Influences of physical and biogeochemical variability of the central Red Sea during winter.

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Key Points:

- In situ and remote sensing observations shows for first time the persistence of a cyclonic eddy in the CRS for 4 months
- An enhancement of the surface chlorophyll does not always indicate an increase in water column phytoplankton biomass and/or productivity
- Subduction, as indicated by the DCM, occurred at the interface between an anticyclonic eddy and cyclonic eddy in the Central Red Sea

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Abstract

The central Red Sea (CRS) has been characterized by significant eddy activity throughout the year. Weakened wintertime stratification contributes to enhanced vertical exchange. In winter 2014-2015, an extended glider time series in the CRS captured this variability. Surface cooling and stronger winds resulted in deepening of the mixed layer (ML) to nearly 90 m. The vertical distributions of density and oxygen suggest that the ML did not penetrate into the nutricline. However, mixing events dispersed phytoplankton from the deep CHL maximum (DCM) throughout the ML increasing nearsurface chlorophyll. Following the mixing events a mesoscale cyclonic eddy (CE) grew and intensified causing weakening of stratification and a decrease in the ML depth within the eddy. Where the CE interfaced with an adjacent anticyclonic eddy (AE), the CE DCM subducted beneath the shallower AE DCM leading to a local integrated chlorophyll maximum. Low salinity water containing relative high chlorophyll and CDOM concentrations, originating from the Gulf of Aden, appeared in late winter. Mesoscale eddy activity resulted in an 160 m upward displacement of the nutricline to ~60 m, well within the euphotic layer. Remote sensing imagery indicates that these eddies contribute to horizontal dispersion, including exchange between the open sea and coastal coral reefs. When the phytoplankton is distributed through the ML, clear diel variability was evident in the temporal CHL distribution. Because not all of the biogeochemical responses were apparent at the surface, sustained glider observations were essential to understand the temporal and spatial scales and their impact on these processes.

1. Introduction

The Red Sea is a long, narrow basin spanning a distance of nearly 2000 km from the strait of Bab al Mandab in the south to the southern tip of the Sinai peninsula in the north. Surrounded by hot deserts, its primary linkage to the global ocean is through the constricted strait of Bab el Mandeb, connecting it to the Arabian Sea through the Gulf of Aden. The Red Sea is characterized by negligible fresh water input from rainfall and terrestrial runoff, high temperatures and high salinity due to solar heating and significant evaporation rates (Smeed 1997; Sofianos 2002; Berumen et al. 2019) and is considered to be a predominantly oligotrophic sea (Grasshoff 1969; Edwards and Head 1987). In addition, the Red Sea contains one of the most diverse marine ecosystems in the world with important economic and environmental impacts primarily due to its coral reefs (Belkin 2009; Longhurst 2007; Raitso et al. 2013). Coral reefs, often referred to as “marine rainforests”, contain one of the most diverse ecosystems in the oceans, yet they are fragile and vulnerable to oceanic warming and other anthropogenic pressures (Cantin et al. 2010).

Wind and thermohaline forces drive the large-scale circulation of the Red Sea (Patzert 1974; Phillips 1966). Until recently, in situ observations of the Red Sea have been limited (Bower et al. 2013; Bower and Farrar 2015; Kürten et al. 2019; S.Morcos 1974; Qurban et al. 2014; Sofianos and Johns 2007; Zarokanellos et al. 2017a,b) and our knowledge regarding the variability of the basin-scale circulation has relied primarily on ocean models and remotely sensed observations (Clifford et al. 1997; Sofianos and Johns 2002; Sofianos 2003; Tragou et al. 1999; Yao et al. 2014b,a; Zhan et al. 2014). Numerical simulations and remotely sensed observations indicate that much of the time, mesoscale eddies often prevail over much of the basin and can propagate both zonally and meridionally (Raitso et al. 2013; Yao et al. 2014a; Zhan et al. 2014; Chen et al. 2014). Chen et al. (2014) showed that mesoscale eddies play a significant role in the transport of heat, salt and biological and chemical constituents in the Red Sea. Eddies provide valuable ecosystem services for the reef system, pelagic species, and marine mammals (Shulzitski et al. 2016). Although mesoscale processes seem to have a substantial effect on the Red Sea ecosystem, the interaction between eddies and the coastal region, and their role in the modulation of phytoplankton productivity and biomass within the Red Sea have remained poorly understood.

In stratified ocean regions like the CRS where vertical exchange is typically limited,

mesoscale features likely play a fundamental role in regulating phytoplankton dynamics via local nutrient fluxes (Longhurst 2007). Eddy activity influences both optical and biological properties in the oceans (Kürten et al. 2016; Mahadevan 2014; McGillicuddy et al. 1998; Siegel et al. 2008; Pearman et al. 2017). Eddies contribute to the vertical transport of nutrients into the upper layer of the water column (Martin and Richards 2001; McGillicuddy et al. 2003). Eddy transport has already been shown to affect biogeochemical variability in the CRS (Zarokanellos et al. 2017a). Other processes can also contribute to biogeochemical fluxes in the upper layer, including seasonal mixing, eddy interactions, and cross-shelf eddy interaction. Until now, only remote sensing studies have been utilized to examine the seasonal variability of phytoplankton (Acker et al. 2008; Racault et al. 2015; Raitsos et al. 2013; Gittings et al. 2018). Despite remote sensing's ability to provide synoptic imagery of ocean variables, persistent clouds, atmospheric aerosols, sun-glint, and sensor saturation overland limit data acquisition for the Red Sea (Racault et al. 2015). However, satellite imagery suggests that there is an ecological connection between the coastal zone and the central open sea (Raitsos et al. 2017). This is significant, especially for the coral communities that live at the basin edge (Acker et al. 2008). Conversely, satellite remote sensing is unable to inform us regarding the subsurface distribution phytoplankton as indicated by chlorophyll (CHL), as ocean color measurements are limited to the first optical depth.

The Red Sea is experiencing increased anthropogenic pressure due to hydrocarbon exploration, aquaculture, urban and industrial development, proposed deep-sea mining, climate change, and tourism. One likely ecological response to increased environment stress will be change in biological productivity. Therefore, long-term observations of biogeochemical variables related to primary productivity and carbon cycling are essential. Primary production is not only the base for the marine food chain but it also a fundamental pathway in the biogeochemical flux of carbon (Green et al. 2014). Autonomous underwater vehicles (AUVs) provide a tool to sustain observations of biogeochemical properties over a range of spatial and temporal scales, increasing our knowledge and enabling us to reduce uncertainties associated with carbon budgets (Cetinić et al. 2012; Perry et al. 2008). Eddies and fronts can enrich the upper layer with nutrients, enhancing primary production often resulting in increased particle concentration reflecting particulate organic carbon (POC), increased CHL and oxygen production (Li et al. 2012; Ohman et al. 2012; Stramma et al. 2013). Horizontal transport and eddy stirring can contribute to the carbon flux in a given region (McGillicuddy 2016). The distributions of biogeochemical

variables including CHL, suspended particles, nutrients and oxygen demonstrate the coupling between various scales of physical forcing and variability with biogeochemical dynamics (e.g. Mahadevan et al. 2012; Mignot et al. 2014). The use of bio-optical variables from both remote sensing and in-situ observations has yielded new insights into the biogeochemical functioning of Red Sea (Acker et al. 2008; Brewin et al. 2015; Gittings et al. 2018; Racault et al. 2015; Raitso et al. 2013; Tiwari et al. 2018b; Zarokanellos et al. 2017b; Kheireddine et al. 2020)

In a phytoplankton dominated regime a number of optical properties will covary with CHL concentration (Bricaud et al. 1988; Siegel et al. 2005), but other non-algal constituents are capable of absorbing and scattering light, such as detrital particles, CDOM, and suspended solids, which all contribute independently to ocean color (Loisel et al. 2002, 2007; Maritorena and Siegel 2005; Tiwari et al. 2018a). The present study aims to address the influence of wind mixing, mesoscale eddies, and lateral advection on the optical properties. Furthermore, this study examines how these physical drivers modulate the optical properties in the surface and subsurface of the CRS during the winter period.

