Author's Accepted Manuscript

Characterization of light absorption by chromophoric dissolved organic matter (CDOM) in the upper layer of the Red Sea

Malika Kheireddine, Mustapha Ouhssain, Maria Ll. Calleja, Xosé Anxelu G. Morán, Y.V.B Sarma, Surya P. Tiwari, Burton H. Jones

www.elsevier.com

PII: S0967-0637(17)30100-0
DOI: <https://doi.org/10.1016/j.dsr.2018.02.001>
Reference: DSRI2878

To appear in: *Deep-Sea Research Part I*

Received date: 19 March 2017
Revised date: 28 January 2018
Accepted date: 4 February 2018

Cite this article as: Malika Kheireddine, Mustapha Ouhssain, Maria Ll. Calleja, Xosé Anxelu G. Morán, Y.V.B Sarma, Surya P. Tiwari and Burton H. Jones, Characterization of light absorption by chromophoric dissolved organic matter (CDOM) in the upper layer of the Red Sea, *Deep-Sea Research Part I*, <https://doi.org/10.1016/j.dsr.2018.02.001>

This is a PDF file of an unedited manuscript that has been accepted for publication. As a service to our customers we are providing this early version of the manuscript. The manuscript will undergo copyediting, typesetting, and review of the resulting galley proof before it is published in its final citable form. Please note that during the production process errors may be discovered which could affect the content, and all legal disclaimers that apply to the journal pertain.

Characterization of light absorption by chromophoric dissolved organic matter (CDOM) in the upper layer of the Red Sea

Malika Kheireddine^{1*}, Mustapha Ouhssain^{1,2,3}, Maria Ll. Calleja¹, Xosé Anxelu G. Morán¹,
Y.V.B Sarma¹, Surya P. Tiwari¹, Burton H. Jones¹

¹King Abdullah University of Science and Technology (KAUST), Red Sea Research Center (RSRC), Biological and Environmental Sciences & Engineering Division (BESE), Thuwal, 23955-6900, Saudi Arabia

²UMS 829 - Observatoire Océanologique - Centre National de la Recherche Scientifique, Villefranche-sur-Mer, France

³Université Pierre et Marie Curie, Paris VI, Paris, France

Correspondence to: Malika Kheireddine (malika.kheireddine@kaust.edu.sa)

Abstract

The absorption coefficient of chromophoric dissolved organic matter (CDOM) is a major variable used in developing robust bio-optical models and understanding biogeochemical processes. Over the last decade, the optical properties of CDOM in the open sea have been intensely studied. However, their variations in clear water are poorly documented, particularly in the Red Sea, owing to the absence of in situ measurements. We performed several cruises in the Red Sea to investigate the spatial distribution of the absorption coefficient of CDOM. The spectral absorption coefficients were determined from 400 nm to 740 nm using a WETLabs ac-s hyper-spectral spectrophotometer. In general, we found a latitudinal gradient in the CDOM

absorption coefficient at 443 nm ($a_{\text{CDOM}}(443)$) from south to north that is likely influenced by the exchange of water through the strait of Bab-el-Mandeb and the thermohaline circulation of the Red Sea. However, high $a_{\text{CDOM}}(443)$ values were observed in the northern Red Sea due to the existence of a sub-mesoscale feature that may induce an increase in phytoplankton production and lead to CDOM production. The $a_{\text{CDOM}}(443)$ covaried with the chlorophyll a concentration ([Chl a]), despite a high scatter. Furthermore, the $a_{\text{CDOM}}(443)$ for a given [Chl a] concentration was higher than those predicted by global ocean bio-optical models. This study advances our understanding of CDOM concentration in the Red Sea and may help improve the accuracy of the algorithms used to obtain CDOM absorption from ocean color.

Keywords

Red Sea, Chromophoric dissolved organic matter, light absorption, spatial distribution.

Introduction

Historically known as “gelbstoff” (Kalle, 1966) and, in English, as “yellow substances” (Shifrin, 1988) due to its high humic matter content, the chromophoric dissolved organic matter (CDOM) is the colored portion of the dissolved organic matter (DOM) in the ocean (Laane and Koole, 1982; Thurman, 1985; Williams and Druffel, 1988; Laane and Kramer, 1990). Marine dissolved organic carbon (DOC) represents the largest reservoir of reduced carbon in the ocean (Hansell, 2002) and plays an essential role in the global carbon cycle (Mopper and Kieber, 2002). CDOM is one of the main factors that affects the transmission of light in seawater (Siegel et al. 2002; Nelson and Siegel, 2013). This is crucial for evaluating underwater light availability in seawater, which is an essential variable for several marine organisms (Siegel et al. 1995; Siegel and Michaels, 1996; Vodacek et al. 1997; Nelson et al. 1998; Blough and Del Vecchio, 2002).

CDOM affects structures of the marine food web (Smith et al. 1992; Herndl et al. 1993; Häder et al. 1998) by influencing both primary and microbial production. In coastal waters, the distribution and concentration of CDOM are mainly controlled by terrestrial runoff and riverine inputs (Kalle, 1938; Højerslev, 1982; Carder et al. 1989; Twardowski and Donaghay, 2001; Blough and Del Vecchio, 2002; Del Vecchio and Blough, 2004; D'Sa et al. 2009), whereas in open waters, CDOM has primarily autochthonous origins (Nelson et al. 1998, 2004; Nelson and Siegel, 2002; Steinberg et al. 2004; Ortega-Retuerta et al. 2009; Shank et al. 2010; Organelli et al. 2014). Many studies have shown that CDOM is produced through diverse biological processes (Nelson et al. 1998, 2004; Nelson and Siegel, 2002; Steinberg et al. 2004; Ortega-Retuerta et al. 2009; Shank et al. 2010). CDOM results from both microbial production and degradation processes (Nelson et al. 1998, 2004; Steinberg et al. 2004; Yamashita and Tanoue, 2004), and it can be degraded by photobleaching in surface waters (Vodacek et al. 1997; Nelson et al. 1998, 2004; Siegel et al. 2002, 2005; Twardowski et al. 2002; Del Vecchio and Blough, 2004). Generally, CDOM in the open ocean is controlled by biological processes and photobleaching. The spatial distribution of CDOM is influenced by transport, convection, upwelling and other processes that affect water mass circulation.

Measuring the CDOM light absorption spectrum, $a_{\text{CDOM}}(\lambda)$, to study its distribution and abundance in the open ocean has been a widely used approach over the past decade (Babin et al. 2003; Siegel et al. 2002; Nelson and Siegel, 2013; Organelli et al. 2014). The slope of CDOM absorption from 400 to 500 nm, S_{CDOM} , can be used to indicate changes in the CDOM composition by incorporating ratios of humic to fulvic acids (Carder et al. 1989; Blough and Green, 1995; Vodacek et al. 1997; Moran et al. 2000; Helms et al. 2008). Changes in S_{CDOM} are detected when CDOM is produced by bacteria (Ortega-Retuerta et al. 2009; Yamashita et al.

2013) compared to CDOM generated by others pelagic organisms. Furthermore, numerous field studies (Vodacek et al. 1997; Nelson et al. 1998; Del Vecchio and Blough, 2002, 2004; Twardowski et al. 2002; Swan et al. 2009; Bricaud et al. 2012) have shown an increase in S_{CDOM} associated with a decrease in $a_{CDOM}(\lambda)$ in response to the solar bleaching of CDOM in surface waters. Because CDOM mostly absorbs solar radiation in the UV and blue regions of the spectrum (Carder et al. 1989; Kirk, 1976, 1999; Visser, 1983), CDOM absorption and the resulting remote sensing reflectance may result in inaccuracies when determining the chlorophyll a concentration from satellite imagery analysis, particularly when using both empirical and semi-analytical algorithms (Morel, 1980; Carder et al. 1991; Siegel et al. 2005; Morel and Gentili, 2009a). This generally results in an overestimation of the chlorophyll a concentration in seawater when the CDOM concentrations are high. For example, from both ocean color and in situ data, Morel and Gentili (2009b) noticed that the CDOM concentration in the Mediterranean Sea was twice as high as those measured at the same latitude and chlorophyll a concentration in the Atlantic Ocean. Southeast of the Mediterranean Sea, the Red Sea is a marginal, semi-enclosed oligotrophic basin. Brewin et al. (2015) suggested that a refractory component of CDOM that is resistant to photo-bleaching might accumulate over time.

The Red Sea is one of the saltiest and warmest deep seas in the world (Belkin, 2009; Longhurst, 2007; Raitos et al. 2011, 2013). Its typical seawater temperature varies from 21°C to 28°C in the north and from 26°C to 32°C in the south (Nandkeolyar et al. 2013). The Red Sea is characterized by little precipitation, small riverine discharge (Patzert, 1974) and strong evaporation rates (Sofianos and Johns, 2003). In addition, the Red Sea is represented by horizontal circulation that is mainly dominated by eddies (Yao et al. 2014a, b). The southern part of the Red Sea is connected to the Arabian Sea (Northern Indian Ocean) via the strait of Bab-el-

Mandeb. In the north, the Red Sea connects to the Mediterranean Sea via the Suez Canal, although the exchange with the Mediterranean Sea is much less than that with the Gulf of Aden. The Red Sea is considered a large marine ecosystem (Belkin, 2009), and it supports coral reefs, mangroves and seagrass beds that provide crucial habitats for a large variety of marine organisms (Berumen et al. 2013; Almahasheer et al. 2016). The Red Sea is distinguished by strong latitudinal gradients of temperature, salinity and inorganic nutrient concentrations from south to north, primarily dependent on the exchange of water through the strait of Bab-el-Mandeb and the thermohaline circulation of the Red Sea. During the Arabian Sea southwest monsoon season (June to September), the monsoon-driven wind reversal modifies the circulation dynamics of the Bab-el-Mandeb strait, leading to an influx of cool, less salty, and nutrient-rich Gulf of Aden Intermediate Water (GAIW) from the Indian Ocean into the southern Red Sea, increasing phytoplankton biomass and primary production (Neumann and McGill, 1962; Sofianos and Johns, 2007; Raitsos et al. 2013, 2015; Churchill et al. 2014; Dreano et al. 2016; Wafar et al. 2016; Gittings et al. 2017). During the winter monsoon (October to April), the reversal of the wind direction induces a northward advection of cooler, fresher, nutrient-rich surface waters into the Red Sea from the Gulf of Aden.

The Red Sea is deemed oligotrophic mainly due to the depletion of nutrients in the upper layer. This nutrient depletion becomes especially important in the central and northern Red Sea in response to northward advection. The northern Red Sea is marked by significant physical variability, with deep mixed layers occurring in the winter and strong stratification in the summer. These dynamics drive the seasonal changes in trophic levels from oligotrophy in summer to bloom conditions in winter-spring (Raitsos et al. 2013). Based on remotely sensed data, the maximum chlorophyll *a* concentrations in the north-central Red Sea rise during winter

but remain the lowest throughout the Red Sea during this time. Raitsos et al. (2013) suggested that anti-cyclonic eddies may be a potential mesoscale feature responsible for the higher productivity observed in the region (particularly during summer). The eddies occupying the central Red Sea are bounded by coastal coral reef structures and could influence the Red Sea ecosystem via the exchange of nutrients, chlorophyll a, and other organic matter between the open sea and coastal reefs.

The interaction of wind and thermohaline processes facilitate an intrusion of cold, low-salinity and nutrient-rich waters into the upper layers of the Southern Red Sea from the Gulf of Aden, which makes this part of the Red Sea a more productive region than the northern and central parts. The seasonal cycle of phytoplankton biomass in the Red Sea appears to be controlled by physical processes (winter mixing, mesoscale eddies, horizontal advection and the influx of water from Bab-el-Mandeb) that determine the availability of nutrients to the euphotic layer. However, dust deposition and nutrient input from coral reefs may also contribute to the productivity of the Red Sea. Essential information about the seasonal cycle of the phytoplankton biomass in the Red Sea is provided in the recent publications of Raitsos et al. (2013, 2015), Triantafyllou et al. (2014) and Racault et al. (2015).

Bricaud et al. (2010) and Morel and Gentili (2009a) proposed empirical global bio-optical relationships between the inherent optical properties (IOPs) and some biogeochemical parameters, such as the chlorophyll a concentration, in oligotrophic waters. These relationships can then provide the ability to assess the variability of biogeochemical parameters from in situ measurements or ocean color.

