Global correlations between winds and ocean chlorophyll


Received 2 July 2010; revised 23 August 2010; accepted 1 October 2010; published XX Month 2010.

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1. Introduction

Phytoplankton productivity and biomass in the world ocean are limited by nutrient (N, P, Si, Fe) concentrations modulated by vertical mixing and seasonal variability in daily insolation. Phytoplankton productivity drives the oceanic biological pump and therefore has the potential to affect global atmospheric CO₂ levels. Changes in atmospheric CO₂ and the associated climate forcing can in turn impact phytoplankton productivity by changing ocean stratification, circulation, and pH. A number of authors have proposed the definition and use of quasi-stable biogeochemical provinces as a means of assessing basin scale oceanic productivity and biogeochemical characteristics. These provinces were traditionally based on measurements from ship-based platforms with the obvious consequence that the observed properties were dramatically undersampled in both space and time and the resulting boundaries were not well defined. Global time series of satellite measurements provide a significant amount of data to classify ocean environments as different biogeochemical provinces and to monitor the interannual and long-term changes in province boundaries. A number of different methods have been proposed to differentiate between biogeochemical provinces. These include the annual variability in phytoplankton pigment concentration; remotely sensed chlorophyll concentration, sea surface temperature, and the fixed boundaries of Longhurst’s provinces; a bioinformatic clustering algorithm using water-leaving radiance at two wavelengths and the sea surface temperature; a bioinformatic classification of ocean bio-optical signatures; Ocean ecosystems are governed by physical forcing, including winds, and studies of the relationships between winds and ocean biology have a long history. However, global, high-resolution data sets of winds and phytoplankton data have not been available until recently. Here we use the correlation between time series of satellite-derived winds and surface chlorophyll-a concentration to map the main biogeochemical provinces in the world ocean based on the dominant mechanisms responsible for the variability in phytoplankton biomass.

2. Data and Methods

Chlorophyll-a concentrations (Chl-a, mg m⁻³) were obtained from NASA’s Ocean Color Web site (McClain, 2009) (see Table 1 for Web links and references) and from the European Space Agency’s GlobColour project. For this analysis, we used remotely sensed level 3 (i.e., binned and mapped) monthly and daily Chl-a data sets that were derived using standard case 1 water algorithms (O’Reilly et al., 1998; Morel and Maritorena, 2001). Any single ocean color sensor has a limited daily coverage resulting from gaps between the swaths, Sun glint, and cloud cover. Merging data from multiple sensors, if data from more than one sensor are available, will increase the coverage due to the combination of patchy and uneven daily coverage from sensors viewing...
the ocean at slightly different times and geometries. Comp-
equivalent to the commonly used anomaly of log(whenever possible (April 2002 to October 2009) and data
Surface winds Cross
series were created by first by calculating the average value
$vU$ have approximately twice the mean global coverage and
during periods when only the single sensor data were avail-
\[2001\]. SST monthly anomalies were
product based on standard case 1 water algorithms should
in other words,
5 day
created by subtracting the monthly mean value from the
consistent over a wide range of conditions \[Maritorena
products of individual sensors, the large-scale Chl
91 distributions produced by the major color missions are
consistent over a wide range of conditions \[Morel et al.
3. Wind Speed and Chlorophyll Correlations
\[7\] For climatological ocean mixed-layer depth, we used
monthly climatological fields computed by de Boyer
Montégut et al. [2004] from hydrographic data. Intera-
nannual variability in the mixed-layer depth was obtained from
model results \[Behrenfeld et al., 2005\]. The model mixed-
layer depths perform well in the tropics, and hence we only
use them in the tropics and not in the extra-tropics, where
validation data are comparatively sparse. Monthly MLD
anomalies were created by subtracting the monthly mean
value from the current monthly value.