2. Data Sources and Methodology

2.1 Glider Observations

Data from two Seaglider missions were obtained as part of the ongoing research effort in the CRS (Figure 1a; Eriksen et al. 2001). The glider was deployed along a 250 km long L-shaped pattern for a period of 91 days during the winter period from December 2, 2014 to March 2, 2015 (Figure 1b & Table 1). The offshore turning point of the L-shaped pattern targeted toward the center of a recurrent eddy that previous remote sensing and in situ studies have indicated (Raitso et al. 2013; Zhan et al. 2014; Zarokanellos et al. 2017b). Each glider deployment began offshore of the Shi'b Nazar reef system (located at 22°16'12 N, 38° 57'44 E; reef not shown in Figure 1b), and the glider was piloted northwestward to the offshore turning point (C) where it turned northeastward toward the northern end point (D) near Yanbu. The glider then returned to the southern waypoint retracing the same track, completing the round trip in about three weeks. Each Seaglider was equipped with a suite of sensors that measured temperature and salinity (Sea-Bird Electronics' custom CT sensor with an unpumped conductivity cell), dissolved oxygen (Optode 4831 sensor, Aanderaa), CHL, CDOM and phycocyanin (FL3-IRB sensor, WETLabs), and a three-channel optical backscatter sensor (λ : 532, 650 and 880 nm; BB3 IRB sensor,

WETLabs). We present only the optical backscatter at 532 nm, as the overall patterns for 650 nm and 880 nm wavelengths were very similar to those at 532 nm. In both glider missions, the optical face of the FL3 and BB3 sensor appeared clean with little to no fouling after each deployment. Any fouling that occurred probably happened during the brief time that the gliders were in the euphotic zone (<6%).

Potential temperature (hereafter, pot. temperature) is the temperature at depth corrected for pressure relative to the ocean surface. Both Seaglidors were programmed to profile from the surface to 1000 m at a nominal vertical speed of 0.18 ± 0.02 m/s and a horizontal speed of approximately 20-24 km day⁻¹. Each dive cycle required 3-5 hours to complete depending on the operational target depth, environmental conditions, and engineering control settings. However, for this study we only consider the upper 400 m layer of the water column.

The pre- and post-calibration activities were carried out at KAUST's Coastal and Marine Resources Core Lab (CMRCL) glider facility and included all ballasting, adjustment operations, and optical sensor dark counts measurements taken before and after each glider deployment. More information regarding the manufacturer's calibration of the ECO Puck's fluorometric measurements could be found in Cetinić et al., (2009). Moreover, a comparison was made with CTD cast from hydrographical expedition in the study area, with the assumption that below 500 meters the water mass changes slowly relative to the time scale of sampling such that the T-S plots from both modes of sampling should tightly overlay each other (Zarokanellos et al. 2017a).

Evaporation rates within the Red Sea are of the order of 2 m year⁻¹ (Sofianos 2002). The high evaporation rates contribute to high salinity and significantly affect the upper layer mixing (Lee et al. 2000). The mixed layer depth (MLD) was defined as the vertical depth range in which the density variation was less than or equal to 0.03 kg m^{-3} (de Boyer Montégut 2004). The MLD was calculated using 1 m binned vertical profile data. Spikes from the raw hydrographic measurements were removed. In addition, CTD data were examined for thermal lag effect and no evidence of its effect was observed.

FL3 and BB3 sensors were manufacturer calibrated. Negative spikes and instrument dark signal were removed from the fluorescence profiles. In addition, a global bias correction factor of two was applied for the fluorescence profiles based on the global comparison of fluorometer and extracted CHL measurements (Roesler et al. 2017). Euphotic depth (Zeu) was estimated using the vertical chlorophyll distribution and a model of chlorophyll-dependent light attenuation

(Morel and Maritorena 2001).

Regions with high irradiance like the Red Sea often experience chlorophyll fluorescence quenching during the daylight hours (Behrenfeld 1999; Kinkade et al. 1999). Chlorophyll fluorescence quenching tends to occur near the surface, but it can affect fluorescence below the upper few meters, especially in waters with a low downwelling irradiance attenuation coefficient, K_d . Seasonal variability of photochemical quenching has been observed, particularly during spring and summer when solar radiation increases, with the intensity of fluorescence quenching reduced at greater depth. CHL concentrations were corrected for quenching where needed, based on the mixed layer depth, using as a reference the MLD (Sackmann et al. 2008; Xing et al. 2012).

Raw scattering counts of optical backscatter minus dark counts were converted to total volume scattering $\beta(124, \lambda)$ using a factory-calibrated scale factor. The data from sensors were converted to the engineering units by using the manufacturer calibrations:

$$|\beta(\theta_c), \text{Chla}, \text{CDOM}| = \text{scale factor} \times (\text{output} - \text{dark count}) \quad (1)$$

$$\text{Scale Factor} = \frac{a}{(\text{output}_a - \text{dark counts})} \quad (2)$$

or,

$$\text{Scale Factor} = E \left(\frac{b_b(\lambda)}{\text{dark counts}} \right) * T(\beta(\theta_c)/b_b(\lambda)) \quad (3)$$

where scale factor (SF) from (2) or (3) is used in (1) for the computation of CHL (mg m^{-3}), CDOM (ppb), or total volume scattering for a given wavelengths ($\beta(\theta_c)$; $\text{m}^{-1} \text{sr}^{-1}$), respectively. In (2), the known standard concentration, a , is the constituent concentration of either CHL or CDOM and output_a is the measured signal output for each standard concentration. Furthermore, in equations (3) is the total backscattering coefficient (m^{-1}). Dark counts is the sensor signal in the absence of light, which has been measured placing black electrical tape over the detector, and output is the measured signal output during field sampling. E and T in (3) are used to denote experimental and theoretical fractions, respectively. A detailed description of these optical computations is found on the official Seabird website (<https://www.seabird.com/auv-rov->

sensors/eco-puck/family?productCategoryId=55352274651).

From $\beta(\theta_c)$, the volume scattering of particles (β_p) was calculated by subtracting the volume scattering of water (β_w) which is a function of temperature and salinity (Zhang et al. 2009; Morel 1974). The particulate backscattering coefficients (b_{bp} ; m^{-1}) from β_p and total backscattering coefficients (b_b ; m^{-1}) was then calculated from β_p according to Boss and Pegau (2001):

$$b_{bp}(532) = 2\pi\chi\beta_p(532) \quad (4)$$

where χ is 1.1.

POC was estimated from b_{bp700} using the relationship described by Cetinić et al. (2012). Because b_{bp700} was not directly measured, the spectral slope (γ) that was calculated from the 3 measured wavelengths (532, 650 and 880 nm) was then used to estimate b_{bp700} from the nearest measured wavelength at 650 nm (Loisel et al. 2006; Tiwari and Shanmugam 2013). The resulting calculation of b_{bp700} from the measured b_{bp650} is given by

$$b_{bp700} = b_{bp532} * \left(\frac{700}{650}\right)^\gamma \quad (5)$$

The calculated γ values are consistent with the results presented by Loisel et al. (2016) where low chlorophyll waters are associated with small-sized particles and high values γ (mean= 1.7214; STD $\pm$ 0.97).

