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# Increasing coccolithophore abundance in the subtropical North Atlantic from 1990 to 2014

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## Abstract

As environmental conditions evolve with rapidly increasing atmospheric CO<sub>2</sub>, biological communities will change as species reorient their distributions, adapt, or alter their abundance. In the surface ocean, dissolved inorganic carbon (DIC) has been increasing over the past several decades as anthropogenic CO<sub>2</sub> dissolves into seawater, causing acidification (decreases in pH and carbonate ion concentration). Calcifying phytoplankton, such as coccolithophores, are thought to be especially vulnerable to ocean acidification. How coccolithophores will respond to increasing carbon input has been a subject of much speculation and inspired numerous laboratory and mesocosm experiments, but how they are currently responding in situ is less well documented. In this study, we use coccolithophore pigment data collected at the Bermuda Atlantic Time-series Study (BATS) site together with satellite estimates (1998–2014) of surface chlorophyll and particulate inorganic carbon (PIC) to show that coccolithophore populations in the North Atlantic Subtropical Gyre have been increasing significantly over the past two decades. Over 1991–2012, we observe a 37 % increase in euphotic zone-integrated coccolithophore abundance at BATS. We further demonstrate that variability in coccolithophore abundance here is positively correlated with variability in DIC (and especially the bicarbonate ion) in the upper 30 m of the water column. Previous studies have suggested that coccolithophore photosynthesis may benefit from increasing CO<sub>2</sub>, but calcification may eventually be hindered by low pH<sub>T</sub> (< 7.7). Given that DIC has been increasing at BATS by ~ 1.4 μmol kg<sup>-1</sup> yr<sup>-1</sup> over 1991 to 2012, we speculate that coccolithophore photosynthesis and perhaps calcification may have increased in response to anthropogenic CO<sub>2</sub> input.

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# 1 Introduction

Coccolithophores are the most abundant type of calcifying phytoplankton in the ocean. Belonging to the phytoplankton group known as haptophytes, coccolithophores comprise a substantial fraction of the primary producers in many diverse marine environments from cold, sub-polar waters to warm, tropical waters (Thierstein and Young, 2004). Coccolithophores produce calcium carbonate ( $\text{CaCO}_3$ ) shells that sink to the deep ocean forming chalk deposits and thus are important for global biogeochemical cycling of carbon and climate feedbacks. Further, coccolithophores comprise the base of many marine food webs and are widespread throughout the world ocean (Boyd et al., 2015). Changes in coccolithophore abundance could therefore have far-reaching effects from the ecosystem level to global carbon cycling (Boyd et al., 2015).

Ocean acidification from the gradual oceanic absorption of anthropogenic  $\text{CO}_2$  has been projected to impact future coccolithophore populations, possibly disrupting the formation and/or dissolution of their  $\text{CaCO}_3$  shells (e.g., Doney et al., 2009). As  $\text{CO}_2$  is absorbed from the atmosphere, it reacts with water releasing hydrogen ions ( $\text{H}^+$ ) and increasing dissolved inorganic carbon (DIC). Excess  $\text{H}^+$  ions and more DIC in the water column lead to a decrease in the carbonate ion ( $\text{CO}_3^{2-}$ ) concentration in the ocean (for an overview of  $\text{CO}_2$ -carbonate chemistry see Bates et al., 2014). Lower  $\text{CO}_3^{2-}$  concentrations decrease the saturation state of  $\text{CaCO}_3$  and could result in lower calcification rates in coccolithophores. Recently, Bach (2015) proposed that a substrate-inhibitor ratio may be a better indicator of potential biocalcification rates. Bicarbonate ions ( $\text{HCO}_3^-$ ) are the substrate for calcification in most calcifying organisms, but high concentrations of  $\text{H}^+$  ions can limit calcification. Therefore, a decreasing  $[\text{HCO}_3^-]/[\text{H}^+]$  ratio may eventually hinder calcification, rather than low  $\text{CO}_3^{2-}$  concentrations. A  $\text{CO}_3^{2-}$  saturation state of less than one would still, however, cause the dissolution of  $\text{CaCO}_3$  shells. Speculation of how coccolithophores will respond to increasing DIC and acidification (i.e., the balance between  $\text{CaCO}_3$  production vs. dissolution) has been the subject of many laboratory and mesocosm studies (e.g., Iglesias-Rodriguez et al., 2008; Schlüter et al.,

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2014; Riebesell et al., 2007). These, however, have yielded mixed results, highlighting the complexity of biological responses to these changing oceanic conditions.

Numerous laboratory studies indicate that acidification of oceanic waters leads to a decrease in calcification rates for coccolithophores (Riebesell et al., 2000; Scian-  
dra et al., 2003; Delille et al., 2005). However, some recent studies show that coccolithophores have no response or even increase calcification in response to increasing CO<sub>2</sub> (e.g., Iglesias-Rodriguez et al., 2008; Schlüter et al., 2014). Indeed, responses to elevated CO<sub>2</sub> by different species of coccolithophores vary in all directions, making the extrapolation of these laboratory results to natural populations challenging. How calcifying phytoplankton will react to continually increasing CO<sub>2</sub> may vary from region to region in the world's oceans, depending on phytoplankton assemblages (e.g., the dominant species or strain of coccolithophore), available nutrients, and temperature (Sett et al., 2014). Coccolithophore responses to ocean acidification may be species specific or even vary within species (different *morphotype* responses, see Beaufort et al., 2011).

Differences in physiological mechanisms could play an important role in determining relative phytoplankton abundances under increasing DIC and acidification. For instance, nutrient uptake rates vary between phytoplankton species depending on the affinity of the transport mechanism for its substrate (e.g., phosphate ion). Laboratory experiments on the widespread coccolithophore *Emiliana huxleyi* have shown this species to have an efficient phosphate uptake system with a low half-saturation constant for phosphate, making it a superior competitor in phosphate-limited oceanic regions (Riegman et al., 2000). These results have been supported by field data from the subtropical Pacific Ocean (Cortés et al., 2001) and the NE Atlantic Ocean (Tyrrell and Taylor, 1996). For inorganic carbon uptake for photosynthesis, however, *E. huxleyi* displays a relatively high half-saturation constant compared to other phytoplankton (Riebesell, 2004), indicating that coccolithophores may benefit from increasing atmospheric CO<sub>2</sub> absorbed into the ocean. However, there are few in situ studies of coc-

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cololithophore responses to increasing anthropogenic carbon (e.g., Freeman and Lovenduski, 2015).

To test the hypothesis that coccolithophores may be responding positively to additional CO<sub>2</sub> inputs, we employed data from the Bermuda Atlantic Time-Series (BATS), a long running oceanic time series in the North Atlantic Subtropical Gyre (Sargasso Sea) located at approximately 31.7° N, 64.2° W (Fig. 1). At this site, twice-monthly and monthly hydrographic and biogeochemical measurements have been made since the late 1980s (Fig. 1; Lomas et al., 2010; Bates et al., 2014). The BATS environment is characterized by Ekman down-welling and convergence, which results in an oligotrophic setting (Sarmiento and Gruber, 2006). While this area displays strong summer stratification (mixed layer depth ~ 35 m), seasonal overturning results in a deep mixing of the water column during winter (mixed layer depth ~ 250 m; Steinberg et al., 2001; Lomas et al., 2013). Though oligotrophic oceanic gyres have relatively low productivity compared to other areas of the oceans, they cover vast areas and thus are important on a global scale (Sarmiento and Gruber, 2006). Furthermore, the strong summer stratification experienced in these regions could be indicative of future trends, as increased stratification of the water column is projected with global warming (Gruber, 2011; Giovannoni and Vergin, 2012; Cabré et al., 2015). Thus, understanding phytoplankton dynamics subject to these environmental conditions is essential for accurately forecasting future ocean biogeochemistry.

