3 Biophysical processes in the Western Tropical Indian Ocean

3.1 Introduction

The chlorophyll \( \text{a} \) (chl) distribution and productivity in Western Tropical Indian Ocean (WTIO) is poorly understood mainly because of the lack of \textit{in situ} data. The International Indian Ocean Expedition (IIOE) carried out several cruises during late 1950s and early 1960s and attempted to describe the primary production and plankton distribution in Indian Ocean (\textit{Krey and Babenerd} (1976); \textit{Ryther et al.} (1966)). However, large part of observations were made from northern Indian Ocean and coastal areas and lacked the seasonal coverage. Studies during the Indian Joint Global ocean Flux Studies (JGOFS) mainly concentrated on the northern and central Arabian Sea and highlighted the existence of temporal variability in the chl distribution and productivity during the different monsoon phases (\textit{Bhattathiri et al.} (1996); \textit{Smith et al.} (1998)).

A cursory look in to the climatological satellite chl data indicates that in general, WTIO is oligotrophic, except at certain bloom locations. However, presence of a deep chl maximum (DCM) was noted in many studies at a depth of around 50 m (\textit{Ryther et al.} (1966);\textit{Owens et al.} (1993)). DCM is a wide spread feature of oligotrophic waters located towards the base of the photic zone and near the nutricline. This layer is the result of light cut off from above and nutrient supply from below (\textit{Huisman et al.} (2006); \textit{John J. Cullen} (1982)). Thus the surface oligotrophy and the presence of DCM points to the fact that a major fraction of total productivity in this area confines to the DCM, and hence understanding the vertical distribution of chl and dynamics of DCM are very important in the WTIO.

In the tropical Indian Ocean, phytoplankton distribution is modified by seasonally reversing monsoon winds extending to 10°S (\textit{Wiggert et al.} (2006)). The winds are clockwise/strong during summer monsoon (SM) but anticlockwise/weak
during winter monsoon (WM). Temporal and spatial variability of primary productivity in the Arabian sea is assumed to be resulting directly from monsoonal forcing of the upper ocean (Bartolacci and Luther (1999)). The spatial structure and reversing nature of monsoon winds regulate mixing by mechanical stirring, convective over turning, downwelling and upwelling. These processes regulate the nutrient flux to the euphotic zone and hence the chl distribution (Banse (1987); McCreary et al. (1996)). Mixing process also influence overall light condition encountered by the phytoplankton (McCreary et al. (1996)). During summer monsoon, strong north eastward wind stress variation across the axis of Findlater jet causes down welling and deep mixed layer in southern Arabian Sea (Bauer et al. (1991)). In contrast, during winter monsoon comparatively weak south westward wind stress cause weak upwelling in southern Arabian Sea. However, not many studies were carried out to understand the influence of this in the chl distribution.

The equatorial band of Indian Ocean is characterized by annual mean westerly winds, but during summer monsoon the winds are northward across the equator. There are no easterlies at the equator and hence no upwelling. During the inter monsoon seasons the equatorial Indian Ocean is characterized by an eastward propagating semiannual Wyrtky jet (Wyrtki (1973)). In the equatorial Indian Ocean, the Wyrtki jets maintain a deep thermocline in the east and a shallow thermocline in the west (Murtugudde et al. (2000)). This characteristic structure of the equatorial thermocline (as well as the nitracline) leads to prominent phytoplankton blooms in the west whereas the biological activity remains relatively low in the east.

McCreary et al. (1993), identified a zonal region of upwelling from 4°S to 14°S forced by negative windstress curl resulting in Ekman divergence. This shallow thermocline region is bound by south equatorial current in the south, equatorial counter current in the north during winter and the equatorial jets during the monsoon transition periods. Hermes and Reason (2008) termed this region as, Seychelles – Chagos thermocline ridge (SCTR, 5°S-10°S, 50-80°E). Yokoi and Tozuka (2008) also showed that even if wind stress curl is positive, upwelling is active, either due to the beta effect or the influence of Rossby waves. This region is biologically and climatologically important because upwelling is active and/or the thermocline here is shallow and susceptible to the atmospheric forcing at different time scales. Resplandy et al. (2009) opined that the surface blooms occurring in the SCTR may be due to the entrainment of nutrients into the mixed layer and/or the distribution of phytoplankton from the DCM to the surface.
South of this thermocline region is subduction region associated with the highly
oligotrophic subtropical gyre with minimal chl concentration (Morel, A. et al.
(2010)).

Thus there exists different regimes in the WTIO with distinct bio-physical char-
acteristics. However observations are sparse in this region to carry out studies
on the bio-physical coupling that exists in these regimes. In the present chapter,
the chl distribution and its coupling with the physical processes in the WTIO
are detailed. The chapter is organized in the following manner. In section 3.2,
a brief description of data sets used and analysis carried out are presented. In
section 3.3, the variability of chl distribution and influencing physical param-
eters at different time scales (seasonal/diurnal) are described. In section 3.5
conclusions drawn from the study are presented.

<table>
<thead>
<tr>
<th>Cruise</th>
<th>CTD</th>
<th>Chlorophyll</th>
<th>Nutrients</th>
<th>Additional Sensors on CTD</th>
</tr>
</thead>
<tbody>
<tr>
<td>TIO2008</td>
<td>CTD (Seacat SBE19 plus portable) attached to Idronut water sampler</td>
<td>Spectrophotometric method</td>
<td>Skalar auto analyzer</td>
<td>PAR (QSP230 Biospherical)</td>
</tr>
<tr>
<td>TIO2009</td>
<td>CTD (Seacat SBE19 plus portable) attached to Idronut water sampler</td>
<td>Spectrophotometric method</td>
<td>Skalar auto analyzer</td>
<td>PAR (QSP230 Biospherical)</td>
</tr>
<tr>
<td>TIO2014</td>
<td>CTD (911 Plus SBE)</td>
<td>Fluorometric method</td>
<td>Skalar auto analyzer</td>
<td>PAR (QSP200 Biospherical) Fluorance sensor (Turner)</td>
</tr>
</tbody>
</table>

Table 3.1: Instrumentation, chlorophyll and nutrient estimation methods used
during different expeditions to Western Tropical Indian Ocean.

### 3.2 Data and sample collection

This study used hydrographic measurements taken from the WTIO during the
tropical Indian Ocean expeditions, TIO2008 (during the winter monsoon from
13th November to 15th December 2008 (WM)), TIO2009 (during the summer
monsoon from 1st June to 8th July 2009 (SM)), TIO2014 (during the SM from
8th June to 17th June 2014). TIO2008 and TIO2009 expeditions were made on
Figure 3.1: November-December averaged climatology of (a) chl (mg/m$^3$) derived from the monthly SeaWiFS data for the period of December 1997 to December 2010. The black contours represent the November-December averaged climatology of delayed time weekly SLA (October 1992 to April 2012 from AVISO live access server). (b) Wind curl (Pa/m) derived from QuikSCAT winds (August 1999 to October 2009). The vectors represent the wind stress (Pa). (c) Ocean currents derived from OSCAR for a period October 1992 to October 2012. The vectors represent the velocity components. June monthly climatology of (d) chl (mg/m$^3$) derived from the monthly SeaWiFS data. The black contours represent the November-December averaged climatology of delayed time weekly SSHA. (e) Wind curl (Pa/m) derived from QuikSCAT winds. The vectors represent the wind stress (Pa). (f) Ocean currents derived from OSCAR. The vectors represent the velocity components.
board ORV Sagar kanya and TIO2014 expedition was carried out onboard the ORV Sagar Nidhi. During both the TIO2008 and TIO2009 CTD stations were occupied at every 1° interval from 8°N to 20°S along 65°E. During TIO2014 a 10 day time series was conducted at 8°S 67°E and CTD casts were taken at every 6 hour interval from 8th June to 17th June 2014. Sea water samples were collected from 11 depths (0, 10, 20, 30, 50, 75, 100, 120, 150, 180 and 200 m) at 2° intervals from 8°N to 18°S for chl and nutrient measurements. The details of the CTD used and the nutrient and chl estimation methods are given in Table 3.1. It can be noticed that in the thesis chl values from both spectrophotometric (Strickland and Parson (1972)) and flurometric (AU10 Flourometer: Turner designs Inc., USA) method has been used according to the logistics available during the cruises. Previous studies suggested that if the purpose of chl estimation is an estimate of phytoplankton biomass then both spectrophotometric and flurometric methods are quick and reliable (Holm hansen et al. (1965); Murray et al. (1986)). However the later one has more sensitivity, it can measure a lower value up to ~0.02 mg/m³ (Pinckney et al. (1994)). The MJO (Madden Julian Oscillation) index (Wheeler and Hendon (2004)) showed values corresponding to phase 6 during both TIO2008 and TIO2009, indicating suppressed convection in the Indian Ocean. The Madden Julian Oscillation (MJO) induced wind bursts were absent or less evident during the study period (TIO2008 and TIO2009).