Although robust statistical relationships (Bricaud et al. 2010; Morel and Gentili, 2009a) were established between the CDOM absorption coefficients and chlorophyll a, representative of the

open ocean, no direct measurements of CDOM absorption have been collected in the Red Sea until now. The Red Sea waters and the regional empirical bio-optical relationships can be more affected by little precipitation, small riverine discharge and desert dust events than other regions are. The recent study of Brewin et al. (2015) assessed the bio-optical characteristics of the Red Sea, and the authors hypothesized that the Red Sea has a higher CDOM concentration per unit of chlorophyll a than do other regions. They suggested that the overestimation of the chlorophyll a concentration in the Red Sea, as retrieved from the standard bio-optical algorithms used to process ocean color remote sensing measurements, could have been due to an excess of CDOM absorption per unit of chlorophyll a. However, this hypothesis has never been tested.

The aim of this study was to examine if an excess of CDOM exists in Red Sea waters per unit of chlorophyll a. Particular attention was given to the CDOM absorption coefficients because these coefficients have not been measured in previous studies. Using in situ measurements of CDOM absorption collected during different cruises performed in the Red Sea (from October 2014 to April 2015), we investigated the spatial variations in CDOM concentration, as derived from measurements of spectral light absorption coefficients and the CDOM spectral slope values from the surface down to 200 meters depth. In addition to documenting CDOM distribution and its relationship with chlorophyll a, this study aims to advance the state of the art regarding CDOM light absorption coefficients and provide key information to develop regional algorithms such that the chlorophyll a and CDOM absorption coefficients can be retrieved accurately.

Datasets and methods

Study area and in situ measurements---

Data were acquired during four research campaigns performed in the Red Sea between October 2014 and April 2015 on the research vessel Thuwal: in the central Red Sea (CRS) from 16 to 28

October 2014 and 29 March to 03 April 2015 and in the southern Red Sea (SRS) from 8 to 21 February 2015 and in the northern Red Sea (NRS) from 17 to 28 April 2015 (Table 1). To differentiate the two cruises performed in the CRS in October 2014 and March-April 2015, they are denoted as CRS-01 and CRS-02, respectively. Overall, 43 stations were sampled: 23 in the NRS, 12 in the CRS and 8 in the SRS (Fig. 1).

Table 1: Location, abbreviation, period, and number of stations for each campaign used in this study

Campaign	Platform	Location	Abbreviation	Period	Number of stations
Nutrient cycle cruise 1	<i>RV Thuwal</i>	Central Red Sea	CRS-01	16-28 Oct. 2014	6
Jazan cruise	<i>RV Thuwal</i>	Southern Red Sea	SRS	8 -21 Feb. 2015	8
Nutrient cycle cruise 2	<i>RV Thuwal</i>	Central Red Sea	CRS-02	29 Mar. to 03 Apr. 2015	6
Duba cruise	<i>RV Thuwal</i>	Northern Red Sea	NRS	17- 28 Apr. 2015	23
Total					43

Figure 1: Locations of stations (black dots) sampled during 4 cruises between October 2014 and April 2015 in the Red Sea. The stations of the central axis of the Red Sea are represented by red dots. The green squares correspond to the stations of the time series project. The northern Red Sea (NRS), central Red Sea (CRS) and southern Red Sea (SRS) are also indicated on the map.

To provide a general overview of light absorption by CDOM in the Red Sea,

the datasets from these four cruises were assembled and analyzed. Discrete seawater samples for determining phytoplankton pigments were collected using a rosette system of 10 L Niskin bottles, typically at 10 depths from surface to 200 m (5, 10, 20, 40, 50, 60, 70, 80, 150 and 200 m). Temperature and salinity profiles were obtained using a SBE 9 (Sea-Bird Electronics) Conductivity-Temperature-Depth (CTD) probe. The first optical depth that corresponds to the surface layer observed by satellite ocean color instruments (Gordon and McCluney, 1975) is a function of the attenuation of the downwelling Photosynthetically Available Radiation (PAR) ($1/K_{PAR}$), which is obtained as Z_e divided by 4.6 (Morel, 1988). The euphotic layer depth, Z_e , reflects the depth where only 1% of the PAR ($\ln(1\%) = -4.6$) of its surface value prevailed and was derived from chlorophyll a concentration profiles (Morel and Maritorena, 2001).

Phytoplankton pigments ---

Discrete sampling was achieved during each cruise from surface to 200 m. Seawater samples were acquired from Niskin bottles and filtered through 25-mm-diameter Whatman GF/F filters (0.7 μm nominal porosity). During the cruise, these samples were stored in liquid nitrogen and later transferred to -80°C in the laboratory. Phytoplankton pigment contents were assessed using High Performance Liquid Chromatography (HPLC). Filters were ground in 3 ml of 100% methanol together with 1-mm-diameter glass beads by a cell homogenizer. The holder for the tube was cooled in a freezer before use. The extract was centrifuged for ten minutes at 7500 rpm and cooled at -5°C ; the supernatants were then filtered through a 0.2- μm Teflon syringe filter. The extracts were analyzed within 24 hours by HPLC using a complete 1260 Agilent Technologies system, in accordance with the protocol described in Ras et al. (2008). The total chlorophyll a concentration is calculated as the sum of the concentrations of chlorophyll a, chlorophyllide a and divinyl chlorophyll a. The sum of concentrations of chlorophyll a, divinyl

chlorophyll a, chlorophyllide a, pheophorbide a and pheophytin a is adopted here as a proxy of phytoplankton biomass and noted as [Chl a].

CDOM absorption---

Hyperspectral CDOM absorption ($a_{\text{CDOM}}(\lambda)$, where λ is the wavelength), temperature and salinity data were collected using a WET Labs ac-s hyper-spectral spectrophotometer and a SBE 49 FastCAT (Sea-Bird Electronics) CTD probe, respectively, from 0.5 m (closest value without getting bubbles) to 200 m depth. The water passed through a 0.2- μm filter (Whatman Polycap 75 TC) positioned at the intakes of the ac-s sensor. The filter was changed every 3 to 4 profiles within each cruise. The ac-s instrument provided spectral $a_{\text{CDOM}}(\lambda)$ at 81 wavelengths from 400 to 740 nm. A Milli-Q water calibration was carried out prior to and after each cruise. The uncertainty obtained from the standard deviation of several measurements of calibration coefficients values varied spectrally from 0.001 to 0.004 m^{-1} for absorption and from 0.002 to 0.006 m^{-1} for attenuation, with greater uncertainties observed in the blue region of the spectrum. Larger uncertainties in calibration values for the c-tube were obtained; thus, we only used measurements collected through the a-tube of the ac-s to obtain $a_{\text{CDOM}}(\lambda)$ (attenuation measurements were usually approximately 0.004 m^{-1} lower to those of absorption). We calculated CDOM absorption ($a_{\text{CDOM}}(\lambda) = a_{\text{filtered}}(\lambda) - a_{\text{pure-water}}(\lambda)$) and performed a residual temperature correction using the temperature correction tables (Sullivan et al., 2006) to account for the eventual dissimilarities in temperature between the filtered sample and the pure-water sample (Slade et al. 2010). Note that some spectra were discarded because they were either too noisy or contained obvious spectral artifacts.

The spectral slope of $a_{\text{CDOM}}(\lambda)$ (S_{CDOM} , nm^{-1}) was determined using a non-linear regression of the CDOM absorption spectrum from 400 to 500 nm (Twardowski et al. 2004).

In addition, data from two stations located in the CRS were collected as part of a time series project (green squares in Fig. 1). Seawater collection was performed monthly at the same depths described previously between surface and 200 m. Seawater samples were filtered using 0.2 μm polycarbonate membranes (Whatman Nuclepore, 47 mm). They were stored in pre-combusted (450°C, 4 h) brown glass bottles and placed in the refrigerator (4°C). Within 12 to 24 hours, the CDOM samples were analyzed in the laboratory. Prior to the analysis, CDOM samples were kept at room temperature to avoid any bias caused by variations in temperature between the sample and Milli-Q water (reference). CDOM absorbance spectra were obtained in every nanometer from 200 to 800 nm using a double-beam Ultraviolet-Visible spectrophotometer (Varian Cary 5000) with a 10-cm quartz cuvette. A baseline was obtained with Milli Q water in both beams, and each sample was analyzed facing Milli Q water as a blank. Then, $a_{\text{CDOM}}(\lambda)$ was computed from absorbance spectra using the following formula:

$$a_{\text{CDOM}}(\lambda) = 2.303A(\lambda)/l \quad (1)$$

where $A(\lambda)$ corresponds to the absorbance of the CDOM sample at a given wavelength λ and l is the optical path-length of the quartz cuvette in meters. As suggested by Babin et al. (2003), a standard correction was performed for each CDOM absorbance spectrum by subtracting the mean absorbance from 680 to 690 nm to the complete spectrum to minimize the effects of both temperature and salinity. This spectral range approximately 685 nm was chosen because of its negligible CDOM absorption and very small temperature and salinity effects on water absorption (Pegau et al. 1997). The absorption spectral shape of CDOM is commonly modeled as an exponential function (Bricaud et al. 1981):

$$a_{\text{CDOM}}(\lambda) = a_{\text{CDOM}}(\lambda_0)e^{-S(\lambda-\lambda_0)} \quad (2)$$

where $a_{CDOM}(\lambda)$ corresponds to the absorption coefficient at a given wavelength λ , $a_{CDOM}(\lambda_0)$ is the absorption coefficient at reference wavelength λ_0 , and S is the spectral slope in the spectral range from λ_0 to λ with $\lambda_0 < \lambda$. As done previously, S_{CDOM} was determined using a non-linear regression of the CDOM absorption spectra from 400 to 500 nm (Twardowski et al. 2004). CDOM measurements from the time series project in CRS (green squares in Fig. 1) that were collected at nearly the same time (within 60 minutes) as the deployment of the ac-s meter (Dec., Feb., Apr., and May 2014-2015) were used to assess the agreement of the $a_{CDOM}(\lambda)$ spectra obtained by the ac-s with those obtained from the spectrophotometer. Overall, 37 measurements were acquired from depth-discrete profiles (5-20 m, 30-50 m, 70-90 m, 120-150 m).

Results

Comparison of $a_{CDOM}(\lambda)$ and S_{CDOM} obtained with ac-s and spectrophotometer---

We compared the $a_{CDOM}(\lambda)$ spectra determined by the ac-s sensor ($a_{CDOM-ac-s}(\lambda)$) to the spectra determined by the spectrophotometer ($a_{CDOM-spectro}(\lambda)$) (Fig. 2A). The two quantities are significantly correlated ($R^2=0.89$, $N=37$, $p<0.0001$). We also compared the estimates of S_{CDOM} obtained from spectrophotometer measurements ($S_{CDOM-spectro}$) and those from ac-s measurements ($S_{CDOM-ac-s}$), as shown in Figure 2B. They are highly correlated ($R^2=0.97$, $N=37$), and the slope did not significantly differ from the 1:1 line ($p<0.0001$). These results suggest that the slope values obtained from the ac-s sensor and the spectrophotometer are in close agreement.

Figure 2: (A) Comparison of $a_{CDOM}(443)$ measured with the ac-s and spectrophotometer instruments. Correlation between the two quantities was highly significant ($R^2=0.89$, $N=37$, $p<0.0001$): $a_{CDOM_spectro} (443) = 0.92 * a_{CDOM_ac_s} (443) - 0.001$. (B) Comparison of S_{CDOM} computed within the range 400-500 nm from the spectrophotometer or ac-s measurements. The relationship between both quantities was also high ($R^2=0.97$, $N=37$, $p<0.0001$): $S_{CDOM_spectro} = 0.95 * S_{CDOM_ac_s} + 0.0009$. The black solid line represents the 1:1 line.

Hydrographic environment---

Because CDOM can be influenced by complex physico-biological processes along the water column, we present only the major hydrographical features based on measurements of temperature, salinity and [Chl a] at the stations through the central axis of the Red Sea (red dots in Fig. 1) for each cruise (Figs. S1, S2, S3 and S4 in supplementary material).

The Red Sea is characterized by general latitudinal gradients in salinity and temperature due to evaporation because water with lower salinity (<37) intrudes from the Gulf of Aden to the SRS (Figs. S1, S2, S3 and S4 in supplementary material). The study area was oligotrophic, and there was good agreement with the [Chl a] gradients expected along the Red Sea; the [Chl a] concentrations are usually higher in the south than they are in the north. We characterized the spatial distribution of [Chl a] by the existence of a deep chlorophyll a maximum (DCM) located at approximately 40-95 m in the NRS and CRS (Figs. S1C, S2C, S3C in supplementary material) and approximately 20-50 m in the SRS (Fig. S4C in supplementary material).