Pigment analyses have been used to study the distribution, abundance, and assemblages of natural phytoplankton populations. Using high performance liquid chromatography (HPLC) to identify the presence and concentration of signature pigments, researchers can obtain relative components of chlorophyll *a* (Chl *a*) from phytoplankton (Letelier et al., 1993; Wright and van den Enden, 2000; Van Lenning et al., 2004). Coccolithophores, a haptophyte algae, are identified using signature pigments for haptophytes (mainly 19'-hexanoyloxyfucoxanthin), of which coccolithophores are a main component (Thierstein and Young, 2004), particularly at BATS (Haidar and Thierstein, 2001). While HPLC pigment analyses can provide a site-specific record of phytoplank-

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ton relative abundance, satellite-based records can provide information at larger spatial scales.

Ocean color remote sensing, through the Sea-viewing Wide Field-of-view Sensor (SeaWiFS; 1997–2010) and the Moderate Resolution Imaging Spectroradiometer (MODIS) Aqua (2002–present) platforms, has revolutionized our understanding of the ecological processes of the upper ocean on a variety of spatial and temporal scales. Satellite-estimated Chl *a* concentration has been used as a proxy for phytoplankton abundance and biomass since 1978. While most phytoplankton are not very effective light scatterers relative to their surroundings, coccolithophores produce CaCO<sub>3</sub> shells that are highly reflective. An understanding of coccolithophore-specific water-leaving radiances and the calcite-specific backscattering cross section allows for the concentration of coccolithophore particulate inorganic carbon (PIC) to be estimated via remote sensing (Gordon et al., 2001; Balch et al., 2005; Balch and Utgoff, 2008). Therefore, satellite estimates of PIC provide a suitable proxy for coccolithophore abundance.

In this study, we combine pigment data from BATS along with PIC and Chl *a* measurements from the satellite record to assess recent trends in phytoplankton dynamics in the North Atlantic subtropical gyre, with a focus on coccolithophores. We show that coccolithophore populations in the North Atlantic are increasing in abundance. Correlations suggest that they have been responding positively to increasing inorganic carbon from anthropogenic inputs in the upper mixed layer.

## 2 Methods

### 2.1 Data sources

Pigment measurements were obtained from the BATS website (<http://bats.bios.edu>), resampled at regular monthly intervals using a linear interpolation between measurements, and converted to relative Chl *a* components from different phytoplankton groups as in Letelier et al. (1993). Briefly, each phytoplankton group is associated with signa-

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lation analysis on linearly detrended and deseasonalized (1 year boxcar smoothing) anomalies. Correlations between the main Chl *a* phytoplankton fractions present at BATS (*Prochlorococcus*, *Synechococcus*, haptophytes, and diatoms) were calculated for a variety of oceanic measurements hypothesized to influence phytoplankton abundance: DIC, alkalinity, inorganic nitrogen concentrations, temperature, salinity, and carbonate chemistry variables (see above section on carbonate chemistry). We also explored possible correlations between Chl *a* phytoplankton fractions and the mixed layer depth (MLD). We used two methods for calculating MLD. MLD<sub>sigma</sub> was determined to be where the potential density anomalies ( $\sigma\text{-}\theta$ ) at depth displayed a  $> 0.125 \text{ kg m}^{-3}$  difference from surface waters, while MLD<sub>temp</sub> was calculated as the location of  $0.5 \text{ }^\circ\text{C}$  change in temperature from the surface (Monterey and Levitus, 1997). Mean density and temperature measurements within the top 10 m of the water column were used for “surface” values. We also tested correlations with the monthly mean North Atlantic Oscillation (NAO) index, obtained from NOAA National Weather Service Climate Prediction Center (<http://www.cpc.ncep.noaa.gov/data/teledoc/nao.shtml>).

### 2.2.2 Trends

In order to quantify temporal trends in pigment concentrations, biogeochemical measurements, and satellite data, we calculated the slope of a straight line that best fit the time-series in a least-squares sense. When comparing trends in pigments with trends in biogeochemical measurements at BATS, we used the average values over the top 30 m of the water column. Satellite Chl *a* and PIC trend analysis was performed on a grid cell basis.

### 3 Results

#### 3.1 Chl $a_{\text{hapto}}$ at BATS

Haptophytes (i.e., coccolithophores; Thierstein and Young, 2004; Haidar and Thierstein, 2001) are present throughout the euphotic zone at BATS and display a pronounced seasonal cycle. Concentrations of Chl  $a$  from haptophytes (Chl  $a_{\text{hapto}}$ ) surpassed  $100 \mu\text{g m}^{-3}$  during periods of high abundance, with  $\sim 35 \mu\text{g Chl } a_{\text{hapto}} \text{ m}^{-3}$  during periods of relatively low abundance, such as between 2000 and 2004 (Fig. 1b). The bulk of haptophyte pigments occurred around  $\sim 80 \text{ m}$  of depth, but pigments were also abundant in the upper 30 m, especially during spring. Haptophyte abundance is low below depths of  $\sim 140 \text{ m}$ . During the last six years of the dataset, Chl  $a_{\text{hapto}}$  was more concentrated, especially in the upper 30 m of the water column (Fig. 1b).

Haptophytes comprise roughly 30% of the Chl  $a$  in the upper 30 m of the water column at BATS (Fig. 2), a percentage that persists from the start of measurements ( $\sim 1990$ ) to the end of our dataset ( $\sim 2012$ ). However, a period of low haptophyte abundance occurred between 2000 and 2004, reducing their relative contribution to Chl  $a$  to 15%. *Synechococcus* Chl  $a$  is variable, ranging from 20 to 70% of the total Chl  $a$  in the upper 30 m. Unfortunately, signature pigment concentrations necessary to calculate *Synechococcus* Chl  $a$  were missing from the BATS dataset at the beginning of our time-series as well as during a five-year segment from  $\sim 1997$  to  $\sim 2002$  (shown as hatched area in Fig. 2). In contrast to generally high *Synechococcus* and haptophyte abundance, *Prochlorococcus* and diatoms contribute relatively small fractions to the total Chl  $a$  in the upper 30 m. Correlations with potentially influential oceanographic drivers can aid to explain variable abundance of different phytoplankton groups in these surface waters.

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## 3.2 Correlations of Chl *a* components with oceanographic measurements at BATS

Correlation coefficients between detrended, deseasonalized anomalies of Chl  $a_{\text{hapto}}$ , Chl *a* from *Synechococcus* (Chl  $a_{\text{syn}}$ ), Chl *a* from *Prochlorococcus* (Chl  $a_{\text{pro}}$ ), and Chl *a* from diatoms (Chl  $a_{\text{diatoms}}$ ) and anomalies in other oceanographic measurements/indices are presented in Fig. 3. Chl  $a_{\text{hapto}}$  shows strong positive correlations with DIC and  $\text{HCO}_3^-$  (explaining nearly 20 % of the variability). Chl  $a_{\text{pro}}$  showed similar, but somewhat weaker, correlations with DIC and  $\text{HCO}_3^-$ . Conversely, *Synechococcus* displayed a strong negative correlation with DIC and  $\text{HCO}_3^-$  (Fig. 3). All phytoplankton groups, except *Synechococcus*, were positively correlated with nitrate ( $\text{NO}_3^-$ ) variability (measurements also include nitrite,  $\text{NO}_2^-$ ;  $\text{NO}_3^- + \text{NO}_2^-$  is referred to hereinafter as  $\text{NO}_3^-$ ).  $\text{CO}_3^{2-}$  concentration, the saturation state of aragonite ( $\Omega_{\text{arag}}$ ), and temperature were negatively correlated with haptophyte abundance, opposite to *Synechococcus*. Indeed, *Synechococcus* and haptophytes display inverse correlations for nearly every variable tested, including temperature (Fig. 3).