3.3 Results and Discussion

3.3.1 Climatological WM and SM chl distribution

The climatological WM (November-December averaged) SeaWiFS chl is shown in Figure 3.1 a. The figure shows that in general WTIO (40°E-80°E, 10°N-20°S) is an oligotrophic region with chl concentration <0.2 mg/m³. During WM, elevated surface chl (> 0.5 mg/m³) is noted only in the coastal regions. Wind stress vectors suggest an anticlockwise pattern from 10°N to 10°S and further south, the winds were southeasterlies (Figure 3.1b). Wind curl was positive from 10°N to 4°S and further south till 16°S it is negative which again changed to positive till 20°S. The positive (negative) wind stress curl in the northern (southern) hemisphere indicate wind induced upwelling in the open Ocean. The major feature noticed in the WM climatology is the presence of eastward Wyrtki jet along the equatorial belt and westward flowing South equatorial current (Figure 3.1c) . In the SM, the chl distribution is similar to WM except for the very low (undetectable levels)
noted in the eastern equatorial belt and patches of blooms noted from 4°S to 12°S (Figure 3.1d). winds in the SM changed to clockwise and winds were stronger than that in WM. The wind stress curl is also negative from 10°N to 16°S and it become positive further south (Figure 3.1e). In the SM the equatorial eastward currents get weakened but SEC became stronger and the core shifted further south to 12°S (Figure 3.1e ).

3.3.2 Background hydrographic settings during TIO2008 and TIO2009

3.3.2.1 Chlorophyll and nitrate distribution

To understand the biophysical coupling in the WTIO, the vertical distribution of chl, nitrate, temperature and salinity data collected during the TIO2008 and the TIO2009 were used. The TIO2008 and TIO2009 expeditions represent the consecutive winter and summer monsoon (WM and SM respectively). For brevity henceforth in this chapter the TIO2008 and TIO2009 will be mentioned as WM and SM respectively. Figure 3.2a and b shows the cruise track overlaid on the WM and SM satellite chl respectively. It can be seen that the WM surface chl has similar distribution to the climatological November-December surface chl. Along the WM transect, the surface chl concentration varied between 0.04 and 0.3 mg/m$^3$, and the lower concentrations were recorded south of 10°S (Figure 3.2c). Notable features in the vertical structure of chl was the presence of patches of high surface concentrations (0.2–0.3 mg/m$^3$) between 8°N and 6°N, 2°N and 6°S, and at 12°S, and the occurrence of the DCM at an approximate 50 m depth from 8°N to 6°S, which deepened to approximately 150 m towards the south. The concentration of chl in the DCM also varied with location. It can also be seen that in the northern stations where the DCM was shallow, the chl concentration was ~0.6 mg/m$^3$ whereas in southern stations where the DCM was deep, it was ~0.3 mg/m$^3$. In general, the chl concentrations in the DCM were two times higher compared to that of the surface.

Because nitrate is a major macronutrient for the growth of phytoplankton, its distribution in the region requires detailed study. In the oligotrophic open oceans, the thermocline/nitracline depth is important because well-lit surface layers are generally devoid of nitrates for primary production (John J. Cullen (1982); Huisman et al. (2006)). As the thermocline/nitracline deepens, it results in a weak and deep DCM. The DCM becomes weak because of the exponential decay of
Figure 3.2: (a) The WM (13\textsuperscript{th} November to 15\textsuperscript{th} December 2008) CTD stations overlaid (★) on the November–December 2008 monthly SeaWiFS chl (b) SM (1\textsuperscript{st} June to 8\textsuperscript{th} July 2009) CTD(★) stations overlaid on the 2009 June monthly SeaWiFS chl. Latitudinal sections of (c) chl, (d) nitrate (e), temperature, and (f) salinity during the WM. The continuous and dashed black curves represent the depth of the nitracline (2 \textmu M isolines of nitrate) and the MLD (calculated using the 0.2 kg/m\textsuperscript{3} density criterion), respectively. Euphotic depth is shown by the black boxes (■). The dotted 23 °C isotherm contour represents the thermocline.
light with depth (Kirk (1994)) and becomes deep due to insufficient nitrate availability above the nitracline. During the WM, the upper 20 m water column from 8°N to 10°S was devoid of nitrate, and further south, the nitrate deficient water column extended up to 50 m (Figure 3.2d). Between 8°N and 10°S, the nitracline was located between 20 and 50 m whereas south of 10°S, it deepened up to 175 m; the nitrate concentration decreased drastically in the upper 200 m. Shoaling of surface nitrate isolines centered at 4°N and 8°S was also noticed. From the above analysis, it becomes clear that the DCM occurs at the base of the euphotic zone where nitrate is abundant. This concurrence underlines the fact that macro nutrient distribution is a major factor in maintaining the DCM. The temperature section exhibited three regions of varying thermocline depth, the first one near 75 m extending from 8°N to 4°S, the second one at 25 m centered at 8°S in SCTR, and the third one around 110 m south of SCTR (Figure 3.2e). Thus, from 8°N to 12°S, the undulations in the nitracline followed that of the thermocline, and south of 12°S, the location of the nitracline was deeper than the thermocline whereas at 14°S, it shoaled to approximately 60 m. The salinity section showed great spatial variability from north to south (Figure 3.2f). The maximum salinity (~36) was noted at 8°N and minimum at the SCTR where the MLD was shallowest just at the base of freshened surface layer (~34). Previous studies suggested that MLD has a role on the surface chl distribution (Vinayachandran and Saji (2008); Waliser et al. (2005)). Near the equator, where the high surface chl (~0.3 mg/m³) was recorded, the location of the MLD (~70 m) was deeper than the nitracline (~50 m). In the SCTR, where the MLD was shallower (6–20 m) than the nitracline (~25 m), the surface chl concentrations were low (~0.1 mg/m³).

During the SM, the vertical distribution of chl exhibited a pattern similar to that of the WM. However the satellite chl data was contaminated by the cloud cover (Figure 3.2b). The surface chl concentration varied between 0.13 and 0.4 mg/m³ (Figure 3.3a), and patches of high surface chl concentration were seen between 6°N and 4°N, 6°S and 10°S, and at 16°S. The DCM at most of the stations was located between 50 and 75 m but deepened up to 120 m in the southernmost stations. Compared with the WM from 8°N to 10°S, the chl concentration in the DCM was comparatively weaker (~0.4 mg/m³) whereas south of 10°S, it was higher (0.24–0.47 mg/m³). As in the WM, the chl values in the DCM were approximately two times higher than that of the surface. The euphotic zone at 4°N, 2°N, 2°S and 4°S was extended up to 70–75 m in depth (Figure 3.3a). In the SCTR region, the euphotic zone was ~70 m deep but deepened to 95 m from 11°S to 18°S. This trend was similar to that observed during the WM. The
Results and Discussion

Chapter 3.