During our cruises, this general trend was interrupted by shallow temperature and salinity waters due to the influence of eddies, mixing events, and localized upwelling.

In the northern Red Sea, the doming of temperature and salinity isolines at 27.3° N is indicative of an upwelling of deeper waters, which is typical of a cyclonic eddy (cold core eddy) (Fig. S1A, B in supplementary material). This feature is accompanied by an increase in the mixed layer depth (MLD) to ~ 80 m, as calculated using the threshold method with a density gradient criterion of 0.125 kg m^{-3} (Levitus, 1982; Boyer and Levitus, 1994) and a shallower DCM (~ 50 m at 27° N - 27.5° N, Fig. S1 in supplementary material).

In October 2014 (CRS-01), we observed a strong thermal and density stratification in the central Red Sea in the upper 100 meters until approximately 23° N, inducing a deepening of the DCM (Fig. S2 in supplementary material). During the second cruise conducted in this region in March 2015 (CRS-02), the isotherm and isohaline surfaces generally tilted upward from south to north, resulting in cooler, saltier and denser water present nearer the surface toward the north (Fig. S3 in supplementary material).

In the SRS, the upper layer was characterized by relatively cold ($< 26^{\circ}\text{C}$), less saline (< 38.25) and less dense ($< 27 \text{ kg m}^{-3}$) water advecting from the south ($16.75^{\circ} - 17.4^{\circ} \text{ N}$) (Fig. S4 in supplementary material). More information about the spatio-vertical distribution of the phytoplankton pigments during the NRS, CRS-02 and SRS can be found in Kheireddine et al. (2017).

CDOM spectral features---

The average and standard deviation of CDOM absorption spectra were measured during each cruise, and the S_{CDOM} was calculated as the fit from Equation (2), as shown in Figure 3. The CDOM absorption of each sample decreased exponentially as the wavelengths increased.

The spectral shape and magnitude of $a_{\text{CDOM}}(\lambda)$ varied by ~70% in the NRS, ~45% in the CRS-01, ~40% in CRS-02 and ~25% in the SRS from surface to 200 m depth (Fig. 3A to D). The overall changes in S_{CDOM} from surface to 200 m are illustrated in Figure 4. The slope ranged from 0.009 to 0.030 nm^{-1} , with an average value of $0.0170 \text{ nm}^{-1} \pm 0.0035 \text{ nm}^{-1}$. The S_{CDOM} statistics for each region of the Red Sea are provided in Table 2.

Table 2: Red Sea and region-by-region statistics for the S_{CDOM} (nm^{-1})

<i>Acronym</i>	<i>Time</i>	<i>Region</i>	<i>N</i>	<i>Depths (m)</i>	<i>Min(S_{CDOM})</i>	<i>Max(S_{CDOM})</i>	<i>Mean(S_{CDOM})</i>	<i>Range (S_{CDOM}) Red Sea (Brewin et al. 2015)</i>	<i>Mean (S_{CDOM}) Global ocean (Morel and Gentili, 2009)</i>	<i>Standard Deviation S_{CDOM}</i>	<i>Variation coefficient, %</i>
<i>NRS</i>	17-28 Apr. 2015	North Red Sea	198	1 st opt. depth 0.5-200	0.0099 0.0090	0.0247 0.0265	0.0177 0.0171			0.0031 0.0039	18.05 22.60
<i>CRS-01</i>	16-28 Oct. 2014	Central Red Sea	52	1 st opt. depth 0.5-200	0.0114 0.0114	0.0244 0.0253	0.0172 0.0180			0.0036 0.0025	20.78 14.12
<i>CRS-02</i>	29 Mar.-3 Apr. 2015	Central Red Sea	60	1 st opt. depth 0.5-200	0.0144 0.0127	0.0224 0.0231	0.0170 0.0170	0.0130-0.0380	0.0180	0.0023 0.0025	13.52 14.67
<i>SRS</i>	8-21 Feb. 2015	South Red Sea	57	1 st opt. depth 0.5-200	0.0135 0.0126	0.0300 0.0300	0.0192 0.0170			0.0056 0.0037	29.08 21.43
<i>All</i>	---	Red Sea	367	1 st opt. depth 0.5-200	0.0099 0.0090	0.0300 0.0300	0.0173 0.0172			0.0035 0.0035	20.15 19.85

Figure 3: Averages of CDOM absorption spectra (solid line), plus standard deviations (gray shaded area) measured in the northern Red Sea (NRS), central Red Sea (CRS) and southern Red Sea (SRS). The corresponding exponential fit referenced to 443 nm is represented by the dotted line in each panel. CDOM absorption decreased exponentially with increasing wavelength.

Figure 4: Frequency distributions of $a_{\text{CDOM}(443)}$ (A, B, C, D) and S_{CDOM} (E, F, G, H) estimated for the sub-regions of the Red Sea. The median of $a_{\text{CDOM}(443)}$ and S_{CDOM} values is represented by a gray dashed line in each panel. Variations in $a_{\text{CDOM}(443)}$ are large. Waters in SRS show high values varying from 0.03 and 0.10 m^{-1} , whereas in the NRS and CRS, $a_{\text{CDOM}(443)}$ varies from 0.01 to 0.06 m^{-1} . Rare high values ($> 0.1 \text{ m}^{-1}$) can be observed in the NRS within the cyclonic eddy. S_{CDOM} ranged from 0.009 to 0.030 nm^{-1} , with an average value of $0.0170 \pm 0.0035 \text{ nm}^{-1}$.

CDOM absorption characteristics in the Red Sea---

The latitudinal and vertical variations of $a_{\text{CDOM}(443)}$ and the spectral slope of CDOM absorption for each cruise are presented in Figs. 5, 6, 7 and 8. In this study, we used the CDOM absorption at 443 nm, which is commonly adopted to describe the amount of CDOM in the water because most inversion algorithms generate CDOM absorption at this specific wavelength.

Figure 4 illustrates the overall changes in $a_{\text{CDOM}}(443)$ observed from surface to 200 m. The variations in $a_{\text{CDOM}}(443)$ can reach almost two orders of magnitude. Waters in the SRS showed high values of $a_{\text{CDOM}}(443)$ that varied from 0.03 and 0.10 m^{-1} along the water column (0-200 m) (Figs. 4 and 8A), whereas $a_{\text{CDOM}}(443)$ varied from 0.01 to 0.06 m^{-1} in the NRS and the CRS (Figs. 4 and 5A, 6A, 7A). However, we also observed rare values higher than 0.10 m^{-1} in the NRS, although the highest median value was observed in the SRS (Figs. 4E, H and 5A, 8A). The distribution of $a_{\text{CDOM}}(443)$ along the water column was relatively homogeneous. Some patches of high values of $a_{\text{CDOM}}(443)$, however, were along the water column in all regions (Figs. 5, 6, 7, 8). The $a_{\text{CDOM}}(443)$ statistics for each region of the Red Sea are provided in Table 3.

The variations of S_{CDOM} across sections in the NRS and CRS were surprisingly large. High S_{CDOM} values (from 0.018 to 0.025 nm^{-1}) were surprisingly found from 80 to 200 m, and low values (from 0.011 to 0.017 nm^{-1}) were found at surface (Figs. 5B, 6B, 7B). However, as expected, we found low values in the deeper layer and high values at the surface in the CRS and NRS (Figs. 5B, 6B, 7B). In some stations, the slope values were homogeneously distributed along the water column, approximately 0.014 nm^{-1} in the NRS (Figs. 5B, 26.5° N – 27.5° N) and 0.016 nm^{-1} in the CRS (Fig. 6B, 23.5° N and Fig. 7B, 20° N and 24° N). In the SRS, the variability of S_{CDOM} was less than that in the CRS and NRS. In this region, the S_{CDOM} values were higher from surface to 70 m (0.017 to 0.03 nm^{-1}) and lower, approximately 0.016 nm^{-1} , from 70 m to 200 m (Fig. 8B). However, a patch of higher values in the bottom was observed along the axis (Fig. 8B, 17° N – 17.5° N).

Our observations indicate that CDOM absorption could be roughly estimated as

$$a_{\text{CDOM}}(\lambda) = a_{\text{CDOM}}(443)e^{-0.017(\lambda-443)} \quad (3)$$

Figure 5: Section of $a_{\text{CDOM}}(443)$ (m^{-1}) and S_{CDOM} (nm^{-1}) along the central axis of the northern Red Sea in April 2015 (NRS) within 0.5-200 m. Contours plot of (A) $a_{\text{CDOM}}(443)$ (m^{-1}) and (B) S_{CDOM} (nm^{-1}). The vertical black dots represent sampling points of each profile. Vertical variation in $a_{\text{CDOM}}(443)$ does not reflect the patterns in [Chl a]. $a_{\text{CDOM}}(443)$ along the water column is generally homogeneous. Patches of high $a_{\text{CDOM}}(443)$ are observed within the cyclonic eddy at 27.3° N. The distribution of $a_{\text{CDOM}}(443)$ appears to be determined by the distributions of temperature and salinity. S_{CDOM} is highly variable, with a tendency toward high values (0.018 - 0.025 nm^{-1}) from 80 to 200 m and low values (0.010 - 0.017 nm^{-1}) at the surface.

Figure 6: Section of $a_{CDOM}(443)$ (m^{-1}) and S_{CDOM} (nm^{-1}) along the central axis of the central Red Sea in October 2014 (CRS-01) within 0.5-200 m. Contours plot of (A) $a_{CDOM}(443)$ (m^{-1}) and (B) S_{CDOM} (nm^{-1}). The vertical black dots represent sampling points of each profile. Vertical variation in $a_{CDOM}(443)$ does not reflect the patterns in [Chl a]. $a_{CDOM}(443)$ along the water column is generally homogeneous. S_{CDOM} is highly variable, with a tendency toward higher values from 80 to 200 m and lower values in the upper layer.

Figure 7: Section of $a_{CDOM}(443)$ (m^{-1}) and S_{CDOM} (nm^{-1}) along the central axis of the central Red Sea in March-April 2015 (CRS-02) within 0.5-200 m. Contours plot of (A) $a_{CDOM}(443)$ (m^{-1}) and (B) S_{CDOM} (nm^{-1}). The vertical black dots represent the sampling points of each profile. Vertical variation in $a_{CDOM}(443)$ does not reflect the patterns in [Chl a]. $a_{CDOM}(443)$ along the water column is generally homogeneous. S_{CDOM} is highly variable, with a tendency toward higher values in deep water (100-200 m).

Figure 8: Section of $a_{\text{CDOM}}(443)$ (m^{-1}) and S_{CDOM} (nm^{-1}) along the central axis of the southern Red Sea in February 2015 (SRS) within 0.5-200 m. Contours plot of (A) $a_{\text{CDOM}}(443)$ (m^{-1}) and (B) S_{CDOM} (nm^{-1}). The vertical black dots represent the sampling points of each profile. Vertical variation in $a_{\text{CDOM}}(443)$ does not reflect the patterns in [Chl a]. $a_{\text{CDOM}}(443)$ along the water column is generally homogeneous. In this region, S_{CDOM} values were higher from the surface to 70 m (0.017 - 0.03 nm^{-1}) and lower, approximately 0.016 nm^{-1} , from 70 to 200 m.

Figure 9 shows a scatterplot of S_{CDOM} against $a_{\text{CDOM}}(443)$. In general, S_{CDOM} was inversely related to $a_{\text{CDOM}}(443)$, although data were scattered throughout the water column, both within the first optical depth ($R^2=0.17$, $N=104$, $p<0.01$) and in deeper waters ($R^2=0.14$, $N=367$, $p<0.01$).

The large scatter observed in this relationship can be attributed to the fact that the sampling has been done over different seasons.

Figure 9: Scatterplot of the slope of CDOM absorption spectra, S_{CDOM} , versus $a_{CDOM}(443)$ for all depths. Data collected within the first optical depth are represented by crosses. S_{CDOM} was inversely related to $a_{CDOM}(443)$ despite the fact that data were scattered throughout the water both within the first optical depth ($R^2=0.17$, $N=104$, $p<0.01$) and in deeper waters ($R^2=0.14$, $N=367$, $p<0.01$).