Temperature is negatively correlated with Chl  $a_{\text{hapto}}$  (as well as Chl  $a_{\text{pro}}$  and Chl  $a_{\text{diatoms}}$ ) and positively correlated with Chl  $a_{\text{syn}}$  (Fig. 3). Further, when the North Atlantic Oscillation (NAO) index is in a positive phase, temperatures over this region of the North Atlantic are generally warmer (Visbeck et al., 2001). Therefore, in line with the temperature correlations, the NAO index is negatively correlated with Chl  $a_{\text{hapto}}$ , opposite again to Chl  $a_{\text{syn}}$ . Mixed layer depth (MLD), an indicator of both temperature and nutrient availability, shows corresponding correlations with Chl *a* components. When the MLD is deeper, there is more Chl  $a_{\text{hapto}}$  and less Chl  $a_{\text{syn}}$ . Both methods of calculating MLD ( $\text{MLD}_{\text{temp}}$  and  $\text{MLD}_{\text{sigma}}$ ) showed similar correlations (Fig. 3).

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### 3.3 Fluctuations in abundance of different phytoplankton groups

Haptophytes and *Synechococcus* generally show opposing correlations with the variables tested (Fig. 3). This is supported by the opposing dominance of either *Synechococcus* or haptophytes throughout the timeseries (Fig. 4). Indeed Chl  $a_{\text{hapto}}$  and Chl  $a_{\text{syn}}$  show a significant negative correlation ( $p < 0.00001$ ). During periods of low haptophyte abundance (e.g., 2000–2004), *Synechococcus* dominate the water column. Later in the time-series, however, *Synechococcus* populations decline and haptophyte abundance increases. *prochlorococcus* at BATS reside mostly in the deep chlorophyll maximum at  $\sim 100$  m of depth and are a relatively minor component of the Chl  $a$  in the upper 30 m of the euphotic zone. Diatom abundance, usually associated with cold, high nutrient environments, is sporadic and generally low in this oligotrophic oceanic region of the subtropical North Atlantic (Fig. 4).

In general, *Synechococcus* and haptophytes are the most abundant phytoplankton groups at BATS (Figs. 2 and 4). However, the category classified as “other” (includes chrysophytes, dinoflagellates, prasinophytes, among others) in Fig. 3 comprises a large portion ( $> 30\%$ ) of the phytoplankton biomass during certain periods, indicating certain conditions may favor neither *Synechococcus* nor haptophytes. The opposing correlations *Synechococcus* or haptophytes exhibit with DIC,  $\text{HCO}_3^-$ ,  $\text{CO}_3^{2-}$ ,  $\text{NO}_3^-$  and temperature could lead to interesting trends in phytoplankton abundance in an ocean increasingly influenced by anthropogenic climate change.

### 3.4 Trends in Chl $a_{\text{hapto}}$ and total Chl $a$ at BATS

A time-series of Chl  $a_{\text{hapto}}$  at BATS shows that haptophytes (i.e., coccolithophores; Fig. 5a, b), have been increasing significantly since 1990 ( $p < 0.01$  for 30 m integral;  $p < 0.001$  for 140 m integral). Mean concentration of Chl  $a_{\text{hapto}}$  in the upper 30 m of the water column has increased by  $0.848 \mu\text{g m}^{-3} \text{yr}^{-1}$  (standard error = 0.332), corresponding to a 68 % increase over the course of the BATS time-series (1991–2012; an overall increase of  $17.8 \mu\text{g m}^{-3}$ ), while the 140 m integral of Chl  $a_{\text{hapto}}$  has increased

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by  $0.103 \text{ mg m}^{-2} \text{ yr}^{-1}$  (standard error = 0.0307) corresponding to a 37 % increase (an overall increase of  $2.2 \text{ mg m}^{-2}$ ). We assess the sensitivity of these trends to interannual variability by performing trend calculations for a range of start and end years over the time-series. The resulting trend pattern shown in Fig. 7a shows mostly positive trends in haptophyte abundance, except for end years in the 2000–2004 period. Total chlorophyll *a* (Chl  $a_{\text{total}}$ ) also shows a significant positive trend over the time-series ( $p < 0.05$  for 30 m integral;  $p < 0.001$  for 140 m integral; Fig. 6; upper right corner of Fig. 7c). Figure 7c shows trends in Chl  $a_{\text{total}}$  for a range of start and end years, displaying a different pattern than that of the trends in the Chl  $a_{\text{hapto}}$  component of Chl *a*. For instance, for end years in the 2000–2004 period, Chl  $a_{\text{total}}$  shows positive trends (but nonsignificant), whereas Chl  $a_{\text{hapto}}$  shows significant negative trends (Fig. 7). Mean Chl *a* in the upper 30 m primarily exhibits negative trends in the later part of the time-series and most trends in Chl *a* are nonsignificant (Fig. 7c). Unfortunately, missing *Synechococcus* pigment data did not allow for long-term trend analysis of this group of phytoplankton (see hatched area in Fig. 2 and white area in Fig. 4).

In line with the results of our correlation analysis, the trends in  $\text{HCO}_3^-$  for various start and end years show a similar pattern to the trends in Chl  $a_{\text{hapto}}$  (Fig. 7b): mostly positive trends with slightly negative (but nonsignificant) trends in  $\text{HCO}_3^-$  concentration for end year in the 2000–2004 period, a low point in haptophyte abundance. Conversely, the trend pattern for the substrate-inhibitor ratio,  $[\text{HCO}_3^-]/[\text{H}^+]$ , is distinctly different from that of Chl  $a_{\text{hapto}}$ , exhibiting all negative trends (Fig. 7d). Trends in calcifying haptophyte (coccolithophore) abundance can be further corroborated with particulate inorganic carbon (PIC) measurements from the satellite record.

### 3.5 Relating BATS pigment data to satellite products

Significant correlations were detected between Chl  $a_{\text{hapto}}$  (30 m integral) measured at BATS and PIC derived from each satellite. The SeaWiFS-derived PIC correlated somewhat better with Chl  $a_{\text{hapto}}$  than MODIS-derived PIC (SeaWiFS PIC-Chl  $a_{\text{hapto}}$   $p =$

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0.0075,  $r^2 = 0.19$  vs. MODIS PIC-Chl  $a_{\text{hapto}}$   $p = 0.050$ ,  $r^2 = 0.12$ ). This difference is likely inherent in the different algorithms used to estimate PIC from each satellite (see following paragraph). Nevertheless, these correlations demonstrate correspondence between Chl  $a_{\text{hapto}}$  measurements and satellite PIC, both of which denote relative coccolithophore abundance.

Two radiance-based PIC algorithms can be used to relate water-leaving radiance to calcite absorption and scattering properties: a two-band algorithm (Balch et al., 2005) and a three-band algorithm (Gordon et al., 2001). The North Atlantic subtropical gyre exhibits relatively low PIC concentrations year-round. During the SeaWiFS/MODIS overlap period (2003–2007), PIC estimated from the two satellites revealed stark differences (Fig. 5c), possibly explained by the differences in algorithm performance in this region (i.e., sensitivity to low/background PIC concentrations). The low correspondence between the two estimates of PIC prevented the generation of a single, merged PIC time-series. We therefore report trends in PIC separately over the respective satellite eras (Fig. 8).

