Physical drivers of interannual chlorophyll variability in the eastern subtropical North Atlantic

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Abstract. Interannual chlorophyll variability and its driving mechanisms are evaluated in the eastern subtropical North Atlantic, where elevated surface chlorophyll concentrations regularly extend more than 1500 km into the central subtropical North Atlantic and modulate the areal extent of the North Atlantic’s lowest chlorophyll waters. We first characterize the considerable interannual variability in the size of the high chlorophyll region using Sea-WiFS satellite data. We then evaluate the relationship between satellite chlorophyll and Sea Surface Height (SSH), which are anticorrelated in the study region, most likely as a result of the inverse relationship between SSH and nutricline depth. To put these results in a longer temporal context, we study a hindcast simulation of a global ocean model with biogeochemistry (GFDL’s MOM4.1 with TOPAZ biogeochemistry), after evaluating the model’s skill at simulating chlorophyll and SSH relative to observations. In the simulation, the variability seen during the satellite era appears to be imbedded in a much larger multidecadal modulation. The drivers of such variability are assessed by evaluating all the terms in the nutrient budget of the euphotic zone. Because diffusive processes are not a dominant control on nutrient supply, stratification is not a good indicator of nutrient supply. Rather, vertical advection of nutrients, strongly tied to Ekman pumping, is the leading driver of variability in the size of the high chlorophyll region and the productivity within the study area.
1. Introduction

The eastern subtropical North Atlantic accommodates one of the most important eastern boundary current ecosystems in terms of total annual primary production [Carr, 2002; Carr and Kearns, 2003]. The upwelling system is bounded in the west by the subtropical gyre, a low-productivity region characterized by downwelling. In a warming environment, conditions in the subtropical gyre and in the coastal upwelling region may evolve differently. Coastal upwelling rates have been hypothesized to increase due to an increased land-sea pressure gradient [Bakun, 1990; Bakun et al., 2010]. On the other hand, increased stratification in oligotrophic gyres (a possible consequence of warming) is thought to reduce upward mixing of nutrients and decrease photosynthesis [McClain et al., 2004; Gregg et al., 2005; Behrenfeld et al., 2006; Polovina et al., 2008]. However, the link between declines in surface chlorophyll and increased stratification has most often been inferred from correlations between chlorophyll and sea surface temperature, leaving the underlying physical mechanisms unexplored. Specifically, an expansion of the lowest-chlorophyll regions has been seen in satellite chlorophyll data [Polovina et al., 2008], but it is not clear that the position of all of the oligotrophic boundaries are governed by vertical mixing. Furthermore, a debate exists about whether recently observed primary production decline is a climate-change induced trend [Polovina et al., 2008; Irwin and Oliver, 2009] or part of a multidecadal oscillation [Martinez et al., 2009]. A recent study of satellite ocean color data and output from three biogeochemical models suggests that distinguishing between climate-change driven trends and natural variability will require at least 40 years of continuous satellite chlorophyll measurements [Henson et al., 2010].
Here we focus on the physical mechanisms that give rise to variability in the size of the 
highly productive eastern boundary upwelling region in the subtropical North Atlantic, 
the most spatially and temporally variable of the four major eastern boundary current 
ecosystems [Carr, 2002]. Because the size of the high chlorophyll region governs the east-
ward extension of the neighboring oligotrophic region, our work also sheds light on the 
hypothesized link between the size of the oligotrophic region and vertical stratification 
[McClain et al., 2004; Gregg et al., 2005; Behrenfeld et al., 2006; Polovina et al., 2008]. 
The basin’s eastern limb shows one of the strongest zonal gradients of chlorophyll in the 
world’s ocean [e.g. Figure 2 in McClain, 2009], where the boundary between upwelling 
and downwelling favourable wind stress curl is found. The large-scale wind-stress field 
drives downwelling in the subtropical gyre, which results in a deep pycnocline and nutri-
cline, and consequent low surface chlorophyll (<0.07 mg m$^{-3}$). Along the basin’s eastern 
margin, the wind forcing causes divergence of the horizontal ocean currents as a result of 
both nearshore Ekman transport and offshore curl-driven Ekman pumping [McClain and 
Firestone, 1993]. The upward velocities induced by this coastal Ekman transport and 
Ekman pumping bring nutrient-rich waters to the euphotic layer, through a combination 
of along-isopycnal and diapycnal transfer, sustaining high chlorophyll concentrations at 
the basin’s eastern margin [Pelegrí et al., 2006].

Surface chlorophyll in the eastern subtropical North Atlantic has a strong seasonal cy-
cle; yet, interannual variability can be as large as the amplitude of the seasonal changes. 
Figure 1 illustrates the range of variability by comparing the difference between the av-
erage summer and winter chlorophyll concentrations to the difference between February 
2000, which saw maximum chlorophyll concentrations in the satellite record, and Febru-
ary 1998, which saw minimum concentrations. While recent advances have been made in understanding the differences among various eastern boundary upwelling systems [e.g. Démarcq, 2009; Lachkar and Gruber, 2011], the focus has been on the narrow near-shore regime where coastal upwelling is the principal source of nutrients. Our study region extends from the coast to the edge of the low chlorophyll subtropical gyre and its focus is on interannual to decadal-scale variability, which is at or beyond the limit of the time scale observed by satellites. We thus turn to a global ocean model coupled to a state of the art biogeochemistry model, which simulates 49 years of ocean variability using historical forcing from the CORE-reanalysis [Griffies et al., 2009; Large and Yeager, 2009]. The use of models allows in-depth study of several hypothesized physical drivers of chlorophyll variability such as changes in stratification and the large scale wind field.

The model’s skill at simulating spatial and temporal variability of the size of the high chlorophyll region is first tested by comparing its output to nine years of Sea-viewing Wide Field-of-View Sensor (SeaWiFS) data (section 3) and satellite measurements of sea surface height (SSH, section 4). We examine links between chlorophyll and SSH to explore the premise that SSH gives an insight to possible mechanisms explaining chlorophyll variability. We then assess the drivers of chlorophyll variability in the model output during the second half of the twentieth century. Specifically, we test various hypotheses that might explain biomass variability in section 5 through the assessment of each term in a nutrient budget. At the end of section 5, we consider the large-scale physical drivers influencing the delivery of nutrients, including shifts in the wind-driven boundary of the subtropical gyre and large-scale climate forcing. The conclusions are presented in section 6.
2. Data and Methods

2.1. Satellite Data

To validate the model’s ability to represent variability in our study region, we use satellite chlorophyll and altimetric observations. We use Level 3 SeaWiFS monthly chlorophyll downloaded from http://oceancolor.gsfc.nasa.gov at 9 km resolution for the period November 1997 to December 2006. What we call satellite SSH is the absolute sea surface dynamic topography produced by Ssalto/Duacs and distributed by Aviso, with support from CNES (downloaded from http://www.aviso.oceanobs.com/duacs), for the same time period. It corresponds to merged data from Topex/Poseidon and ERS satellites. We averaged the original weekly data to monthly resolution, and both data sets were regridded onto the model grid.