Table 3: Overall and region-by-region statistics for $a_{CDOM}(443)$ (m^{-1})

Acronym	Time	Region	N	Depths (m)	Min($a_{CDOM}(443)$)	Max($a_{CDOM}(443)$)	Mean($a_{CDOM}(443)$)	Standard Deviation a_{CDOM}	Variation coefficient, %
NRS	17-28 Apr. 2015	North Red Sea	198	1 st opt. depth 0-200	0.0108	0.1490	0.0350	0.0273	78.07
					0.0108	0.1603	0.0355	0.0249	70.07
CRS-01	16-28 Oct. 2014	Central Red Sea	52	1 st opt. depth 0-200	0.0074	0.0489	0.0239	0.0123	51.49
					0.0074	0.0525	0.0334	0.0116	34.61
CRS-02	29 Mar	Central	60	1 st opt.	0.0095	0.0722	0.0242	0.0158	65.09
					0.0095	0.0424	0.0238	0.0070	29.07

	.-3 Apr. 201 5	Red Sea		<i>depth</i> 0-200					
SRS	8-21 Feb. 201 5	South Red Sea	57	<i>1st</i> <i>opt.</i> <i>depth</i> 0-200	0.0346	0.0657	0.0493	0.0123	24.93
					0.0126	0.0996	0.0553	0.0141	25.57
All	---	Red Sea	36	<i>1st</i> <i>opt.</i> <i>depth</i> 0-200	0.0074	0.1490	0.0335	0.0234	69.73
			7		0.0074	0.1603	0.0370	0.0218	59.17

Bio-optical relationships---

The scatterplot between S_{CDOM} and [Chl a] for all cruises performed in the Red Sea is shown in Fig. 10. The relationship between S_{CDOM} and [Chl a] was not clear when all samples at all depths (0 - 200 m) were considered ($R^2=0.06$, $N=367$, $p<0.1$) or when the analysis was restricted to surface values only ($R^2=0.07$, $N=106$, $p<0.1$) (Fig. 10).

Figure 10: Variations of the CDOM exponential slope (S_{CDOM}) as a function of [Chl a] for the 0-200 m water column. Data collected within the first optical depth are represented by crosses. No

clear relationships between S_{CDOM} and $[\text{Chl } a]$ were apparent at all depths ($R^2=0.06$, $N=367$, $p<0.1$) or within the first optical depth ($R^2=0.07$, $N=104$, $p<0.1$).

The $a_{\text{CDOM}}(443)$ values significantly covaried with $[\text{Chl } a]$, although there was important scatter ($R^2=0.13$, $N=367$, $p<0.01$) (Fig. 11A). When only the values within the first optical depth were considered, the scatter was reduced, but $a_{\text{CDOM}}(443)$ was still loosely related to $[\text{Chl } a]$ (Fig. 11B) ($R^2=0.28$, $N=104$, $p<0.001$). Organelli et al. (2014) also observed a large scatter within the relationship between $a_{\text{CDOM}}(443)$ and $[\text{Chl } a]$ and attributed it to the time lag between the CDOM and chlorophyll maxima. The large scatter observed here may have been due to our sampling strategy (over different seasons from October 2014 to April 2015).

Our results did not agree well with the existing global relationships. All points in Figure 11B fell above the relationships proposed by Bricaud et al. (2010) and Morel and Gentili (2009a).

However, when we compared our dataset to the recent model developed for the Red Sea by Brewin et al. (2015), we found an improved match (Fig. 11B).

Figure 11: Variations of $a_{CDOM}(443)$ as a function of [Chl *a*] for the 0-200 m water column (A) and only within the first optical depth (B). The $a_{CDOM}(443)$ values significantly covaried with [Chl *a*], although with important scatter ($R^2=0.13$, $N=367$, $p<0.01$). When restricted to the first optical depth, the scatter was reduced but $a_{CDOM}(443)$ was still loosely related to [Chl *a*] ($R^2=0.28$, $N=104$, $p<0.001$). In panel (B), the relationships from Morel and Gentili (2009a), Bricaud et al. (2010) (red line) and the model plus the standard deviation developed by Brewin et al. (2015) (black and gray lines, respectively) are displayed. All points fell above the relationships proposed by Bricaud et al. (2010) and Morel and Gentili (2009a), but when we compared our dataset to the recent model developed for the Red Sea by Brewin et al. (2015), we found an improved match.

Discussion

This article presents the first study of direct in situ measurements of $a_{CDOM}(\lambda)$ obtained in the Red Sea. We showed that the overall range of variation in $a_{CDOM}(443)$ and S_{CDOM} was surprisingly large but that the range of values for $a_{CDOM}(443)$ and S_{CDOM} are within the ranges found in the literature for other oligotrophic environments (Bricaud et al. 1981; Blough et al. 1993; Barnard et al. 1998; Bricaud et al. 2010, Morel et al. 2010, Boss et al. 2013; Nelson and Siegel, 2013 and Organelli et al. 2014, Brewin et al. 2015).

Bio-physical environment---

We observed that [Chl *a*], temperature, and salinity greatly varied from south to north, as reported in previous studies (Sofianos and Johns, 2007; Yao et al. 2014a, b; Racault et al., 2015), and that the salinity regimes were a result of the presence of various distinct water masses interacting in the Red Sea (Figs. S1, S2, S3, S4 in supplementary material). We observed that cool ($< 26^{\circ}\text{C}$) and fresher (salinity < 38.25) water intruded from the southern part of the Red Sea (Fig. S4 in supplementary material), where the temperature and salinity of the surface water markedly decreased. Prior studies have shown that water from the Gulf of Aden is an important source of inorganic nutrients for the southern Red Sea (Churchill et al. 2014; Yao and Hoteit, 2015). The input of nutrients may enhance phytoplankton biomass, ultimately contributing to

increased CDOM production (Raitsos et al. 2013, 2015; Churchill et al. 2014; Yao and Hoteit, 2015).

The presence of mesoscale features (i.e., cyclonic and anticyclonic eddies, frontal zone) during our cruises may also impact the physico-chemical and biological properties along the water column, ultimately influencing the concentration and the spatial dispersion of CDOM in the Red Sea. Both modeling and observational studies (Clifford et al. 1997; McGillicuddy and Robinson, 1997; Sofianos and Johns, 2007; Papadopoulos et al. 2015; Kheireddine et al. 2017) have suggested that the occasional persistence of cyclonic eddies entrains deep nutrient-rich waters towards the surface layers of the northern Red Sea, thus considerably inducing the development of phytoplankton biomass. The cyclonic eddy observed in the NRS induced the subsurface elevation of isopycnal surfaces by 125-150 meters, indicative of eddy-driven upwelling. This upwelling may vertically transport nutrients towards the surface, contributing to higher phytoplankton biomass (Fig. S1C in supplementary material) and to higher CDOM concentrations (Fig. 5A). Specifically, Moran et al. (2000) showed that cyclonic eddies may produce episodic increases in CDOM in the upper layer of the water column, likely due to CDOM production by autotrophs and heterotrophs. Anticyclonic eddies have the capacity to transport deep, nutrient-rich water upwards at their edges (Mizobata et al. 2002), thereby affecting the plankton population distributions both above and below the euphotic layer (Abraham, 1998; Levy et al. 2001) and modifying the CDOM concentrations.

Variability in CDOM absorption properties ---

The variations in temperature and salinity determine the distribution of water masses, which may, in turn, influence the CDOM concentration and distribution in the Red Sea. Because salinity can be a good proxy of water masses in the Red Sea, relationships between $a_{CDOM}(443)$

and salinity were examined (Fig. 12A). The $a_{\text{CDOM}}(443)$ values generally show a tendency to decrease when going from the south (fresher water) to the north (saltier water) of the Red Sea (Fig. 12), suggesting that water masses may affect the CDOM distribution in the Red Sea.

Sources of CDOM patchiness can be attributed to mesoscale eddies, frontal zones, diffusive mixing and to biological processes (i.e., microbial activities). Numerous biological processes can induce a production of CDOM in the water column, e.g., phytoplankton release (Yentsch and Reichert, 1962; Sieburth and Jensen, 1969; Carder et al. 1989; Siegel and Michaels, 1996), zooplankton grazing (Momzikoff et al. 1994; Steinberg et al. 2004) and bacterial production (Nelson and Siegel, 2002; Organelli et al. 2014; Matsuoka et al. 2015). We showed that the a_{CDOM} values were loosely related to [Chl a] (Fig. 11). The large scatter in this relationship was expected because the vertical variations in $a_{\text{CDOM}}(443)$ do not reflect the patterns observed in [Chl a]. While a DCM was observed in all sections, no similarities to the vertical distribution of $a_{\text{CDOM}}(443)$ were observed. Organelli et al. (2014) observed a time shift between CDOM and [Chl a] maxima. In this study, measurements were taken during different periods (October 2014 to April 2015), which might have affected the relationship observed here.

Figure 12: $a_{CDOM}(443)$ as a function of the salinity (A) and latitude °N (B) for the different regions of the Red Sea described in the manuscript. The arrow indicates the direction of the latitudinal gradient. $a_{CDOM}(443)$ values generally show a clear tendency to decrease when going from the south to the north of the Red Sea.

The lack of a strong relationship between CDOM concentration and [Chl a] was addressed in previous studies (Nelson et al. 1998; Rochelle-Newall et al. 1999; Rochelle-Newall and Fisher, 2002), and it was suggested that direct release by phytoplankton is not a significant source of CDOM.

Previous studies suggested that CDOM could be preferentially formed by microbial heterotrophic activity rather than by being instantly released by phytoplankton biomass (Organelli et al. 2014; Matsuoka et al. 2015). Steinberg et al. (2004) also showed that zooplankton can directly release CDOM and could be a source of substrate for bacteria. As CDOM is a pool that is a combination of multiple organic compounds affected by different biotic and abiotic production and degradation pathways, identifying the dominating mechanism in each system is difficult and thus remains unresolved. In this study, we assumed that zooplankton could contribute considerably to the CDOM cycle, either by the release/excretion of CDOM or by supporting the microbial activity by furnishing substrate for microorganisms (Nelson and

Siegel, 2002; Steinberg et al. 2004). Recently, Kheireddine et al. (2017) showed pigment degradation related to grazing within the cyclonic eddy in the NRS and SRS, suggesting the presence of zooplankton. The Red Sea contains a diverse marine ecosystem that includes coastal coral reef, mangrove and seagrass communities, all of which sustain a large amount of biological productivity (Munday et al. 2008; Nagelkerken et al. 2008; Berumen et al. 2013; Almahasheer et al. 2016). Numerous studies showed increased CDOM concentrations near coral reefs, mangroves and seagrass beds (Boss and Zaneveld, 2003; Stabenau et al. 2004; Shank et al. 2010). Benthic processes may influence CDOM concentration and its distribution throughout the Red Sea. The Red Sea also receives significant inputs from dust (Prospero et al. 2002; Ginoux et al. 2012; Al-Taani et al. 2015; Prakash et al. 2015). Frequent dust outbreaks and dust storms have been observed in the Red Sea. Satellite observations (http://neo.sci.gsfc.nasa.gov/view.php?datasetId=MODAL2_M_AER_OD) reveal that Saharan dust events occurred in March-April 2015 in the entire Red Sea. During our three cruises, most of the stations where samples were collected occurred when significant atmospheric dust that may have affected the ecosystem was present over the Red Sea. Several studies showed that dust deposition can have an effect on bacterial ecology that is associated with an increase in dissolved organic matter and, hence, its CDOM fraction (Reche et al. 2009; Lekunberri et al. 2010; Morales-Baquero et al. 2013). Mladenov et al. (2011) reported that dust depositions may impact the dissolved organic matter compartment. De Vicente et al. (2012) directly showed that the chromophoric components from dust deposition could somewhat increase the concentration in CDOM and affect its spatial distribution. Numerous authors have indicated that variations in the spectral slope could provide further information on the CDOM pool characteristics, reflecting the nature and origin of the CDOM (Carder et al. 1989, Vodacek et al. 1997). The spectral slope was

identified as an indicator of photochemical degradation processes and microbial activities that are expected to vary both vertically and horizontally (Nelson and Siegel, 2013). Therefore, the slope is used as a substitute to analyze the variations in CDOM composition. The slope values obtained in this study are consistent with values stated in Brewin et al. (2015) for the Red Sea and with the mean value for the global ocean suggested by Morel and Gentili, (2009). Our results confirm that S_{CDOM} demonstrated large variability and showed, at some stations, high values at depth. This unusual distribution could result from the CDOM photo-reactivity, which might be dependent on the time of exposure to light. Yamashita et al. (2013) and Swan et al. (2012) found that the photochemical lability of CDOM was dependent on both the accumulated exposure to light and the light history of the water column. Swan et al. (2012) also observed that CDOM susceptibility to solar-induced destruction can vary according to the photo-oxidative and biogeochemical conditions of the water column (nitrate concentrations). Because of the biophysical characteristics encountered in the Red Sea, the spatio-vertical distribution observed in a_{CDOM} and S_{CDOM} can be affected, as observed in this study.

