Deep-Sea Research II 130 (2016) 14–27

Contents lists available at ScienceDirect

Deep-Sea Research II

journal homepage: www.elsevier.com/locate/dsr2

Regular article

Regional chlorophyll *a* algorithms in the Arctic Ocean and their effect on satellite-derived primary production estimates

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ARTICLE INFO

Available online 7 May 2016

Keywords:
Arctic Ocean
Chukchi Sea
Primary production
Remote sensing
Ocean color algorithm
Phytoplankton
Case 2 water
CDOM
Pigment packaging

ABSTRACT

The Arctic is warming at approximately twice the global rate in response to anthropogenic climate change, resulting in disappearing sea ice, increased open water area, and a longer growing season (IPCC, 2013). This loss of sea ice has resulted in a 30% increase in annual net primary production (NPP) by Arctic Ocean phytoplankton between 1998 and 2012 (Arrigo and van Dijken, 2015). To quantify NPP, many algorithms require input of chlorophyll *a* (Chl *a*) concentration, which serves as a biomass proxy for phytoplankton. While satellites provide temporally and spatially extensive data, including Chl *a*, the standard global ocean color algorithms are prone to errors in Arctic Ocean waters due to higher than average phytoplankton pigment-packaging and chromophoric dissolved organic matter (CDOM) concentrations. Here, we evaluate retrievals of Chl *a* using existing ocean color algorithms, test and develop a new empirical ocean color algorithm for use in the Chukchi Sea, and evaluate the effect of using different satellite Chl *a* products as input to an NPP algorithm. Our results show that in the Chukchi Sea, Chl *a* was overestimated by the global algorithm (MODIS OC3Mv6) at concentrations lower than 0.9 mg m^{-3} because of contamination by CDOM absorption, but underestimated at higher concentrations because of pigment packaging. Only within the in situ Chl *a* range of 0.6–2 mg *a* m^{-3} was the satellite retrieval error by the OC3Mv6 algorithm below the ocean color community goal of $<35\%$. Using coincident in situ Chl *a* concentrations and optical data, a new linear empirical algorithm is developed (OC3L) that yields the lowest statistical error when estimating Chl *a* in the Chukchi Sea, compared to existing ocean color algorithms (OC3Mv6, OC4L, OC4P). When we estimated regional NPP using different Chl *a* satellite products as input, three distinct bio-optical provinces within the Arctic Ocean emerged. These provinces correspond to the inflow shelves, interior shelves, and outflow shelves+deep basin as defined by Carmack et al. (2006). Eleven sub-regions within the Arctic Ocean were grouped into each of these three provinces based on their mean value for R, the ratio of blue to green remote sensing reflectance (R_{RS}). Our results suggest that three algorithms tuned to each of the three bio-optical provinces may be sufficient to capture the bio-optical heterogeneity within the Arctic Ocean. Currently, only within the inflow shelf province do we feel confident that Chl *a* and NPP can be accurately estimated by satellite using the OC3L algorithm. The interior and outflow shelf+basin provinces require development of ocean color algorithms specific to their respective bio-optical conditions.

1. Introduction

The rate of anthropogenic warming in the Arctic is approximately twice the global rate, a phenomenon known as polar amplification, with the majority centered over the Arctic Ocean (IPCC, 2013). Rising temperatures in this region have already triggered profound environmental changes. Over the last few decades, sea ice has decreased in concentration, volume, and duration, with summer sea ice predicted to disappear completely by mid-century (IPCC, 2013; Overland and Wang, 2013). This loss of sea ice has resulted in a 30% increase in annual net primary production (NPP) by Arctic Ocean phytoplankton between 1998 and 2012 (Arrigo and van Dijken, 2015).