2.2. Model Output

The global ocean general circulation model used is Version 4 of the Geophysical Fluid Dynamics Laboratory’s Modular Ocean Model (MOM4) [Griffies et al., 2008], forced with the Common Ocean-Ice Reference Experiment (CORE) data set [Griffies et al., 2009; Large and Yeager, 2009]. We use version 2 of the CORE reanalysis effort (CORE2), which includes six-hourly interannual varying meteorological fields for the period 1958 - 2006 (10 m air temperature, humidity, air density, zonal wind, meridional wind, and sea level pressure). The CORE2 winds are based on the 2-degree resolution NCEP reanalysis data set [Kalnay et al., 1996]. The speed and direction are adjusted using QuikSCAT satellite scatterometer wind vectors following Chin et al. [1998] (Figures 2a and b show the wind stress curl filed from QuikSCAT data [Risien and Chelton, 2008] and from CORE2 winds). The original resolution of approximately $2^\circ \times 2^\circ$ is interpolated to the model grid. Daily
varying shortwave and longwave radiative fluxes are available from 1983 and monthly varying precipitation since 1979; prior to those years, these variables are specified from the climatological annual cycle. Continental runoff, imposed as a climatological annual mean, is part of the CORE2 configuration. It is based on flows at river mouths but highly smoothed at a scale of approximately 4 degrees to give a salinity signature that is seen in climatologies. The ocean model has fifty levels in the vertical direction, a longitudinal resolution of 1° and a latitudinal resolution varying between 1° in the extratropics and 1/3° on the equator. Steric effects (i.e. effects of temperature, salinity and pressure on seawater buoyancy) are not included in our estimates of sea level variability. The simulation was initialized from hydrographic mean properties taken from the World Ocean Atlas 2001 [Conkright et al., 2002]. It was spun-up for 348 years with forcing from a climatological year calculated using the mean CORE data during the years 1958-1977 before the final loop with the 49 years of CORE interannual variability was integrated.

The biogeochemical component is given by the model Tracers for Ocean Phytoplankton with Allometric Zooplankton (TOPAZ), which simulates prognostically all major nutrient elements (N, P, Si and Fe) [Dunne et al., 2010]. The ecosystem is based in three classes of phytoplankton. The small class dominates under nutrient limitation; this size class resists sinking. Large phytoplankton represent diatoms and other phytoplankton that bloom and sink quickly. Finally, diazotrophs fix dinitrogen gas directly. Phytoplankton growth rates are modeled as a function of variable chlorophyll to carbon ratios and colimited by nutrients and light. The model includes river run-off estimates of geographically-variable $NO_3$, $NH_4$ and labile dissolved organic nitrogen concentrations from Green et al. [2004] and globally fixed river concentrations for iron, alkalinity and dissolved inorganic carbon.
A more detailed description of the biogeochemical model structure can be found in Dunne et al. [2010] and Henson et al. [2009].

2.3. Methods

We use the distance from the coast to the 0.2 mg m\(^{-3}\) isoline of surface chlorophyll \(D_{CHL}\) to characterize the size of the high chlorophyll region between the oligotrophic subtropical gyre and the African coast. Larger distances indicate a larger area with chlorophyll higher than 0.2 mg m\(^{-3}\). Studies on chlorophyll variability in coastal upwelling systems generally have defined the areas of high productivity as those having >1 mg m\(^{-3}\) [e.g. Nixon and Thomas, 2001; Carr, 2002; Demarcq et al., 2007; Lathuilière et al., 2008], while studies on the oligotrophic subtropical gyres have focused on surface waters having less than 0.07 mg m\(^{-3}\) to limit the study region [e.g. McClain et al., 2004; Polovina et al., 2008]. By studying the area defined by waters exceeding 0.2 mg m\(^{-3}\) we cover an intermediate region between the oligotrophic subtropical gyre and the chlorophyll-rich coastal regime. We repeated the analysis using other chlorophyll isolines that fall within the same region (from 0.07 to 0.3 mg m\(^{-3}\)), and found little change in the qualitative and quantitative results reported here. It is not possible to do the same analysis with the 1 mg m\(^{-3}\) isoline, as the model chlorophyll concentrations are lower than the satellite values near the coast, such that there are some latitudes that never see concentrations of 1 mg m\(^{-3}\) over the course of the year (see the Supplementary Material for further details). We evaluate the time evolution of chlorophyll distances as a function of latitude as well as their meridional average.

The temporal variability of SSH is also examined as a function of latitude and as an average over the rectangular study region. The differences in the mean SSH values between
model and observations arises from the reference level used (Figure 2a and b). Remotely sensed absolute dynamic topography, here simply named satellite SSH, is obtained by adding the sea level anomaly to a mean dynamic topography, calculated using a combination of altimetry and in-situ data, and subtracting a model geoid. In the numerical model, the surface of the ocean at rest is taken as the zero level and SSH at a given time step is given as a departure from this level.

We evaluate three physical mechanisms that may cause interannual changes in the size of the high chlorophyll area through variable nutrient supply into the study region: coastal Ekman transport, Ekman pumping, and mixing of subsurface nutrients into the euphotic layer. Coastal Ekman transport and Ekman pumping are not calculated explicitly in ocean general circulation models. Wind stress components are introduced in the momentum horizontal equations, and the vertical velocity (either forced by Ekman processes or not) is a result of any ensuing divergence of the horizontal currents. Therefore, we diagnose the Ekman transports from the wind stress fields forcing the model and compare them to the modeled vertical transports. The vertical volume transport due to coastal Ekman transport \( V_{\text{coast}}, \text{m}^3\text{s}^{-1} \) is calculated as

\[
V_{\text{coast}} = - \int_L \frac{\tau_s}{f \rho_w} dy, \tag{1}
\]

where \( \tau_s \) is the alongshore component of the wind stress (N m\(^{-2}\), positive northwards), \( L \) (m) is the coastline length, \( f \) is the Coriolis parameter (s\(^{-1}\)) and \( \rho_w \) is the water density (kg m\(^{-3}\)), taken as a constant value of 1025 kg m\(^{-3}\). Vertical velocities driven by Ekman pumping \( w_e, \text{m s}^{-1} \) are calculated as

\[
w_e = \text{curl}_z \left( \frac{\tau}{f \rho_w} \right), \tag{2}
\]
where $\mathbf{\tau}$ is the horizontal wind stress vector at the sea surface and the subscript $z$ indicates the vertical component of the curl. The vertical transport ($V_{\text{curl}}$, m$^3$ s$^{-1}$) related to this process is obtained by integrating the vertical velocities ($w_e$) over the area of the study region (10 to 24°N, 40°W to coast).

