



Seasonal and interannual variability of chlorophyll-*a* and associated physical synchronous variability in the western tropical Pacific



Xueyan Hou^{a,b}, Qing Dong^{a,*}, Cunjin Xue^a, Shuchao Wu^{a,b}

^a Key Laboratory of Digital Earth Science, Institute of Remote Sensing and Digital Earth, Chinese Academy of Sciences, Beijing 100094, China

^b University of Chinese Academy of Sciences, Beijing 100049, China

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ABSTRACT

Based on long-term satellite-derived ocean data sets and methods of empirical orthogonal function and singular value decomposition, we investigated the spatiotemporal variability of the chlorophyll-*a* concentration (CHL) on seasonal and interannual timescales in the western tropical Pacific associated with physical ocean variables of sea surface temperature (SST), sea level anomaly (SLA) and sea surface wind (SSW), and the El Niño Southern Oscillation (ENSO) index. The bio-physical synchronous variation on interannual timescale was also confirmed in terms of the scales of variability and oscillation periods in the time-frequency space using the methods of Fourier transform, Morlet wavelet transform, and wavelet coherence analysis. On a seasonal timescale, the first two modes of the monthly mean CHL fields described the consecutive spatiotemporal variation in CHL in the western tropical Pacific. CHL reached the maximum during late winter–early spring and minimum during summer–early autumn with the exception of the northeast of Papua New Guinea and the Solomon Islands. The CHL bloom in boreal winter–spring was closely associated with cold SST, high sea level along the North Equatorial Countercurrent meanders, and strong wind. On an interannual timescale, the variability of CHL exhibited a close correlation with SST, SLA, SSW, and ENSO. During El Niño, CHL increased in the oligotrophic western basin of the warm pool associated with cold SST, low SLA, and strong westerly winds but decreased in the mesotrophic eastern basin of the warm pool in association with warm SST, high SLA, and weak easterly trade winds. There may exist time-lag for the bio-physical covariation, i.e., CHL and SST varied simultaneously within 1 month, and CHL variations led SLA by approximately 0–3 months but lagged wind speed by about 1 month. In the time-frequency domain, the interannual variability in CHL and physical ocean variables had high common power, indicating that the variability scales and oscillation periods of CHL were significantly related to those of SST, SLA, and ENSO index. The significant anti-phase relationships were also shown between CHL and SST, CHL and SLA, and CHL and multivariate ENSO index through the wavelet coherence analysis.

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1. Introduction

Variability in ocean phytoplankton, which plays a vital role in global climate change by influencing the marine carbon cycle (Behrenfeld et al., 2006; Mackey et al., 1995; McPhaden et al., 2006), has received increasing attention with the accumulation of satellite-derived data sets of ocean variables over the past decades. The chlorophyll-*a* concentration (CHL) is the key biological variable for estimating the phytoplankton biomass and ocean primary production (Gregg et al., 2003; Martinez et al., 2009). We will use CHL and phytoplankton interchangeably hereafter. As an indicator of ocean phytoplankton, CHL is closely related to the physical ocean environments. Many studies have focused on the observation and variability in CHL and its relationship with ocean physical factors and climate variability (Leonard and McClain, 1996; McClain et al., 2002; Park et al., 2011; Radenac et al., 2012). The tropical Pacific

is the birthplace of El Niño Southern Oscillation (ENSO) (McPhaden et al., 2006). The western tropical Pacific enclosed by the 28 °C surface isotherm (Fig. 1b) is an important component of the climate system due to its strong air-sea interactions (Messié and Radenac, 2006; Radenac et al., 2013). The ocean primary production has been found to be low in the western tropical Pacific but with strong biological variability associated with ENSO (Radenac et al., 2013; Messié and Chavez, 2012).