Chl  $a$  measured at BATS (30 m integral) and Chl  $a$  derived from satellite were significantly correlated ( $p < 0.01$ ,  $r^2 = 0.16$ ). In this case, Chl  $a$  measured by satellite displayed good correspondence between the two satellite eras and could be merged into one time-series. Following the regression technique of Brown and Arrigo (2012), we generated one continuous record of Chl  $a$  from 1998 to 2014 by applying linear regression over the 2003–2007 SeaWiFS/MODIS overlap period to predict these variables from 2008 to 2014 (Fig. 6c).

### 3.6 Regional trends in satellite PIC and Chl $a$

Linear trends in PIC derived from satellite observations are positive for most of the North Atlantic subtropical gyre (Fig. 8). Nearly all significant trends ( $p < 0.05$ ) in PIC concentration are positive, especially during the MODIS era (Fig. 8c, d; 1998–2007 for SeaWiFS and 2003–2012 for MODIS). However, unlike at the BATS site, Chl  $a$  does

not appear to be increasing in most of the gyre (Fig. 9). There are slight positive trends in Chl *a* around the BATS region (Fig. 9a), but these are not statistically significant (Fig. 9b). Indeed, most of the North Atlantic subtropical gyre shows a slight negative trend in Chl *a* or no trend at all. A trend of a subset of the satellite Chl *a* from 1998 to 2012 shows a slight, but nonsignificant, upward trend in Chl *a* in the BATS region (Fig. 1;  $0.0009 \text{ mg m}^{-3} \text{ yr}^{-1}$ ,  $p > 0.05$ ), just as the corresponding gridbox for Chl *a* at BATS in Fig. 7.

## 4 Discussion

### 4.1 Phytoplankton dynamics at BATS

In this study, we observed that coccolithophore populations, based on pigment data for haptophytes, are increasing at BATS and are positively correlated with DIC and  $\text{HCO}_3^-$  (Figs. 1b, 3, 7a). We observed opposite correlations for DIC and  $\text{HCO}_3^-$  with *Synechococcus*, the other major member of the phytoplankton community at BATS. Some studies have suggested that photosynthesis and growth of the coccolithophore, *E. huxleyi*, is carbon limited and could possibly benefit from increasing  $\text{CO}_2$  (Riebesell, 2004; Bach et al., 2013). Since  $\text{CO}_{2(\text{aq})}$  and  $\text{HCO}_3^-$  concentrations increase with increasing DIC/ocean acidification, both photosynthesis and calcification could be stimulated in coccolithophores, which primarily use  $\text{CO}_2$  for photosynthesis and  $\text{HCO}_3^-$  for calcification (Bach et al., 2013). The results presented in this study support the hypothesis that coccolithophores are responding positively to increasing carbon availability, perhaps increasing their competitive ability in oligotrophic settings such as BATS. However, a threshold  $\text{H}^+$  ion concentration could be reached with further ocean acidification, eventually constraining coccolithophore growth.

*Synechococcus*, on the other hand, was negatively correlated with increasing carbon in the upper mixed layer and positively correlated with temperature. In laboratory experiments, *Synechococcus* showed only a slight, non-significant increase in growth

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rate under elevated CO<sub>2</sub> conditions, but increased growth 2.3 fold with increasing CO<sub>2</sub> and temperature (Fu et al., 2007). Sea surface temperature in the upper 30 m at BATS has not increased significantly over the time period of this study (0.04 °C yr<sup>-1</sup> trend,  $p = 0.22$ ). However, positive temperature anomalies were recorded during the

5 2000–2004 period, a period of increased *Synechococcus* abundance (and low coccolithophore abundance). Conditions that could favor *Synechococcus* may eventually arise with further warming, increasing the competitive ability of *Synechococcus*.

In order to examine redundancy in our correlations, we performed multiple linear regressions between Chl  $a_{\text{hapto}}$  and several of the driver variables with which it showed

10 the strongest correlations (not shown). When DIC and HCO<sub>3</sub><sup>-</sup> were regressed together with Chl  $a_{\text{hapto}}$ , all the statistical power of DIC was removed, indicating that HCO<sub>3</sub><sup>-</sup> is the primary driver of the two for Chl  $a_{\text{hapto}}$  variability. Further, when temperature and NO<sub>3</sub><sup>-</sup> were regressed together with Chl  $a_{\text{hapto}}$ , the statistical effect of temperature was removed. This indicates that temperature is not a controlling factor for coccolithophore variability, but rather is a proxy for nutrient concentration in relation to coccolithophore growth. A multiple linear regression of Chl  $a_{\text{hapto}}$  with both HCO<sub>3</sub><sup>-</sup> and NO<sub>3</sub><sup>-</sup>

15 explained > 50% of the variance in Chl  $a_{\text{hapto}}$  ( $r^2 = 0.52$ ). Both HCO<sub>3</sub><sup>-</sup> and NO<sub>3</sub><sup>-</sup> have increased significantly over this time period ( $p < 0.001$ ). However, NO<sub>3</sub><sup>-</sup> measurements are highly variable and near zero, making their accuracy questionable, and the trend is only significant if the last two years are included in the time-series (there were particularly high NO<sub>3</sub><sup>-</sup> measurements during 2011 and 2012). This is in contrast to HCO<sub>3</sub><sup>-</sup>, which shows largely positive trends over this time period matching quite well with those of Chl  $a_{\text{hapto}}$  (Fig. 7).

The positive trend in HCO<sub>3</sub><sup>-</sup> concentration in the upper mixed layer of the water column at BATS is most likely due to increasing absorption of anthropogenic CO<sub>2</sub> from the atmosphere. The upper 30 m of the water column is particularly inundated with anthropogenic CO<sub>2</sub> in the North Atlantic (Sabine et al., 2004; Bates et al., 2012). From

20 1991 to 2012, DIC concentration in the upper 30 m at BATS increased by a rate of

1.4  $\mu\text{mol kg}^{-1} \text{ yr}^{-1}$ , which is roughly the expected rate of increase given the rise in atmospheric  $\text{CO}_2$  (see Chapter 10 in Sarmiento and Gruber, 2006). Increasing inorganic carbon supply could also be accompanied by warmer sea surface temperatures, increased stratification, and decreased nutrient supply over the next century (Cabré et al., 2015). Enriched coccolithophore growth by this additional carbon, as well as other predicted oceanic changes with global warming, could lead to shifts in phytoplankton community structure at BATS.

Coccolithophores are not the only phytoplankton that may be responding positively to additional inorganic carbon. *Trichodesmium*, the filamentous  $\text{N}_2$ -fixing cyanobacteria, has been shown to increase growth and  $\text{N}_2$  fixation under increasing  $\text{CO}_2$  (Hutchins et al., 2007), yet other drivers such as sea surface temperature, nutrients, and species diversity tend to exert more control on their growth in situ (Snow et al., 2015; Gradoville et al., 2014). *Trichodesmium* has been reported to be a common component of the phytoplankton assemblages in the subtropical North Atlantic (Carpenter et al., 2004; Agawin et al., 2013; Orcutt et al., 2001), but was not specifically resolved in this study. *Trichodesmium* contain a similar suite of pigments as *Synechococcus* (Carpenter et al., 1993; Andersen et al., 1996), and therefore could be included in the  $\text{Chl } a_{\text{syn}}$  fraction of our calculations. This would aid to explain the negative correlation between  $\text{Chl } a_{\text{syn}}$  and  $\text{NO}_3^-$  (Fig. 3), and further explain why  $\text{Chl } a_{\text{syn}}$  is more abundant than  $\text{Chl } a_{\text{hapto}}$  in the upper water column during warmer, more stratified periods.