Nitrate concentrations in the surface waters were also low, but patches of up to 1 µM concentration were present at certain locations (Figure 3.3b). From 8°N to 2°N, the nitracline was approximately 50 m deep but shoaled to 20 m at 8°S and again deepened to 150 m at 18°S. Shoaling of the nitracline was observed centered at 6°N, which resulted in the enrichment of the euphotic zone by 2 µM. In general, the water column below a depth of 50 m was rich in nitrate (>10 µM) up to 12°S, but south of 12°S, this value was approximately 5 µM, significantly higher than that of the WM. North of the SCTR, the nitrate concentration was higher in the subsurface layer during the SM compared to WM. Although the temperature section showed a pattern similar to that of WM, the thermocline was deeper in the northern stations and shallower in the southern stations during the SM (Figure 3.3c). The thermocline depth was ~ 100 m in the northernmost

Figure 3.3: Latitudinal sections of (a) chl, (b) nitrate, (c) temperature, and (d) salinity during SM. The continuous and dashed black curves represent the depth of the nitracline and the MLD, respectively. The euphotic zone is shown by the black boxes (■). The dotted 23 °C isotherm contour represents the thermocline.
stations, 35 m centered at 6°S in the SCTR and extended up to 110 m south of the SCTR. The shallow thermocline south of the SCTR would explain the comparatively high subsurface nitrate in this region compared with that of WM. The MLD along the track varied between 27 m (6°S) and 82 m (16°S). During SM, the MLD was deeper than that of the WM except at the equator, 2°S, 14°S and 18°S. The marked difference between WM and SM salinity section was that during SM salinity reduced in the northern stations above the equator and salinity increased in the SCTR (Figure 3.3d). This observation both in WM and SM clearly indicates that along the transect, the surface chl distribution was influenced by the MLD, and the DCM was influenced by the thermocline. The importance of physical forcing in regulation of the MLD and the thermocline is further explained below by analyzing the basin wide wind field followed by the SLA.

3.3.2.2 Winds

The weekly QuikSCAT winds averaged over the WM cruise period (Figure 3.4a) showed that in the southern Arabian Sea (8°N to 5°N), the winds were northeasterly with an average speed of ~ 4 m/s. The wind stress curl was positive, which resulted in an average upward Ekman pumping velocity of 0.5 m/day. This positive Ekman pumping velocity was active for 30 days in this region (time series figure not shown). The equatorial region (4°N to 4°S) was characterized by weak westerlies (~ 5 m/s), and hence, the contribution from wind-induced mechanical mixing to the deepening of MLD can be ruled out. In the SCTR centered at 8°S, a cyclonic circulation with an upward Ekman pumping velocity reached up to 1.5 m/day, and the average wind speed in this region was ~ 6 m/s. South of the SCTR, the winds were easterly with a velocity of ~ 7 m/s. The Ekman pumping velocity from 11°S to 13°S showed a positive value of ~ 0.25 m/day. South of 13°S, the Ekman pumping direction was downward with an average velocity of 0.25 m/day. The SLA averaged over the cruise period (Figure 3.4b) showed negative values in the northern-most stations and the SCTR region, indicating a shallow thermocline, whereas near the equator and south of the SCTR, the SLA was positive, indicating a deeper thermocline. An anticyclonic feature centered at 14°S was noted.

During SM weekly QuikSCAT winds averaged for the cruise period (Figure 3.4c) indicated southwesterly winds with a speed of 6 m/s in the southern Arabian Sea (8°N–5°N). In this region, the Ekman pumping was directed downward, and
the average velocity over the cruise period was -0.2 m/day. At 4°N, the Ekman pumping was directed upward (~0.5 m/day), and at the equator, the southerly winds crossed the equator at a speed of 6 m/s. South of the equator, the winds were southeasterly in direction with an average speed of 8 m/s, and the Ekman
pumping was downward and weak (~0.25 m/day). The SLA averaged over the cruise period (Figure 3.4d) was negative only between the equator and at 8°S along the cruise track. The depression in the thermocline observed at 5°S (Figure 3.3c) and could be due to the cyclonic eddy. From the SLA plot, the presence of mesoscale features was evident south of the SCTR. In addition to wind, which will directly alter the thermocline/nitracline depth, remote forcing through baroclinic waves in the ocean can also affects the thermocline/nitracline.

3.3.2.3 Sea level anomaly

A longitude time plot of the SLA averaged over 6°N to 8°N indicated a westward propagating event of negative SLA with a phase velocity of approximately 17 cm/s (Figure 3.5a), which is comparable to the phase speed of the second-mode annual baroclinic Rossby waves radiated from the western boundary of the Indian subcontinent (Subrahmanyam et al. (2001); Brandt et al. (2002)). The phase speed of the Rossby waves was calculated from the slope of the wave pattern. The westward propagation of the negative SLA was also observed between 3°N and 5°N (Figure 3.5b) with a velocity of 36 cm/s, the same as that of the second-mode baroclinic theoretical phase speed at 5°N. The SLA averaged from 3°S to 2°N showed an eastward-propagating Kelvin wave with a velocity of 84 cm/s (Figure 3.5c). These Kelvin waves are generated in the western boundary in September and reach the eastern boundary during the November–December period (Peter and Mizuno (2000); Rao et al. (2010)). The SLA averaged over 4°S to 9°S showed a westward-propagating signal with a velocity of 25 cm/s. However, this propagation appeared to be damped at 70°E (Figure 3.5d). Earlier studies indicated the breakdown of Rossby waves in the SCTR region due to the effect of topography (Matano et al. (2002)) or interaction with local wind forcing (Wang et al. (2001)). A recent modeling study by Hermes and Reason (2008) suggest that topography has little effect on the damping of Rossby waves in the SCTR region. During the WM, a positive Ekman pumping velocity was noted (Figure 3.4a) in the SCTR region. Thus, the damping could be due to the destructive interference of upward vertical velocity associated with the local Ekman pumping and the downward vertical velocity associated with the Rossby waves. The SLA averaged over 10°S to 13°S (Figure 3.5e) and 14°S to 18°S (Figure 3.5f) showed a westward propagation of the positive SLA that was slower than that of the free Rossby waves but comparable to the speed of the ocean–atmosphere-coupled Rossby waves (Jury and Huang (2004)). Westward propagation of chl anomalies with similar speeds have been also noticed in the south tropical Indian
Figure 3.5: Time longitude plot of the SLA (a) averaged over 6°N to 8°N, (b) averaged over 3°N to 5°N, (c) averaged over 3°S to 2°N, (d) averaged over 9°S to 4°S, (e) averaged over 13°S to 10°S, and (f) averaged over 14°S to 18°S. The thick black line represents the cruise period, and the dashed line represents the slope of the wave pattern.
Ocean (White et al. (2004)), indicating that coupled ocean–atmosphere Rossby waves could represent an important biophysical process in this region.

A westward propagation of Rossby waves with speeds comparable to those of the WM (Figure 3.5a) in the longitude time plot of the SLA averaged over 8°N to 6°N during the SM was also noticed (Figure 3.5a). Furthermore, a negative SLA patch was also observed by the end of the cruise. Westward propagation of Rossby waves was evident from 5°N to 3°N (Figure 3.5b), but between 2°N and 3°S, eastward propagating Kelvin waves were observed (Figure 3.5c). The SLA values averaged over 4°S to 9°S (Figure 3.5d) and 10°S to 13°S (Figure 3.5e) showed no westward propagation. However, from 14°S to 18°S, the westward propagating Rossby waves were again observed with speeds comparable to those observed during the WM (Figure 3.5f).