The oxygen sensor was manufacturer calibrated. Lacking direct in situ oxygen measurements, the oxygen measurements were adjusted by a correction factor based on the median oxygen saturation in the surface layer. The correction factor was the ration of the median oxygen saturation concentration for the upper 10 m to the median measured concentration within the upper 10 meters of the water column. The entire water column data was then corrected by multiplying the reported oxygen concentrations by the correction factor. The correction provided consistency between the current glider data set and the vertical distribution for this region in the World Ocean Atlas (Garcia et al. 2019). Apparent oxygen utilization (AOU) was computed as the difference between oxygen saturation and observed oxygen concentrations ($\mu\text{mol kg}^{-1}$). Oxygen saturation, or solubility, was computed from salinity and temperature using the TEOS-10 GSW Oceanographic Toolbox (IOC et al. 2010)

The relationship between apparent oxygen utilization (AOU) and nitrate described by Naqvi et al. (1986a). Churchill et al. (2014) provided a more recent evaluation consistent with

Naqvi et al.'s original evaluation. The relationship reveals that nitrate concentrations above 1 $\mu\text{M/L}$ occur when the oxygen concentration is less than 180-185 $\mu\text{mol kg}^{-1}$ (not shown). Thus, the 180 $\mu\text{mol kg}^{-1}$ oxygen isopleth is used as an index of the nitracline.

2.2 Atmospheric Reanalysis Products

The upper layer circulation of the CRS is strongly affected by the monsoonal wind patterns. Between fall and spring, the winds tend to converge in the south central to central Red Sea and then transition to a more uniform pattern of northwesterly winds throughout the Red Sea during summer (Abualnaja et al. 2015; Sofianos and Johns 2003; Zarokanellos et al. 2017b). Although northwesterly winds prevail over the CRS throughout the year (Papadopoulos et al. 2013), it is unclear how air temperature, wind stress, relative humidity and net air-sea heat exchange (Q_{net}) affect the MLD in the Red Sea. Atmospheric reanalysis observations are used to investigate the relationship between the atmospheric forcing and MLD in the CRS. The Q_{net} is defined as: $Q_{\text{net}} = Q_{\text{Sw}} + Q_{\text{Lw}} + Q_{\text{s}} + Q_{\text{L}}$, where Q_{Sw} is the shortwave radiation, Q_{Lw} the longwave radiation, Q_{s} the sensible heat flux, and Q_{L} the latent heat flux. Wind stress was calculated from wind speed using the method of Large and Pond (1981). We use daily mean values for the deployment period from December 2, 2014 to March 2, 2013 retrieved from NASA's Modern Era Retrospective Analysis for Research and Applications (Rienecker et al. 2011; Al Senafi et al. 2019) with a resolution $2/3^\circ$ by $1/2^\circ$. The spatial average of the air temperature, wind stress, relative humidity and Q_{net} has been estimated over the study area (between 21.9° and 24.1° N and from 38.1° to 38.8° E).

2.3 Satellite data

To place the glider observations into the larger context of the CRS two relatively clear daily MODIS Level-3 images with 4 km horizontal resolution were obtained from NASA's Ocean Color Web (<http://oceancolor.gsfc.nasa.gov/>) for 2 and 7 February 2015. Sea level anomaly (SLA) observations are used to detect and characterize the dynamical characteristics in the CRS during the observation period. The SLA observations are from Copernicus Marine Environmental Monitoring Service (CMEMS) website (https://resources.marine.copernicus.eu/?option=com_csw&view=details&product_id=SEALEVEL_GLO_PHY_L4_REP_OBSERVATIONS_008_047)

2.4 Statistical analysis

Pearson correlation analysis was applied to examine the relationship of integrated CHL, CDOM, b_{bp532} , and POC in the ML with the atmospheric forcing (air temperature, wind stress, and relative humidity). The transect line was subdivided into three portions (AB, ABC, and CD) as indicated in Figure 1b. Then mean values for CHL, b_{bp532} , CDOM, POC, and POC:CHL within the mixed layer were calculated for each subdivision. In addition, principal component analysis (PCA) was used to investigate the linear relationship between variables. In order to standardize the data, because the variables have different units and magnitudes, the PCA was performed using the correlation matrix instead of the covariance matrix. The following data were used in the PCA analysis: temperature, salinity, density, oxygen, CDOM, b_{bp532} , and CHL. The data were divided in four depths categories to explore the vertical distribution and the interactions of the physical and biochemical processes in the CRS. We partitioned data into the following depth bins (surface-to maximum depth, surface-20 m, 20-100 m and 100-200 m).

3 Results and Discussion

3.1 Temporal evolution

3.1.1 Atmospheric conditions in the central Red Sea during winter

Model analyses of the Red Sea indicate the importance of the atmospheric forcing in the Red Sea circulation (Clifford et al. 1997; Sofianos and Johns 2003; Yao et al. 2014b,a). The location of the convergence zone between northerlies and southerlies in the central Red Sea plays a crucial role in the forcing of Red Sea circulation and surface waves during winter when winds in the southern half of the Red Sea are southerly (Langodan et al. 2015). Air temperature, wind stress and relative humidity reveal the arrival of cold, intense northerly winds (Figures 2a and 2b). Glider temperature 5m beneath the surface follows a similar pattern with the air temperature, with an average time lag of 8 days based on cross-correlation analysis (Figure 2a; cross-correlation analysis not shown). Wind stress of more than 0.005 N m^{-2} is associated with negative Q_{net} and results in deepening of the ML (Figures 2b-2d). The average Q_{net} was negative ($-129.6 \pm 148.6 \text{ W m}^{-2}$) during the observation period (in this study negative Q_{net} flux refers to heat transferred from the ocean to the atmosphere; Figure 2c)

Air temperature initially declined to a minimum on January 11 ($\sim 19^\circ \text{C}$) following a period of strong winds and declining humidity (Figures 2a and 2b). Coincident with air

temperature minimum and strong wind stress, the ML reached its maximum observed depth, 88m, during the deployment period (Figures 2a and 2d). The ML then shallowed during the next 10 days to between 10 and 20m depth. Two subsequent, brief atmospheric events (centered on February 12 and February 20) were also characterized by cooler air temperature, higher wind stress, and higher Q_{net} loss (Figures 2a-c). During the first February event, the ML deepened to more than 30m ($Q_{net} \sim -310 \text{ W m}^{-2}$) and during the second event to more than 60m ($Q_{net} \sim -410 \text{ W m}^{-2}$). Following the second event, air temperature increased to more than 24°C and wind stress decreased to $\sim 0.0045 \text{ N m}^{-2}$. Despite the warming air temperature and reduced wind stress, the ML remained at about 60m (Figures 2a and 2b). We thus conclude that other processes such as lateral advection contributed to maintenance of this local deeper ML. We examined the dependence of the MLD on wind stress, air temperature and relative humidity to understand the importance of atmospheric forcing during the winter period. An overall correlation between the MLD with wind stress ($|r|=0.42$; $p<0.05$), air temperature ($|r|=0.32$; $p<0.05$), relative humidity ($|r|=0.22$; $p<0.05$) and Q_{net} ($|r|=0.51$; $p<0.05$) is weak. Despite the weak correlation of MLD with the components of atmospheric forcing, the ML depth ranged between 8m and 88m (mean $=34.8\text{m} \pm 24.4\text{m}$). The MLD exceeded 80m (Figure 2d) for a few days during 3 periods: 22 December 2014, 29 December 2014, and 10 January 2015. During each period, phytoplankton was uniformly distributed throughout the mixed layer (Figure 2e).

3.1.2 Impact of atmospheric forcing in the upper layer

The glider time series of salinity and CHL reveals the influence of atmospheric forcing in the upper layer. During this period, salinity and CHL variability ($<100\text{m}$) was affected by both atmospheric forcing and mesoscale oceanographic features (Figures 2d and 2e). In particular, the temporal evolution of surface salinity and CHL (5m) reveals significant changes that in an extreme case reached up to 0.81 psu and 0.2 mg m^{-3} , respectively, over a two day period. The $180 \mu\text{M}$ oxygen isopleth (indicative of the nitracline depth) ranged from as shallow as $\sim 60 \text{ m}$ to a maximum depth of $>200\text{m}$ (Figure 2e). Despite the large variations in nitracline depth, Zeu only varied between 83.5m to 126.2m (mean $104.4\text{m} \pm 7\text{m}$; Figure 2e).