Whatever the exact components of  $\text{Chl } a_{\text{syn}}$ ,  $\text{Chl } a_{\text{hapto}}$  generally shows opposing abundance with this group. We hypothesize that when the *Synechococcus* component is abundant (perhaps due to a positive temperature anomaly), photosynthesis accompanying an increase in *Synechococcus* draws down DIC (thus,  $\text{HCO}_3^-$ ). Low DIC provokes carbon limitation of the coccolithophore population, hindering their competitive ability. If coccolithophores are becoming more competitive at BATS due to a lessening of carbon limitation, then they could continue to exert greater competitive stress on *Synechococcus*, which appear to be competing with coccolithophores for a similar niche. However, if the surface waters continue to warm in this region of the Atlantic,

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as predicted (Cabr  et al., 2015), then *Synechococcus* could regain its competitive edge. Furthermore, declining  $H^+$  ion concentrations could eventually constrain coccolithophore growth (Bach et al., 2013, 2015).

The *Prochlorococcus* group, designated by Chl  $a_{pro}$ , shows similar correlations with oceanic driver variables as for Chl  $a_{hapto}$  (Fig. 3). *Prochlorococcus* have been shown to lack a  $CO_{2(aq)}$  uptake mechanism and therefore rely on  $HCO_3^-$  uptake for photosynthesis (Badger and Price, 2003), possibly explaining similar behavior to coccolithophores (also positively correlated with  $HCO_3^-$ ), which use  $HCO_3^-$  for calcification. *Prochlorococcus* reside mainly in the deep chlorophyll maximum, comprising a rather small portion of the photosynthetic biomass in the upper 30 m at BATS (Figs. 2, 4). However, since  $\sim 2005$ , *Prochlorococcus* have been more common in the upper 30 m, resulting in an overall positive trend in Chl  $a_{pro}$  over the entire time-series ( $p < 0.001$ ).

Consistent with colder, high nutrient environments in which diatoms are normally found, Chl  $a_{diatoms}$  showed a strong positive correlation with  $NO_3^-$  and a negative correlation with temperature (Fig. 3). If predicted trends in sea surface temperature and nutrient supply with further stratification are realized, then diatoms could become a reduced component of the phytoplankton assemblage at BATS (Cabr  et al., 2015). Combining fine scale phytoplankton dynamics from BATS with the satellite record can help to elucidate what changes are occurring over large spatial scales.

## 4.2 Trends Chl $a$ and Chl $a_{hapto}$ in the subtropical gyre

Unlike at BATS, Chl  $a$  in the North Atlantic subtropical gyre is not increasing (Fig. 9). PIC, on the other hand, shows mainly positive trends over the whole gyre, in agreement with data from BATS. Together with an absence of Chl  $a$  trend, this implies that coccolithophores are increasing in abundance relative to other types of phytoplankton in the subtropical gyre. Accompanying this conclusion, however, are uncertainties associated with the satellite-derived PIC estimates.

Radiance-based algorithms for deriving PIC from satellite reflectance data are formulated to capture the light scattering properties of the numerically dominant coc-

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colithophore, *E. huxleyi* (Gordon et al., 2001; Balch et al., 2005). PIC concentrations in the North Atlantic subtropical gyre are comparatively low, generally  $\sim 2.7 \text{ mg m}^{-3}$ , compared to other coccolithophore bloom regions, which have PIC concentrations between 10 and  $100 \text{ mg m}^{-3}$  (Balch et al., 2005). The low concentrations of PIC observed in the North Atlantic subtropical gyre could be within background error or nearing the sensitivity threshold of the instrument. Errors in satellite-derived PIC can arise from atmospheric correction, inclusion of other suspended minerals (such as silica; “opal contamination”), and/or the influence of chlorophyll or colored dissolved organic matter (see Balch et al., 2005). However these problems can be minimized by binning in space and time, as we have done in this study (using monthly, 9 km data rather than daily, 4 km data). Ratios relating  $\text{Chl } a_{\text{hapto}}$  to PIC can further elucidate confidence in satellite-derived PIC estimates.

Data on the amount of Chl *a* per coccolithophore cell allows the calculation of cell concentration of coccolithophores in the surface waters at BATS. Using a value of  $0.26 \text{ pg Chl } a \text{ per cell}$  (Haxo, 1985), mean coccolithophore abundance in the upper 30 m is  $1.43 \times 10^6 \text{ cells L}^{-1}$  (corresponding to the mean value of  $37.3 \text{ } \mu\text{g Chl } a_{\text{hapto}} \text{ m}^{-3}$  in the upper 30 m). Employing a ratio of PIC to coccolith of 0.26 (Balch et al., 1992) and considering 15 coccoliths per cell (a minimal monolayer of coccoliths covering the cell) under nutrient replete conditions and 100 coccoliths per cell under severe nutrient limitation (Paasche, 1998), we arrive at a PIC concentration range of 0.56 to  $3.73 \text{ mg PIC m}^{-3}$ . This range corresponds well to the average satellite-derived PIC concentration in the BATS region (Fig. 1a) over the study period:  $2.71 \text{ mg PIC m}^{-3}$  for SeaWiFS (standard deviation = 0.50) and  $2.66 \text{ mg PIC m}^{-3}$  (standard deviation = 0.39) for MODIS. The relatively high satellite-derived PIC concentration further suggests that coccolithophores may be experiencing nutrient limitation at BATS (and producing additional coccoliths in response; Paasche, 1998). Therefore, even given the multiple sources of error involved with satellite-derived PIC estimates, we feel the strong predominance of positive trends in PIC, along with the BATS  $\text{Chl } a_{\text{hapto}}$  data, suggests that coccolithophores are proliferating in this region (Fig. 8).

### 4.3 Trends in coccolithophore abundance: present and future

This is not the only study documenting an increase in coccolithophore abundance in the North Atlantic. Antia et al. (2001) and Deuser et al. (1995) observed decreases in the opal:carbonate ratio in the North Atlantic during the 1980s and 1990s. Further, using Continuous Plankton Recorder ship measurements, Rivero-Calle et al. (2015) document an increase in coccolithophore occurrence from  $\sim 2$  to  $> 20\%$  in the North Atlantic from 1965 to 2010, which they attribute to increasing  $\text{CO}_2$  concentrations. The data region in their study extends from  $\sim 40$  to  $\sim 65^\circ \text{N}$  (subpolar gyre), just north of the subtropical gyre region focused on in this study. Thus, these studies combined add robustness to the conclusion that coccolithophores in the North Atlantic are increasing in abundance and are likely stimulated by additional carbon from anthropogenic sources.

Ocean acidification will likely eventually hinder the growth and calcification of coccolithophores, however. Recently, Bach (2015) and Bach et al. (2015) introduced the “substrate–inhibitor” ratio, describing the dependence of calcification on  $\text{HCO}_3^-$  (the substrate) and  $\text{H}^+$  (the inhibitor) concentrations. When this ratio falls below a critical level (i.e., intercellular to extracellular  $\text{H}^+$  concentration ratio too low) then coccolithophore calcification will be hindered, unless they evolve a mechanism for coping with low pH (Bach et al., 2015). Bach et al. (2013) demonstrate that pH starts to have a negative impact below  $\text{pH}_7 7.7$ , whereas the BATS average  $\text{pH}_7$  is  $\sim 8.1$ . Thus, critical pH levels will not likely happen for several thousand years (Bach et al., 2015). Other factors besides carbonate chemistry, such as light availability, temperature, and nutrients, likely influence coccolithophore growth in present day oceans.