We investigate the relative importance of nutrient supply to the top 80 m in the study region through lateral advection, vertical advection and mixing. The 80 m depth level is close to the average euphotic layer depth, calculated as the depth of the 1 Watt m$^{-2}$ irradiance. Phosphate ($PO_4$) is used for the calculations, although it would be nearly equivalent to use nitrate as their temporal evolution is highly correlated in the model ($R=0.99$, $p<<0.01$, for monthly mean concentrations in the top 80 m over the study region). The phosphorus cycle is simpler than the nitrogen cycle, as the latter includes $N_2$ fixation at the surface and is strongly affected by denitrification. Therefore, fewer terms are required to close the phosphorous budget. Regarding silicic acid, this nutrient does not actually limit growth in the TOPAZ biogeochemical model; rather, it only increases the depth scale of penetration of exported material by increasing the amount of biogenic opal, which acts as a ballast for organic matter. The time rate of change of phosphate (mol m$^{-2}$ s$^{-1}$) is given by:

$$\frac{\partial PO_4}{\partial t} = -\left( \frac{\partial (uPO_4)}{\partial x} + \frac{\partial (vPO_4)}{\partial y} + \frac{\partial (wPO_4)}{\partial z} \right) + \text{mixing} + jPO_4. \quad (3)$$

The mixing term in equation (3) encompasses all subgrid-scale processes which result in either vertical or horizontal nutrient fluxes, including the advection due to the parameterization of submesoscale eddies according to the Gent and McWilliams [1990] scheme. Vertical mixing is parameterized in various schemes: a specified background diffusivity with a coefficient of $10^{-5}$ m$^2$ s$^{-1}$ diffuses tracers throughout the domain; the K-Profile
Parameterization scheme from Large et al. [1994] is important in and near the surface boundary layer and represents convective mixing; and parameterizations of tidal mixing from Simmons et al. [2004] and Lee et al. [2006] are important near the sea floor. The last term in equation (3), $jPO_4$, represents the biological sources minus the sinks of $PO_4$, i.e. the remineralization of particulate and dissolved phosphorus minus phosphate uptake by phytoplankton. The archived model data allows the calculation of a balanced nutrient budget everywhere except the surface grid cell. At the surface, a term used to filter SSH anomalies to reduce barotropic motions was accidentally excluded from the archived data. As a consequence, about 3% of the $PO_4$ variance in the phosphate time series is not explained by the sum of the budget terms in the surface grid cell only. Also, note that the nutrient budget terms were archived from a shorter simulation (1959 to 2004) that was in all other ways identical to the simulation used for all other analyses, and therefore years 1958, 2005 and 2006 are missing.

The study region is bounded by latitudes 10 to 24° N, from 40° W to the coast, in the top 80 m of the water column. The variables studied have different spatial dimensions, i.e. chlorophyll distance varies only as a function of latitude, chlorophyll concentrations and SSH vary as a function of latitude and longitude, and $PO_4$ concentrations as a function of latitude, longitude and depth. Throughout this work, all spatial averages and integrations of the variables are made within these boundaries, unless specified otherwise.

3. Satellite - Model Chlorophyll Comparison

To test the model’s ability to reproduce the observed spatial and temporal patterns of chlorophyll variability, we compare the distance to the 0.2 mg m$^{-3}$ isoline of chlorophyll in the model to the satellite data for the period November 1997 to December 2006 (a more
elaborate comparison between the model and observations is provided in the Supplementary Material). The modeled climatological year (Figure 3a) shows same seasonality as satellite data (Figure 3d). The overall distances in the modeled climatological year are 400 km greater than those estimated from the satellite, but both model and observations have an averaged amplitude of the seasonal cycle of 780 km. In an average year, the size of the high chlorophyll region is largest in winter and spring and shrinks to its minimum in August and September. In both the model and observations, the vicinity of Cape Blanc (latitudes 18 to 23°N) the area with chlorophyll above 0.2 mg m\(^{-3}\) occupies an east-west coastal band that is greatest during the first quarter of the year but it remains wider than 400 km throughout the year. In summer, this elevated band near Cape Blanc is limited to a narrow meridional region in the SeaWifs data, with a gap in the monthly climatology due to the presence of clouds in these months. In contrast, to the south of Cape Blanc, the region of high chlorophyll vanishes completely in summer. The correlation coefficient for the chlorophyll distance monthly climatology varies between 0.94 at 13.7°N and 0.73 at 22.7°N, the meridional average correlation being R=0.86.

Although the chlorophyll seasonal cycle in this area is very important [Lathuilière et al., 2008], especially in the coastal band, interannual changes represent a significant contribution to the total variability in the eastern subtropical North Atlantic (Figures 1 and 3). In early 1998, the first year of the satellite record (and the corresponding year in the model), there is essentially no area with chlorophyll in excess of 0.2 mg m\(^{-3}\). The biggest positive anomaly of the observational record occurs at the end of 1998 and beginning of 1999: at this time, the 0.2 mg m\(^{-3}\) chlorophyll isoline extends across almost the entire basin, an anomaly of 1000 km over the seasonal mean (Figure 3c). Although the precise
mechanism for this event remains largely unknown, Pradhan et al. [2006] noted that it is correlated with the switch from an El Niño to a La Niña. The model reproduces the strong chlorophyll rise and sea level drop between February 1998 and February 1999, but misses the peak values when the chlorophyll was elevated across the width of the whole basin. We interpret the correspondence in the temporal pattern to suggest that the model simulates many of the dynamical responses to atmospheric variability. The mismatch in the peak values leads us to hypothesize a forcing mechanism not included in the model: a likely candidate is the input of iron-rich aerosol dust, as hypothesized by Pradhan et al. [2006]. As the study region is downstream of the world’s largest dust source [Goudie and Middleton, 2006], it is thought that the delivery of dust can have an important impact on productivity and chlorophyll [Duarte et al., 2006] and dust is also known to influence the algorithms that retrieve chlorophyll from the satellite images [Moulin et al., 2001]. Using aerosol optical depth as a proxy for dust (from SeaWiFS data), we found that 1998 was the dustiest year during the SeaWiFS 1997 - 2007 period (not shown), which might have contributed to the 1998-1999 SeaWiFS satellite anomaly. Our simple analysis of SeaWiFS aerosol optical depth is consistent with a more sophisticated composite time series of tropical North Atlantic aerosols constructed from satellite, in situ, and proxy data by Evan and Mukhopadhyay [2010]. The relationship of aerosols to chlorophyll extends beyond the scope of this work, as our model prescribes wet and dry dust deposition from the monthly climatology of Ginoux et al. [2001] and thus no interannual variability in aerosols exists. The absence of this factor in the model allows us to focus on chlorophyll variability caused by ocean and atmosphere dynamics directly.
Even excluding the very high positive anomaly mentioned above, the width of the high chlorophyll region displays large interannual variability, with the standard deviation of the monthly anomalies equal to 102 km in the satellite data and 144 km in the model output. This interannual variability is about half as large as the seasonal cycle, which has a standard deviation of 260 km for satellite data and 279 km for model data. The simulated distances show the high chlorophyll region being largest from 1999 to the end of 2003 (predominantly positive anomalies, Figure 3f and g) and smallest from 2004 to 2006. The SeaWiFS anomalies are dominated by the 1998/1999 event. The anomalies of the high chlorophyll region averaged from 10 to 24°N in the model and SeaWiFS are correlated beyond the 1% level (R=0.42, Figure 3g). Though significant, this correlation coefficient shows that the model does not simulate a large part of the satellite-observed chlorophyll variability. However, both observed and simulated chlorophyll are linked to sea surface height (Figure 2c and d), suggesting that the model realistically simulates some of the physical drivers influencing biology, as explored in the next section.

4. Sea Surface Height and Chlorophyll

Processes that influence SSH are thought to also impact chlorophyll variability. The link between SSH and chlorophyll has been hypothesized on the basis that SSH anomalies reflect anomalies of the thermocline depth [Stammer, 1997; Mayer et al., 2001] and the thermocline and nutricline are often coincident [Wilson and Coles, 2005; Signorini et al., 1999]. Therefore, SSH anomalies may reflect movement of the nutricline towards or away from the euphotic zone with an attendant influence on productivity. Because nutrient observations are sparse in space and time, we can evaluate the nutricline-SSH relationship on interannual time scales only in the model. In the simulation, there is a strong inverse
relationship between monthly anomalies of SSH and nutricline depth (defined as the depth of the maximum vertical gradient in $NO_3$) averaged over the study region ($R=-0.71$). As hypothesized, when SSH is depressed, the nutricline is closer to the surface in the simulation, and we would expect chlorophyll concentrations to be elevated (Figure 2c and d).