Based on the atmospheric forcing and the ocean response, we have separated the time series into three distinct periods: a) the first period, from early December through mid-January, was characterized by mixing events to more than 50m as shown in Figure 2d. While deepening

of the ML was often associated with increased wind stress, not every wind stress increase yielded deepening of the ML (Figures 2b and 2d); b) during the second period, from 11 January through 10 February 2015, an apparent cyclonic eddy (CE) intensified as evidenced by the elevation of 27.5 kg m^{-3} isopycnal in the offshore region from 135m in late December to less than 20m by mid-January. During this period, Q_{net} gradually increased and became positive at the end of the period (Figure 2c). Simultaneously, the ML became progressively shallower (Figure 2d). During the last period, the CE is still evident, but warmer, less saline water appears in the upper layer suggesting advection of water into the region from the south. This warmer, low salinity water first appeared when the glider was southbound along the last complete transit of the glider pattern (see section 3.2; Figures 3i and 3j).

Admittedly, the feature that is identified as a cyclonic eddy could simply be a region of persistent geostrophic shear, and the appearance of it as an eddy could be an artifact of the glider trajectory across this frontal boundary. However, sea surface altimetry indicates that this is part of a larger cyclonic circulation centered to the northwest of the EEZ (discussed later in the manuscript).

3.2 Physical and chemical variability

The glider time series captured the evolution of salinity, potential temperature, stratification (Brunt–Väisälä Frequency; BVF) and oxygen structure in the upper 300 m through the winter period (Figure 3) for the section ACD (Figure 1b). The in situ observations illustrate the importance of a) winter mixing, b) the physical and chemical characteristics of the CE, c) and the interaction of the eddy with warm, low salinity intrusion water.

During the period of winter mixing, the glider observations reveal that the mixed layer, characterized by weak stratification ($N^2 \sim 0$), extended to a maximum depth of nearly 90m (Figure 3). This weak stratification extended to the depth of the permanent pycnocline, located between the 26 and $27.5 \text{ kg m}^{-3} \sigma_\theta$ surfaces (Figure 3c). Winter mixing thus oxygenated the upper layer to the depth of the mixed layer. During this period, a small cyclonic feature centered on the glider's offshore turning point at $\sim 23.3^\circ \text{N}$ was apparent. The effect of the eddy was to elevate the pycnocline reducing the MLD and lifting lower oxygen water into the upper layer (Figure 3d). A patch of high oxygen ($185\text{--}205 \text{ mol kg}^{-1}$), high salinity ($40.3\text{--}40.38 \text{ psu}$) and comparatively warm ($24.8\text{--}25.6^\circ \text{C}$) water was embedded in the center of the CE between the isopycnals of 26 and

27.5 (Figures 3a-b and 3d).

After the winter mixing period, the CE appears to intensify (Figures 3e and 3f). Within the upper 100m, temperature decreased ($\Delta\theta \sim -2^\circ\text{C}$) and salinity increased ($\Delta S \sim 0.3$) in a period of about 10 days (not shown here). The undulations of the pycnocline, isohalines and oxygen isopleths reflect the influence of CE on the local hydrography (Figures 3f-j). In the upper 100 m, the core of the CE was saltier ($S > 40$) and cooler ($\theta < 24.5^\circ\text{C}$) than the water to either side of the feature. By early February, a lens of warmer ($> 25^\circ\text{C}$), less salty (< 39.8) water appeared south of the eddy center. We assume that lower salinity water originated from the Gulf of Aden, given that no other known freshwater sources contribute to the Red Sea. Isopycnal slopes are consistent with an anticyclonic eddy (AE) circulation but the observations are insufficient to wholly resolve this feature. During the eddy intensification period, the stratification in the pycnocline was significantly weaker than in late December (Figures 3c and 3g). The appearance of the warm, low salinity water at the end of January restratified the water column, and significantly reduced the mixed layer depth. The seasonal pycnocline as indicated by the BVF followed the 27 kg m^{-3} isopycnal. The high BVF associated with 28 kg m^{-3} isopycnal suggests that this density surface is good indicator of the permanent pycnocline. During this period, the CE had an important role in modulating the vertical oxygen distribution in the CRS (Figure 3h). The highest oxygen concentrations ($> 200\text{ }\mu\text{mol kg}^{-1}$) occurred at the periphery of the CE where the maximum concentrations penetrated to 90m. The thickness of the high oxygen layer ($> 200\text{ }\mu\text{mol kg}^{-1}$) decreased from 90m on the periphery to less than 60m in the core of the CE. Based on Naqvi et al. (1986) and Churchill et al., (2014) NO_x concentrations of $\geq 1\text{ }\mu\text{M}$ begin to appear at an AOU of about $20\text{ }\mu\text{mol kg}^{-1}$, corresponds to an oxygen concentration of $\sim 180\text{ }\mu\text{mol kg}^{-1}$. This oxygen concentration typically occurs at densities $\geq 28\text{ kg m}^{-3}$. Within the CE we see an elevation of this isopycnal from more than 190m to less than 125m (Figure 3h). In the periphery of the CE, high oxygen concentrations ($> 180\text{ }\mu\text{mol kg}^{-1}$) extended to a depth of approximately 200 m.

In February, low salinity, warmer water intruded into the region within the upper 100m in the southern part up to 23.1°N latitude (Figures 3i and 3j). This warmer, fresher water is consistent with entrainment of water from the northward transport of Gulf of Aden water as was described for an eastern boundary current (EBC) by Zarokanellos et al., (2017a and 2017b). The CE entrains the low salinity along its periphery and likely contributes to the northward distribution of this fresher water (Figure 3j). Zarokanellos et al. (2017b) observed northward

transport of low salinity water via an EBC coexisting with an AE in an earlier study in spring 2014. In that case, the AE either blocked or redirected the EBC offshore. During the low salinity intrusion period, the stratification (BVF) increased in the upper 100 m (Figure 3k). The warmer, fresher water contributed to increased stratification, as it was less dense than other water present in the area. Within the CE, the 28 kg m^{-3} isopycnal rises from 235 m to ~ 115 m in the core of the eddy. The CE lifts low oxygen, assumed to be nutrient-rich water based on $180 \text{ } \mu\text{mol kg}^{-1}$ oxygen isopleth, toward the upper layer (Figure 3l). The reduction of near surface oxygen (maximum $\text{O}_2 < 205 \text{ } \mu\text{mol kg}^{-1}$) in the center of CE compared with the eddy intensification period is consistent with decreasing saturation with increasing temperature (Figure 3l).

3.3 Water masses characteristics

Water masses characteristics underwent significant transition during the winter. The pot. temperature-salinity (Θ/S) diagram color-coded with CHL from both glider missions is shown in Figure 4. The separate panels represent the following periods: a) winter mixing, b) the eddy intensification phase, and c) the interaction of the eddy with the warm, low salinity intrusion water. The most distinctive feature of the Θ/S distribution is the very abrupt transition from a relatively diffuse Θ/S distribution in the upper layer prior to mid-January to a tight, linear Θ/S pattern in the upper layer after mid-January (Figure 4b). This abrupt transition happened during the intensification of the CE, when the 27.5 kg m^{-3} isopycnal rose near the surface. Figure 4c shows the distinct arrival of lower salinity water containing higher CHL concentrations ($> 0.3 \text{ mg m}^{-3}$) in the salinity range between 39 and 39.4 and density anomalies between 26 and 26.5 kg m^{-3} , consistent with Gulf of Aden (GAW) (Churchill et al. 2014; Zarokanellos et al. 2017a). The very linear transition between the surface water and the water at the density 27 kg m^{-3} is indicative of diapycnal mixing between the incoming water and the extant deeper water. The water was detectable from the southern end of the glider transect until $\sim 23^\circ \text{N}$, where it intersected with the CE (Figure 4i). However, given that the most likely winter inflow at Bab al Mandab is from Gulf of Aden surface water (Smeed 2004; Sofianos et al. 2002; Yao et al. 2014b) and the linearity of the mixing line between the deeper water and the surface water we interpret this as Gulf of Aden surface water, not the intermediate water that was observed in summer and early fall.