### 3.3.3 The physical control of chl distribution during WM and SM

The significant features observed during both WM and SM cruise included the presence of intermittently elevated surface chl concentrations and a prominent DCM (Figure 3.2c, and Figure 3.3a) that deepened and weakened towards the south.

#### 3.3.3.1 Southern Arabian Sea

In the southern Arabian Sea (5°N–8°N), the DCM was shallow and strong during the WM compared with observations during the SM. The weakening and deepening of the DCM during the SM was primarily due to the deepening of thermocline/nitracline and the low surface PAR values. The deep thermocline during the SM compared with the shallow thermocline during the WM represents a combined effect of Ekman pumping and Rossby wave propagation. During the SM, downwelling occurs due to the anti-cyclonic wind stress curl south of the Findlater jet (Figure 3.4c) (Bauer et al. (1991)). However, during the WM, when the winds are northeasterly, upwelling occurs in the southern Arabian Sea due to cyclonic wind stress curl (Figure 3.4a). The longitude time plot of the SLA (Figure 3.5a) showed westward-propagating second-mode downwelling Rossby waves during the SM and upwelling Rossby waves during the WM radiating from the western Indian continent, which were subsequently modified by
the wind curl over the central Arabian Sea (Brandt et al. (2002); Subrahmanyam et al. (2001); Rao et al. (2010)).

Earlier modeling studies have shown that the Rossby waves radiating from the western Indian continent are remotely connected to the equatorial and Bay of Bengal regions (McCreary et al. (1993); Yu et al. (1991)). Shankar et al. (2004) and Gopalakrishna et al. (2008) pointed out that local wind stress curl off the southwest coast of India and the remote forcing from the south of Sri Lanka are more important to the Rossby wave radiation than the equatorial forcing. Rao et al. (2010) have shown that of the two pairs of alternate upwelling (the first occurring during January–March and the second one occurring during August–September) and downwelling (the first occurring during April–June and the second occurring during October–December) Kelvin waves that propagate eastward along the equator and hit the Sumatra coast and bifurcate, only the second downwelling Kelvin wave reaches the southeastern Arabian Sea. Thus, the westward propagation of upwelling Rossby waves during the WM (which are initiated during August-September) is not a result of signals from the equator whereas the downwelling Rossby waves observed during the SM (which are initiated during October - December) could represent the combined effect of the local wind stress curl off the south-west coast of India, remote forcing from south of Sri Lanka and equatorial signals. However, not all of the biological responses observed along the transect were due to seasonal forcing. The surface chl enhancement observed in the region of 6°N during the SM was associated with a cyclonic eddy (Figure 3.4d). This cyclonic eddy shoaled the nitrate isolines and resulted in high production because the MLD was sufficiently deep to mix the nitrate to the surface. The SLA averaged over 8°N to 6°N (Figure 3.5a) also showed negative values, indicating the presence of a cyclonic eddy by the end of the cruise period.

3.3.3.2 Equatorial band

The significant features observed in the equatorial band (4°N–4°S) were the low surface chl concentration and a weak DCM during the SM compared with the WM, even though the DCM depth was the same during both seasons. During the WM, the thermocline/nitracline shoaled at 4°N, which is consistent with the strong upward Ekman pumping (>1 m/day). The beta effect has a profound influence when the wind becomes zonal in the equatorial regions. During the SM at 4°N, although the upward Ekman pumping was close to 1 m/day (Figure 3.4c), the thermocline was deep, and the surface layers were not enriched with nutri-
ents. This is attributed to the downwelling second-mode baroclinic Rossby waves that originated from the eastern boundary of the Indian Ocean as a reflected Rossby wave (Peter and Mizuno (2000)) in addition to the low-salinity surface waters (Figure 3.3d). Shankar et al. (2004) suggested that the low-salinity Bay of Bengal waters intrude into the Laccadive Sea (~8°N–14°N and ~70°E–74°E) from early November to January and migrate westward along with the westward-propagating Rossby waves. Hence, the observed freshening could be due to the recirculating relics of low-salinity waters intruded during the winter monsoon. Furthermore, positive anomalous precipitation was also observed centered at 4°N (Figure 3.6d). It was noted that the high-salinity water corresponding to the Arabian Sea High Saline Water (ASHSW) mass was present in the subsurface layer. This low-salinity surface layer and the high-salinity subsurface water maintained a strong stratification that induced a high static stability (Figure 3.7b). Thus, the deepening of the thermocline associated with the westward propagating positive SLA and the presence of low-salinity surface waters inhibited upwelling due to the upward Ekman pumping. This situation documents the overwhelming role of two different processes that triggered at remote locations on a local forcing; the first was a westward-propagating Rossby wave, and the second was the advection of water masses, both of which originated in distinct geographic locations in the east. At this point, it is recalled that at 6°N, where the shoaling of nitrate isolines was noticed during SM, the salinity was high, and no low saline surface layer was present.

During the WM, the MLD in the equatorial region south of 4°N was extended up to the depth of nitracline (Figure 3.2c) at most of the stations and reached up to the DCM at the equator and 2°S. The deep MLD extending to the nitracline or the DCM serves in two ways, one by entraining the nutrients to the surface and thereby increasing the productivity, and the other by distributing the phytoplankton itself in the mixed layer (Vinayachandran and Saji (2008); Waliser et al. (2005)). This deepening of the MLD resulted in nitrate fertilization of the near surface layers and an elevated surface chl. The deep MLD observed in the region cannot be explained by the wind speed (~ 5 m/s ). This situation could have been caused by the vertical mixing associated with the highly sheared Wyrtki jet (Figure 3.6a). It was also observed from Figure 3.5c that although the background SLA was negative with the eastward propagating Kelvin wave signal during the cruise period, a mesoscale feature with a positive SLA was also encountered. However, during the SM, the thermocline/nitracline was deep, and the MLD was not deep enough to reach the nitracline, which could have resulted...
Results and Discussion

3.3.3.3 Seychelles Chagos thermocline Ridge

The SCTR is characterized by year round upwelling due to the upward Ekman pumping (Murtugudde et al. (2000); Rao and Sivakumar (2000); Schott et al. (2002)). During the WM, the Ekman pumping velocity was positive with a value greater than 1 m/day and was active for longer than a month, which rendered the thermocline shallower (25–45 m) and centered at 8°S. Despite the high upward Ekman pumping (Figure 3.4a), the upwelled waters remained at subsurface, the SST was 28 °C (Figure 3.2e) and the surface chl concentrations were low (Figure 3.2c).

![Figure 3.6:](image)

Figure 3.6: (a) OSCAR current velocity overlaid by the current vectors during WM (b) monthly averaged rain rate anomaly for WM. (c) OSCAR current velocity overlaid by the current vectors during the SM and (d) monthly averaged rain rate anomaly for the SM. The black straight line along 65°E represents the cruise track.