In agreement with earlier studies performed in the Pacific Ocean and the Mediterranean Sea, S_{CDOM} did not show any relationship with [Chl *a*] (Fig. 10) (Bracchini et al. 2010; Bricaud et al. 2010). However, contrary to previous reports (Organelli et al. 2014), S_{CDOM} values did not covary with the [Chl *a*] concentration when the analysis was restricted to surface values alone ($R^2=0.07$, $N=106$, $p<0.1$) (Fig. 10). This result could be partly attributed to the different time periods of sampling in the whole basin (October 2014 to April 2015).

As reported in previous studies (Carder et al. 1989; Nelson and Siegel, 2002), S_{CDOM} covaried inversely with CDOM absorption (Fig. 9), even though the relationship is highly scattered due to sampling done over different seasons. This tendency is consistent with earlier observations in

diverse oligotrophic and mesotrophic environments (Hayase and Tsubota, 1985; Summers et al. 1987; Carder et al. 1989; Pages and Gadel, 1990; Nelson and Siegel, 2002; Yacobi et al. 2003; Organelli et al. 2014). This result suggests that CDOM is a complex mixture of heterogeneous macromolecules (humic and fulvic acids, contributions of semi-labile and refractory fractions, or the photobleaching of some CDOM components) that coexist in different amounts within the phytoplankton compartment and that they may induce changes in S_{CDOM} values (Carder et al. 1989). The high spectral slope values associated with low values of $a_{CDOM}(443)$ at the surface may be due to the photobleaching of CDOM. Many studies have shown that the spectral slopes tend to rise with CDOM photodegradation and fall when the age of water masses increases in the deep layers of the ocean (Nelson et al. 2007), where the refractory components accumulate.

This study evaluates the variations in the absorbance spectra of dissolved organic matter in the visible range. We cannot determine whether the variations in S_{CDOM} values were produced by changes in the proportions of different chromophores due to changes in water masses or by chemical or biotic modifications of the CDOM. As previously reported, CDOM can have many sources (release from phytoplankton, production by bacteria, coral reefs, mangroves, or dust deposition, water masses). We can thus postulate that CDOM may be transported to the intermediate and the deep layers of the water column by mixing and sinking, and can also be transported to the surface or intermediate waters by many physical processes (mixing, cyclonic and anticyclonic eddies). Recently, Raitsos et al. (2017) showed that the water masses form coral reefs, which may be associated with high CDOM concentrations, were dragged away from coral reefs and transported from the west coast to the east coast and vice-versa due to cyclonic and anticyclonic eddies in the Red Sea. Additionally, Brewin et al. (2015) emphasized that the variations in CDOM are in good agreement with the findings for the interactions between

CDOM and marine physical processes (Blough et al. 1993; Nelson and Siegel, 2013). By comparing monthly climatological values of the CDOM index with modelled seasonal variations in the mixed layer depth from Yao et al. (2014a) in the northern Red Sea, Brewin et al. (2015) identified that the increase in the mixed layer depth entrains cool deep waters enriched in CDOM within the surface mixed layer. Furthermore, CDOM may be produced by heterotrophs in the dark ocean, as documented by Catalá et al. (2015). To thoroughly describe the origins, composition and fate of CDOM in the Red Sea, additional approaches and datasets, such as those that examine the ultraviolet wavelengths together with the visible absorbance spectra of CDOM and those that evaluate the changes in specific fluorophores within the CDOM pool, are needed.

Implications for retrieving [Chl a] concentrations in the first optical depth of the Red Sea from ocean color remote sensing measurements---

Ocean color remote sensing is one of the most important observational approaches for studying marine ecosystems at large spatio-temporal scales. Nevertheless, to accurately estimate the chlorophyll a, the ocean color algorithm must perform robustly. Brewin et al. (2015) tested the performance of different satellite chlorophyll a algorithms and found that the standard NASA OC4 (O'Reilly et al. 2000) and OCI (Hu et al. 2012) algorithms consistently overestimate chlorophyll a when they are correlated with in situ measurements. They therefore hypothesized that the high concentration of CDOM per unit of chlorophyll a could be responsible for the overestimation of the chlorophyll a concentration in the Red Sea using ocean-color algorithms, and they presented a bio-optical model that accounts for the regional characteristics of the Red Sea. However, this hypothesis was not based on direct measurements of CDOM absorption; thus, Brewin et al. (2015) reported that supplementary in situ measurements are necessary to verify if the Red Sea is characterized by high CDOM absorption per unit of chlorophyll a. We have

shown that when using in situ measurements, the Red Sea has elevated CDOM absorption for a given chlorophyll a concentration in comparison with the Mediterranean Sea and subequatorial Pacific waters. For instance, the CDOM absorption measured in the Red Sea over the [Chl a] range $0.01 - 0.6 \text{ mg.m}^{-3}$ is 20% to 550% higher than that measured in the Mediterranean Sea (Organelli et al. 2014) (Fig. 11). The empirical relationships established between the CDOM absorption coefficients and chlorophyll a concentrations (Morel and Gentili, 2009a, Bricaud et al. 2010) for different regions of the global ocean were compared with that obtained for the Red Sea, and they reveal that CDOM concentration in the Red Sea was higher than that typically expected for oligotrophic waters, thus verifying the hypothesis proposed by Brewin et al. (2015). However, we observed that some deviations from the model established for the Red Sea (Fig. 11B) persist, indicating that the model would require adjustments. Note that the model developed by Brewin et al. (2015) was derived from a dataset of indirect measurements of CDOM concentration and did not include all seasons; this likely explains the uncertainty that we observed in the derivation of $a_{\text{CDOM}}(\lambda)$. Thus, in situ measurements of CDOM absorption collected in the Red Sea including all environmental conditions with variable photobleaching situations will be required to develop a better-adapted version of the model established by Brewin et al. (2015).

Numerous studies (Carder et al. 1991; Lee et al. 1996; Morel and Gentili, 2009a) have provided evidence that a high CDOM concentration within the first optical depth could affect the estimation of phytoplankton concentration from ocean color remote sensing. In oligotrophic waters such as the Red Sea, algorithms to retrieve a single parameter, such as chlorophyll a, are based on empirical relationships. CDOM absorption is greater at shorter wavelengths (ultraviolet-blue region of the absorption spectrum) and decreases exponentially toward longer

wavelengths. As with CDOM, phytoplankton cells absorb highly in the shorter wavelengths, with a peak near 443 nm. Consequently, ocean color algorithms to retrieve [Chl a] based on empirical relationships that do not consider the excess of CDOM concentration will overestimate chlorophyll a when they are used in the Red Sea (Brewin et al. 2015). Therefore, future research must focus on improving the ocean color algorithm so it accurately retrieves the CDOM, chlorophyll a and other variables for the Red Sea.

Acknowledgments

The authors express their gratitude to the scientists, officers and crews of the research vessel Thuwal and the Coastal and Marine Resources Core Lab for logistical support and assistance onboard during the fieldwork. V. Vellucci, E. Organelli and E. Boss are cordially thanked for their advice and valuable discussions of the results presented here. We also thank David Antoine and Annick Bricaud for providing CDOM absorption data collected in the Mediterranean Sea (BOUSSOLE project, “BOUée pour l’acquiSition d’une Série Optique à Long termE project”) and in the Pacific Ocean (BIOSOPÉ cruise, “Biogeochemistry and Optics South Pacific Experiment”), respectively. Robert Brewin and six anonymous reviewers are warmly thanked for their constructive comments on a previous version of the manuscript. Most of the figures presented here were drawn using Ocean Data View software (Schlitzer, R., Ocean Data View, <http://odv.awi.de>, 2012). This study is funded by the King Abdullah University of Science and Technology (KAUST), Kingdom of Saudi Arabia. The data presented in this study are available from the authors upon request (malika.kheireddine@kaust.edu.sa). The data are archived at the Red Sea database (RSDB) (<https://lthdatalib.kaust.edu.sa/rsdbUI/>).

References

- Abraham, E.R., 1998. The generation of plankton patchiness by turbulent stirring. *Nature*. 391(6667), 577-580, <http://dx.doi.org/10.1063/1.4751329>.
- Almahasheer, H., Aljowair, A., Duarte, C. M., Irigoien, X., 2016. Decadal stability of Red Sea mangroves. *Estuar. Coast. Mar. Sci.* 169, 164-172, <http://dx.doi.org/10.1016/j.ecss.2015.11.027>.
- Al-Taani, A.A., Rashdan, M., Khashashneh, S., 2015. Atmospheric dry deposition of mineral dust to the Gulf of Aqaba, Red Sea: Rate and trace elements. *Marine Poll. Bull.* 92(1-2), 252-258, <http://dx.doi.org/10.1016/j.marpolbul.2014.11.047>.
- Babin, M., Stramski, D., Ferrari, G.M., Claustre, H., Bricaud, A., Obolensky, G., Hoepffner, N., 2003. Variations in the light absorption coefficients of phytoplankton, nonalgal particles, and dissolved organic matter in coastal waters around Europe. *J. Geophys. Res.* 108 (C7), 3211, <http://dx.doi.org/10.1029/2001JC000882>.
- Barnard, A.H., Pegau, W.S., Zaneveld, J.R.V., 1998. Global relationships of the inherent optical properties of the oceans. *J. Geophys. Res.* 103(C11), 24955-24968, <http://dx.doi.org/10.1029/98jc01851>.
- Belkin, I.M., 2009. Rapid warming of Large Marine Ecosystems. *Prog. Oceanogr.* 81(1-4), 207-213, <http://dx.doi.org/10.1016/j.pocean.2009.04.011>.
- Berumen, M.L., Hoey, A.S., Bass, W.H., Bouwmeester, J., Catania, D., Cochran, J.E.M., Khalil, M.T., Miyake, S., Mughal, M.R., Spaet, J.L.Y., and Saenz-Agudelo, P., 2013. The

status of coral reef ecology research in the Red Sea. *Coral Reefs*. 32(3), 737-748, <http://dx.doi.org/10.1007/s00338-013-1055-8>.

Blough, N.V., Del Vecchio, R., 2002. Chromophoric DOM in the coastal environment. *Biogeochemistry of Marine Dissolved Organic Matter*, Edited by Hansell, D.A., Carlson, C.A., Academic Press, San Diego, pp. 509–546.

Blough, N.V., Green, S.A., 1995. Spectroscopic characterization and remote sensing of nonliving organic matter. R.G. Zepp, Ch. Sonntag (Eds.), *The Dahlem Workshop on the Role of Nonliving Organic Matter in the Earth's Carbon Cycle*, Berlin, 1993, Wiley, New York (1995), pp. 23–45.

Blough, N.V., Zafiriou, O.C., Bonilla, J., 1993. Optical absorption spectra of waters from the Orinoco River outflow: Terrestrial input of colored organic matter to the Caribbean. *J. Geophys. Res.* 98(C2), 2271-2278, <http://dx.doi.org/10.1029/92jc02763>.

Boss, E., Gildor, H., Slade, W., Sokoletsky, L., Oren, A., Loftin, J., 2013. Optical properties of the Dead Sea. *J. Geophys. Res.* 118(4), 1821-1829, <http://dx.doi.org/10.1002/jgrc.20109>.

Boss, E., Zaneveld, J.R.V., 2003. The effect of bottom substrate on inherent optical properties: Evidence of biogeochemical processes. *Limnol. Oceanogr.* 48(1), 346-354, http://dx.doi.org/10.4319/lo.2003.48.1_part_2.0346.

Boyer, T.P., Levitus, S., 1994. Quality control and processing of historical temperature, salinity and oxygen data. NOAA Technical Report NESDIS 81. U.S. Department of Commerce, Washington, DC, USA, 65 pp.