The ecosystem in the western tropical Pacific is not uniform and could be separated into oligotrophic waters and mesotrophic waters by the 0.1 mg m⁻³ isoline of surface chlorophyll-*a* (Fig. 1a), which was revealed in previous studies (Antoine et al., 1996; Messié and Radenac, 2006; Radenac et al., 2013). Mesotrophic waters with CHL levels above 0.1 mg m⁻³ are observed from 155°E eastward along the equator and near the coasts of New Guinea and the Solomon Islands in the west (Fig. 1a). Oligotrophic waters with CHL levels of less than 0.1 mg m⁻³ are between the mesotrophic waters, particularly in the subtropical gyres (Radenac et al., 2013; Thomas et al., 2012).

* Corresponding author. Tel.: +86 10 82178121.
E-mail address: qdong@radi.ac.cn (Q. Dong).

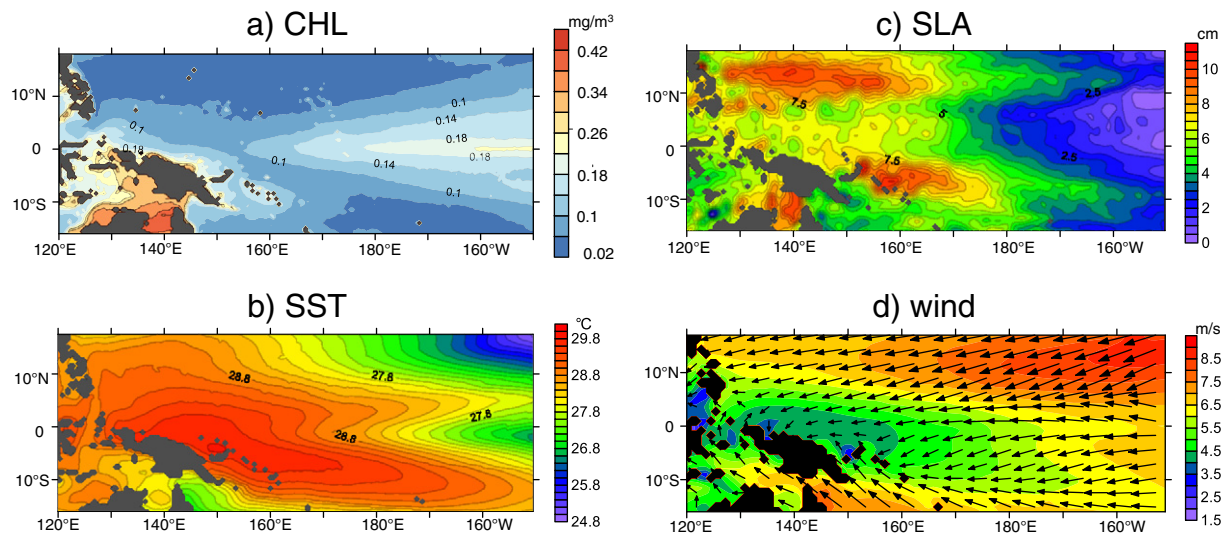


Fig. 1. The temporal mean of (a) chlorophyll concentration (CHL) averaged over 1998–2010, (b) sea surface temperature (SST) averaged over 1998–2009, (c) sea level anomaly (SLA) averaged over 1998–2010, (d) sea surface wind averaged over 1998–2010.

Phytoplankton growth is constrained by the availability of light and nutrients (nitrogen, iron, and phosphate) (Behrenfeld et al., 2006). For solar insolation, it is a primary driver of phytoplankton growth and temperature-driven stratification. In terms of the nutrient availability in the upper ocean, it is related to various ocean physical environment processes, such as deep nutrient-rich water upwelling and coastal nutrient-rich water advection. Therefore, the phytoplankton distribution and biomass are correlated with variations in the physical ocean environments, primarily the sea surface temperature (SST), sea level anomaly (SLA), and sea surface winds. In addition, variability in CHL is also closely correlated with ENSO (Chavez et al., 1999; Gierach et al., 2012; Kahru et al., 2010; Messié and Radenac, 2006; Radenac et al., 2012; Sasaoka et al., 2002; Uz and Yoder, 2004).