3. Wind Speed and Chlorophyll Correlations
\[8\] Chl-a, a measure of phytoplankton biomass and
biological productivity, is expected to depend on the avail-
ability of light and nutrients. Light levels are high at the ocean surface, so phytoplankton will experience strong light and potentially rapid growth if they are confined to the top of the ocean and unable to mix down to deeper levels. Nutrient levels tend to be high in the deep ocean, and processes that allow deep water to mix into the surface euphotic zone will supply nutrients to support biological growth (although nutrients can also have atmospheric or continental origins). Wind is one factor governing upwelling and vertical mixing into the upper ocean, and wind also influences the depth of the ocean’s surface mixed layer. In Figure 1a we show the correlation between monthly anomalies of wind speed and Chl-$a$. Figure 1 shows statistically significant patterns over the world ocean with positive correlations in the tropics and subtropical gyres and negative correlations in the subpolar gyres, the eastern tropical Pacific, and the eastern tropical Atlantic. Correlations that are not statistically significant at the 95% level are shaded gray in Figure 1. While the correlations are not large enough to explain all Chl-$a$ variability, the large-scale patterns in Figure 1a imply that wind speed fluctuations have a measurable influence on Chl-$a$. It is essential that the correlations be calculated between the Chl-$a$ values normalized by monthly climatological means and the wind anomalies, as discussed in section 2. Otherwise the annual cycle would dominate the correlations (e.g., at midlatitudes, the typically strong winds and low Chl-$a$ in the winter would produce a strong negative correlation).
increase this flux by increasing vertical mixing, likely by breaking down the stratification built by plankton growth and biomass are primarily limited by the solar flux, inducing upwelling of colder subsurface waters) and by increasing evaporation. Areas with positive correlations between monthly wind speed anomalies and SST are particularly deep in the regions of intermediate or mode water formation along the northern flank of the Antarctic Circumpolar Current in the Indian and Pacific sectors. Within these regions, Chl is negatively correlated with wind speed and SST is negative over most of the world ocean. This pattern is consistent with the hypothesis that in most regions increased wind speed cools the surface by vertical mixing (by breaking down the stratification built by solar flux, inducing upwelling of colder subsurface waters) and by increasing evaporation. Areas with positive correlation between monthly wind speed anomalies and SST anomalies, e.g., in the eastern tropical Pacific and eastern tropical Atlantic, are dominated by the interannual modes of variability [Murtugudde et al., 1999; Signorini et al., 1999] and will be discussed in section 4.

[10] Formally, the strength of wind-induced upwelling and mixing is governed by wind stress rather than wind speed. We therefore also calculated the correlations between monthly anomalies of Chl and anomalies of squared or cubed wind speed (not shown), both of which produced correlation patterns that were essentially identical to those shown in Figure 1a.

[11] The large-scale correlation patterns between wind speed and Chl suggest different limiting factors for phytoplankton biomass in different parts of the world ocean. Positive correlations that occur through much of the tropics and subtropics are consistent with the idea that phytoplankton growth and biomass are primarily limited by the upward flux of nutrients and that stronger winds can increase this flux by increasing vertical mixing, likely by deepening the mixed layer. The idea that stronger winds deepen the mixed layer is supported by Figure 1b, which shows that the correlation between monthly anomalies of wind speed and SST is negative over most of the world ocean. This pattern is consistent with the hypothesis that in most regions increased wind speed cools the surface by vertical mixing (by breaking down the stratification built by solar flux, inducing upwelling of colder subsurface waters) and by increasing evaporation. Areas with positive correlation between monthly wind speed anomalies and SST anomalies, e.g., in the eastern tropical Pacific and eastern tropical Atlantic, are dominated by the interannual modes of variability [Murtugudde et al., 1999; Signorini et al., 1999] and will be discussed in section 4.

[12] In contrast to the mostly tropical and subtropical regions with positive correlations between wind speed and Chl, the subpolar gyres of the North Pacific, North Atlantic, and Southern Ocean have an opposite response, with strong winds corresponding to reduced Chl concentrations, and this points to a different process compared with wind speed/SST interactions. The negative correlation between wind and Chl can be explained if strong winds tend to increase the depth of vertical mixing sufficiently to remove phytoplankton from the euphotic zone and therefore reduce their average light levels. Indeed, these regions of negative correlation correspond well to regions with deep late winter/early spring mixed layers, as shown in Figure 2. By the same token, deeper mixed layers and colder temperatures render winds less effective in cooling SSTs, resulting in weak correlations between the two [Seager and Murtugudde, 1997].