### 4.4 Potential implications

Increases in coccolithophore abundance in the North Atlantic could have far-reaching ecological, biogeochemical, and climate effects. A shift in phytoplankton community structure could change trophic dynamics, ultimately resulting in ecosystem shifts (Pörtner et al., 2014). For example, though the evolutionary purpose of coccolithophore

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shells is unclear, some studies speculate they could protect against grazing (see chapter on functions of coccoliths in Winter and Siesser, 1994). Shifts to relatively more coccolithophores in a phytoplankton assemblage could reduce trophic energy available for grazers. Coccolithophore shells also function as a ballast material, sinking faster due to increased weight of the  $\text{CaCO}_3$  shell, and sequestering organic matter in the deep ocean (Sarmiento and Gruber, 2006; for a recent study of export at BATS, see Lomas et al., 2010). Increases in coccolithophore abundance may have a positive impact on export production, thus a negative feedback on increasing atmospheric  $\text{CO}_2$ . In addition to bringing carbon to the deep ocean, coccolithophores produce the marine trace gas dimethyl sulfide (DMS; Keller, 1989), which affects cloud formation and climate. Therefore, overall increases in coccolithophore abundance could increase marine DMS production. Furthermore, DMS production by the coccolithophore, *E. huxleyi*, has been shown to increase with increasing temperature and ambient  $\text{CO}_2$  (Arnold et al., 2013). Thus, changes in coccolithophore abundance could have a multitude of effects on marine ecosystems in the North Atlantic, as well as global carbon cycling and climate. These effects could be further amplified if other ocean basins show similar shifts in phytoplankton composition.

In the subtropical North Atlantic, the upper mixed layer contains particularly high levels of anthropogenic  $\text{CO}_2$  (Sabine et al., 2004; Bates et al., 2012). We speculate that this rise in DIC is contributing to the increases in coccolithophore abundance documented in this study. However, it is not clear if phytoplankton communities in other similar oceanic ecosystems, e.g., the North Pacific subtropical gyre, will show similar changes as atmospheric  $\text{CO}_2$  concentrations continue to increase and inundate the upper mixed layer. The aforementioned ecosystem and carbon cycle effects of coccolithophore increases could become even more prevalent in the world's ocean, or, alternatively, coccolithophore growth could be further modulated by temperature, nutrients, and light. In any case, monitoring the response of natural coccolithophore populations to increasing DIC/ocean acidification is essential for understanding effects of anthropogenic carbon emissions on the world's oceans.

## 5 Conclusions

In this study, we documented an increase in coccolithophore abundance in the subtropical North Atlantic. Coccolithophores appear to be responding positively to additional inorganic carbon in the upper mixed layer of water column. These results complement those of Rivero-Calle et al. (2015), who also document an increase in coccolithophore populations in the North Atlantic, albeit farther north, stimulated by anthropogenic CO<sub>2</sub> emissions. Increasing coccolithophore abundance is contrary to what numerous laboratory studies have predicted, highlighting the importance of in situ observations. Growth of coccolithophores could, however, be eventually inhibited by decreasing pH and/or other environmental effects of climate change.

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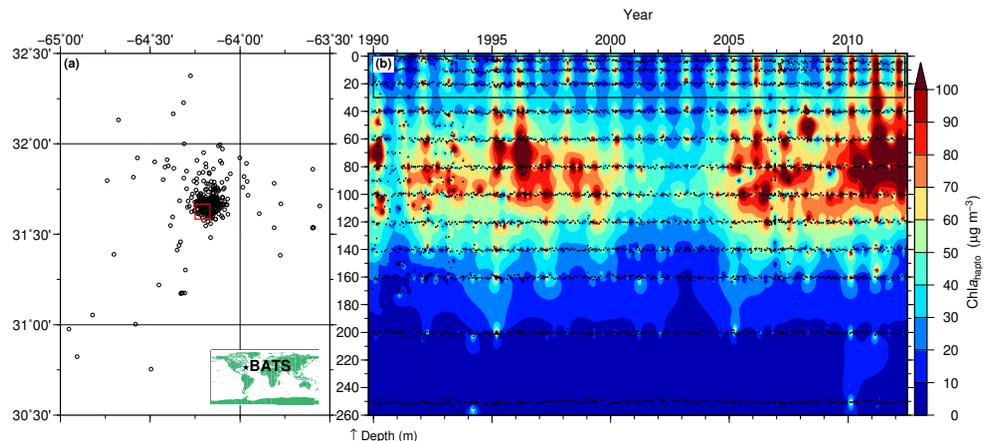
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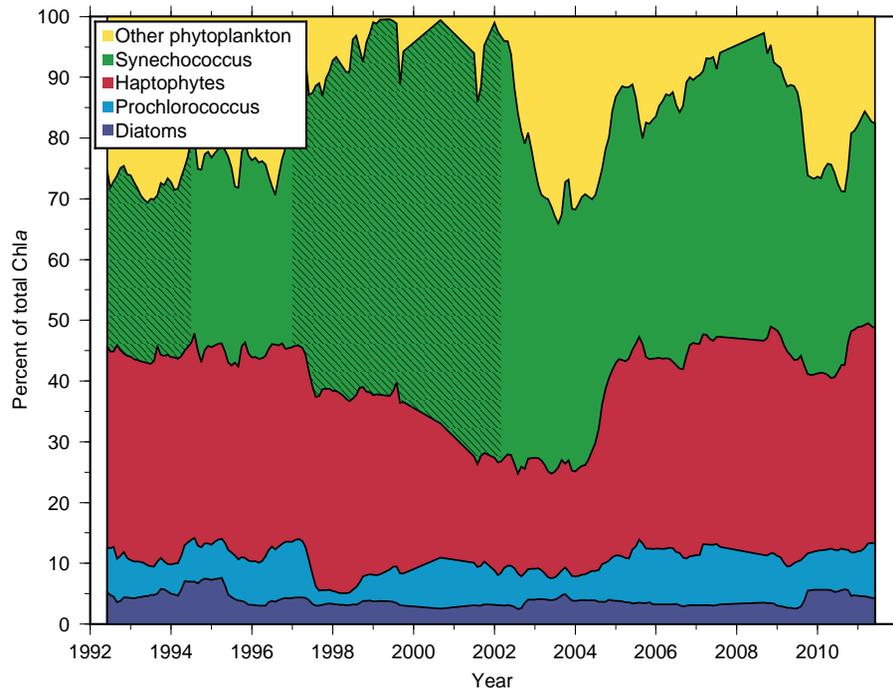
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**Figure 1.** Horizontal **(a)** and vertical **(b)** distribution of pigment measurements taken at BATS (black dots) overlaid on a contour plot of HPLC-measured Chl  $a_{\text{hapto}}$  at BATS. The red box in **(a)** shows the PIC grid cell containing the most BATS measurements (PIC data shown in Fig. 8), while the black box in **(b)** shows the upper 30 m of measurements used for PIC-Chl  $a_{\text{hapto}}$  correlations and correlations presented in Fig. 3.

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**Figure 2.** Percent of Chl *a* from main phytoplankton groups at BATS from 1992 to 2012 in the top 30 m of the water column derived from signature pigment and Chl  $a_{\text{total}}$  concentrations, deseasonalized with a 1 year boxcar filter (purple = diatoms, blue = *Prochlorococcus*, red = haptophytes, green = *Synechococcus*, yellow = other phytoplankton). Hatched area indicates missing pigment data for *Synechococcus*.