However, the correlation between sea ice decline and increased annual NPP may not be the only indicator of how NPP will continue to change (Tremblay and Gagnon, 2009; Bélanger et al., 2013). As the Arctic Ocean is predominately a light-limited system in spring, NPP will likely continue to increase and possibility shift earlier in the year as open water area and the length of the phytoplankton growing season increase (Walsh et al., 2005; Kahru et

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http://dx.doi.org/10.1016/j.dsr2.2016.04.020
0967-0645/

al., 2011; Arrigo and van Dijken, 2011). Since presently the magnitude of the phytoplankton bloom is ultimately limited by nutrient availability (Walsh et al., 2005; Tremblay and Gagnon, 2009), changes in climate that would lead to increased nutrient delivery to the euphotic zone, either through increases in Bering Strait flow, shelfbreak jets, wind-driven upwelling, or nutrients from rivers, could sustain NPP past its typical seasonal decline (Woodgate et al., 2012; Tremblay et al., 2012; Pickart et al., 2013). However, other environmental changes could reduce NPP in Arctic waters. For instance, increased freshwater flux through precipitation, ice melt, and river outflow would enhance stratification and strengthen the halocline, thus impeding upward delivery of nutrients (Peterson et al., 2006). Increased cloud cover, another consequence of climate change, limits incoming solar radiation and has been shown to decrease NPP in some regions, perhaps dampening the overall effect of higher light availability due to increased open water area (Bélanger et al., 2013).

Because the Arctic Ocean ecosystem has a relatively simple food web with few trophic links, NPP directly affects higher trophic level organisms (Grebmeier et al., 2006). Bottom-up changes in NPP may even trigger nonlinear responses, coined 'trophic amplifications' (Chust et al., 2014; Kirby and Beaugrand, 2009). For example, earlier phytoplankton blooms may cause a timing-mismatch for migratory organisms, such as whales and seabirds, that rely on predictable timing for feeding, which in turn could impact subsistence harvesting (ACIA, 2005; Kahru et al., 2011; Wassman, 2011). Equally, if earlier and larger phytoplankton blooms outpace grazers, carbon export and burial, benthic denitrification, and advection of excess phosphorus to the Atlantic Ocean may increase (Arrigo et al., 2008; Søreide et al., 2010). Thus, in order to understand the multi-layered effects of Arctic Ocean biogeochemistry and ecology as the climate continues to warm, it is imperative to accurately monitor changes in the magnitude and timing of NPP.

Given the limited sampling opportunities in the harsh Arctic environment, satellite remote sensing is a necessary tool for comprehensive spatial and temporal monitoring of the Arctic Ocean. The concentration of chlorophyll *a* (Chl *a*), the primary photosynthetic pigment in phytoplankton, serves as the proxy for phytoplankton biomass and is the main input for many NPP algorithms. Thus, satellite-based estimates of NPP that require Chl *a* as input are only as good as the accuracy of satellite-derived Chl *a* (Matsuoka et al., 2007; Arrigo et al., 2011; Lee et al., 2015).

The standard NASA Chl *a* product is derived using a global satellite ocean color algorithm. The standard algorithm is parameterized using the SeaBAM dataset that consists primarily of measurements obtained from Case 1 waters with very few data from polar regions or waters with Chl *a* concentrations >3 mg m^{-3} (O'Reilly et al., 1998). While the algorithm represents Case 1 waters reasonably well, it is not suitable for the Arctic Ocean, which exhibits unique bio-optical properties that differ significantly from the global oceans. First, cloud cover is persistent and sun angles are very low in polar regions, even in summer. Polar phytoplankton acclimate to perpetually low light conditions by increasing Chl *a* content per cell. While this 'pigment packaging' effectively increases total light absorbed per cell compared to phytoplankton in lower latitude waters, it lowers light absorption on a per Chl *a* basis (a^*_{ph}), resulting in an underestimation of Chl *a* by the global algorithm (Mitchell, 1992; Matsuoka et al., 2007; 2011; Wang and Cota, 2003). As importantly, Arctic rivers discharge high concentrations of chromophoric dissolved organic matter (CDOM) which absorbs strongly in the blue but weakly in the green wavelengths similar to phytoplankton pigments, causing an overestimation of Chl *a* by the global algorithm when CDOM levels are high (Mitchell, 1992; Matsuoka et al., 2007; Ben Mustapha et al., 2012).