[13] Early spring mixed layers are plotted in Figure 2, because they are typically the deepest mixed layers observed during the year. The mixed layer becomes shallower during the summer, and water that was contained within the mixed layer in spring is below the base of the mixed layer in summer. Regions with deep spring mixed layers may have shallow summer or fall mixed layers, but the water just below these shallow mixed layers will typically have low stratification. Therefore, small anomalous increases in summer wind can easily remix the upper ocean and reestablish deep mixed layers, which would tend to decrease Chl, relative to the seasonal average. Both in the North Pacific and North Atlantic the correlations between Chl and wind speed anomalies change from negative to positive where the spring mixed layer becomes shallower than about 70 m along a generally equatorward gradient. Also, in comparison with the North Pacific, the North Atlantic has deeper mixed layers and more negative correlation between Chl and wind speed anomalies. In the Southern Ocean, spring mixed layers are particularly deep in the regions of intermediate or mode water formation along the northern flank of the Antarctic Circumpolar Current in the Indian and Pacific sectors. Within these regions, Chl is negatively correlated with
wind speed, whereas to the north, where mixed layers are shallower, Chl-α is positively correlated with wind speed. A number of patches with positive wind speed/Chl-α correlations occur inside the Southern Ocean, e.g., the Scotia Sea, northeast from the Antarctic Peninsula, and these regions tend to have shallower spring mixed layers. Within these two general scenarios, we acknowledge that the factors limiting phytoplankton growth can change during the annual cycle, particularly in temperate and polar environments. In addition, at high latitudes the light levels during winter are not sufficient to derive Chl-α from water-leaving radiances, and therefore the correlation patterns at high latitudes are naturally weighted toward the phytoplankton growth period. The correlation patterns are typically spatially variable or not significant in these regions (the Arctic and high-latitude Southern Ocean), where the growing season covers both newly mixed and stratified periods, and near-surface stratification is often controlled by the effects of salinity (e.g., through melting of ice).

Figure 3 shows the corresponding correlation patterns between monthly anomalies of Chl-α and the anomalies of eastward (uU) or northward (vU) wind pseudostress. Some areas, e.g., eastern North Atlantic (20°N–30°N) off Africa, have a relatively deep mixed layer (~100 m) in the spring but still show positive correlation between wind speed anomalies and Chl-α anomalies (Figure 1a). It is likely that the fertilizing effect of the Saharan dust associated with strong winds drives the positive correlation in the subtropical North Atlantic. This supposition is supported by the observed negative correlation with the eastward wind pseudostress anomaly (Figure 3a), i.e., positive correlation with the stronger than usual westward winds that bring the Saharan dust over the Atlantic. Atmospheric dust deposition may also affect some other areas such as the Arabian Sea, the South China Sea, and the Patagonian shelf.

As demonstrated by Figures 1 and 3, the correlation patterns between monthly anomalies of Chl-α or SST on the one hand and monthly anomalies of wind speed or wind components on the other show different patterns, each conveying a certain relationship. These correlation patterns as well as the correlation patterns between monthly Chl-α anomaly and the nonnormalized (i.e., without taking the
anomaly) wind speed and wind components were used in the means cluster analysis using the correlation coefficients between the following means anomaly and eastward wind pseudostress anomaly, eastern tropical Pacific and Atlantic. In a biogeographic means cluster analysis [Hartigan and Wong, 1979] the cluster shown with green pixels in Figure 4 is found Chl anomaly and anomaly, anomaly and eastward wind pseudostress anomaly, (3) monthly Chl-a anomaly and wind speed anomaly, (4) monthly Chl-a anomaly and eastward wind pseudostress, (5) monthly Chl-a anomaly and northward wind pseudostress, (6) monthly Chl-a anomaly and wind speed, and (7) monthly SST anomaly and wind speed anomaly.

4. Equatorial Upwelling Patterns

The most complex features including the highest correlations between Chl-a and winds can be seen in the equatorial regions. Here the western Pacific (WEPAC), with the highest positive correlation between Chl-a and wind speed is adjacent to the eastern Pacific (EEPAC) with the highest negative correlation between Chl-a and wind speed. We selected areas in the Western Equatorial Pacific (WEPAC, 2°N–2°S, 160°E–160°W) and Eastern Equatorial Pacific (EEPAC, 2°N–2°S, 140°W–100°W) to show the relationships in more detail.