Bracchini, L., Tognazzi, A., Dattilo, A.M., Decembrini, F., Rossi, C., Loiselle, S.A., 2010. Sensitivity analysis of CDOM spectral slope in artificial and natural samples: an application in the central eastern Mediterranean Basin. *Aquat. Sci.* 72, 485–498, <http://dx.doi.org/10.1007/s00027-010-0150-y>.

Brewin, R.J.W., Raitsos, D.E., Dall'Olmo, G., Zarokanellos, N., Jackson, T., Racault, M.-F., Boss, E., Sathyendranath, S., Jones, B.H., Hoteit, I., 2015. Regional ocean-colour chlorophyll algorithms for the Red Sea. *Remote Sens. Environ.* 165, 64-85, <http://dx.doi.org/10.1016/j.rse.2015.04.024>.

Bricaud, A., Babin, M., Claustre, H., Ras, J., Tieche, F., 2010. Light absorption properties and absorption budget of Southeast Pacific waters. *J. Geophys. Res.* 115, C08009, <http://dx.doi.org/10.1029/2009JC005517>.

Bricaud, A., Ciotti, A.M., Gentili, B., 2012. Spatial-temporal variations in phytoplankton size and colored detrital matter absorption at global and regional scales, as derived from twelve years of SeaWiFS data (1998-2009). *Global Biogeochem. Cycles.* 26, GB1010, <http://dx.doi.org/10.1029/2010GB003952>.

Bricaud, A., Morel, A., Prieur, L., 1981. Absorption by dissolved organic matter of the sea (yellow substance) in the UV and visible domains. *Limnol. Oceanogr.* 26 (1), 43–53, <http://dx.doi.org/10.4319/lo.1981.26.1.0043>.

Carder, K.L., Hawes, S.K., Baker, K.A., Smith, R.C., Steward, R.G., Mitchell, B.G., 1991. Reflectance model for quantifying chlorophyll a in the presence of productivity degradation products. *J. Geophys. Res.* 96(C11), 20599–20611, <http://dx.doi.org/10.1029/91jc02117>.

Carder, K.L., Steward, R.G., Harvey, G.R., Ortner, P.B., 1989. Marine humic and fulvic acids: Their effects on remote sensing of ocean chlorophyll. *Limnol. Oceanogr.* 34(1), 68–81, <http://dx.doi.org/10.4319/lo.1989.34.1.0068>.

Catalá, T.S., Reche, I., Álvarez, M., Khatiwala, S., Guallart, E.F., Benítez-Barrios, V.M., Fuentes-Lema, A., Romera-Castillo, C., Nieto-Cid, M., Pelejero, C., Fraile-Nuez, E., Ortega-Retuerta, E., Marrasé, C., Álvarez-Salgado, X.A., 2015. Water mass age and aging driving chromophoric dissolved organic matter in the dark global ocean. *Global Biogeochem. Cycles.* 29(7), 917–934, <http://dx.doi.org/10.1002/2014gb005048>.

Churchill, J.H., Bower, A.S., McCorkle, D.C., Abualnaja, Y., 2014. The transport of nutrient-rich Indian Ocean water through the Red Sea and into coastal reef systems. *J. Mar. Res.* 72(3), 165–181, <https://doi.org/10.1357/002224014814901994>.

Clifford, M., Horton, C., Schmitz, J., Kantha, L.H., 1997. An oceanographic nowcast/forecast system for the Red Sea. *J. Geophys. Res.* 102(C11), 25101-25122, <http://dx.doi.org/10.1029/97jc01919>.

De Vicente, I., Ortega-Retuerta, E., Morales-Baquero, R., Reche, I., 2012. Contribution of dust inputs to dissolved organic carbon and water transparency in Mediterranean reservoirs. *Biogeosciences*, 9(12), 5049-5060, <http://dx.doi.org/10.5194/bg-9-5049-2012>.

Del Vecchio, R., Blough, N.V., 2002. Photobleaching of chromophoric dissolved organic matter in natural waters: kinetics and modeling. *Mar. Chem.* 78, 231–253, [http://dx.doi.org/10.1016/S0304-4203\(02\)00036-1](http://dx.doi.org/10.1016/S0304-4203(02)00036-1).

Del Vecchio, R., Blough, N.V., 2004. Spatial and seasonal distribution of chromophoric dissolved organic matter and dissolved organic carbon in the Middle Atlantic Bight. *Mar. Chem.* 89, 169–187, <http://dx.doi.org/10.1016/j.marchem.2004.02.027>.

Dreano, D., Raitsos, D.E., Gittings, J., Krokos, G., Hoteit I., 2016. The Gulf of Aden Intermediate Water Intrusion regulates the Southern Red Sea Summer Phytoplankton Blooms. *PloS One*, 11 (12): e0168440, <https://doi.org/10.1371/journal.pone.0168440>.

D'Sa, E.J., DiMarco, S.F., 2009. Seasonal variability and controls on chromophoric dissolved organic matter in a large river-dominated coastal margin. *Limnol. Oceanogr.* 54(6), 2233-2242, <http://dx.doi.org/10.4319/lo.2009.54.6.2233>.

Estrada, M., Marrase, C., Latasa, M., Berdalet, E., Delgado, M., Riera, T., 1993. Variability of deep chlorophyll maximum characteristics in the Northwestern Mediterranean. *Mar. Ecol. Prog. Ser.* 92(3), 289-300, <http://dx.doi.org/10.3354/meps092289>.

Ginoux, P., Prospero, J.M., Gill, T.E., Hsu, N.C., Zhao, M., 2012. Global-scale attribution of anthropogenic and natural dust sources and their emission rates based on MODIS Deep Blue aerosol products. *Rev. Geophys.* 50, <http://dx.doi.org/10.1029/2012rg000388>.

Gittings, J.A., Raitsos, D.E., Racault, M-F., Brewin, R.J., Pradhan, Y., Sathyendranath, S., Platt, T., 2017. Seasonal Phytoplankton Blooms in the Gulf of Aden revealed by Remote Sensing. *Remote Sens. Environ.*, 189 (2017), 56–66, <http://dx.doi.org/10.1016/j.rse.2016.10.043>.

Gordon, H.R., McCluney, W.R., 1975. Estimation of the depth of sunlight penetration in the sea for remote sensing. *Appl. Opt.* 14(2), 413-416, <http://dx.doi.org/10.1364/ao.14.000413>.

Häder, D.P., Kumar, H.D., Smith, R.C., Worrest, R.C., 1998. Effects on aquatic ecosystems. *Photochem. Photobiol.* 46(1-3), 53-68, [http://dx.doi.org/10.1016/s1011-1344\(98\)00185-7](http://dx.doi.org/10.1016/s1011-1344(98)00185-7).

Hansell, D.A., Carlson, C.A., Suzuki, Y., 2002. Dissolved organic carbon export with North Pacific Intermediate Water formation. *Global Biogeochem. Cycles*. 16(1), <http://dx.doi.org/10.1029/2000gb001361>.

Hayase, K., Tsubota, H., 1985. Sedimentary humic acid and fulvic acid as fluorescent organic materials. *Geochim. Cosmochim. Ac.* 49(1), 159-163, [http://dx.doi.org/10.1016/0016-7037\(85\)90200-5](http://dx.doi.org/10.1016/0016-7037(85)90200-5).

Helms, J.R., Stubbins, A., Ritchie, J.D., Minor, E.C., Kieber, D.J., Mopper, K., 2008. Absorption spectral slopes and slope ratios as indicators of molecular weight, source, and photobleaching of chromophoric dissolved organic matter. *Limnol. Oceanogr.* 53 (3), 955–969, <http://dx.doi.org/10.4319/lo.2008.53.3.0955>.

Herndl, G.J., Mullerniklas, G., Frick, J., 1993. Major role of ultraviolet-B in controlling bacterioplankton growth in the surface layer of the ocean. *Nature*, 361(6414), 717-719, <http://dx.doi.org/10.1038/361717a0>.

Højerslev, N., 1982. Yellow substance in the sea, in *The role of solar ultraviolet radiation in marine ecosystems*. J. Calkins, Ed., *The Role of Solar Ultraviolet Radiation in Marine Ecosystems*, Plenum, New York, 1982, pp. 263-281.

Hu, C., Lee, Z., Franz, B., 2012. Chlorophyll a algorithms for oligotrophic oceans: A novel approach based on three-band reflectance difference. *J. Geophys. Res.* 117, C01011, <http://dx.doi.org/10.1029/2011JC007395>.

Kalle, K., 1938. Zum problem der meereswasserfarbe. *Ann. Hydrol. Mar. Mitt.* 66, 1-13.

Kalle, K., 1966. The problem of the gelbstoff in the sea. *Oceanogr. Mar. Biol. Annu. Rev.* 4(9), 1-104.

Kheireddine, M., Ouhssain, M., Claustre, H., Uitz J., Gentili, B., Jones, B.H., 2017. Assessing Pigment-Based Phytoplankton Community Distributions in the Red Sea. *Front. Mar. Sci.*, 4:132, <https://doi.org/10.3389/fmars.2017.00132>.

Kirk, J.T.O., 1976. Yellow substance (Gelbstoff) and its contribution to the attenuation of photosynthetically active radiation in some inland and coastal south-eastern Australian waters. *Mar. Freshwater Res.* 27(1), 61-71, <https://doi.org/10.1071/MF9760061>.

Kirk, J.T.O., 1999. Multiple scattering of a photon flux: implications for the integral average cosine of the underwater light field. *Appl. Opt.* 38(15), 3134-3140, <https://doi.org/10.1364/AO.38.003134>.

Laane, R., Koole, L., 1982. The relation between fluorescence and dissolved organic carbon in the Ems-Dollart estuary and the western Wadden Sea. *Neth. J. Sea Res.* 15(2), 217-227, [https://doi.org/10.1016/0077-7579\(82\)90005-9](https://doi.org/10.1016/0077-7579(82)90005-9).

Laane, R., Kramer, K.J.M., 1990. Natural fluorescence in the North Sea and its major estuaries. *Neth. J. Sea Res.* 26(1), 1-9, [https://doi.org/10.1016/0077-7579\(90\)90052-i](https://doi.org/10.1016/0077-7579(90)90052-i).

Lee, Z.P., Carder, K.L., Peacock, T.G., Davis, C.O., Mueller, J.L., 1996. Method to derive ocean absorption coefficients from remote-sensing reflectance. *Appl. Opt.* 35(3), 453-462, <https://doi.org/10.1364/ao.35.000453>.

Lekunberri, I., Lefort, T., Romero, E., Vazquez-Dominguez, E., Romera-Castillo, C., Marrase, C., Peters, F., Weinbauer, M., Gasol, J.M., 2010. Effects of a dust deposition event on coastal marine microbial abundance and activity, bacterial community structure and ecosystem function. *J. Plankton. Res.* 32(4), 381-396, <http://dx.doi.org/10.1093/plankt/fbp137>.

Levitus, S., 1982. Climatological atlas of the world ocean, NOAA Professional Paper N°13, U.S. Govt. Printing Office, Washington, D. C., 173 pp.

Levy, M., Klein, P., Treguier, A.M., 2001. Impact of sub-mesoscale physics on production and subduction of phytoplankton in an oligotrophic regime. *J. Mar. Res.* 59(4), 535-565, <http://dx.doi.org/10.1357/002224001762842181>.

Longhurst, A., 2007. Toward an ecological geography of the sea. *Ecological Geography of the Sea*, pp. 1-17.

Matsuoka, A., Bricaud, A., Benner, R., Para, J., Sempere, R., Prieur, L., Belanger, S., Babin, M., 2012. Tracing the transport of colored dissolved organic matter in water masses of the Southern Beaufort Sea: relationship with hydrographic characteristics. *Biogeosciences*, 9(3), 925-940, <http://dx.doi.org/10.5194/bg-9-925-2012>.

Matsuoka, A., Hill, V., Huot, Y., Babin, M., Bricaud, A., 2011. Seasonal variability in the light absorption properties of western Arctic waters: Parameterization of the individual components of absorption for ocean color applications. *J. Geophys. Res.* 116, <http://dx.doi.org/10.1029/2009jc005594>.

Matsuoka, A., Ortega-Retuerta, E., Bricaud, A., Arrigo, K.R., Babin, M., 2015. Characteristics of colored dissolved organic matter (CDOM) in the Western Arctic Ocean: relationships with microbial activities. *Deep Sea Res. Part II*, 118, 44–52, <http://dx.doi.org/10.1016/j.dsr2.2015.02.012>.