3.4 Bio-optical variability

Remotely sensed ocean color has shown that nearsurface chlorophyll concentrations increase in the CRS during winter (Raitso et al. 2013). Glider observations indicate that when the mixed layer deepens the mixing disperses the CHL from the DCM throughout the ML (Figure 5a). When this occurs, the signal detected by ocean color imagery increases. Within this glider dataset the correlation between the integrated CHL concentration (0-200 m) and the nearsurface chlorophyll concentration, within one optical depth of the surface, is not significant (not shown). During the mixing period, the patterns of b_{bp532} ($\sim 10^{-3}$) generally follow those of CHL concentration (Figure 5b). Thus, CHL concentration covaried with b_{bp532} as shown in Figures 5a and 4b. Within the CE, the elevated pycnocline inhibits the mixed layer penetration, and CHL and b_{bp532} remain in a subsurface maximum, a transposition of the DCM associated with elevation of the pycnocline. The CHL: b_{bp532} ratio was low where the CHL mixed in the water column (Figure 5c). Moreover, a high CHL: b_{bp532} ratio occurred within the subsurface CHL maximum when the euphotic layer was stratified. In the deep mixing case, the deep ML (88 m) intersected with the depth of the $180 \mu\text{mol kg}^{-1}$ isopleth of oxygen suggesting that nutrients could have been entrained into the ML. We note that during the period of deeper mixing temporal banding was evident in the CHL and b_{bp532} distributions (Figure 2e). These fluctuations are evidence of the daily water column productivity, but not the focus of this discussion. During the mixed period, low CDOM concentration (<0.5 ppb) were found in the ML (Figure 5d). Below the ML, CDOM generally increased with depth, with the maximum concentration appearing in the oxygen minimum zone (OMZ; not shown here). The $180 \mu\text{mol kg}^{-1}$ isopleth of oxygen appears to delineate the boundary below which CDOM begins to increase (Figure 5d).

As the CE intensified, rapid changes in the density field reflect changes in the CE structure. The highest CHL concentration occurred in the center of the mesoscale CE eddy where the MLD shoaled to ~ 20 m (Figure 5f). Deepening of CHL and b_{bp532} to more 100 m at the southern periphery (left side of figure) of the CE, suggest subduction of the eddy pycnocline between 27.5 and 28 km m^{-3} isopycnal beneath the adjacent upper layer (Figures 5f-g). The b_{bp532} within the 100m of the CE was higher than in the upper 100m of the surrounding water column. During this period, the highest POC concentration value occurred in the core of the CE reaching up to 24 mg C m^{-3} (not shown here). A higher CHL: b_{bp532} ratio was observed in the DCM as indicated at Figure 5g. In the broad sense, CDOM was negatively correlated with dissolved

oxygen so that for oxygen concentrations $<180 \mu\text{mol kg}^{-1}$ CDOM concentrations increased above 1 ppb (Figures 5d, 5h and 5i). However, when the low salinity water mass appeared in the region, a band of relatively high CDOM (>1 ppb) was occurred in the upper layer of the low salinity water mass between the density surfaces of 25.75 to 26.5 kg m^{-3} . The presence of CDOM in this water mass is further confirmation of its origin in the Gulf of Aden, as this is consistent with the results presented by Zarokanellos et al. (2017b) where low salinity and elevated CHL and CDOM at intermediate depths characterized Gulf of Aden Intermediate Water (GAIW) in the CRS.

During the warm, low salinity intrusion period, higher CHL concentration ($>0.4 \text{ mg m}^{-3}$) was observed between 50 and 90 m, where the density anomaly was approximately 26 kg m^{-3} and in the upper part of the high elevated CDOM region described above (Figure 5i). Figure 5j shows that higher $b_{\text{bp}532}$ ($>10^{-3} \text{ m}^{-1}$) was also associated with this warmer, fresher water mass.

In the core of the CE where the density surfaces were shallowest, higher particle concentrations ($b_{\text{bp}532} > 1.2 \times 10^{-3} \text{ m}^{-1}$) and CHL (0.4 mg m^{-3}) were found in the density range of 27.25-27.75 kg m^{-3} (Figures 5i and 5j). Through this period, CHL and $b_{\text{bp}532}$ covaried each other (Figures 5i and 5j). Subduction is indicated by the downward projection of CHL, $b_{\text{bp}532}$, and low CDOM water along the 27.5 kg m^{-3} isopycnal beneath the eddy interface southward from the eddy center to nearly 200m depth (Figures 5i and 5j). As a result, the highest water column-integrated chlorophyll concentrations occurred at interface of the intrusion and the CE where the DCM from the intrusion water (located on the 26 kg m^{-3} isopycnal) and from the cyclonic eddy (DCM between 27 and 27.5 kg m^{-3} isopycnal) overlapped. These results are consistent with the growing body of literature that demonstrates the importance of eddies and fronts in the physically-driven, downward flux of organic matter beneath the euphotic zone (Ashjian et al. 2006; Jones et al. 1991; Kadko et al. 1991; Shih et al. 2015; Stukel and Ducklow 2017; Washburn et al. 1991). The CHL: $b_{\text{bp}532}$ ratio was higher in the DCM layer and at intermediate depths coinciding with the 26 kg m^{-3} isopycnal (Figure 5k). Low CDOM concentration (<0.5 ppb) is present at intermediate depths in the center of the CE (Figure 5l). In addition, the 180 $\mu\text{mol kg}^{-1}$ oxygen isopleth followed the vertical distribution of the upper bound of the CDOM maxima (Figure 5l).

3.5 Physical and biological variability in time and space

The CRS has been characterized as having high mesoscale activity and yet is likely the most oligotrophic region of the Red Sea (Raitsos et al. 2013; Zhan et al. 2014). The glider observations demonstrate significant physical and biogeochemical changes in a relatively short period during the winter. Vertical profiles of the mean and standard deviation (STD) of salinity, pot. temperature, CHL, CDOM and b_{bp532} illustrate the magnitude of the variability that was observed in this period (Figure 6). The data was parsed into three time-periods as described earlier in the paper. The bulk of the variability, particularly in optical variables, occurred within the upper 200m (Figure 6). This is consistent with previous observations of eddy structure in CRS where evidence of the eddy penetrated to at least 200m (Zarokanellos et al. 2017b). Throughout the winter mixing period, the upper layer is more homogeneous, particularly in temperature, salinity, and b_{bp532} . The greatest variability was observed during the eddy intensification period. This variability was likely due to seasonal transition and significant horizontal exchange through the interaction of complex eddy activity. The upper layer variability was again reduced when the intrusion of Gulf of Aden water constrained the variability of the upper layer.