The goal of this study is to evaluate the performance of different ocean color remote sensing Chl *a* algorithms and how they impact estimates of NPP in the Arctic Ocean. To do so, we evaluated satellite retrievals of both spectral remote sensing reflectance and surface Chl *a* using a robust in situ bio-optical and hydrographic dataset from the ICESCAPE 2010 and 2011 programs. Next, using this in situ dataset, we tested the performance of three existing ocean color algorithms, including the current global standard MODIS OC3Mv6 and the Arctic-specific OC4L and OC4P algorithms. In addition, we developed our own empirical ocean color algorithm (OC3L) for the Chukchi Sea using ICESCAPE data. Lastly, we tested the impact of these different Chl *a* products on estimates of NPP for the entire Arctic Ocean and its sub-regions.

2. Methods

2.1. ICESCAPE in situ field data

To evaluate the performance of the standard global ocean color algorithm used in the Chukchi Sea and broader Arctic Ocean, satellite retrievals of Chl *a* and remote sensing reflectance were compared to coincident in situ biological and optical measurements. During the ICESCAPE field program, these field measurements were made across the Chukchi Sea between Bering Strait and the western Beaufort Sea from 18 June through 16 July 2010 and from 28 June to 24 July 2011 onboard the USCGC *Healy* (Fig. 1a).

Radiometric measurements: In situ spectral remote sensing reflectance ($R_{RS}(\lambda)$, sr^{-1}) was calculated from underwater vertical profiles of spectral upwelling radiance (L_u), measured by a free fall Profiling Reflectance Radiometer (Biopherical Instruments Inc. PRR800/810), and spectral downwelling surface irradiance ($E_s(0^+)$), measured by a mast-mounted radiometer (PRR810) at 19 spectral channels. A typical PRR deployment consisted of three replicate casts and coincided with a midday CTD and Niskin bottle cast.

Water leaving radiance (L_w) was obtained by linearly extrapolating the regression of log-transformed L_u versus depth to just below the surface ($L_u(0^-)$) using the measured attenuation coefficient for upwelling radiance (K^*_{Lu}). $L_u(0^-)$ was then propagated through the water–air interface to obtain L_w using the equation (Morel and Gentili, 1991; 1993; Reynolds et al., 2001)

$$L_w(0^+) = 0.544\, L_u(0^-) \quad (1)$$

(the wavelength, λ, has been omitted from the equation for simplicity). The average R_{RS} of the replicate casts at a given wavelength was calculated as the average ratio of the upwelling radiance just above the water surface ($L_w(0^+)$) to $E_s(0^+)$ using the formula (Morel and Gentili, 1991, 1993; Reynolds et al., 2001)

$$R_{RS} = \frac{L_w(0^+)}{E_s(0^+)} \quad (2)$$

Chlorophyll a: Surface Chl *a* concentrations were measured at all 313 stations sampled during ICESCAPE (Fig. 1a). Samples were filtered onto 25 mm Whatman GF/F filters (nominal pore size 0.7 μm), placed in 5 mL of 90% acetone, and extracted in the dark at 3 °C for 24 h. The concentration of Chl *a* and phaeopigments was measured fluorometrically using a Turner 10-AU fluorometer (Turner Designs, Inc.) (Holm-Hansen et al., 1965). The fluorometer was calibrated using a pure Chl *a* standard (Sigma-Aldrich). We chose to use fluorometric Chl *a* instead of HPLC Chl *a* because there are significantly more fluorometric samples available for the satellite matchup analysis. Also, fluorometric Chl *a* allows for direct comparisons to the previous analyses of Cota et al., Wang et al., and Matsuoka et al.