McGillicuddy, D.J., Robinson, A.R., 1997. Eddy-induced nutrient supply and new production in the Sargasso Sea. *Deep Sea Res. Part. I* 44(8), 1427-1450, [http://dx.doi.org/10.1016/s0967-0637\(97\)00024-1](http://dx.doi.org/10.1016/s0967-0637(97)00024-1).

Mizobata, K., Saitoh, S.I., Shiimoto, A., Miyamura, T., Shiga, N., Imai, K., Toratani, M., Kajiwara, Y., Sasaoka, K., 2002. Bering Sea cyclonic and anticyclonic eddies observed during summer 2000 and 2001. *Prog. Oceanogr.* 55(1-2), 65-75, [http://dx.doi.org/10.1016/s0079-6611\(02\)00070-8](http://dx.doi.org/10.1016/s0079-6611(02)00070-8).

Mladenov, N., Sommaruga, R., Morales-Baquero, R., Laurion, I., Camarero, L., Diéguez, M.C., Camacho, A., Delgado, A., Torres, O., Chen, Z., Felip, M., Reche, I., 2011. Dust inputs and bacteria influence dissolved organic matter in clear alpine lakes. *Nature Communications*, 2, <http://dx.doi.org/10.1038/ncomms1411>.

Momzikoff, A., Dallot, S., Gondry, G., 1994. Distribution of seawater fluorescence and dissolved flavins in the Almeria-Oran front (Alboran Sea, western Mediterranean Sea). *J. Mar. Syst.* 5, 361–376, [https://doi.org/10.1016/0924-7963\(94\)90056-6](https://doi.org/10.1016/0924-7963(94)90056-6).

Mopper, K., Kieber, D.J., 2002. Photochemistry and the cycling of carbon, sulfur, nitrogen and phosphorus. *Biogeochemistry of marine dissolved organic matter*, pp. 455-507.

Morales-Baquero, R., Pulido-Villena, E., Reche, I., 2013. Chemical signature of Saharan dust on dry and wet atmospheric deposition in the south-western Mediterranean region. *Tellus B*, 65, <http://dx.doi.org/10.3402/tellusb.v65i0.18720>.

Moran, M.A., Sheldon, W.M., Zepp, R.G., 2000. Carbon loss and optical property changes during long-term photochemical and biological degradation of estuarine dissolved organic matter. *Limnol. Oceanogr.* 45(6), 1254-1264, <http://dx.doi.org/10.4319/lo.2000.45.6.1254>.

Morán, X.A.G., Estrada, M., Gasol, J.M., Pedros-Alio, C., 2002. Dissolved primary production and the strength of phytoplankton bacterioplankton coupling in contrasting marine regions. *Microbial Ecol.* 44(3), 217-223, <http://dx.doi.org/10.1007/s00248-002-1026-z>.

Morán, X.A.G., Taupier-Letage, I., Vazquez-Dominguez, E., Ruiz, S., Arin, L., Raimbault, P., Estrada, M., 2001. Physical-biological coupling in the Algerian Basin (SW Mediterranean): Influence of mesoscale instabilities on the biomass and production of phytoplankton and bacterioplankton. *Deep Sea Res. Part. I* 48(2), 405-437, [http://dx.doi.org/10.1016/s0967-0637\(00\)00042-x](http://dx.doi.org/10.1016/s0967-0637(00)00042-x).

Morel, A., 1980. In-water and remote measurements of ocean color. *Bound. Layer Meteor.* 18, 177-201, <http://dx.doi.org/10.1007/BF00121323>.

Morel, A., 1988. Optical modeling of the upper ocean in relation to its biogenous matter content (case I waters). *J. Geophys. Res.* 93(C9), 10749-10768, <http://dx.doi.org/10.1029/JC093iC09p10749>.

Morel, A., Claustre, H., Gentili, B., 2010. The most oligotrophic subtropical zones of the global ocean: similarities and differences in terms of chlorophyll and yellow substance. *Biogeosciences*, 7(10), 3139-3151, <http://dx.doi.org/10.5194/bg-7-3139-2010>.

Morel, A., Gentili, B., 2009a. A simple band ratio technique to quantify the colored dissolved and detrital organic material from ocean color remotely sensed data. *Remote Sens. Environ.* 113(5), 998-1011, <http://dx.doi.org/10.1016/j.rse.2009.01.008>.

Morel, A., Gentili, B., 2009b. The dissolved yellow substance and the shades of blue in the Mediterranean Sea. *Biogeosciences*, 6(11), 2625-2636.

Morel, A., Maritorena, S., 2001. Bio-optical properties of oceanic waters: A reappraisal. *J. Geophys. Res.* 106(C4), 7163-7180, <http://dx.doi.org/10.1029/2000jc000319>.

Munday, P.L., Jones, G.P., Pratchett, M.S., Williams, A.J., 2008. Climate change and the future for coral reef fishes. *Fish and Fisheries*, 9(3), 261-285, <http://dx.doi.org/10.1111/j.1467-2979.2008.00281.x>.

Nagelkerken, I., Blaber, S.J.M., Bouillon, S., Greene, P., Haywood, M., Kirtong, L.G., Meyneke, J.-O., Pawlik, J., Penrose, H.M., Sasekumaran, A., Somerfield, P.J., 2008. The habitat function of mangroves for terrestrial and marine fauna: a review. *Aquat. Bot.* 89(2), 155-185, <http://dx.doi.org/10.1016/j.aquabot.2007.12.007>.

Nandkeolyar N., Raman, M., Sandhya Kiran, G., Ajai, ., 2013. Comparative Analysis of Sea Surface Temperature Pattern in the Eastern and Western Gulfs of Arabian Sea and the Red Sea in Recent Past Using Satellite Data. *Int. J. Oceanogr.* vol. 2013, 16, <http://dx.doi.org/10.1155/2013/501602>.

Nelson, N.B., Carlson, C.A., Steinberg, D.K., 2004. Production of chromophoric dissolved organic matter by Sargasso Sea microbes. *Mar. Chem.* 89,273–287, <http://dx.doi.org/10.1016/j.marchem.2004.02.017>.

Nelson, N.B., Siegel, D.A., 2002. Chromophoric DOM in the open ocean. In: Hansell, D.A., Carlson, C.A. (Eds.), *Biogeochemistry of Marine Dissolved Organic Matter*. Academic, San Diego, Calif ,pp.547–578.

Nelson, N.B., Siegel, D.A., 2013. The global distribution and dynamics of chromophoric dissolved organic matter. *Annu. Rev. Mar. Sci.* 5,447–476, <http://dx.doi.org/10.1146/annurev-marine-120710-100751>.

Nelson, N.B., Siegel, D.A., Carlson, C.A., Swan, C.M., 2010. Tracing global biogeochemical cycles and meridional overturning circulation using chromophoric dissolved organic matter. *Geophys. Res. Lett.* 37, L03610, <http://dx.doi.org/10.1029/2009GL042325>.

Nelson, N.B., Siegel, D.A., Carlson, C.A., Swan, C., Smethie Jr, W.M., Khatiwala, S., 2007. Hydrography of chromophoric dissolved organic matter in the North Atlantic. *Deep Sea Res. Part I* 54, 710–731, <http://dx.doi.org/10.1016/j.dsr.2007.02.006>.

Nelson, N.B., Siegel, D.A., Michaels, A.F., 1998. Seasonal dynamics of colored dissolved material in the Sargasso Sea. *Deep Sea Res. Part I* 45, 931–957, [http://dx.doi.org/10.1016/s0967-0637\(97\)00106-4](http://dx.doi.org/10.1016/s0967-0637(97)00106-4).

Neumann, A.C., McGill, D.A., 1962. Circulation of the Red Sea in early summer. *Deep Sea Res.* 8, 223-235, [https://doi.org/10.1016/0146-6313\(61\)90023-5](https://doi.org/10.1016/0146-6313(61)90023-5).

Olaizola, M., Ziemann, D.A., Bienfang, P.K., Walsh, W.A., Conquest, L.D., 1993. Eddy-induced oscillations of the pycnocline affect the floristic composition and depth distribution of phytoplankton in the subtropical Pacific. *Mar. Biol.*, 116(4), 533-542, <http://dx.doi.org/10.1007/bf00355471>.

O'Reilly, J.E., Maritorena, S., Siegel, D., O'Brien, M.C., Toole, D., Mitchell, B.G., et al., 2000. Ocean color chlorophyll a algorithms for SeaWiFS, OC2, and OC4: Technical report. In S.B. Hooker, and E.R. Firestone (Eds.), *SeaWiFS postlaunch technical report series. SeaWiFS postlaunch calibration and validation analyses, part 3, Vol. 11.* (pp. 9–23). Greenbelt, Maryland: NASA, Goddard Space Flight Center.

Organelli, E., Bricaud, A., Antoine, D., Matsuoka, A., 2014. Seasonal dynamics of light absorption by chromophoric dissolved organic matter (CDOM) in the NW Mediterranean Sea (BOUSSOLE site). *Deep Sea Res. Partt. I.* 91, 72-85, <http://dx.doi.org/10.1016/j.dsr.2014.05.003>.

Ortega-Retuerta, E., Frazer, T.K., Duarte, C.M., Ruiz-Halpern, S., Tovar-Sánchez, A., Arrieta, J.M., Reche, I., 2009. Biogeneration of chromophoric dissolved organic matter by bacteria and krill in the Southern Ocean. *Limnol.Oceanogr.* 54(6), 1941–1950, <http://dx.doi.org/10.4319/lo.2009.54.6.1941>.

Ortega-Retuerta, E., Siegel, D.A., Nelson, N.B., Duarte, C.M., Reche, I., 2010. Observations of chromophoric dissolved and detrital organic matter distribution using remote sensing in the Southern Ocean: Validation, dynamics and regulation. *J. Marine Syst.* 82(4), 295-303, <http://dx.doi.org/10.1016/j.jmarsys.2010.06.004>.

Pages, J., Gadel, F., 1990. Dissolved organic matter and UV absorption in a tropical hyperhaline estuary. *Sci. Total Environ.* 99(1-2), 173-204, [http://dx.doi.org/10.1016/0048-9697\(90\)90222-g](http://dx.doi.org/10.1016/0048-9697(90)90222-g).

Papadopoulos, V.P., Zhan, P., Sofianos, S.S., Raitzos, D.E., Qurban, M., Abualnaja, Y., Bower, A., Kontoyiannis, H., Pavlidou, A., Asharaf, T.T.M., Zarokanellos, N., Hoteit, I., 2015. Factors governing the deep ventilation of the Red Sea. *J. Geophys. Res.* 120, 7493–7505, <http://dx.doi.org/10.1002/2015JC010996>.

Patzert, W.C., 1974. Wind-induced reversal in Red Sea circulation. *Deep Sea Res.* 21(2), 109-121, [http://dx.doi.org/10.1016/0011-7471\(74\)90068-0](http://dx.doi.org/10.1016/0011-7471(74)90068-0).

Pegau, W.S., Deric, G., Zaneveld, J.R.V., 1997. Absorption and attenuation of visible and near-infrared light in water: Dependence on temperature and salinity. *Appl. Opt.*, 36(24), 6035– 6046, <https://doi.org/10.1364/AO.36.006035>.

Prakash, P.J., Stenchikov, G., Kalenderski, S., Osipov, S., Bangalath, H., 2015. The impact of dust storms on the Arabian Peninsula and the Red Sea. *Atmos. Chem. Phys.* 15(1), 199-222, <http://dx.doi.org/10.5194/acp-15-199-2015>.

Prospero, J.M., Ginoux, P., Torres, O., Nicholson, S.E., Gill, T.E., 2002. Environmental characterization of global sources of atmospheric soil dust identified with the Nimbus 7 Total Ozone Mapping Spectrometer (TOMS) absorbing aerosol product. *Rev. Geophys.* 40(1), <http://dx.doi.org/10.1029/2000rg000095>.

Racault, M.-F., Raitsos, D.E., Berumen, M.L., Brewin, R.J.W., Platt, T., Sathyendranath, S., Hoteit, I., 2015. Phytoplankton phenology indices in coral reef ecosystems: Application to ocean-color observations in the Red Sea. *Remote Sens. Environ.* 160, 222-234, <http://dx.doi.org/10.1016/j.rse.2015.01.019>.

Raitsos, D.E., Brewin, R.J.W., Zhan, P., Dreano, D., Pradhan, Y., Nanninga, G. B., Hoteit, I., 2017. Sensing coral reef connectivity pathways from space. *Scientific Reports*, 7, 9338, <http://dx.doi.org/10.1038/s41598-017-08729-w>.