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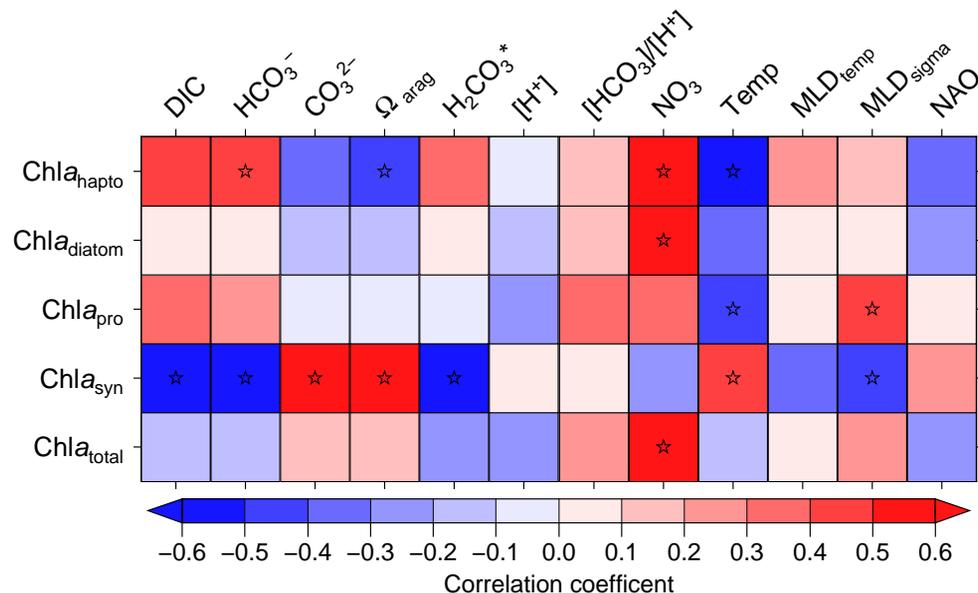
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**Figure 3.** Correlation coefficients between Chl *a* components and various oceanographic measurements made in the upper 30 m of the water column at BATS, NAO index, calculated mixed layer depth (MLD; see Methods) and derived carbonate chemistry parameters. Stars indicate the absolute value of the correlation coefficient is greater than 0.4.

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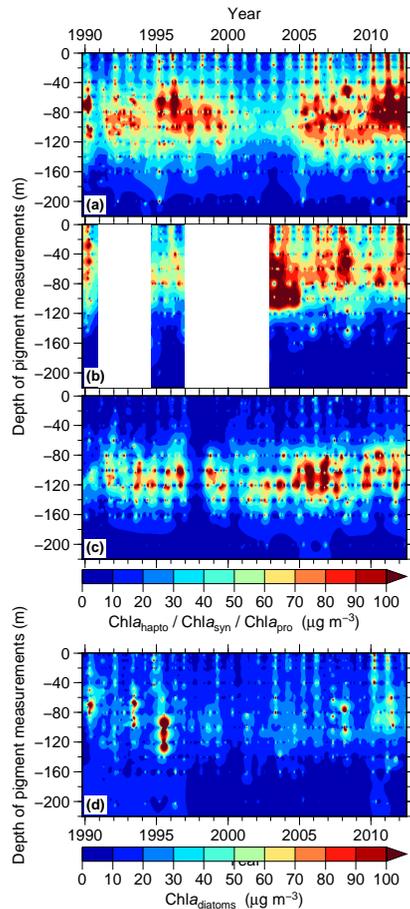
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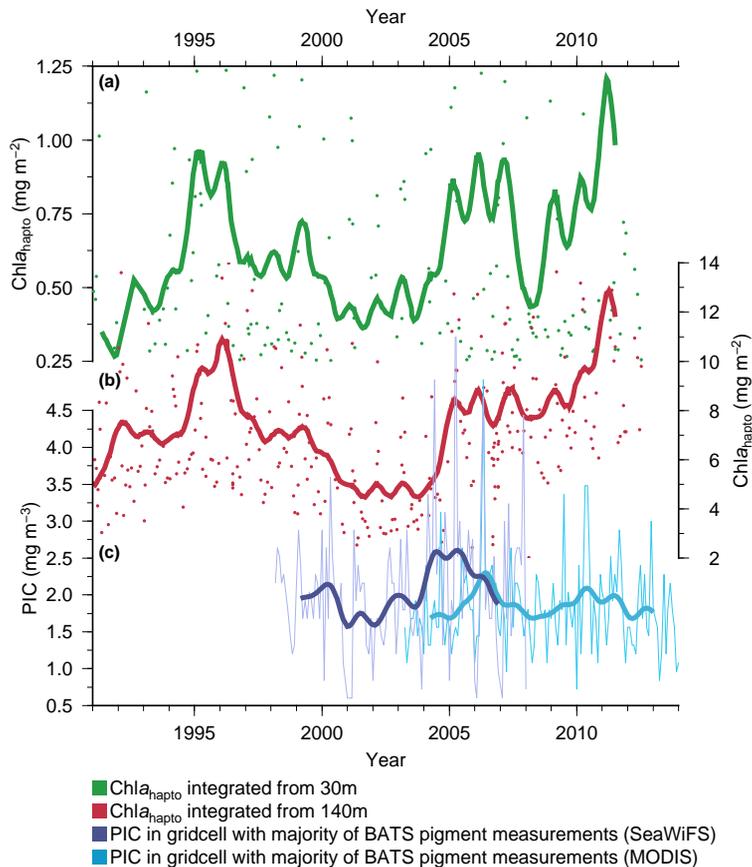
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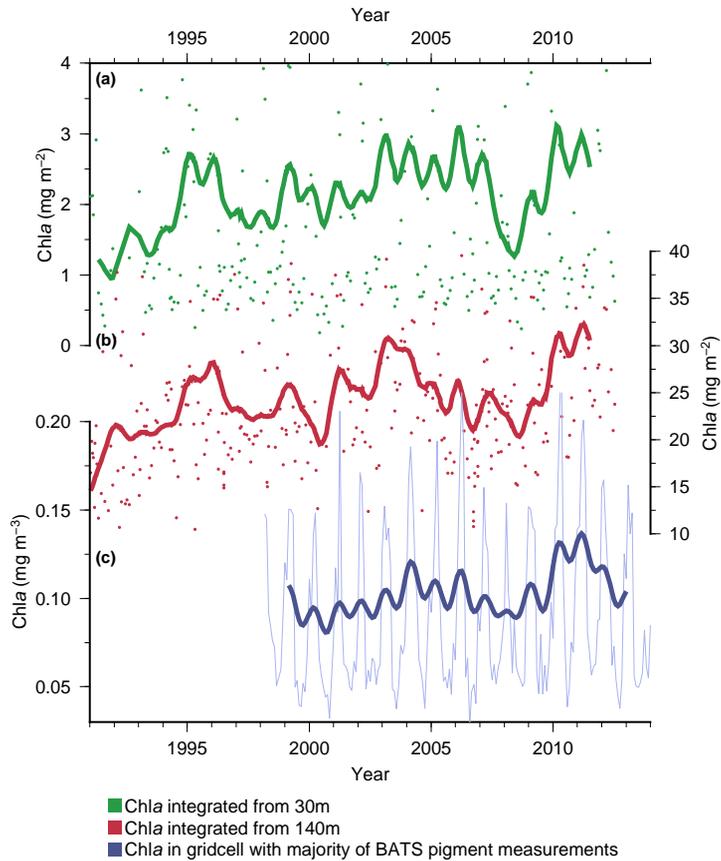