Figure 7 shows the statistics for the variables of CHL, b_{bp532} , and CDOM for six latitudinal transects (line AB) and four longitudinal onshore-offshore transects (line CD) between the upper 20 m CHL versus the integrated CHL of 200 m. The surface-20m CHL shows two-fold variability, while the integrated chlorophyll shows less than 25% variability. Figure 7 shows the statistics for the variables of CHL, b_{bp532} and CDOM for six latitudinal transects (line AB) and four longitudinal onshore-offshore transects (line CD). The increase of the CHL in the surface-20m zone results from deep mixing which entrains the DCM into the mixed layer but without a net increase of the integrated water column CHL. This mixing is evident in the CHL distribution and MLD shown in Figure 4a. During the eddy intensification period, CHL decreased in the upper 20 m, but without a significant change in the integrated water column CHL (Figures 4a and 7). However, along the line CD surface-20m b_{bp532} , integrated CDOM and integrated b_{bp532} increased during the eddy intensification phase. These increases are likely due to enhanced along coast transport. During the interaction between the eddy with the warm, low salinity intrusion, both surface-20m and 200m integrated b_{bp532} increased on line segment AB along with a small increase in 0-200 integrated CHL. Despite the increase of integrated CHL the

surface-20m CHL continued to decrease (Figures 7a-b). These results indicate that b_{bp532} and CHL are not necessarily well correlated. Advection of Gulf of Aden Water containing elevated CDOM contributed to the CDOM variability in the region (Zarokanellos et al 2017b). Both surface and integrated CDOM were higher from mid-January onwards (Figures 7c and 7f). Near surface (0-20m) CDOM was highest on CD-4 during the interaction between the eddy with warm, low salinity intrusion. Thus, GASW appeared in the CRS in the late winter bringing with elevated b_{bp532} . CHL and CDOM The lack of correlation between CHL and b_{bp532} could result from photoacclimation following the stratification imposed by the intrusion of the low salinity water.

3.6 Environmental factors controlling the optical properties

As described above, significant physical and optical variability occurred during the winter period. To further characterize this variability, principal component (henceforth PC) analysis was performed to determine the data's main modes of variability. For the PC analysis, the data were subdivided into four depth regions a) the upper 200m, b) surface layer to 20m, c) DCM region (20-100 m), and d) sub-euphotic region (100-200 m). We aim to reduce the effects of the vertical variability of the physical and biological factors. The first two PCs accounted for 90.3% of the variance for the upper 200m. PC1 represents water mass variability associated with the vertical structure given the negative relationship between temperature and salinity and the negative loading of CDOM, whose concentration generally increases with depth (Table 2). PC2 is dominated by the biogeochemical variables with little loading by the physical variables of temperature, salinity, and density. The negative loading of salinity and oxygen on PC2 contrasts with positive loadings for CHL, CDOM, and b_{bp532} , perhaps indicative of a nutrient source since oxygen and nitrate concentration correlate negatively in the larger context of the Red Sea (Naqvi et al. 1986). Given that salinity is also a factor, it appears that this component distinguishes the water masses in the upper layer contributing to the phytoplankton biomass. The DCM is found in two density regions, on the 26 kg m^{-3} and the $27\text{-}27.5 \text{ kg m}^{-3}$ depending on whether the source water is from the south on the lesser isopycnal or from the central and north found on the denser region. Both the surface-20 m, and the 20-100m subsets show similar loadings for these 2 components, and both lie in the depth range affected by the low salinity intrusions. The depth region below 100m retains the relationships of the upper 100 m, but there is a decline in the

loadings of the second component, presumably due to the reduced influence of the water masses in structuring the biogeochemical variables in the transition between the surface layer, and the mesopelagic zone

3.7 Processes determining optical variability in the central Red Sea

3.7.1 Mixed Layer

The Red Sea is a marginal sea affected by strong evaporation. When evaporation is accompanied by significant cooling, water column stratification weakens contributing to deeper vertical mixing. If the mixing penetrates the nutricline, nutrients will be entrained into the mixed layer and water column productivity is expected to increase. We investigate the effect of atmospheric forcing on the biological distributions of CHL, CDOM, b_{bp532} , and estimated POC within the ML by estimating the correlation coefficients between the atmospheric (air temperature, relative humidity, and wind speed) and biological variables (CHL, CDOM, b_{bp532} , and estimated POC) from both glider missions (Table 3). The negative relationship of air temperature and relative humidity with CHL (-0.46 & -0.40) and CDOM (-0.47 & -0.34) is consistent with weakened stratification resulting from cooling and evaporation. CHL and CDOM are weakly correlated with the wind speed (0.32 & 0.30, respectively). Significant correlations exist between wind speed with b_{bp532} and POC.

Mean values of CHL, CDOM, b_{bp532} , POC, and POC:CHL in the mixed layer were calculated separately for the two legs of the glider pattern (Table 4). The mean CHL and b_{bp532} concentration in the mixed layer was 0.12 mg m^{-3} and $8.77 \cdot 10^{-4} \text{ m}^{-1}$ respectively during the observation period. CHL concentrations along segments AB-1 to AB-3 increased by nearly 50% when deep mixing and mesoscale eddy activity were present. CHL increased along the segment CD-2 when an intrusion of GAIW occurred (Figure 3b). POC:CHL ratio has been used to differentiate whether autotrophic or heterotrophic processes dominate the pelagic functioning (Bentaleb 1998; Cifuentes et al. 1988). The system is interpreted to be autotrophic when POC:CHL is less than 200 and is dominated by microbial heterotrophic remineralization when the POC:CHL ratio is greater than 200-300 (Bentaleb 1998; Cifuentes et al. 1988). Based on these assumptions, the observed region of the CRS during this winter period was an autotrophic system. However, during the second week of February, the system shifted to some extent to heterotrophic, where the eddy interacted with the warm, low salinity GAW.

3.7.2 Mesoscale eddies

Mesoscale eddies have already been identified as important contributors to the Red Sea circulation (e.g. Bower and Farrar 2015; Quadfasel et al. 1993; Zhai and Bower 2013). While much of the published work emphasizes the physical aspects of these eddies, the results presented in this paper indicate their importance to the biogeochemical distributions within the Red Sea. As demonstrated, the eddies affect the vertical distributions of phytoplankton, nutrients and dissolved oxygen within the CRS. Cyclonic eddies appear to contribute through elevation of nutrients and CDOM into the upper layer in their centers, and through subduction of subsurface phytoplankton along frontal boundaries as demonstrated at the interface between the cyclonic and anticyclonic features present during the winter of 2014-2015 (Figure 4). Previous work in other seasons has also shown that mesoscale eddies affect the distribution of CHL, CDOM and dissolved oxygen which can be a proxy for available nutrients (Naqvi et al. 1986a; Zarokanellos et al. 2017a,b).

Remote sensing analyses indicate that the eddy activity contributes to horizontal CHL dispersion including the exchange between open sea and coastal environments (Raitso et al. 2017). An ongoing question has been whether the surface ocean color signal associated with eddies are truly chlorophyll signatures or perhaps CDOM or particulate organic material that originates from coastal regions. Based on the comparison of the in-situ glider observations with the remotely sensed ocean color from this period, the remotely sensed chlorophyll signature is consistent with the glider observations (Figures 8a and 8b). But the glider observations reveal additional information that is not immediately evident in the ocean color image (Figure 5i). The boundary between the high chlorophyll feature, apparently anticyclonic, and the low chlorophyll region offshore is a boundary between two counter-rotating eddies. The offshore feature is a cyclonic feature, and the coastal feature is anticyclonic. Within these two features, the deep chlorophyll maxima are found on distinct isopycnals. As previously described, the DCM is found on the 27.5 kg m^{-3} isopycnal in the cyclonic feature, while in the anticyclone southern edge of the study area, the DCM lies on the 26.0 kg m^{-3} isopycnal. The lower salinity found in the southern feature suggests that this water mass had advected northward from the southern Red Sea (Figure 3j). At the interface between the two water masses, the DCM associated with the lighter water mass is lifted toward the surface. Thus, the maximum chlorophyll around the periphery of this feature results from shallowing of the DCM and the interaction of the shallower DCM with

the mixed layer. Although the DCM on the denser 27.5 kg m^{-3} isopycnal was lifted toward the surface in the center of the cyclonic feature, it does not rise sufficiently near to the surface to be detected in ocean color imagery. Overall, the integrated chlorophyll was higher in the region underlying the anticyclonic feature (Figure 8d), but this appears to be due to the subduction of the DCM from the denser isopycnals beneath the lower salinity water, thus providing a double DCM within the context of a vertical profile. In addition to simple subduction of the 27.5 kg m^{-3} isopycnal and its associated DCM beneath the lower salinity water from the south, the cross-isopycnal distribution of chlorophyll to nearly 200m beneath the interface between the two eddies suggest that particle sinking may contribute to part of this downward flux.