These observations could be due to the Sverdrup balance attained by the radiating westward-propagating Rossby waves for which Ekman pumping does not
Results and Discussion Chapter 3.

result in upwelling (McCready et al. (2009)). However, a Rossby wave radiation was not observed during this period. At the same time, we observed a westward-propagating Rossby wave from the eastern boundary interacting with the SCTR, which is in agreement with the findings of Hermes and Reason (2008) and Hermes and Reason (2009), thus constraining the effect of wind-driven Ekman pumping. It was also noted that during the WM, the SCTR was capped by a low-salinity water mass (Figure 3.2f). Wyrtki (1973) referred to this low-salinity layer as the Tropical Surface Water derived from the fresh water of the Pacific entering the Indian Ocean through the Indonesian Archipelago and carried westward by the south equatorial current (SEC). The excess of precipitation over evaporation in the eastern Inter Tropical Converging Zone of the Indian Ocean (New et al. (2007); Vialard et al. (2008)) also contributes to the Tropical Surface Water. During the WM, the core of the freshened layer was centered at 10°S (Figure 3.2f) whereas during the SM, this core was centered at 14°S. It was of interest to observe that the core of the SEC was also located at 10°S and 14°S during the WM (Figure 3.6a) and SM (Figure 3.6c) respectively, most likely indicating that the fresh surface layer was associated with the SEC. Anomalous high precipitation also occurred at 10°S 65°E during the WM (Figure 3.6b) and 14°S 65°E during the SM (Figure 3.6d). The Brunt Vaisala frequency section during the WM (Figure 3.7a) indicated a high stability in the SCTR at a depth of 25 m and at the interface of a freshened surface layer and high-salinity deeper water (Figure 3.2f). In the SCTR, weak winds and surface freshening resulted in a shallow MLD (7–20

\[ \text{Figure 3.7: Latitudinal sections of the Brunt–Vaisala frequency during (a) the WM and (b) the SM.} \]
Results and Discussion

Chapter 3.

m) because the surface freshening induces a high static stability (Figure 3.7a). Consistent with this observation, low surface chl values were also observed during the WM at the SCTR. Thus, similar to the observations at 4°N during the SM, the upwelling and surface chl were governed by radiating Rossby waves and surface low-salinity waters originating in the east.

The Ekman pumping velocity was weak and negative, and the thermocline was deeper (35–70 m) in the SCTR during the SM, but the surface chl values were higher (Figure 3.4). The MLD was also deep (28–47 m) during this season and extended to the nitracline (~25 m), which entrained nutrients to the upper layer and resulted in high surface chl values. The fresh water observed at the SCTR during the WM shifted to the south during the SM as the SEC followed a more southward course. This situation precluded the presence of low-salinity surface waters in the SCTR, weakening the static stability of the water column (Figure 3.7a) and thereby deepening the MLD.

3.3.3.4 South Tropical Indian Ocean

The area south of the SCTR is climatologically characterized by the deep thermocline and low-nutrient waters (Longhurst (2006)) associated with the subtropical gyre system. The significant feature observed south of the SCTR (11°S–18°S) was a deepening of the DCM to approximately 120 m during both the WM and SM (Figure 3.2c and Figure 3.3a). However, this deepening was abrupt during the WM, and the same was true for the thermocline (Fig. 2c) and nitracline (Fig. 2b). The observation revealed anti-cyclonic eddies during the WM and SM (Figure 3.4b and d). Anti-cyclonic eddies depress the thermocline and nitracline, making the surface layer oligotrophic and thereby deepening the DCM. These eddies can also entrap waters that are distinct from the ambient waters, which could be the reason why pockets of high chl and nitrate were observed. However, the time longitude diagrams of the SLA showed westward propagation with the phase speed of ocean–atmosphere-coupled Rossby waves (Figure 3.5e and f).

In general, the DCM depth in the WTIO was well correlated with the thermocline/nitracline depth during the WM and SM (R = 0.61/0.89 and 0.71/0.63 during the WM and SM, respectively, with a 95% significance level for 12 degrees of freedom, figure not shown). However, the chl concentration at the DCM was not well correlated with thermocline/nitracline depth during both the WM and SM (R = −0.30/−0.54 and −0.30/−0.22 during WM and SM, respectively). This situation could have occurred because the chl concentration is being influ-
A Special emphasis to the physical control of chlorophyll in the Seychelles Chagos thermocline Ridge

Chapter 3.

enced by several other parameters, including the PAR, nutrients, stability, and grazing, among others.

3.4 A Special emphasis to the physical control of chlorophyll in the Seychelles Chagos thermocline Ridge

As discussed in previous section, being a year round upwelling region and center to the Indian Ocean variability (Vialard et al. (2008)) at different time scales, a special emphasis has given for the SCTR chl variability.

Figure 3.8: Monthly climatological SeaWiFS surface chlorophyll distribution (mg/m$^3$) in the tropical Indian Ocean overlaid by the monthly climatological thermocline depth shallower than 70m. The thermocline depth contour interval is 5 m. The western Box represents the WSCTR and eastern box represent ESCTR. Stars represent the islands.
Figure 3.9: Annual march of various parameters averaged over WSCTR (black line, 50°E - 62°E, 5°S-10°S) and ESCTR (red line, 63°E - 75°E, 5°S-10°S) (a) SeaWiFS surface chl (mg/m³) (b) Thermocline depth (m) (c) Mixed layer depth (m) (d) distance between MLD and thermocline (m) (e) magnitude of Wind stress (N/m²) (f) Ekman pumping (m/day) (g) Temperature (°C) (h) Salinity (i) Rain rate (mm/hr) (j) friction velocity (m/s) (k) Buoyancy flux (m²/s³) (l) Haline buoyancy flux (m²/s³) (m) Thermal Buoyancy (m²/s³) (n) Net heat flux (W/m²) (o) PAR (Einstein/m²/Day).
3.4.1 Physical control on the chlorophyll variability in seasonal time scale

Figure 3.8 shows the climatological thermocline depth < 70 m overlaid on the climatological SeaWiFS chl concentration. It can be seen that major part of this shallow thermocline region is restricted to the SCTR region. The shallow thermocline indicates active year round upwelling and presence of nutrient rich thermocline waters within the euphotic zone in the region (McCreary et al. (2009)). Even though this region is characterized by year round shallow thermocline, elevated surface chl concentrations (> 0.2 mg/m$^3$) are noted only from June to October. The western part of the SCTR (WSCTR, 50°E to 62°E) has higher chl concentration than the eastern part of SCTR (ESCTR, 63°E to 75°E). To understand the response of seasonal surface chl to the various physical parameters, we constructed climatology of different physical properties and averaged it over the WSCTR and ESCTR region. Details of climatology is given in Chapter 2.

Seasonal variability of averaged SeaWiFS surface chl concentration in WSCTR and ESCTR is shown in Figure 3.9a. The remarkable feature is a weak semiannual signal in WSCTR/ESCTR with a primary maximum in July-August (~0.26/~0.16 mg/m$^3$) and secondary maximum in January (~0.14/~0.12 mg/m$^3$), again two minima (~0.12/~0.1 mg/m$^3$) are noted in March - April and December. The amplitude of seasonal variability of surface chl in WSCTR is more than that in ESCTR. It is clear from the seasonal variation of thermocline depth averaged over WSCTR and ESCTR (Figure 3.9b), that the strong semiannual variation in thermocline depth is not captured in the surface chl. The shallow thermocline (~50 m – 54 m) in WSCTR and ESCTR during January did not result in a chl peak as observed during August. Resplandy et al. (2009) opined that shallow thermocline is a necessary but not a sole condition for elevated surface chl in the SCTR, a deep MLD is also required to bring the subsurface nutrients to surface. Figure 3.9c shows the monthly climatological MLD derived from the objectively analyzed ARGO data in the tropical Indian Ocean averaged over WSCTR and ESCTR. It can be seen that the MLD is deep (30-40 m) from June to October and is shallow (< 30 m) from November to May coinciding with the high and low chl period respectively. Previous studies showed that surface chl concentration in this region is regulated by the MLD (Wilson and Qiu (2008); George et al. (2013)). The seasonal variation of chl followed a pattern similar to that of MLD. The distance between the base of mixed layer and heart of thermocline (Figure 3.9d) indicates its lowest values during June - August in WSCTR region.
A Special emphasis to the physical control of chlorophyll in the Seychelles
Chagos thermocline Ridge
Chapter 3.