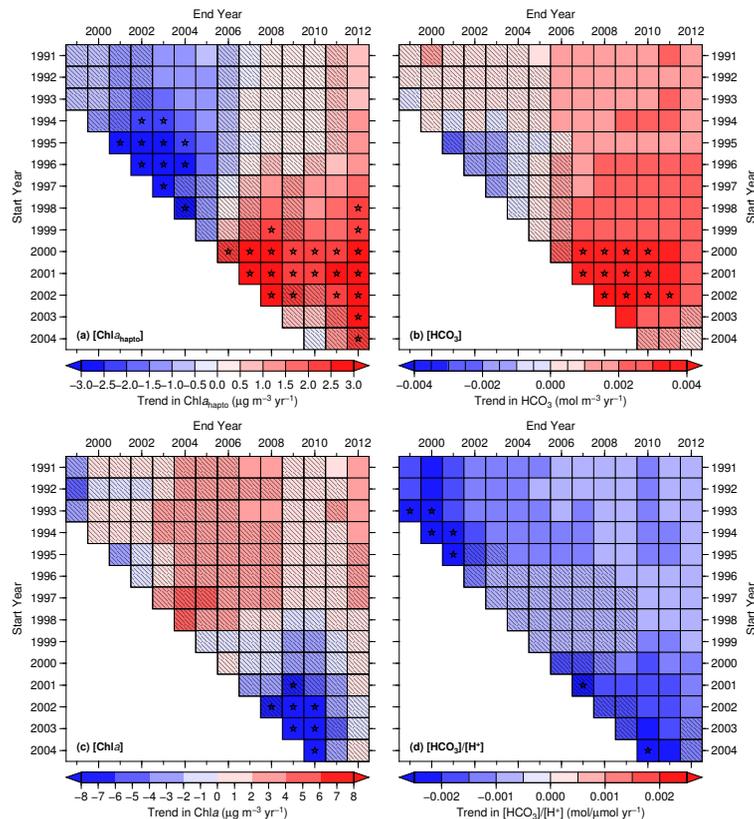
**Figure 4.** Temporal evolution of the vertically-resolved Chl *a* concentration from the main phytoplankton present at BATS derived from signature pigments from 1990 to 2012: **(a)** Chl *a*<sub>hapto</sub>, **(b)** Chl *a*<sub>syn</sub>, **(c)** Chl *a*<sub>pro</sub>, and **(d)** Chl *a*<sub>diatoms</sub>.



**Figure 5.**  $\text{Chl } a_{\text{hapto}}$  measured at BATS and satellite-derived PIC.  $\text{Chl } a_{\text{hapto}}$  integrated from 30 m is shown in (a), while  $\text{Chl } a_{\text{hapto}}$  integrated from 140 m depth is shown in (b). PIC data shown in (c) was obtained from the 5 min satellite grid cell with the most BATS measurements (see Fig. 1). Bolder lines represent a 2 year Gaussian filter on the data.



**Figure 6.** Chl  $a_{\text{total}}$  measured at BATS and satellite-derived Chl *a*. Chl  $a_{\text{total}}$  integrated from 30 m is shown in (a), while Chl  $a_{\text{total}}$  integrated from 140 m depth is shown in (b). Chl *a* data shown in (c) was obtained from the 5 min satellite grid cell with the most BATS measurements (see Fig. 1). Bolder lines represent a 2 year Gaussian filter on the data.



**Figure 7.** Linear trends for a range of start and end years in **(a)**  $\text{Chl } a_{\text{hapto}}$ , **(b)**  $\text{HCO}_3^-$ , **(c)**  $\text{Chl } a$ , and **(d)** ratio of  $\text{HCO}_3^- : \text{H}^+$ . All trends are based on mean concentrations measured at BATS in the upper 30 m of the water column. Boxes with hatch lines demarcate nonsignificant trends. Stars indicate absolute values of trends are greater than  $0.2 \mu\text{g m}^{-3} \text{yr}^{-1}$  for  $\text{Chl } a_{\text{hapto}}$ ,  $0.003 \text{ mol m}^{-3} \text{yr}^{-1}$  for  $\text{HCO}_3^-$ ,  $6 \mu\text{g m}^{-3} \text{yr}^{-1}$  for  $\text{Chl } a$ , and  $0.002 \text{ mol}/\mu\text{mol yr}^{-1}$  for  $[\text{HCO}_3^-]/[\text{H}^+]$ .

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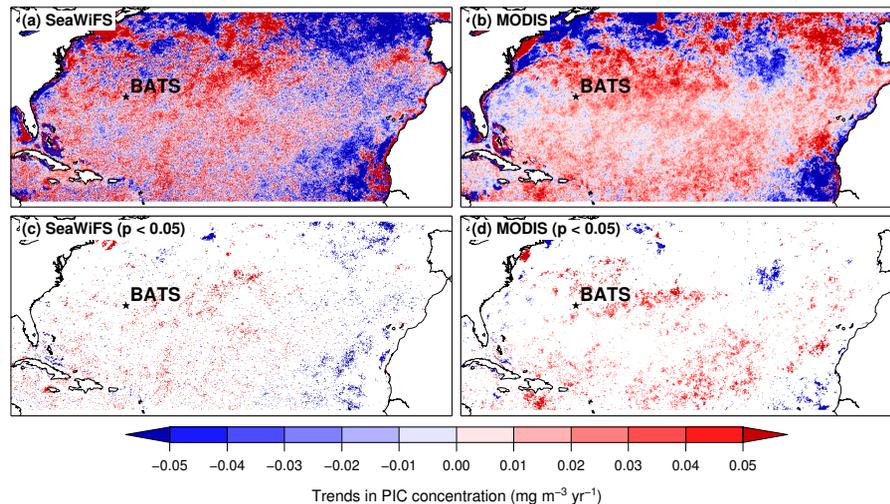
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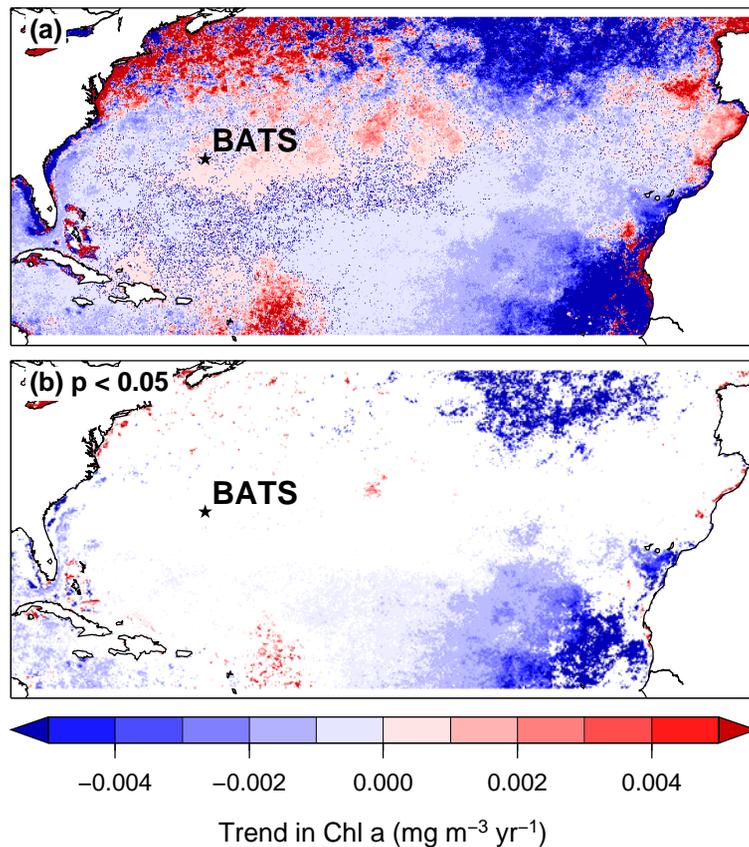
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**Figure 8.** Trends in PIC concentration derived from satellite data for (a) and (c) SeaWiFS (1998–2007), and (b) and (d) MODIS (2003–2014). Bottom panels shows significant trends ( $p < 0.05$ ).

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**Figure 9.** Trends in Chl a concentration derived from merged satellite records: SeaWiFS and MODIS (1998–2014). The bottom map (b) shows significant trends ( $p < 0.05$ ).

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