3.7.3 Lateral advection

To exam the contribution of lateral advection in the Red Sea circulation, it is vital to understand the pathways and the temporal variability of the Gulf of Aden water. It is well known that GASW enters the Red Sea from September to May under the influence of the northeast monsoon (Murray and Johns 1997; Smeed 2004; Sofianos and Johns 2002; Yao et al. 2014b). During the southwest monsoon, the GAIW enters when winds in the southern Red Sea are toward the south-southeast. GAIW is known to be low in oxygen, high in inorganic nutrients, and elevated in CHL and CDOM (Churchill et al. 2014; Sofianos and Johns 2007; Wafar et al. 2016; Zarokanellos et al. 2017a). The advection of low salinity water into the study region appears from the TS perspective to be GASW that has entered during the winter period. However, embedded within the TS structure, both CDOM and CHL concentrations are elevated along the 26.0 kg m^{-3} . A possible explanation of this could be that under strong wind forcing near the strait of Bab en Mandab flow of Gulf of Aden Water into the Red Sea may entrain some GAIW as well as GASW. Thus, the winter injections of water of the Gulf of Aden may also enhance the nutrient flux and productivity of the Red Sea but not to the extent of the summer GAIW intrusions.

The GAIW intrusions have seasonal and episodic character (Churchill et al. 2014; Sofianos and Johns 2007; Zarokanellos et al. 2017a). However, the spatial distribution of GAIW and its contribution to primary production within the basin remains unclear. Once it has entered the Red Sea, northward advection of GAIW modifies its physical and optical characteristics due to solar insolation, strong evaporation, and mixing with ambient Red Sea water (Figures 3k-l and

5k-l). The nutrients become diluted and depleted as the GAIW advects northward. The GAIW is typically located between the 26 and 26.5 kg m⁻³ isopycnals and it has been characterized by high concentration of CDOM (>0.9 ppb) and CHL (>0.4 mg m⁻³; (Zarokanellos et al. 2017a)). The high CHL concentrations in the DCM, may also result from eddy activity that can elevate subpycnocline nutrients into the photic layer. Both GASW and GAIW were detected below 23 °N, during the interaction between of the eddy with warm, low salinity intrusion. These two water masses differ not only in their thermohaline characteristics, but also show distinct optical signatures.

3.8 Persistence of the eddy structure.

As presented above, the evidence of an offshore eddy structure first appeared in late December during the beginning of the second glider deployment (Figure 3a-d). As the glider continued to transit this pattern, the apparent eddy intensified and remained a dominant feature into early February (Figure 3i-l). At the end of March 2015, a Scanfish tow, part of another research effort (Kürten et al. 2019), was performed along the central axis of the Red Sea from 24°N to approximately 20°N (Figure 9). When the salinity section from the Scanfish tow is compared with the salinity distribution from early February, a very similar pattern is observed. Both sets of observations show doming of isohalines, consistent with a cyclonic eddy circulation centered near the point where the glider track intersects with the Scanfish track (Figure 9b). The mean SLA for the period from 1 December 2014 through 31 March 2015 is consistent with the observed in situ patterns showing a high SLA centered at about 22.5°N, 38°E indicative of an anticyclonic geostrophic circulation, and a low SLA region centered at 24°N, 36°E on the western side of the basin (Figure 9a). Thus, the glider and the Scanfish captured a portion of the southeast quadrant of the cyclonic eddy. Both the glider and Scanfish surveys show lower salinity near the surface on the northern side of their sections of the feature indicative of entrainment of the lower salinity water from the south around the periphery of the cyclonic feature. Weekly average snapshots of SLA from AVISO show that this feature was present throughout the observational period (See supplementary figure 1). Thus, this eddy structure was persistent for nearly four months from early December through March, much longer than the statistical average of about 6 weeks reported by Zhan et al. (2014).

4. Summary and Conclusions

In-situ and remote sensing observations have been used to investigate the response of physical and biological processes to atmospheric forcing during the winter period. Our findings provide new in situ observations of the physical and biological responses in the CRS. Not surprisingly, glider and remote sensing observations show that atmospheric forcing contributed to enhanced vertical mixing and mesoscale eddies and other advective processes contributed to the physical and biogeochemical variability.

During the winter of 2014-2015, surface cooling and strong winds caused mixing to approximately 90m depth. The intense mixing resulted homogenized of physical, chemical, and bio-optical distributions in the upper layer. Phytoplankton from the DCM were distributed throughout the upper mixed layer increasing the concentration of nearsurface chlorophyll, and thus the chlorophyll detected by MODIS. Despite the increase in nearsurface chlorophyll, the integrated (0-200m) CHL remained essentially constant. Following strong mixing in December a mesoscale CE grew in size and resulted in uplift of isopycnals (Figures 3a and 3b). Patches of high CHL and b_{bp532} occurred within the center of the CE (Figures 4a-b). As the eddy intensified, stratification weakened and the MLD decreased (Figures 3f).

In February, the DCM and isopycnals deepened between 22.3 and 23°N, consistent with an intensification of the anticyclonic circulation in this region. At the interface between the AE and CE, two chlorophyll maxima were evident. One lying on the 26 kg m⁻³ isopycnal within the shallow, warm, less saline water, and the other in the density region between the 27 and 27.5 kg m⁻³. We infer from this that the water from the DCM layer in the CE subducted beneath the intrusion of the fresher, warmer near-surface water of the AE. At the interface where the two CHL maxima overlay each other, the integrated CHL concentration and b_{bp532} were higher than to either side of the interface. We attribute this increase to the “simple” physical overlaying of the water from the two sources rather than a nutrient driven water column increase. During the interaction between the eddy with the warm, low salinity intrusion, the northward transport along the northeastern boundary of the eddy contributed to a persistent flow of the warmer, low salinity water around the periphery of the cyclonic circulation. The water mass properties, including bio-optical characteristics are consistent with Gulf of Aden source water. The interplay between the CE and warm, low salinity intrusion, controls the distribution of the Gulf of Aden water along with its chemical and optical characteristics within the CRS.

Better understanding of the physical dynamics and ecological response of the region requires continued long-term monitoring. In order to better understand the biogeochemical dynamics of the region, further efforts require elucidation of the primary productivity response to seasonal, mesoscale and longer-term processes. Autonomous underwater vehicles (AUVs) equipped with the appropriate sensor suite enable monitoring of physical and biogeochemical properties over a range of temporal and spatial scales, helping to achieve these goals. The direct and indirect effects of isopycnal displacement on phytoplankton distribution and productivity, close alignment of physical and biological properties at frontal boundaries, and variation of physical and biogeochemical features of eddies, underscore the importance of high-resolution in situ measurements at appropriate scales. It is well known that eddies and fronts contribute nutrients into the upper layer enhancing primary production, phytoplankton biomass and particle concentration (Li et al. 2012; Ohman et al. 2012; Stramma et al. 2013). But it has been shown in this set of observations, not all increases in surface chlorophyll represent increased phytoplankton biomass and/or productivity in the water column. Continuing to study these processes in the globally extreme environment of the Red Sea will contribute to a better understanding of the potential response of the global ocean to the current trends in climate.

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Figure Captions

Figure 1: a) Bathymetric map of the Red Sea b) Bathymetric map of the study region with the two glider missions. The black line between points A and B represents the glider mission that was conducted between 02/12/2014 to 14/12/2014 and the red line show the glider mission from 17/12/2014 to 02/03/2015. The letters A, B, C and D represent turning points of the glider missions segments.