A mixed layer close to the heart of thermocline can enhance the entrainment of nutrient rich thermocline waters to the surface layers which eventually can increase the surface chl concentration. This may be one of the reasons for the elevated surface chl in WSCTR than that in ESCTR during June - August.

It is clear from seasonal variation of SST and Sea Surface Salinity (SSS) (Figure 3.9g and h) that low SSS and high SST during November - May support strong stratification while high SSS and low SST support the weak stratification from June to October in the SCTR region. The rain rate averaged over the WSCTR and ESCTR region (Figure 3.9 i) also supports the observed variation in SSS, with minimum rain rate in July-August. Stratification is the combined effect of wind and buoyancy forcing. It is of interest to quantify the contribution from wind and buoyancy forcing towards the observed stratification. Figure 3.9 j shows the climatological friction velocity averaged over WSCTR and ESCTR, which is proportional to the energy imparted from the atmosphere to the ocean (Foltz et al. (2003)). It is clearly seen that during the high chl period the friction velocity is at its maximum, and in the WSCTR it is more than that of ESCTR. Figure 3.9k shows the surface buoyancy flux averaged over WSCTR and ESCTR. In WSCTR, the buoyancy flux has its minimum in June and in ESCTR, the minimum is in July. Clearly, the buoyancy flux and winds support the weak stratification in SCTR during the elevated chl period. The saline contribution to the buoyancy flux is shown in Figure 3.9l. It is evident from the figure that the saline contribution to the buoyancy flux is less than that compared to the thermal contribution (Figure 3.9m). The observed dip in buoyancy flux during June - September is associated with low or negative net heat flux in the region (Figure 3.9n) (Yokoi et al. (2012)). The surface PAR averaged over WSCTR and ESCTR ranges from ~36 to ~48 Einstein/m²/day (Figure 7i). Letelier et al. (2004) used the surface PAR value less than 32 (Einstein/m²/day) to represent the less productive winter light condition in the North Pacific subtropical gyre. The satellite derived surface PAR values noted in the WSCTR and ESCTR is always greater than 32 Einstein/m²/day and reached to 48 Einstein/m²/day. Previous studies in the tropical Indian Ocean have suggested that, even though the cloud cover during boreal summer reduces the surface PAR it is not limited for the primary production (Arnone et al. (1998); Prasannakumar et al. (2010)). Further the maximum chl concentration is noted during the boreal summer when the surface PAR was at its lowest. This indicates that the seasonal variability of the surface PAR may not be the governing factor of the observed chl variability.

The depth at which the wind generated turbulence is balanced by the surface
buoyancy can be represented by Monin-Obukhov length (L). The ratio of Monin-Obukhov length (L) and MLD can suggest if the mixed layer is dominated by wind mixing ( > 1) or weak wind mixing confined by buoyancy effects ( < 1) (Anitha et al. (2008)). Figure 3.10 shows the variability averaged over WSCTR and ESCTR. From the figure, it is clear that from April to September, the values are greater than one and mixed layer is dominated by wind mixing both in WSCTR and ESCTR. However, from November to March, the values were less than 1 when the wind energy is less and the net buoyancy flux values is high. This indicates that wind mixing is confined by the strong surface stratification. Analysis of climatological data set suggests that wind mixing has a major role in deepening of MLD, supported by the convective mixing in the boreal summer.

![Figure 3.10: Annual march of ratio of Monin-Obukhov length (L) and mixed layer depth (MLD).](image)

Physical parameters and their relation to the chl in WSCTR and ESCTR showed a similar picture in both the regions. The observed high amplitude of chl variability in WSCTR during June- September could not be explained by the MLD variability alone. Waters around Chagos archipelago (~5°C, ~72°E) and Mascarene plateau (4 - 20°C, ~56°E) are always characterized by elevated chlorophyll in the SCTR (New et al. (2007); Wilson and Qiu (2008)). Some areas of elevated chl in SCTR is tied up with ridges and islands (Stars in Figure 3.8, GEBCO Atlas 2003). WSCTR is characterized by the presence of Mascarene plateau and many small islands (New et al. (2005)). These islands and ridges present in the plateau can help in breaking the internal waves and eventually cause turbulent diffusion of nutrients from the thermocline to the upper layers (Konyaev et al.
A Special emphasis to the physical control of chlorophyll in the Seychelles
Chagos thermocline Ridge (1995); Morozov and d Vlasenko (1996); Wunsch and Ferrari (2004)). During the less stratified period when the winds are strong, this turbulent diffusion may be more, and it would result in high concentration of surface chl is observed from June to September in WSCTR. Another mechanism of increased productivity in WSCTR is eddy shedding. When the currents become strong over Mascarene plateau eddies are generated supplying nutrients to the surface layers (New et al. (2005)).

3.4.2 Turbulence and diurnal variation in the SCTR

One of the ubiquitous features of the world ocean is that it is turbulent in nature. In a stagnant water body heavy particles fall down and attain a terminal velocity determined by the balance between the gravity and the fluid viscosity. Phytoplankton are typically about 2 to 5 percent denser than the ambient water (Ghosal et al. (2000)) and as a consequence they tend to sink in the water column. This sinking could then take them below the euphotic layer, stopping photosynthesis. To overcome sinking, plankton use various techniques. One is swimming, which is an active energy-consuming process to oppose the gravitational sinking. Other phytoplankton can modify their buoyancy, becoming, at least for some time, positively buoyant and float. Others can exploit the turbulence in the mixed layer, using it to stay suspended for longer times (Margalef (1978)). Margalef (1978) proposed a conceptual model of functional morphology for phytoplankton on the basis of nutrient supply and turbulence intensity. It suggested that diatoms, which are usually non-motile and with fast potential growth rates thrive in relatively turbulent, nutrient rich waters. Under these conditions, lack of motility is compensated by re-suspension of cells due to vertical mixing. However, dinoflagellates regulate their position actively. This allows the survival in stratified waters, where motility and migration behavior can override sedimentation. Recent studies indicate that the micro-environment experienced by phytoplankton and other microorganisms are not homogenous (Stocker (2012)). Other than the availability of nutrients, light and physical properties of the water, the dynamics of the micro scale fluid motion influence the phytoplankton (Reynolds (2006)).

Turbulence has another important role in supplying the nutrients to surface layer through vertical diapycnal mixing. Thus turbulence has a very vital role in the ecology of phytoplankton and hence chl distribution. Previous section clearly suggest that SCTR region is characterized by summer chl blooms mainly associated with the wind induced vertical mixing. It is of interest to quantify this vertical
A Special emphasis to the physical control of chlorophyll in the Seychelles
Chagos thermocline Ridge

Chapter 3.

Figure 3.11: (a) June 2014 monthly averaged MODIS chl (mg/m$^3$) (b) weekly SLA from AVISO (m) (c) daily ASCAT wind stress (Pa) (d) daily OSCAR currents (m/s) (e) daily GHRSSST SST (°C) (f) AQUARIUS daily SSS averaged for the period of the time series during TIO2014.

mixing in the region. Nevertheless, recent study by Thushara and Vinayachandran (2014) suggested that diurnal variability has strong influence on the intra seasonal variability. In light of these, a 10 day time series has been conducted in the SCTR at 8°S 67°E from 8 June to 17 June 2014. Although mixing at the upper ocean has been recognized as an important mechanism for the nutrient
supply to productive layers in the SCTR, quantitative estimates of diapycnal nutrient fluxes in those regions are sparse. However, robust estimates of diapycnal transport of nutrients are vital for validating biogeochemical models and carbon export estimates from the euphotic zone.