The most striking correlation between wind speed and Chl-a anomalies is evident in the western tropical Pacific (Figure 5), where there is almost a linear correlation between the anomalies of these two variables, except during the 1998 ENSO that produced exceptionally strong phytoplankton blooms along the equator near 170°E during the transition between El Niño and La Niña in March–June 1998 [Murtugudde et al., 1999; Ryan et al., 2002]. The 1998 La Niña event stands out in Figure 5 as having anomalously high Chl-a even relative to the anomalously high wind speeds. The correlation with wind speed is primarily driven by the negative linear correlation between Chl-a and the zonal wind velocity (negative u, characteristic of upwelling-favorable winds, correlated with positive Chl-a), which occurs both north and south of the equator (Figures 6a and 6c). (Note that in Figure 6 we show zonal and meridional...
wind components rather than anomalies in order to indicate clearly the direction of the wind.)

The mechanisms invoked to explain the particular blooms have included the island effect of the Kiribati islands \[\text{Messié et al.}, 2006\], the erosion of the barrier layer in the western Pacific warm pool \[\text{Murtugudde et al.}, 1999\], and changes in the dynamics of the New Guinea coastal currents \[\text{Ryan et al.}, 2006\]. The 1997–1998 El Niño/Southern Oscillation (ENSO) event was comparatively warm, but since mid-1998, the equatorial

![Figure 5](image_url)

**Figure 5.** Time series of monthly anomalies of (left axis) wind speed \((U, \text{m/s})\) and (right axis) Chl-a (%) in the northern WEPAC box \(2°N–0°N, 160°E–160°W\). \(R = 0.705, P < 0.01\). The Multivariate ENSO index (MEI, http://www.esrl.noaa.gov/psd/data/climateindices/list/) is shown for comparison.

![Figure 6](image_url)

**Figure 6.** Scatter plots of the monthly Chl-a anomaly versus (a and c) eastward wind velocity \((u)\) and (b and d) northward wind velocity \((v)\) in the WEPAC area (a and b) north of the equator and (c and d) south of the equator. The correlations are slightly stronger if anomalies are used instead of the wind velocities themselves, but for simplicity we use the wind velocities here. The black circles are the months during the strong El Niño event of June 1997 to June 1998.
Pacific has experienced only weak El Niño events in 2002–2006, and 2006–2007. The meridional winds are relatively weak (±2 m/s) in the WEPAC area and show no correlation with Chl-a (Figures 6b and 6d). The direction of the north-south wind component is variable but more often poleward both north and south of the equator (positive $v$ in the north and negative $v$ in the south). Since the monsoonal regime intrudes into the western Pacific warm pool region, the meridional component is expected to be noisier than the zonal winds. The decaying stages of the strong ENSO event during the spring of 1998 resulted in the strongest southward winds both north and south of the equator driven by the deep convection that had shifted to the central-equatorial Pacific [Murtugudde et al., 2004].