Figure 2: Time series of observations from 2 December 2014 until 2 March 2015. a) Daily mean of air and sea temperature at 5m from glider ($^{\circ}\text{C}$) b) relative humidity (%) and wind stress (N m^{-2}) c) net heat flux, d) salinity and e) CHL (mg m^{-3}). Isopycnals are indicated with solid white lines in the panel d (contour interval 0.5 kg m^{-3}). The red solid line in panel (d) and (e) indicate mixed layer depth (MLD). The magenta solid line in the bottom panel indicates the Zeu and the yellow solid line indicates the depth of the $180 \mu\text{M}$ isopleth of oxygen.

Figure 3: Seaglider sections plots for potential temperature ($^{\circ}\text{C}$ - Row 1), salinity (Row 2), Brunt–Väisälä frequency ($\text{BVF (N}^2\text{)}$ - Row 3) and dissolved oxygen ($\mu\text{mol kg}^{-1}$ - Row 4) during the winter mixing (a-d), the intensification of the eddy circulation (e-h) and the interaction between the eddy with warm, low salinity intrusion period (i-l). The vertical dash-dot black line indicates the offshore vertex of the transect (C) shown in Figure 1b. Isopycnals are indicated with a solid black line (contour interval = 0.5 kg m^{-3}). The blue line represents the mixed layer depth (MLD) and the white line represents $180 (\mu\text{mol kg}^{-1})$ isopleth.

Figure 4: Pot. temperature–salinity diagrams color coded by CHL (mg m^{-3}) during a) the winter period b) intensification of the eddy circulation and c) interaction between the eddy with the warm, low salinity intrusion.

Figure 5: As in figure 3 but for CHL (mg m^{-3}), $b_{\text{bp}532}$ (m^{-1}), $\text{CHL}:b_{\text{bp}532}$, ad CDOM (ppb). Isopycnals are indicated with a solid white line (contour interval = 0.5 kg m^{-3}). The cyan, black and red solid line represents a) the mixed layer depth (MLD), b) Zeu, the $180 (\mu\text{mol kg}^{-1})$ isopleth, respectively.

Figure 6: Mean vertical profiles of a) Salinity (psu), b) Pot. temperature ($^{\circ}\text{C}$), c) CHL (mg m^{-3}) d) CDOM (ppb) and e) $b_{\text{bp}532}$ (m^{-1}) during the winter mixing (blue line), the intensification of the eddy circulation (red line) and during the interaction between the eddy with the warm, low salinity intrusion (black line). Shaded areas indicate the standard deviation for each period.

Figure 7: Box-and-whiskers plots for latitudinal (AB; panels a-f) and longitudinal (CD; panels g-l) transects for the average 0-20m and integrated 0-200m for CHL (mg m^{-3}), $b_{\text{bp}532}$ (m^{-1}) and CDOM, (ppb) during the winter period. The mean value of CHL, $b_{\text{bp}532}$, and CDOM are shown with the range of variation as well the 25% and 75% quartiles.

Figure 8: Snapshot of the front that developed during the interaction between the eddy with warm, low salinity intrusion. Daily MODIS-Aqua CHL a) 2nd of February 2015, b) 7th of February 2015 c) Average CHL (0-20m; red line) and integrated CHL (0-200m; black line) from Seaglider during the period of the interaction between the eddy with warm, low salinity intrusion water. Red dots indicate the glider location and black vectors the glider derived surface velocity.

Figure 9: a) Average sea level anomaly for the period from 1 December 2014 through 31 March 2015, overlapping the glider deployments and the Scanfish survey from 31 March – 2 April 2015. b) three-dimensional projections of salinity from the last glider section (27 January – 14 February, 2015; black line above indicates track), and from the Scanfish section (31 March – 2 April, 2015; magenta line above indicates the trajectory). The Scanfish line is semi-transparent so that the similarity between in the salinity distribution between the two tracks is apparent. The coastlines are indicated by the thinner black lines at the surface.

Tables

Table 1: Summary of measured parameters and associated sampling period for each glider mission

Glider ID	Seaglider Sensors	Dives	Deployment Period
sg216	Conductivity, Temperature, Pressure, backscatter (532nm, 650nm and 880nm) CHL, CDOM, Phycocyanin and Oxygen	155	2-14 Dec 2014
sg212		547	17Dec 2014 – 2 Mar 2015

Table 2: Matrix with the loadings from the first two PCs. Bold number indicates the dominant environmental parameters in each PC, indicated by the loadings with the highest absolute values. The analysis of the winter period includes data for both glider missions, and for all fourth-depth ranges.

Both missions	The upper 200m		0-20 m		20-100 m		100-200 m	
Variables	PC1	PC2	PC1	PC2	PC1	PC2	PC1	PC2
Temperature	-0.41	-0.07	-0.58	-0.06	-0.48	-0.10	-0.43	-0.15
Salinity	0.40	-0.13	0.48	-0.27	0.48	-0.20	0.42	0.19
Density	0.41	0.01	0.64	-0.03	0.53	0.00	0.44	0.19
Oxygen	-0.37	-0.48	-0.14	-0.50	-0.28	-0.38	-0.41	-0.13
CHL	-0.35	0.41	0.00	0.54	-0.19	0.58	-0.35	0.21
CDOM	0.37	0.48	0.07	0.56	0.24	0.60	0.33	-0.05
b_{bp532}	-0.33	0.59	0.01	0.27	-0.30	0.33	-0.21	0.92

Table 3: Pearson linear correlations between the atmospheric variables (air temperature, relative humidity and wind speed) versus mixed layer CHL, CDOM, b_{bp532} and POC during the study period (both gliders missions). Where p-value is small ($p < 0.05$) then the correlation coefficient r is significant.

Variables	Mean CHL in ML		Mean CDOM in ML		Mean b_{bp532} in ML		Mean POC in ML	
	r	p	r	p	r	p	r	p
Air Temperature	-0.46	<0.01	-0.47	<0.01	0.17	0.1	-0.11	0.30
Relative Humidity	-0.40	<0.01	-0.34	<0.01	0.05	0.65	-0.11	0.28
Wind Speed	0.32	<0.01	0.29	<0.01	0.008	0.94	0.06	0.59

Table 4: Mean mixed layer CHL, CDOM, b_{bp532} , POC and POC:CHL ratio for latitudinal and longitudinal transects during the study period. Grey cells highlight the values that exceed 50% of the mean value.

Mission and Segment	Date	CHL in MLD	CDOM in MLD	b_{bp532} in MLD	POC in MLD	POC:CHL in MLD
Mission 252 - AB	AB 1 - 2014/12/02- 2014/12/08	0.10	0.65	7.97E-04	16.3	86
	AB 2 - 2014/12/08- 2014/12/14	0.10	0.65	8.16E-04	15.2	90
Mission 250 - AB	AB 3 - 2014/12/17- 2014/12/23	0.16	0.72	8.08E-04	25.1	96
	AB 4 - 2015/01/08- 2015/01/14	0.20	0.83	7.47E-04	24.6	77
	AB 5 - 2015/01/14- 2015/01/20	0.15	0.81	7.88E-04	25.0	121
	AB 6 - 2015/02/08- 2015/02/15	0.13	0.87	9.74E-04	29.3	184
Mission 250 - CD	CD 1 - 2014/12/26- 2014/12/30	0.10	0.70	8.11E-04	24.4	95
	CD 2 - 2014/12/30- 2015/01/03	0.13	0.75	7.82E-04	24.3	60
	CD 3 - 2015/01/23- 2015/01/27	0.07	0.69	9.14E-04	30.3	144

	CD 4 - 2015/01/28- 2015/02/04	0.07	0.74	9.78E-04	31.9	157
Both missions	Mean	0.02	0.75	8.77E-04	26.3	122
	STD	0.02	0.09	1.29E-04	6.1	99