### 3.4.2.1 Hydrographic settings during the time series at 8°S 67°E

Figure 3.11 shows the background condition during the time series. During the time series the SCTR latitudinal belt (5°S to 10°S) was characterized by chl concentration ~0.2 mg/m³ and out side of this region in the tropical open Indian Ocean it was < 0.1 mg/m³ (Figure 3.11a). The SLA at time series location was positive indicating deeper thermocline (Figure 3.11b). The winds were south easterly with wind stress of ~0.13 Pa (Figure 3.11c). The OSCAR currents were westward SEC (Figure 3.11d) with speed of ~0.3 m/s. The averaged GHRSSST SST and AQUARIUS SSS were 28 °C and 34.8 respectively (Figure 3.11e and f).

The MLD calculated from the CTD data in the TS location showed strong variability, varying between 20 m to 50 m (Figure 3.12a and b). The SST, SSS and surface chl were consistent with the satellite derived data (Figure 3.12a and b and c). However, the chl depth-time plot shows pulses of elevated surface concentrations (~0.5 mg/m³) especially during the midnight hours (Figure 3.12 c). A strong DCM with chl concentration ~1 mg/m³ was noted throughout the time series just below the MLD. During the time series, in the surface layers nitrate concentration was less than 1 µM (Figure 3.12 d), whereas silicate values were greater than 5 (Figure 3.12 e). During the observation, phosphate concentration was ~ 1µM (Figure 3.12 f) in the surface layer. In general the macro nutrient concentrations increased with depth. The nutrient ratio N:Si was ~1 in the upper 120m indicating favorable condition for diatom growth. Further, the N:P (<16) indicates the active utilization of nitrate by phytoplankton whereas the low Si:P (< 1) suggests that the system was limited by silicic acid.

### 3.4.2.2 Microstructure profiler observations

Following the CTD, microstructure profiler (MSS90L) casts were conducted at every six hourly interval during the time series. Details of the micro-structure
Figure 3.12: Time depth plot of (a) Temperature (b) Salinity (c) chlorophyll (d) $\text{NO}_3+\text{NO}_2$ (e) $\text{SiO}_4$ (f) $\text{PO}_4$ (g) N:Si (h) N:P (i) Si:P. The white asterisk indicates euphotic depth. The black line represents MLD.
Eddy dissipation rate ($\varepsilon$) and eddy diffusivity were derived from the microstructure shear data. During the time series the wind speed was $\sim$10 m/s and major exception was on second and third day when the wind speed reduced to 2m/s (Figure 3.13a). In accordance with the wind, incoming shortwave radiation also showed a dip during the second and third day of the time series (Figure 3.13b). In the ocean, measure of turbulence intensity is the eddy dissipation rate ($\varepsilon$). The dissipation rate depth-time plot (Figure 3.13c) shows that in the upper mixed layer the dissipation rates are of the order of $10^{-6}$ W/kg which is of one order greater than what observed by Callaghan et al. (2014) in the SCTR during winter monsoon. Further down, below the mixed layer $\varepsilon$ was $< 10^{-7}$ W/kg. The eddy diffusivity (Figure 3.13d) in the MLD was of the order of $10^{-2}$ m$^2$/s and below it ranged between $10^{-3}$ m$^2$/s and $10^{-6}$ m$^2$/s. From both Figure 3.13 c and d, it is clear that unlike in the mixed layer, in the pycnocline turbulent mixing is often not very energetic but patchy in both space and time with intermittent bursts of turbulent mixing. These intermittent and patchy spike in diffusivity is what maintaining the DCM with the supply of nutrients from subsurface through vertical mixing. Elevated levels of $\varepsilon$ in the mixed layer were due to wind and night time convection. When the wind is constant over periods longer than one day, the mixed layer often shows a strong diurnal cycle with nighttime convection due to cooling driving active mixing from the surface to the seasonal pycnocline, while during daytime, a shallower restratification may result from radiative heating (Thopre (2005)). However, intermittent bursts of turbulent mixing in the thermocline is mainly attributed to the internal wave breaking.

In ocean it has been suggested that phytoplankton of smaller sizes are more abundant than the bigger ones, although different sizes typically coexist (Barton et al. (2014)). Phytoplankton cell size typically varies from 2 µm to around 200 µm, and some species are known to grow, up to millimeters, by forming chains, filaments, coenobia and colonies (Reynolds (2006)). The picoplankton which are of the size smaller than 2 µm, is below the resolving limit of microstructure profiler. The existence of this size diversity in the water bodies of similar properties are to be fully understood. Major disadvantage of larger phytoplankton compared to the smaller ones are lower specific nutrient affinity, lower specific growth rates and self shading of photosynthetic pigments (Barton et al. (2014)). In spite of this, larger phytoplankton are noted to exist and thrive at different oceanic habitats, and one major hypothesis for this support is the high grazing pressure exerted on the small plankton (Smetacek (1999)). Another mechanism of persistence of
different species or different size scale phytoplankton may be due to the difference in micro environment experienced by the phytoplankton (Margalef (1978)).

Figure 3.13: Time series data at 8°S 67°E of (a) wind speed (m/s) (b) short wave radiation (W/m²) (c) log10 dissipation rate (W/kg) (d) log10 eddy diffusivity (m²/s) (e) Brunt Vaisala frequency (1/s) (f) Averaged log10 dissipation rates averaged in the Mixed layer (W/kg) (blue), form mixed layer depth to 80m (red) and from 80 m to 120 m (green). The shaded part represents 95% significance limit. The black line represents the MLD. The white asterisk indicates the euphotic depth.

It is of interest to discuss the length and velocity scales experienced by these phytoplankton. The energy in any turbulent region is transferred from large scale to fine scale in a cascade process (Kolmogorov (1941)). This ubiquitous fine-scale turbulence is assumed to increase the nutrient uptake (through eddy diffusion) and growth of larger phytoplankton, but smaller cells are mostly unaffected (Mann and Lazier (1996)). According to Kolmogorov’s law, at Kolmogorov length scales viscous force is important and kinetic energy will be dissipated (Kolmogorov (1941)). Thus the length scale of the smallest velocity fluctuation is the Kolmogorov length (L_k) below which viscosity becomes important. If the size of phytoplankton is smaller than L_k, it suggests that the molecular diffusion is important in the nutrient uptake mechanism rather than the eddy
diffusion. Figure 3.14a shows the diurnal variability of $L_k$ at the time series location. Smaller $L_k$ is noted in the MLD (of the order on 1mm) and at deep layer

$L_k$ was of the order of 4mm. From the $L_k$ obtained it is clear that, only the largest cells or colonies approach the size of the smallest turbulent motions, which suggests that turbulence has no effect on direct phytoplankton nutrient uptake. The nutrient distribution in ocean is patchy and turbulent stirring stretches nutrient patches into thin filaments, which are then diffused (Barton et al. (2014)). This nutrient concentration gradients occur down to the Batchelor length scale, $L_B$, (Machado et al. (2014)). This suggests that phytoplankton that grow more than $L_B$ can experience these microscale nutrient gradients and access micro-scale nutrient maxima associated with turbulent micro environment resources heterogeneity. On the other hand, some plankton that can swim, propelled by flagella, at swimming velocity greater than the Kolmogorov velocity microscale, $V_k$, can cross the nutrient gradients and access microscale nutrient maxima. Figure 3.14b shows the diurnal variability of $V_k$ at the time series location. The maximum velocity scale was noted in the MLD (of the order of 1 mm/s) and with depth it
A Special emphasis to the physical control of chlorophyll in the Seychelles Chagos thermocline Ridge

Chapter 3.

decreased. The $L_B$ for nitrate, (Figure 3.14c) also showed a similar trend that of $L_k$ with lowest value at MLD ($10^{-5}$ m) and increased at deeper depth. Buoyancy length ($L_k$) (Figure 3.14d) indicated the maximum possible excursion of phytoplankton by the microscale turbulence. which is important in the flux of phytoplankton out of the euphotic depth. The greater the turbulent energy in the system the larger the buoyancy length and smaller the Kolmogorov length scale (Falkowski and Oliver (2007)). It can be seen that the micro environment experienced by the phytoplankton is different in MLD, DCM and below the euphotic depth layers. However, the lack of phytoplankton species level analysis does not allow to conclude the role of micro environment length/velocity scales on the particular species dominance and existence of different species in the water column.