SST, which reflects the thermal stratification of the ocean and influences the convective mixing in the upper ocean with deep water, has an average temperature of over 28 °C in the western tropical Pacific (Fig. 1b). Thus, this area is also called a “warm pool.” In many areas, SST is generally negatively correlated with CHL because the upwelling of cold nutrient-rich water from below (or mixing with these deep waters) decreases SST and increases surface nutrients (Martinez et al., 2009; Park et al., 2011; Thomas et al., 2012). SLA is a proxy for surface current structure and thermocline depth, which influence the growth of phytoplankton by impacting the nutrient-rich upwelling (Thomas et al., 2012). In the western tropical Pacific, the sea level is relatively higher in the west and lower in the east (Fig. 1c) and is strongly correlated with CHL (Olita et al., 2011; Wilson and Adamec, 2001; Wilson and Adamec, 2002). Additionally, the northeasterly and southeasterly trade winds prevail in the western tropical Pacific (Fig. 1d). The winds affect phytoplankton biomass by influencing wind-driven currents and nutrient-rich deep water upwelling through Ekman pumping (Hou et al., 2014; Murtugudde et al., 1999; Yang et al., 2012).

Numerous previous studies have focused on the variability in CHL associated with the physical ocean environments in different ocean regions (Espinosa-Carreón et al., 2004; Legaard and Thomas, 2006; Tang et al., 2011; Thomas et al., 2012; Uz and Yoder, 2004; Venegas et al., 2008; Volpe et al., 2012). However, a few studies have focused on the western equatorial Pacific, with the exception of Messié and Radenac (2006) who studied the seasonal variability in CHL in the western tropical Pacific, Matsumoto and Furuya (2011) who investigated the interrelationships between phytoplankton dynamics and primary production in the western and central Pacific, and Radenac et al. (2013) who revealed the occurrence and variability of different ecosystems in the equatorial Pacific warm pool. In this study, we focus on the covariability

of CHL and its typical forcing physical variables in the western tropical Pacific (16°S–18°N, 120°E–150°W). Satellite remote sensing provides a way to observe changes in ocean variables using long time series of biological and physical ocean data. As satellites cannot detect CHL in the deep ocean, only the variability of CHL in the surface ocean is analyzed.

Using satellite-based data sets and the empirical orthogonal function (EOF) and singular value decomposition (SVD), we examined the seasonal and interannual variability of CHL as well as the covariability associated with its forcing physical factors. In addition, time-frequency pattern and oscillation periods for both biological and marine physical variables were studied using power spectrum analysis and continuous wavelet transform.

This manuscript is organized as follows. Section 2 describes the data and the methodologies used in this study. Section 3 includes the results and discussion, which are divided into four parts: (1) the results of the dominant seasonal variability of CHL obtained through EOF analysis, (2) the dominant interannual variability of CHL and its forcing physical variables resulting from EOF and time-lagged correlation analysis, (3) the covariability patterns of biological–physical variables on both seasonal and interannual timescales resulting from SVD analysis, and (4) the time-frequency patterns and oscillation periods associated with ENSO obtained through power spectrum analysis and continuous wavelet transform. Finally, the conclusions are provided in Section 4.

2. Data and methodology

2.1. Data

The monthly level 3 standard mapped ocean surface CHL data sets with a spatial grid of 9 km were derived from two satellite ocean colour sensors, namely, Sea-viewing Wide Field-of-view Sensor (SeaWiFS) r2010.0 and Moderate Resolution Imaging Spectroradiometer (MODIS) Aqua r2009.1, on the Ocean Color website (<http://oceancolor.gsfc.nasa.gov/>) (Hooker and McClain, 2000; Park et al., 2011). We used the data covering the period from October 1997 to December 2010, with SeaWiFS from October 1997 to December 2007 and MODIS from January 2008 to December 2010, because of the significant gaps in the SeaWiFS data after December 2007. MODIS has been used as the displacement and continuation of SeaWiFS in terms of the marine CHL and phytoplankton in many previous studies; thus, the consistency between the two missions has been established (Siegel and Franz, 2010; Park et al., 2011; Radenac et al., 2012; Couto et al., 2013; Lee et al., 2014). In

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