Both satellite observations and model results substantiate the hypothesized link between SSH and chlorophyll (Figures 4 and 5). Years with negative SSH anomalies generally correspond with positive chlorophyll distance anomalies and vice versa, indicating an expansion of the high chlorophyll region in years of anomalously low SSH. The SSH-chlorophyll relationship is harder to discern in satellite observations because the 1998-1999 anomaly dominates the chlorophyll record more strongly than the SSH record; yet, the correlation coefficient between anomalies of the average chlorophyll distance and anomalies of the average SSH in the study region ($R=-0.56$) is significant beyond the 1% level. In the model, the relationship is stronger ($R=-0.75$, Table 1) using monthly anomalies over the 49-year simulation.

The SSH-chlorophyll relationship is weaker between 1999 and 2006 (Figure 5c). During this time, the anomalously low SSH appears in the northern part of the domain (north of 14°N, Figure 5b). In this northern region, the SSH-nutricline correlation is weaker because the nutricline is deeper and thus a divergence of the surface waters does not easily mine nutrient-rich subsurface waters, a result shown in at least one other study [Wilson and Coles, 2005].

The correlation between SSH and chlorophyll distance is understood to be due to a shared physical mechanism: the divergence or convergence of mass at a given locale, which
influences both the temporal evolution of SSH [Griffies and Greatbatch, 2012] and the rate of upwelling. A divergence (convergence) of mass above the nutricline causes upwelling (downwelling) and the vertical heaving of the nutricline towards (away from) the euphotic zone. The wind-driven Ekman currents are a major driver of these upwelling velocities: averaged over the study region, a correlation coefficient of 0.84 between simulated vertical velocities at 60 m and Ekman pumping, calculated from the wind stress reanalysis field via equation (2), confirms a strong wind-driven control on upwelling in the model. In reality, temperature and salinity changes also impact steric SSH, but no straightforward conceptual model exists linking steric sea level anomalies to thermocline/nutricline depth anomalies or chlorophyll. Steric effects are not included in the model diagnostic of sea level, so all variability in model SSH is due to the convergence of the vertically-integrated mass transport. The absence of the steric sea level anomalies in the simulated SSH may remove a source of sea level variability that does not systematically influence chlorophyll and be one cause of the stronger SSH-chlorophyll relationship in the model than the observations.

There is no clear trend in the width of the high chlorophyll region over the 9-year satellite chlorophyll record (Figure 4), and previous work noting an apparent expansion of the lowest-chlorophyll North Atlantic waters concurrent with rising sea surface temperatures [Polovina et al., 2008], was likely influenced by the strong 1998/1999 anomaly at the start of the record. In any case, trends observed in the global satellite record have been shown to be small relative to simulated decadal variability [Henson et al., 2010]. The study region is no exception: the subtle downward trend in the size of the high chlorophyll region from 2002 to the end of the simulation is well within the size of the interannual variability and
cannot be interpreted as a part of a long term trend (Figure 5). The dominant temporal signal appears to be a multidecadal oscillation, which is strongly linked to SSH. We next quantify the terms giving rise to the biological temporal variability before looking for a cause of this low-frequency modulation in section 5.4.

5. Mechanisms of Chlorophyll Variability

The size of the high chlorophyll region, as defined by the distance between the coast and the 0.2 mg m\(^{-3}\) chlorophyll isoline, is correlated to the mean chlorophyll concentration in the study region (R=0.97 for model output, R=0.73 for satellite data, both p\(<\)<0.01). Chlorophyll concentrations depend both on biomass and on the chlorophyll to Carbon ratio (Chl:C) in the biomass. Here, we choose to focus on the biomass variability (i.e. the concentration of carbon rather than chlorophyll) to exclude any variability caused by changes in intracellular Chl:C. Therefore, all of the model analysis that follows is based on carbon-biomass. In practice, biomass and chlorophyll are tightly related because higher growth rates (more biomass) and higher Chl:C (more chlorophyll) vary in concert in response to increases in iron and light.

Growth rates and biomass are modeled as a function of irradiance, nutrient availability and temperature. In our study region, irradiance and temperature are not dominant controls on biomass. Prior to 1983, the model irradiance is prescribed as a climatological annual cycle. From 1983 onwards, the model uses variable incoming radiation, but we find no correlation to biomass over the latter time period. The temperature dependence of growth in the model would mean that an increase in temperature increases growth rates, all else being equal. However, in our study region, the average temperature in the top 80 m is negatively correlated to biomass; this inverse relationship likely results from the
role of upwelling in decreasing temperature at the same time it increases nutrient supply.

Because variability in the nutrient supply is primarily responsible for changes in biomass concentration and the size of the high chlorophyll region, we focus our investigation on the different mechanisms of nutrient supply into this region and their relative importance.

5.1. Nutrient Supply

In order to address interannual variability in nutrient supply, we consider each term in the \( PO_4 \) budget (equation 3) for the study region, above 80 m, the average depth of the euphotic layer. Studying these terms provides a mechanistic view of the physical controls on phytoplankton biomass. Because a number of previous studies have linked stratification variability with chlorophyll and biomass variability [McClain et al., 2004; Gregg et al., 2005; Behrenfeld et al., 2006; Polovina et al., 2008], we also compare our biomass time series with a common measure of stratification, the density difference between 200 m and the surface. However, as density variations at 200 m are small, this measure of stratification reflects only density variability at the surface, predominantly driven by temperature changes. Higher temperatures at the surface are indeed correlated with lower biomass, but this gives little indication of the physical mechanisms responsible for the decrease. The density difference between the surface and the base of the euphotic zone (80 m) may be a more suitable measure of the stratification impacting the sunlit layer. The anomalies of this measure of stratification are weakly and actually positively correlated with anomalies of biomass (\( R=0.18, p<<0.01 \), Table 1) and surface chlorophyll concentrations (\( R=0.13, p<0.01 \), Table 1), opposite to the expected correlation if stratification were exerting a leading control on the nutrient supply. Instead, this slight positive correlation suggests
a distinct mechanism controlling phytoplankton variability that is not suppressed under increasing stratification.

The $PO_4$ budget equation (3) includes lateral advection, vertical advection and mixing. Anomalies of the $PO_4$ mixing term only play a relatively important role in the months of February and March (Figure 6d), when convection and vertical diffusion inject $PO_4$ into the northern part of this region. On an interannual timescale, anomalies of $PO_4$ diffusive mixing averaged over the study region explain less than 4% of the biomass variability ($R=0.2$). Conversely, the advective $PO_4$ flux seems to be a key factor (correlation of 0.86 with biomass, explaining 74% of the variability). Amongst vertical, meridional and zonal $PO_4$ fluxes, we find that interannual variability of biomass is most highly correlated to the vertical advection of $PO_4$ to the euphotic zone ($R=0.82$, Table 1), although correlation to the meridional flux is also high ($R=0.73$). Most of the meridional transport of $PO_4$ into the study region takes place through the southern border (95% on a yearly average). This is consistent with waters south of the study region having higher nutrient concentrations than those to the north [Pastor et al., 2012]. Nevertheless, the vertical fluxes are much larger than the meridional fluxes both in their anomaly (Figure 6b and c) and their monthly mean. The mean peak value over the annual cycle for the vertical upwelling term is $6.34 \times 10^{-10}$ mol m$^{-2}$ s$^{-1}$, and is at least a factor of 2.6 and 10 bigger than the meridional and zonal phosphate fluxes, respectively. The vertical fluxes are thus the most important supplier of nutrients into this region and the dominant control on both phytoplankton biomass variability and the areal extent of the high chlorophyll region. Interannual variability of these vertical $PO_4$ fluxes are caused by variability in vertical
velocities rather than changes in the subsurface nutrient reservoir, as measured by the $PO_4$ concentration at 80 m (Figure 7).

