
<td>Dissolved organic matter (incl. Gelbstoff)</td>
</tr>
<tr>
<td>2</td>
<td>443 (blue)</td>
<td>Chlorophyll absorption</td>
</tr>
<tr>
<td>3</td>
<td>490 (blue - green)</td>
<td>Pigment absorption (Case 2), K(490)</td>
</tr>
<tr>
<td>4</td>
<td>510 (blue - green)</td>
<td>Chlorophyll absorption</td>
</tr>
<tr>
<td>5</td>
<td>555 (green)</td>
<td>Pigments, optical properties, sediments</td>
</tr>
<tr>
<td>6</td>
<td>670 (red)</td>
<td>Atmospheric correction (CZCS heritage)</td>
</tr>
<tr>
<td>7</td>
<td>765 (near IR)</td>
<td>Atmospheric correction, aerosol radiance</td>
</tr>
<tr>
<td>8</td>
<td>865 (near IR)</td>
<td>Atmospheric correction, aerosol radiance</td>
</tr>
</tbody>
</table>

of NASA for processing the SeaWiFS data products, incorporated with necessary correction algorithms. The ocean color data products are obtained in Hierarchical Data Format (HDF) and processed using SeaDAS. The final spatial resolution of the data utilized for this study is ~ 9 km and a relevant temporal resolution of 8-days, monthly is used based on the nature of the analysis. The data obtained from the satellite consists of 4 levels depending on the level of processing as shown in the following figure 2.5 for almost all the sensors:

Tables 2.3 and 2.4 presents the details of all the remote sensing platforms used in this thesis.

**Table 2.3:** Satellite Sensor, Parameters measured, their resolutions

<table>
<thead>
<tr>
<th>Sl. No</th>
<th>Sensor</th>
<th>Parameter</th>
<th>Spatial</th>
<th>Temporal</th>
<th>Spectral</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>QuikScat</td>
<td>Winds</td>
<td>0.25° x 0.25°</td>
<td>3 day running mean</td>
<td>14 GHz (Ku band)</td>
</tr>
<tr>
<td>2</td>
<td>TMI</td>
<td>SST</td>
<td>0.25° x 0.25°</td>
<td>3 day running means</td>
<td>10.7 GHz</td>
</tr>
<tr>
<td>3</td>
<td>AVHRR</td>
<td>SST</td>
<td>4 x 4 km</td>
<td>8 day</td>
<td>Thermal IR</td>
</tr>
<tr>
<td>4</td>
<td>SeaWiFS</td>
<td>CHLA</td>
<td>9 x 9 km</td>
<td>8 day</td>
<td>Visible</td>
</tr>
<tr>
<td>5</td>
<td>MODIS</td>
<td>CHLA</td>
<td>9 x 9 km</td>
<td>8 day</td>
<td>Visible</td>
</tr>
<tr>
<td>6</td>
<td>Altimeter</td>
<td>SLA</td>
<td>0.25° x 0.25°</td>
<td>Daily and Weekly</td>
<td>13.6 GHz</td>
</tr>
</tbody>
</table>
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Figure 2.5: Processing levels of satellite data, taken from [Robinson, 2004]

Table 2.4: Satellite data period used for this study and limitations

<table>
<thead>
<tr>
<th>Sl. No</th>
<th>Parameter</th>
<th>Period</th>
<th>Limitations</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Winds</td>
<td>2000 - 2009</td>
<td>Cannot measure during heavy rainfall events</td>
</tr>
<tr>
<td>2</td>
<td>TMI SST</td>
<td>1998 - 2009</td>
<td>Cannot measure during heavy rainfall events</td>
</tr>
<tr>
<td>3</td>
<td>AVHRR SST</td>
<td>1987 - 2007</td>
<td>During Cloud Cover</td>
</tr>
<tr>
<td>4</td>
<td>ScaWiFS CHLA</td>
<td>1998 - 2007</td>
<td>During Cloud Cover</td>
</tr>
<tr>
<td>5</td>
<td>MODIS CHLA</td>
<td>2008 - 2009</td>
<td>During Cloud Cover</td>
</tr>
<tr>
<td>6</td>
<td>SLA</td>
<td>1993 - 2009</td>
<td>Data reliability is less very near to the coast</td>
</tr>
</tbody>
</table>
2.3 Software Tools

To process and analyze the data, suite of software packages were made use of. The following are the list of them:

- Ferret - For major data processing and analysis (www.ferret.noaa.gov/FERRET)
- GrADS - Exclusively used to compute EOF (http://www.iges.org/grads)
- SeaDAS - For retrieving chlorophyll data from SeaWiFS and MODIS level 3 products (http://oceancolor.gsfc.nasa.gov/seadas)
- Origin - For quality line graphs
- Matlab - Exclusively used for wavelet analysis (wavelet analysis toolbox obtained from Colorado University) and Coherence spectra.

2.4 Study Area

By making use of the above mentioned data products and the analysis tools, the spatio-temporal variability of upwelling phenomenon in SEAS (1.9) has been addressed in the present study. This study envisages bringing about an understanding on the upwelling, its generation, forcing factors like wind and remote forcing, heat budget, surface cooling and surface CHLA of SEAS. This region is having immense importance in the climate of India and especially the southwestern coast of India. Through this study, the contribution of satellite data products in effectively understanding the marine environment has been emphasized.