[21] In the EEPAC the correlations between wind speed and Chl-a are opposite to those in the WEPAC area: Chl-a anomaly is inversely related to wind speed anomaly (Figure 1), positively correlated to eastward wind stress anomaly and negatively correlated to northward wind stress anomaly (Figure 3). However, the correlations are weaker than in the WEPAC and the relationships are not linear (Figure 7). The EEPAC is a region dominated by mixed layer-thermocline interactions and nutrient entrainment processes that drive enhanced production, although iron limitation produces high nutrient-low chlorophyll (HNLC) conditions [Coale et al., 1996]. Prevailing winds are westward on the equator but weaken each year during boreal winter or spring as well as during El Niño events (see Figure 7b, which shows the zonal wind rather than its anomaly). At these times westerly wind...
bursts in the western Pacific warm pool region generate equatorial Kelvin waves that propagate eastward along the equator, resulting in warm SSTs in the EEPAC region, a depressed thermocline, and deep mixed layers, which can reduce biological productivity [Aufdenkampe et al., 2002; Chavez et al., 1998]. Ryan et al. [2006] attributed blooms in 1998, 2003, and 2005 to shoaling of the equatorial Pacific thermocline and enhanced iron in the New Guinea coastal upwelling region, which feeds into the equatorial upwelling current. Similarly, Figures 7a–7c indicate anomalous Chl-a blooms in 1998, 1999/2000, 2003, 2006, and 2007 that occur in boreal late winter or spring when westward winds remain weak longer than usual, so that wind speeds are anomalously low for the season (Figures 7a and 7b). These anomalously low winds produce the negative correlation between anomalies of wind speed and Chl-a (Figure 1a), but the wind components u and v are not at their extremes during these events. At the times of these Chl-a blooms, meridional wind stress divergence (Figure 7c) is anomalously high, which is consistent with enhanced upwelling which favors production of cool SST anomalies (Figure 7e) and shallow mixed layer anomalies (Figure 7b). The 1998 and 1999 Chl-a blooms differ in timing and strength from the later blooms but correspond to similar patterns of low wind speed, high wind stress divergence, and low SST. The exceptionally high 1998 peak in Chl-a occurred during the transition from El Niño and La Niña with moderate u toward the west (~3 m s\(^{-1}\), Figure 7b) and weak v toward the north (1.5 m s\(^{-1}\), data not shown) [Murtugudde et al., 1999, 2004; Ryan et al., 2002]. Some of the overall correlations in Figures 1 and 3 might be caused by a single strong event (such as the 1997–1998 ENSO), but as shown by the above analysis, in most cases the correlation is spread over multiple time scales and involves multiple events (Figures 5–7). Cross-spectral analysis between time series of 5 day anomalies of wind speed and 5 day anomalies of Chl-a in the WEPAC area (Figure 4) confirms that significant coherence between wind speed and Chl-a exists over a range of periods from days to years. The broad peak between 1 and 3 months is likely related to the Madden Julian Oscillation [Wheeler and Hendon, 2004; Waliser et al., 2005].

In the tropical Indian Ocean there is no significant equatorial upwelling and the seasonal Chl-a signal is associated with the recirculation in the western equatorial region and the southeastern coast along Java and Sumatra [Murtugudde et al., 1999]. The interannual variability associated with the Indian Ocean Dipole/Zonal Mode produces a boreal fall anomaly in Chl-a [Murtugudde et al., 1999] and is seen as a small region of negative correlation in the far east in Figure 1. In the tropical Pacific and tropical Atlantic phytoplankton blooms are limited by the upward flux of nutrients through the thermocline as shown by the positive effect of higher wind speed in the central-western equatorial regions. Advection of nutrients from the eastern upwelling regions will add to this correlation. This pattern contrasts markedly with the mechanisms that appear to govern the North Pacific and especially the North Atlantic, where phytoplankton are limited by the average light level in the mixed layer, and the prevailing effect of winds is...
negative as wind mixing tends to deepen the mixed layer beyond the euphotic zone depth. (Figure 8)

5. Summary and Conclusions

This study has examined large-scale correlation patterns between winds and surface Chl-a concentrations. These patterns provide a consistent view of the main controlling factors of phytoplankton growth (nutrients or light). For most of the global ocean, high positive correlations occur in regions where the ocean mixed layer is comparatively shallow, and increased winds are readily able to deepen the mixed layer and entrain additional nutrients into the upper ocean. Strong negative correlations tend to occur in regions where late winter mixed layers are deep, and increased winds are likely to deepen the mixed layer enough to push phytoplankton out of the sunlit euphotic zone at the top of the ocean. Because of these distinctions between regions where increased winds appear to reduce light and regions where increased winds appear to increase nutrient availability, regions of strong positive and negative correlation map out distinct biogeochemical regimes within the world’s oceans.

Exceptions occur in the equatorial zone. Although the mixed layer is comparatively shallow along the equator, negative correlations nonetheless occur in the Eastern Equatorial Pacific and Eastern Equatorial Atlantic Oceans. These negative correlations are hypothesized to result from eastward propagating Kelvin wave dynamics that lead to a shoaling of the mixed layer in high wind speed conditions and therefore a remotely forced negative correlation between mixed-layer depth and wind speed.

Acknowledgments. Financial support was provided by the NASA Ocean Biology and Biogeochemistry Program, the NASA Physical Oceanography Program, and the National Science Foundation. We thank the NASA Ocean Color Processing Group, PODAAC at JPL, and NOAA NODC for satellite data.

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