5.2. Coastal Upwelling Versus Offshore Upwelling

Having shown that vertical advection is the leading driver of biomass variability (Figure 6), we next consider two ways in which nutrients are brought to the euphotic layer through vertical advection. One is coastal upwelling, determined in part by the alongshore component of the wind stress. In our study region, the equatorward trade winds generate an offshore Ekman transport of coastal surface waters, leading to nearshore upward velocities that bring subsurface nutrient-rich waters to the surface. The upwelling occurs in a narrow meridional band extending less than 100 km from the shelf break [Barton et al., 1998], with seasonally variable meridional extension. The coastal upwelling further influences the offshore domain where filaments form that transport coastal waters rapidly offshore, as happens when the southward flowing Canary Upwelling Current and poleward Mauritania Current converge [Mittelstaedt, 1991; Pelegrí et al., 2005]. The second vertical nutrient supply mechanism is due to Ekman pumping, driven by wind stress curl that causes divergence of the horizontal flow, and induces vertical velocities at the base of the Ekman layer [Bakun and Nelson, 1991]. The northeast quadrant of the study region, occupied by the subtropical gyre, is characterized by downward curl-driven velocities, whereas the southern part of the domain is characterized by cyclonic wind stress curl and thus upward Ekman suction.

In the study region, the interannual variability of the modeled vertical transports at 60 m in the first cell adjacent to the shelf break is well correlated (R=0.87) to the variability of the Ekman coastal upwelling calculated using equation (1), confirming that the along-
coast wind stress exerts a leading control on variability in coastal upwelling. However, the mean value of upwelling is nearly a factor of four smaller in the model (0.4 Sv) than from the theoretical Ekman upwelling (1.6 Sv). The theoretical upwelling values inferred from the steady, two-dimensional, offshore Ekman transport need not match the model’s vertical velocities, which result from the divergence of the total horizontal velocity field (the sum of geostrophic, wind-induced and other ageostrophic non-steady contributions). A diagnosis of the surface geostrophic velocities from the simulated sea surface height gradients shows convergence of the horizontal geostrophic velocity field (the sum of the along-shore and cross-shore transport) of 0.4 Sv in the first cell along the coast. Such convergence reconciles some of the difference between the simulated vertical velocities along the coast and the Ekman calculation. As a further check that the Ekman transport calculations are reasonable, we also compared the time mean total modeled offshore transport to the sum of the offshore Ekman and geostrophic velocities as in Colas et al. [2008], and found that these agreed to within a tenth of a Sverdrup. The upwelling is also surface intensified in this region, with a mean vertical transport averaged over the first cell adjacent to the shelf break of 0.85 Sv at 20 m. Thus, the mismatch between Ekman and model vertical flux is partly dependent on our chosen depth of analysis. We speculate that the remaining mismatch between modeled and Ekman vertical transport likely arises from convergence of the along-shore transport that is not captured by the geostrophic calculation from the monthly mean sea surface height field, perhaps due to its transient nature as was found by Mason et al. [2012].

Unlike the results for the coastal Ekman upwelling, both the strength of the curl-driven Ekman transport and its variability agree well with the modeled vertical transport av-
eraged at 60 m over the study region. The simulated and theoretical upwelling follow a similar seasonal cycle, peaking in August with values of 3 Sv for model vertical transport and 3.7 Sv for Ekman pumping, and their monthly anomalies are strongly correlated (R=0.84). This vertical transport supplies nutrients to the euphotic zone. In our simulations, the vertical transport of nutrients to the euphotic zone in the offshore domain (i.e. everywhere in the study region offshore of the first cell adjacent to the shelf break) is more than double the offshore transport of nutrients from the first cell adjacent to the shelf break.

Importantly, coastal upwelling in the model might be weak relative to true upwelling in nature, because of both coarseness in the ocean model grid [Marchesiello and Estrade, 2010; Capet et al., 2004] and the atmospheric reanalysis driving it. Thus, although the simulated coastal upwelling of nutrients influencing the offshore domain is about half the curl-driven pumping of nutrients to the euphotic zone, the relative strength of these terms might be sensitive to model resolution. Because of uncertainty in the model’s ability to quantify coastal upwelling, we cannot rule out the possibility that, in nature, the influence of coastal upwelling on the offshore domain might be a more equal partner with the curl-driven Ekman suction in setting the variability in the size of the high chlorophyll region, and we leave this open question to future study with higher resolution models.

5.3. Size of upwelling region versus strength of upwelling velocities

The importance of offshore upwelling to nutrient budgets over our study area is primarily due to the large area over which the wind stress curl is favourable to upwelling. We next turn our attention to the hypothesis of whether chlorophyll variability is controlled by
shifts in the boundary between upwelling and downwelling favourable winds, by changes in the magnitude of the vertical velocities, or a combination of both.

To test this hypothesis, we compare months with anomalously large regions of high chlorophyll (i.e. monthly $D_{CHL}$ more than one standard deviation above the corresponding monthly mean) with those of anomalously small regions (i.e. monthly $D_{CHL}$ more than one standard deviation below the corresponding monthly mean). Figure 8 shows the advective and diffusive terms of the $PO_4$ budget for the mean of the months with anomalously large and small extension of the high chlorophyll region. In both situations, vertical advection of $PO_4$ ($wPO_4$) provides the largest nutrient source. In the high chlorophyll months, as compared to the low chlorophyll months, $wPO_4$ is 2.3 times larger, mixing remains more or less constant and northward transport at $10^\circ$N doubles. Upward $PO_4$ transport and northward transport at $10^\circ$N are linked through the development of the Guinea Dome, a cyclonic circulation in the southern part of the study region [Mazeika, 1967]. As the dome strengthens, upward velocities increase, as does the northward flow at the eastern flank of the dome, located at about $20^\circ$W. The zero wind-stress contour, which sets the boundary between the downwelling-favorable winds of the subtropical gyre and the upwelling-favorable winds of the coastal upwelling and tropical gyre, remains close to coast north of $16^\circ$N; south of this latitude, the zero curl contour shifts to the north in the high chlorophyll years.

While in summer months the boundary between upwelling and downwelling favourable winds is relatively static (Figure 9a), in winter months the contour of zero wind stress curl can shift dramatically: when the extension of the high chlorophyll region is anomalously large during winter months, the area of positive wind stress curl increases by $9 \times 10^5$
km², and this newly-positive area adds a total of 60 mol s⁻¹ of \( PO_4 \) to the photic layer by vertical transport (Figure 9d). On the contrary, when the extension of the high chlorophyll region is anomalously large during summer months, the area of positive wind stress curl increases by only 1.6 x 10⁵ km² and the newly-positive area adds a total of 14 mol s⁻¹ of \( PO_4 \) to the photic layer by vertical transport (Figure 9c). In summer, the shift in the zero wind stress curl line is small but the overall strengthening of the upward velocities is very important. Thus, the size of the high chlorophyll region is influenced both by the strength of the upwelling and the area over which it occurs.