3.4.2.3 Nutrient flux

As described in the previous section SCTR is a year round upwelling region and an important region of biological production due to the nutrient supply to the surface from deeper layers. In the absence of turbulence and instantaneous nutrient consumption upwelling is a reversible transport of fluid (Schafstall et al. (2010)). However, turbulent mixing is the irreversible process which replenishes the nutrients in the surface layer from deeper layer so that nutrients are available for photosynthesis. Here the flux of macro nutrients are quantified to assess the importance of diapycanl nutrient fluxes in the SCTR. The vertical flux of nutrients by turbulent mixing was calculated by multiplying the eddy diffusivity with the vertical nutrient concentration gradients. A detailed description of the calculation is given in chapter2. The time depth plot of nutrients shows that nutricline and base of MLD are partly collocated (Figure 3.12 d, e, f). Strongest vertical gradients of nitrate concentrations occur mostly slightly below the mixed layer. The chl time depth plot shows two major feature, one the very prominent DCM centered at 50 m and peaks in surface chl close to the midnight. Thus, here the diapycnal flux of nutrients just below the mixed layer, between 50 m to 30 m was determined as it appears to be most relevant for the midnight peak in surface chl. Further to understand the maintenance DCM, diapycnal flux of nutrients were calculated between 75 m and 50 m.

The NO$_3$+NO$_2$, SiO$_4$ and PO$_4$ fluxes at the base of MLD (between 30 m and 50 m) and at the DCM (Between 50m and 75m) is shown in the Figure 3.15 a,b,c respectively. Averaged fluxes at the base of mixed layer are 2.03e-3 (1.78e-
A special emphasis to the physical control of chlorophyll in the Seychelles
Chagos thermocline Ridge
Chapter 3.

3- 2.52e-3 µmol / m² / s¹ for NO₃+NO₂, 1.7e-4 (1.48e-4- 2.10e-4) µmol / m² / s¹ for SiO₄, and 1.95e-3(1.69e-3- 2.4e-3) µmol / m² / s¹ for PO₄ (values in brackets denote lower and upper statistical uncertainty limits, see chapter2). Averaged fluxes at the DCM are 2.32e-4 (1.98e-4- 3.09e-4) µmol / m² / s¹ for NO₃+NO₂, 1.01e-5 (8.70e-6- 1.34e-5) µmol / m² / s¹ for SiO₄, and 1.65e-5 (1.41e-4 - 2.19e-4) µmol / m² / s¹ for PO₄. On average the flux values at the base of MLD exceeded the DCM flux by a factor of 10. The increased flux at the base of MLD can be due to the surface wind induced mixing. The large scatter of individual flux estimates suggest that fluxes are dominated by sporadic mixing events. In general, as discussed in previous sections major source of nutrients to the upper layers in the SCTR is wind induced upwelling, zonal advection, vertical mixing and thermocline lifting by planetary waves. However, the wind induced upwelling measured from the Satellite ASCAT wind during the time series was -4.476e-07 m/s indicating downwelling. The averaged SLA during the time series was positive indicating deeper thermocline. This clearly indicates the importance of diapycnal mixing in the region.

Figure 3.15: Time series data at 8°S 67°E of variability of nutrient fluxes at the base of MLD (blue) and across the DCM (red) with time (a) nitrate (b) Silicate (c) Phosphate (d) February 2014 monthly averaged Net primary productivity derived from the AQUA MODIS chl data.

Assuming validity of the Redfield ratio (C:N =~6.6) for phytoplankton, an averaged flux of 2e-3 µmol/ m²/ s¹ will sustain a productivity of 13.2 mg C / m
2/ d. The biological parameter to which this value should directly compare is net production. February 2014 monthly experimental net primary productivity data derived from MODIS AQUA chl showed ~150 mg C / m² / d production of carbon in the time series location (Figure 3.15d). The observed nitrate flux could not support the noted production. This is indicative of the episodic events of diapycnal mixing which sustains the DCM. Furthermore, nutrient flux cannot completely explains midnight elevated chl concentration noted at the time series location. The peaks noted in the chl were not always followed by the peaks in nitrate flux (Figure 3.15a).

3.5 Conclusion

The WTIO is generally oligotrophic, during both the WM and the SM, with a low surface chl concentration. The DCM was prominent, located at a depth of ~ 50–75 m north of 10°S. The surface chl concentration was mainly dependent on the mixed layer depth/presence of mesoscale eddies, and the DCM was determined by the nitrailine/ thermocline depth and the PAR values. The nitracline/thermocline depth was predominantly determined in turn by the Ekman pumping and Rossby waves. A highlight of the study is that different physical processes were observed to control the chl distribution within short distances in the southern Arabian Sea and the equatorial region. For instance, at 4°N during the SM, strong Ekman pumping was inhibited by the presence of Rossby waves and low-salinity surface waters whereas slightly to the north (near 6°N), eddy-induced upward transport was observed. The low-salinity surface waters encountered in this area were of Bay of Bengal origin. Together with the ASHSW at the subsurface levels, these conditions maintained a strong stratification. This stratification and the westward-propagating Rossby waves overwhelmed the local upward Ekman pumping. The influence of the low-salinity surface waters was evident in the SCTR as well. In both cases, the low-salinity waters originated from the east. Precipitation may play a role in the observed surface freshening, but the freshened layer was thick (>25 m), suggesting the significance of freshwater advection. This advection was maintained mainly by the strength and the course of the SEC. Hence, the variability of the SEC may be a major factor that affects the biogeochemistry of the WTIO. The deep mixed layer and the elevated surface chl concentration observed at the equator during the WM despite the weak winds, can be explained by the vertical shear associated with the Wyrtki jet. South of the SCTR, the anti-cyclonic eddies played a major role in deepening
the thermocline and amplifying the low productivity.

Seasonal variation of surface chl concentration in the SCTR showed a weak semi-annual signature. In spite of the year round shallow thermocline prevalent in SCTR, elevated surface chl concentration is noted only from June to October. This is associated with the low stratification and deep mixed layer observed during this period. The analysis of climatological data set suggest that the observed low stratification and deep mixed layer are primarily due to wind mixing during March - October whereas the strong stratification observed in the rest of the year is due to the weak winds and strong surface buoyancy. Another notable observation is that the WSCTR has higher chl concentration than the ESCTR, and this could be due to the presence of Mascarene plateau and many small islands. Further the WSCTR is less stratified than the ESCTR, and the gap between the base of MLD and thermocline is less in WSCTR. It appears that the seasonal surface PAR variability may not have a significant role in the seasonal variability of surface chl in the SCTR.

The time series conducted at the SCTR revealed diurnal variability in the chl distribution. A strong DCM was noted at the location through out with elevated surface chl during midnight hours. Further the diapycnal nutrient flux measured during the time series showed a value not enough to sustain the primary productivity noted at the SCTR indicating that during boreal summer the episodic diapycnal mixing events are the major source of nutrient supply. The different micro environment length and velocity scale showed a large variability both in depth and time during the time series. This variability indicates the strong influence on existence of different phytoplankton species in the region and this warrants a detailed future study of the micro environment dynamics experienced by the phytoplankton.