Raitsos, D.E., Hoteit, I., Prihartato, P.K., Chronis, T., Triantafyllou, G., Abualnaja, Y., 2011. Abrupt warming of the Red Sea. *Geophys. Res. Lett.* 38, <http://dx.doi.org/10.1029/2011gl047984>.

Raitsos, D.E., Pradhan, Y., Brewin, R.J.W., Stenchikov, G., Hoteit, I., 2013. Remote Sensing the Phytoplankton Seasonal Succession of the Red Sea. *Plos One*, 8(6), <http://dx.doi.org/10.1371/journal.pone.0064909>.

Raitsos, D.E., Yi, X., Platt, T., Racault, M.-F., Brewin, R.J.W., Pradhan, Y., Papadopoulos, V.P., Sathyendranath, S., Hoteit, I., 2015. Monsoon oscillations regulate fertility of the Red Sea. *Geophys. Res. Lett.* 42: 855–862, <http://dx.doi.org/10.1002/2014GL062882>.

Ras, J., Claustre, H., Uitz, J., 2008. Spatial variability of phytoplankton pigment distributions in the Subtropical South Pacific Ocean: comparison between in situ and predicted data. *Biogeosciences*, 5(2), 353-369, <http://dx.doi.org/10.5194/bg-5-353-2008>.

Reche, I., Ortega-Retuerta, E., Romera, O., Pulido-Villena, E., Morales-Baquero, R., Casamayor, E.O., 2009. Effect of Saharan dust inputs on bacterial activity and community

composition in Mediterranean lakes and reservoirs. *Limnol. Oceanogr.* 54(3), 869-879, <http://dx.doi.org/10.4319/lo.2009.54.3.0869>.

Rochelle-Newall, E.J., Fisher, T.R., 2002. Chromophoric dissolved organic matter and dissolved organic carbon in Chesapeake Bay. *Mar. Chem.* 77(1), 23-41, [http://dx.doi.org/10.1016/s0304-4203\(01\)00073-1](http://dx.doi.org/10.1016/s0304-4203(01)00073-1).

Rochelle-Newall, E.J., Fisher, T.R., Fan, C., Glibert, P. M., 1999. Dynamics of chromophoric dissolved organic matter and dissolved organic carbon in experimental mesocosms, *Int. J. Remote Sens.* 20(3), 627-641, <http://dx.doi.org/10.1080/014311699213389>.

Shank, G.C., Zepp, R.G., Vahatalo, A., Lee, R., Bartels, E., 2010. Photobleaching kinetics of chromophoric dissolved organic matter derived from mangrove leaf litter and floating *Sargassum* colonies. *Mar. Chem.* 119(1-4), 162-171, <http://dx.doi.org/10.1016/j.marchem.2010.01.003>.

Shifrin, K.S., 1988. Physical optics of ocean water. American Institute of Physics, pp. 285.

Sieburth, J.M., Jensen, A., 1969. Studies on algal substances in the sea. II. The formation of Gelbstoff (humic material) by exudates of Phaeophyta. *J. Exp. Mar. Biol. Ecol.* 3(3), 275-289, [http://dx.doi.org/10.1016/0022-0981\(69\)90051-3](http://dx.doi.org/10.1016/0022-0981(69)90051-3).

Siegel, D.A., Maritorena, S., Nelson, N.B., Behrenfeld, M.J., McClain, C.R., 2005. Colored dissolved organic matter and its influence on the satellite-based characterization of the ocean biosphere. *Geophys. Res. Lett.* 32(20), <http://dx.doi.org/10.1029/2005gl024310>.

Siegel, D.A., Maritorena, S., Nelson, N.B., Hansell, D.A., Lorenzi-Kayser, M., 2002. Global distribution and dynamics of colored dissolved and detrital organic materials. *J. Geophys. Res.* 107(C12), <http://dx.doi.org/10.1029/2001jc000965>.

Siegel, D.A., Michaels, A.F., 1996. Quantification of non-algal light attenuation in the Sargasso Sea: Implications for biogeochemistry and remote sensing. *Deep Sea Res. Part. II.* 43(2-3), 321-345, [http://dx.doi.org/10.1016/0967-0645\(96\)00088-4](http://dx.doi.org/10.1016/0967-0645(96)00088-4).

Siegel, D.A., Michaels, A.F., Sorensen, J.C., Obrien, M.C., Hammer, M.A., 1995. Seasonal variability of light availability and utilization in the Sargasso Sea. *J. Geophys. Res.* 100(C5), 8695-8713, <http://dx.doi.org/10.1029/95jc00447>.

Slade, W.H., Boss, E., Dall'Olmo, G., Langner, M.R., Loftin, J., Behrenfeld, M.J., Roesler, C., Westberry, T.K., 2010. Underway and Moored Methods for Improving Accuracy in Measurement of Spectral Particulate Absorption and Attenuation. *J. Atmos. Ocean. Tech.* 27(10), 1733-1746, <http://dx.doi.org/10.1175/2010jtecho755.1>.

Smith, D.C., Simon, M., Alldredge, A.L., Azam, F., 1992. Intense hydrolytic enzyme activity on marine aggregates and implications for rapid particle dissolution. *Nature*, 359(6391), 139-142, <http://dx.doi.org/10.1038/359139a0>.

Sofianos, S.S., Johns, W.E., 2003. An Oceanic General Circulation Model (OGCM) investigation of the Red Sea circulation: 2. Three-dimensional circulation in the Red Sea. *J. Geophys. Res.* 108(C3), <http://dx.doi.org/10.1029/2001jc001185>.

Sofianos, S.S., Johns W.E., 2007. Observations of the summer red sea circulation. *J. Geophys. Res.* 112(C6), 1-20, <http://dx.doi.org/10.1029/2006jc003886>.

Stabenau, E.R., Zepp, R.G., Bartels, E., Zika, R.G., 2004. Role of the seagrass *Thalassia testudinum* as a source of chromophoric dissolved organic matter in coastal south Florida. *Mar. Ecol. Prog. Ser.* 282, 59-72, <http://dx.doi.org/10.3354/meps282059>.

Steinberg, D.K., Nelson, N.B., Carlson, C.A., Prusak, A.C., 2004. Production of chromophoric dissolved organic matter (CDOM) in the open ocean by zooplankton and the colonial cyanobacterium *Trichodesmium* spp. *Mar. Ecol. Prog. Ser.* 267, 45-56, <http://dx.doi.org/10.3354/meps267045>.

Sullivan, J.M., Twardowski, M.S., Zaneveld, J.R.V., Moore, C.M., Barnard, A.H., Donaghay, P.L., Rhoades, B., 2006. Hyperspectral temperature and salt dependencies of absorption by

water and heavy water in the 400-750 nm spectral range. *Appl. Opt.* 45(21), 5294-5309, <http://dx.doi.org/10.1364/ao.45.005294>.

Summers, R., Cornel, P., Roberts, P., 1987. Molecular size distribution and spectroscopic characterization of humic substances. *Sci. Total Environ.* 62, 27-37, [https://doi.org/10.1016/0048-9697\(87\)90478-5](https://doi.org/10.1016/0048-9697(87)90478-5).

Swan, C.M., Nelson, N.B., Siegel, D.A., Kostadinov, T.S., 2012. The effect of surface irradiance on the absorption spectrum of chromophoric dissolved organic matter in the global ocean. *Deep Sea Res. Part. I.* 63,52–64, <http://dx.doi.org/10.1016/j.dsr.2012.01.008>.

Swan, C.M., Siegel, D.A., Nelson, N.B., Carlson, C.A., Nasir, E., 2009. Biogeochemical and hydrographic controls on chromophoric dissolved organic matter distribution in the Pacific Ocean. *Deep Sea Res. Part. I.* 56(12), 2175-2192, <http://dx.doi.org/10.1016/j.dsr.2009.09.002>.

Thurman, E.M., 2012. Organic geochemistry of natural waters. Martinus Nijhoff/Dr W. Junk, Dordrecht, Netherlands, pp. 497, <http://dx.doi.org/10.1007/978-94-009-5095-5>.

Triantafyllou, G., Yao, F., Petihakis, G., Tsiaras, K.P., Raitos, D.E., Hoteit, I., 2014. Exploring the Red Sea seasonal ecosystem functioning using a three-dimensional biophysical model. *J. Geophys. Res.* 119(3), 1791-1811, <http://dx.doi.org/10.1002/2013jc009641>.

Twardowski, M.S., Boss, E., Sullivan, J.M., Donaghay, P.L., 2004. Modeling spectral absorption by chromophoric dissolved organic matter (CDOM). *Mar. Chem.*, 89: 69-88, <http://dx.doi.org/10.1016/j.marchem.2004.02.008>.

Twardowski, M. S., Donaghay, P.L., 2001. Separating in situ and terrigenous sources of absorption by dissolved materials in coastal waters. *J. Geophys. Res.*, 106(C2), 2545–2560, <http://dx.doi.org/10.1029/1999JC000039>.

Twardowski, M. S., Donaghay, P.L., 2002. Photobleaching of aquatic dissolved materials: Absorption removal, spectral alteration, and their interrelationship. *J. Geophys. Res.*, 107(C8), <http://dx.doi.org/10.1029/1999JC000281>.

Visser, S.A., 1984. Seasonal changes in the concentration and color of humic substances in some aquatic environments. *Freshwater Biol.* 14:79-87, <http://dx.doi.org/10.1111/j.1365-2427.1984.tb00023.x>.

Vodacek, A., Blough, N.V., DeGrandpre, M.D., Peltzer, E.T., Nelson, R.K., 1997. Seasonal variation of CDOM and DOC in the Middle Atlantic Bight: Terrestrial inputs and photooxidation. *Limnol. Oceanogr.* 42(4), 674-686, <http://dx.doi.org/10.4319/lo.1997.42.4.0674>.

Wafar, M., Ashraf, M., Manikandan, K.P., Qurban, M.A., Kattan, Y., 2016. Propagation of Gulf of Aden Intermediate Water (GAIW) in the Red Sea during autumn and its importance to

biological production. J. Mar. Syst. 154, 243-251,
<http://dx.doi.org/10.1016/j.jmarsys.2015.10.016>.

Williams, P., Druffel, E., 1988. Dissolved organic matter in the ocean: comments on a controversy. *Oceanogr.* 1(1), 14-17, <http://dx.doi.org/10.5670/oceanog.1988.33>.

Yacobi, Y.Z., Alberts, J.J., Takacs, M., McElvaine, M., 2003. Absorption spectroscopy of colored dissolved organic carbon in Georgia (USA) rivers: The impact of molecular size distribution. *J. Limnol.* 62(1), 41-46, <http://dx.doi.org/10.4081/jlimnol.2003.41>.

Yamashita, Y., Nosaka, Y., Suzuki, K., Ogawa, H., Takahashi, K., Saito, H., 2013. Photobleaching as a factor controlling spectral characteristics of chromophoric dissolved organic matter in open ocean. *Biogeosciences*, 10(11), 7207-7217, <http://dx.doi.org/10.5194/bg-10-7207-2013>.

Yamashita, Y., Tanoue, E., 2004., In situ production of chromophoric dissolved organic matter in coastal environments. *Geophys. Res. Lett.* 31(14), <http://dx.doi.org/10.1029/2004gl019734>.

Yao, F., Hoteit, I., 2015. Thermocline Regulated Seasonal Evolution of Surface Chlorophyll in the Gulf of Aden. *Plos One*, 10(3), <http://dx.doi.org/10.1371/journal.pone.0119951>.

Yao, F., Hoteit, I., Pratt, L.J., Bower, A.S., Koehl, A., Gopalakrishnan, G., Rivas, D., 2014a. Seasonal overturning circulation in the Red Sea: 2. Winter circulation. *J. Geophys. Res.* 119(4), 2263-2289, <http://dx.doi.org/10.1002/2013jc009331>.

Yao, F., Hoteit, I., Pratt, L.J., Bower, A.S., Zhai, P., Koehl, A., Gopalakrishnan, G., 2014b. Seasonal overturning circulation in the Red Sea: 1. Model validation and summer circulation. *J. Geophys. Res.* 119(4), 2238-2262, <http://dx.doi.org/10.1002/2013jc009004>.

Yentsch, C.S., Reichert, C.A., 1962. The interrelationship between water-soluble yellow substances and chloroplastic pigments in marine algae. *Bot. Mar.* 3, 65-74.