5.4. Nutrient supply in relation to climate variability

Having demonstrated the dependence of the variability in the size of the high chlorophyll region on the strength of the upwelling of nutrients, it is interesting to investigate whether upwelling is related to a known mode of climate variability. El Niño Southern Oscillation (ENSO) has been hypothesized to be a remote driver of chlorophyll variability in the eastern tropical North Atlantic. Pradhan et al. [2006] showed a moderate anticorrelation between the Multivariate el Niño Index (MEI) and the satellite chlorophyll anomaly between 10 and 25° N, and 30°W to the coast over a 6-year time period during which the satellite data were available (\( R = -0.52, p < 0.01 \)). They hypothesized that the correlation was driven by the influence of ENSO on tropical Atlantic meridional wind stress and coastal upwelling. In qualitative agreement, we find a weak, but statistically significant, anticorrelation between the MEI and modeled chlorophyll variability in our study region (\( R = -0.39, p < 0.01 \) for the 12-month running mean of the de-seasonalized chlorophyll distance anomaly). This correlation is computed from the full time series of the modeled chlorophyll and the MEI produced by NOAA’s Earth System Research Laboratory from
reanalysis data of the atmosphere (such as pressure, temperature, winds, cloudiness) and ocean (sea surface temperature) [Wolter, 1987; Wolter and Timlin, 1993]. In addition to explaining very little of the variance in our study region, ENSO variability does not have a strong decadal signal, and therefore cannot explain the multidecadal modulation of simulated chlorophyll distances, with the area of high chlorophyll being large in the 1950s-1960s, shrinking in the 1970s - 1990s, and recovering to the average after 2000.

The dominant statistical mode of tropical Atlantic variability on an interdecadal time scale is the Atlantic Meridional Mode (AMM). The AMM is characterized by the anomalies of the meridional sea surface temperature (SST) gradient across the mean intertropical convergence zone (ITCZ); these SST gradients are intimately linked to the surface winds over the ocean, which flow toward the anomalously warmer hemisphere [Nobre and Shukla, 1996]. Given that the AMM describes a mode in the surface wind field over our study region, it is natural to hypothesize a connection between AMM and the variability in biological productivity, which we have shown is principally wind-driven. Indeed, the positive AMM has been associated with stronger Ekman upwelling in our study region in a coupled model [Doi et al., 2009]. The AMM is calculated by applying Maximum Covariance Analysis (MCA) to the zonal and meridional components of the 10 m wind field over the tropical Atlantic (21°S - 32°N, 74°W-15°E) after removing ENSO-related variability from the wind field. The updated AMM time series, first described in Chiang and Vimont [2004], has been made available at www.esrl.noaa.gov/psd/data/timeseries/monthly/AMM/, where it is calculated with the NCEP/NCAR reanalysis wind field. A map of the wind pattern is available at this website. Because the CORE2 winds driving our model are derived from the NCEP/NCAR reanalysis with the major modification being a time-invariant speed
offset and directional rotation [Large and Yeager, 2009], the modes of variability should be the same for both wind fields. We compare the AMM time series (Figure 10b) to the model vertical velocities at 60 m and to the Ekman pumping from the CORE2 wind stress (Figure 10a). Visually, the time series appear to be at most weakly related at interannual time scales, and this weak relationship is substantiated by a very low correlation coefficient (R=0.1), which is significant at the 1% level only if we ignore autocorrelation in the individual time series. However, the correspondence between the AMM index and the vertical velocities (and wind-stress curl) appears stronger on longer time scales, as the AMM, like the upwelling velocities and chlorophyll distance, is highest in the 1960s, diminished in the 1970s-1990s, and recovers to the mean in 2000. After filtering the AMM and the model vertical velocities with a 60-month boxcar filter, the correlation coefficient rises to 0.56 (which is significant at the 5% level, even assuming only 10 degrees of freedom, given that the filtered time series are autocorrelated for 60 months). Because it is the decadal-scale component of the AMM that is correlated with the upwelling velocity and chlorophyll in our study region, our 48-year simulation is barely long enough to capture one cycle of this low-frequency modulation. It would be intriguing to examine the relationship of this mode of variability with biological productivity using a longer integration of a coupled ocean-atmosphere model.

6. Summary and Conclusions

The high chlorophyll region along the eastern boundary of the subtropical North Atlantic varies strongly on seasonal and interannual time scales. Such variability is highly correlated to the SSH variability averaged over our study region, supporting the hypothe-
sis that SSH variability in this region is caused in large part by divergent surface currents that lead to vertical upwelling and a shallower thermocline and nutricline.

Using an ocean model forced by renanalyzed atmospheric data, we have investigated possible causes of chlorophyll variability in this region and placed the short satellite record in a longer temporal context. Our major findings are threefold. First, the dominant mechanism controlling interannual biomass variability in this eastern boundary region is the advection of nutrients, mainly vertical advection. Because mixing plays a relatively minor role in the nutrient budget, stratification is not a good indicator of chlorophyll concentration or the size of the high chlorophyll region. Relatedly, mixing plays a relatively minor role in the eastward extension of the oligotrophic region, though a frequent assumption is that the strength of vertical mixing drives variability in the size of the oligotrophic gyre [e.g. McClain et al., 2004; Polovina et al., 2008]. Second, variability in the upwelling of nutrients on interannual to interdecadal timescales is governed by changes in vertical velocities rather than variability in the subsurface nutrient reservoir. Third, the upwelling of nutrients brought about by the offshore wind stress curl is critical in determining the size of the high chlorophyll region. There is a tantalizing, yet difficult to substantiate, relationship between this curl driven upwelling and the large scale climate forcing of the AMM.

Given that the source of nutrients to this region is their vertical advective supply and that nutrient-containing particles sink to depths of 1000 m or more [see results of sediment trap studies in the nearby Canary Islands, e.g. Sprengel et al., 2000], the pool of nutrients in the underlying layers would require a long-term input in order to maintain this vertical supply. In these subsurface layers, where the wind-induced vertical transport decreases
and vertical diffusion is small, horizontal subsurface currents must transport nutrient-replete waters towards the upwelling region to close the budget [Pelegri et al., 2006].

The importance of both curl-driven and coastal upwelling has been noted in other eastern boundary regimes. Messié et al. [2009] characterize the curl-driven and coastal upwelling in four of the world’s major eastern boundary systems, but restrict their analysis to a region within 150 km of the coast. Even in this narrow coastal band, curl-driven upwelling may supply as much as 30% of the upward volume transport. Here we have shown that the size of the curl-driven upwelling region can be highly variable, but typically extends much beyond 150 km from the shore in the North Atlantic. Both the area over which the wind stress curl favors upwelling and the strength of the resultant vertical velocities critically influence the size of the high chlorophyll region, using a definition of high chlorophyll ranging from $>0.07$ to $>0.3$ mg m$^{-3}$. In agreement with our analysis, Rykaczewski and Checkley [2008] note that the curl-driven upwelling offshore of the California Current System, though weaker than the coastal upwelling, gives rise to a greater overall upward volume transport because it operates over a much greater spatial scale. Moreover, the average size of the cyclonic wind stress and related upwelling region offshore of California is smaller than the average upwelling region in our study region where it can extend across the entire basin (Figures 8 and 9). The wide extension of the curl-driven upwelling domain is a likely reason that the high chlorophyll region of the eastern subtropical North Atlantic extends further offshore than its counterparts in other basins [Carr, 2002]. Understanding the quantitative impact of the coastal upwelling requires observations and simulations that resolve the mesoscale [Lachkar and Gruber, 2011; Gruber et al., 2011], which we leave for future exploration. Evidence that the tradeoff between the strength of
the coastal and curl-driven upwelling can manifest as variability in the success of different ecosystems [Rykaczewski and Checkley, 2008] provides additional motivation to further explore variability in the strength and spatial extent of both processes.

In conclusion, recent studies have attributed a leading role to stratification in determining the size of the oligotrophic subtropical gyres [e.g. McClain et al., 2004; Polovina et al., 2008], with the view that increased stratification reduces vertical mixing and thus decreases the entrainment of deeper, nutrient-rich waters into the surface layer. While the stratification variability hypothesis has been shown to modulate the poleward extent of the North Pacific oligotrophic gyre [Polovina et al., 2011], in the upwelling region of the eastern subtropical North Atlantic, the correlation between stratification and surface chlorophyll or integrated primary production in the euphotic layer is very weak, because the dominant nutrient supply terms are not due to mixing. Instead, variability in strength of upwelling and the position of the boundary between the upwelling and downwelling domains exert the dominant control on nutrient availability, and stratification variability does not strongly influence the strength of upwelling. Rather, it is the wind that is driving variability in the size of the high chlorophyll domain, and, equivalently, the position of the oligotrophic region’s eastern boundary. Neither the reanalysis wind stress curl nor the simulated productivity that responds sensitively to this wind shows a clear trend over the second half of the 20th century.

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Figure 1. Maps illustrating chlorophyll variability at three timescales. Seasonal variability (a,b) is shown as winter minus summer chlorophyll concentrations. Interannual variability (c,d) is illustrated by February 2000 (February with maximum satellite averaged chlorophyll) minus February 1998 (February with minimum satellite averaged chlorophyll). Interdecadal variability (e) is shown as chlorophyll averaged during the decade 1959 to 1968 minus the average from 1977 to 1986. Left panels correspond to SeaWiFS data, right to model output. Dark grey contour shows the 0.2 mg m$^{-3}$ isoline of chlorophyll for winter, February 2000 and the period 1959 to 1968. Light grey contour shows the same isoline for summer, February 1998 and the period 1977 to 1986.

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Figure 3. Hovmöller plots (latitude vs. time) of the distance between the coast and the 0.2 mg m$^{-3}$ isoline of chlorophyll ($D_{CHL}$). Panels (a, d) show monthly climatology, panels (b, e) monthly means and (c, f) monthly anomalies, for the period of overlapping satellite and model output (November 1997 to December 2006). Panels (a, b, c) show satellite data, while (d, e, f) show model output. Panel (g) is the time series of meridionally averaged (10 to 24$^\circ$N) monthly anomalies, after applying a 3-month low pass filter.
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Figure 5. Modeled monthly anomalies of (a) distance from the coast to the 0.2 mg m$^{-3}$ isoline of chlorophyll and (b) SSH, over the simulated period (1958 to 2006). Panel (c) shows the time series of meridionally averaged (10 to 24°N) anomalies. Note the inverse color axis in (b) and flipped axis for SSH in (c).

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Figure 7. Scatterplots of the anomalies of modeled (a) vertical PO$_4$ advection at 80 m ($w$PO$_4$) and PO$_4$ concentration at 80 m, and (b) vertical PO$_4$ advection at 80 m ($w$PO$_4$) and vertical velocity at 80 m ($w$). Each data point corresponds to a monthly anomaly in each grid cell of the study region.
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Figure 9. Maps showing differences in modeled vertical velocities, $w$ (a,b) and modeled vertical $PO_4$ advection at 80 m, $wPO_4$ (c,d) between months with high $D_{CHL}$ (months that have $D_{CHL}$ one standard deviation above their monthly mean) and months with low $D_{CHL}$ (months that have $D_{CHL}$ one standard deviation below their monthly mean). Maps (a, c) show the differences during summer months (July to September), maps (b, d) show the differences during winter months (January to March). Dark grey contour marks the zero wind-stress curl isoline for the high months, light grey for the low months in all panels. The study region is enclosed by a black rectangle.

Figure 10. Time series of (a) anomalies of curl-driven Ekman upwelling ($V_{curl}$, Sv) and model vertical transport at 60 m ($V_{model}$, Sv), and (b) Atlantic Meridional Mode (AMM) index from Chiang and Vimont (2004).
Table 1. Correlation coefficients between anomalies of modeled chlorophyll concentration (CHL), distance to the 0.2 mg m\(^{-3}\) chlorophyll isoline (\(D_{CHL}\)), sea surface height (SSH), biomass, and vertical, zonal and meridional \(PO_4\) fluxes into the study region (\(wPO_4, uPO_4, vPO_4\), respectively), mixing, density difference between 80 m and the surface (\(\Delta \rho\)) and vertical flux of water due to Ekman pumping (\(V_{curl}\)) and Ekman transport (\(V_{coast}\)), calculated analytically. All correlations are significant at the 1% level.

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<th>(V_{curl})</th>
<th>(V_{coast})</th>
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Supplementary information and figures for "Physical drivers of interannual chlorophyll variability in the eastern subtropical North Atlantic"

M. V. Pastor, J. B. Palter, J. L. Pelegrí, and J. P. Dunne
1. Model Validation

This section offers a detailed assessment of the model’s ability to reproduce observed spatial and temporal patterns of chlorophyll (CHL), sea surface height (SSH), sea surface temperature (SST) and phosphate concentrations at 80 m (PO₄). We use Level 3 SeaWiFS monthly chlorophyll downloaded from http://oceancolor.gsfc.nasa.gov at 9 km resolution. Satellite SSH is produced by Ssalto/Duacs and distributed by Aviso, with support from

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CNES (downloaded from http://www.aviso.oceanobs.com/duacs). We averaged the original weekly data onto monthly. We use an optimally interpolated sea satellite surface temperature (SST) product at one-degree resolution provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from http://www.esrl.noaa.gov/psd. The variables CHL, SSH and SST data are compared with the model for the period November 1997 to December 2006, corresponding to the time period of model/SeaWiFS overlap. Climatological monthly phosphate concentrations from the World Ocean Atlas 2005 [Garcia et al., 2010] are also compared to the simulation-mean monthly phosphate concentrations. All data sets have been regridded into the model grid.

We evaluate the model’s skill in simulating spatial variability by comparing the annual means of the simulated variable to the observed variable. The model’s skill at reproducing temporal variability is evaluated with three metrics: 1) the correlation coefficient (R) between simulated and observed variables averaged over the study region is the statistic most widely used to determine the degree of similarity between two variables; 2) the standard deviation of the modeled variable, normalized to the standard deviation of the observed values ($\sigma_M / \sigma_O$), indicates the degree to which the two variables have the same amplitude of variation; 3) the centered root mean square difference ($\text{RMS}_{\text{centered}}$), or unbiased root mean square, quantifies how far the average error of the simulation is from zero [Taylor, 2001].

Simulated climatological mean chlorophyll concentrations generally compare well to SeaWiFS measurements, although there’s an important underestimation in the first pixel next to coast (Figure S1c). The model simulates an average chlorophyll concentration of 0.12 mg m$^{-3}$ over the study region, while the satellite estimate provides an average of 0.21
mg m$^{-3}$. Over the first pixel next to coast, the climatological model concentration is 0.42 mg m$^{-3}$, while the observed value is 3.32 mg m$^{-3}$. This is a common issue encountered when chlorophyll from biogeochemical models is compared to satellite chlorophyll [see for example Doney et al., 2009; Lachkar and Gruber, 2011]. The difference could be caused by several factors. The model could be underestimating the coastal chlorophyll concentrations due to the coarse spatial resolution. On the other hand, SeaWiFS overestimation of the concentrations is well documented. For example, Gregg and Casey (2004) found that dust plumes spreading from the Saharan Desert represent about 15.3% of positive bias in the Northeast Central Atlantic (10 to 40$^\circ$N). Nevertheless, the modeled and observed climatological location of the 0.2 mg m$^{-3}$ isoline of chlorophyll shows a relative good agreement (Figure S1a and b).

The strong correlation between simulated and satellite-based monthly mean chlorophyll concentrations averaged over the study region (R=0.76, Table S1) indicates the model’s skill at reproducing temporal variability in the chlorophyll concentrations, when the seasonal cycle is included. The standard deviation of simulated chlorophyll is higher than in the satellite values. Correlations for the monthly anomalies are lower (R=0.46). However, their variance shows a good agreement.

As described in Section 2 of the main article, modeled SSH is zero for an ocean at rest and SSH at a given time step is given as the departure from this level. In our Boussinesq simulation, the SSH evolution is due only to non-steric effects. In contrast, remotely sensed Aviso SSH is obtained by adding the sea level anomaly to a mean dynamic topography calculated using a combination of altimetric and in-situ data and a modeled geoid. There is no condition that the global observed SSH average to zero, and its evolution includes
Steric effects. Therefore, the mean simulated and observed values of SSH in our study area are not comparable and have been omitted from Table S1. Figure 2 of main article shows a map of the climatological SSH average from November 1997 to December 2006 for model and satellite. Even without including steric impacts on simulated SSH, the correlation between observed and modeled monthly SSH means averages 0.53 in the study region (Figure S2).

Simulated sea surface temperature shows excellent agreement with observed values (Figure S3). The model has a slight cold bias, as modeled and observed climatological SST averaged over the study region are 24.3°C and 25°C, respectively. Correlation coefficients for both monthly mean SST averaged over the study region and monthly anomalies are high (0.99 and 0.88, respectively), and the variance of the model SST is almost identical to that of observations (Table S1).

No observations of interannual variability of PO$_4$ are available; therefore we compare the climatological modeled PO$_4$ concentrations at 80 m with the corresponding observed climatological values. PO$_4$ concentrations compare generally well, with both modeled and observed climatological mean concentrations averaged over the study region being 0.5 mmol m$^{-3}$ (Figure S4a and b). The latitudinal variation of zonally integrated concentrations is remarkably good (Figure S4d). North of 18°N, the model overestimates the PO$_4$ concentrations by 1.8 mmol m$^{-3}$, and south of 18°N the model underestimates the observations by 1.5 mmol m$^{-3}$ (Figure S4d). The spatial correlation between the two climatologies is 0.95 (Table S1).

In order to further assess the level of interannual variability captured by the model, we also compare 12-month smoothed anomalies in the modeled
SST with the Extended Reconstructed Sea Surface Temperature (ERSST.v3b, http://www.ncdc.noaa.gov oa/climate/research/sst/sst.php), the Kaplan Extended SST V2 (http://www.esrl.noaa.gov/psd/data/gridded/data.kaplan_sst.html) and the Hadley Centre SST data set (HadSST2, http://www.metoffice.gov.uk/hadobs/hadsst2/) over the entire model period for the region of interest (40°W to coast, 10 to 24°N), and the zonal SST gradient across this region (Figure S5). Comparison of the SST anomalies demonstrates that the amount of interannual variability from the early part of the record during which there is relatively sparse data assimilated in the atmospheric reanalysis is similar in scope to the variability from the later part of the record for which there is an ever increasing set of observational constraints. The model’s ability to capture features in the observations does not show any clear evidence of a secular trend. The mismatches between model and observations are highlighted when focusing in on the SST gradient which is nearly always underestimated by the coarse resolution model in not representing the full scope of coastal cold-water-upwelling (Figure S5b). Nevertheless, this mode of analysis too demonstrates the general pattern correspondence between modeled and observational inter-annual variability.

References


Figure S 1. Mean surface chlorophyll concentrations (mg m$^{-3}$) from (a) model and (b) SeaWiFS satellite for the period November 1997 to December 2006. The residual (c) shows model minus satellite values. The black line is the 0.2 mg m$^{-3}$ chlorophyll isoline.

Figure S 2. Map of correlation coefficients between observed and modeled monthly SSH values for the period November 1997 to December 2006.
Figure S 3. Mean sea surface temperature (°C) from (a) model and (b) NOAA Optimum Interpolation V2 satellite data for the period November 1997 to December 2006. The residual (c) shows model minus satellite values.

Figure S 4. Mean phosphate concentrations at 80 m (mmol m$^{-3}$) from (a) model and (b) World Ocean Atlas 2005 for the period November 1997 to December 2006. The residual (c) shows model minus observed values. Panel (d) shows the meridional distribution of phosphate concentrations integrated zonally across the study region.
Figure S 5. Time series of SST averaged over the study region from model output (red line), the Extended Reconstructed Sea Surface Temperature (ERSST.v3b, solid line), the Kaplan Extended SST V2 (dotted line) and the Hadley Centre SST data set (HadSST2, dashed line). Panel (a) shows the anomalies averaged over the study region. Panel (b) shows the zonal SST gradient across the region (Kaplan is not shown as it is only available as an anomaly). All time series have been smoothed with a 12-month filter.
Table S1. Model-observation skill metrics for monthly values and monthly anomalies (noted as a primed quantity) of chlorophyll (CHL, mg m\(^{-3}\)), sea surface height (SSH, cm), and sea surface temperature (SST, °C), and skill metrics of spatial patterns for phosphate concentration at 80 m (PO\(_4\), mmol m\(^{-3}\)). The metrics calculated are the correlation coefficient (R), centered root mean square difference (RMS\(_{\text{centered}}\)), observation (O) and model (M) mean and standard deviations (σ).

<table>
<thead>
<tr>
<th></th>
<th>R</th>
<th>RMS(_{\text{centered}})</th>
<th>mean(_M)</th>
<th>σ(_M)</th>
<th>mean(_O)</th>
<th>σ(_O)</th>
<th>σ(_M)/σ(_O)</th>
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<td>CHL</td>
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<td>0.36</td>
<td>0.21</td>
<td>0.21</td>
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<td>-</td>
<td>0.17</td>
<td>-</td>
<td>-</td>
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<tr>
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<td>-</td>
<td>-</td>
<td>-</td>
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<tr>
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<td>-</td>
<td>-</td>
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<td>-</td>
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<tr>
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<td>25</td>
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<td>1</td>
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<td>-</td>
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<td>0.49</td>
<td>0.54</td>
<td>0.47</td>
<td>0.95</td>
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</table>

*Statistics of spatial patterns