Interannual variability of seasonal phytoplankton blooms in the north polar Atlantic in response to atmospheric forcing

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[1] The year-to-year variability of timing, intensity, and spatial distribution of surface phytoplankton during spring-summer seasons is examined in the north polar region of the Atlantic using satellite-derived chlorophyll a concentration (Chl) over 7 years (1998– 2004). Each year phytoplankton bloom differed in onset, temporal evolution, and intensity. This interannual variability is to a large degree controlled by local weather, as indicated by the analysis of coincident meteorological data. The atmospheric forcing was parameterized in terms of the generation rate of turbulent kinetic energy (TKE_{RT}) supplied from the atmosphere to the ocean. Timing of the bloom is delayed in years with high TKE_{RT} supplied to the ocean in March. In April, Chl (local and regionally averaged) and TKE_{RT} are inversely related to one another. The late winter and early spring atmospheric conditions influence the seasonal Chl patterns. The seasonal (April-August) Chl correlates well with net heat flux, wind energy, and TKE_{RT} in March and April, but the correlation can be negative or positive in different areas of the north polar Atlantic. The correlation between TKE_{RT} and seasonal Chl is positive in the Greenland Gyre (higher seasonal Chl corresponds to higher TKE_{RT} in March–April) and negative in the regions of East Greenland and North Atlantic/West Spitsbergen Currents. The most likely explanation for the positive correlation between TKE_{RT} and Chl within the Greenland Gyre is that higher TKE_{RT} in spring increases the seasonal supply of nutrients into surface waters.

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1. Introduction

[2] Phytoplankton dynamics play an important role in the ocean carbon cycle [e.g., *Longhurst and Harrison*, 1989]. Understanding this role is of particular interest for the north polar Atlantic where deep-water formation is active [e.g., *Aagaard et al.*, 1985; *Budeus et al.*, 1998] and biological processes are characterized by high amplitudes of seasonal rates of productivity [e.g., *Legendre et al.*, 1993; *Smith*, 1994]. Biological pumping has been traditionally thought to be relatively effective in this area, because intensive phytoplankton blooms may not be matched by concomitant secondary production. This could result in the accumulation of particulate organic matter in the surface ocean and its eventual sedimentation into deep waters.

[3] Current research recognizes that we are not yet able to fully reconcile the role of biology in regulating biogeochemical fluxes of carbon in the polar oceans [e.g., *Noji et al.*, 2000]. Among the factors hampering an understanding of complex biological controls has been the lack of timeresolved (seasonal to interannual) biological data, which would allow quantification of phytoplankton biomass and development of blooms and would explain the underlying

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connections with atmospheric forcing and climate. Only recently, with the operation of Sea-viewing Wide Field-of-view Sensor (SeaWiFS) on OrbView-2 satellite, multiyear time series data on surface chlorophyll *a* concentrations (Chl) with excellent temporal and spatial coverage have become available [*Hooker and McClain*, 2000]. Although such satellite-based Chl estimates are subject to some uncertainty (nominal accuracy goal of 30%), space and time coverage over extended scales makes them an invaluable research tool.

[4] In this paper we use the remote sensing data from SeaWiFS to examine the year-to-year variability in the timing and magnitude of phytoplankton blooms in the north polar Atlantic. Northern north Atlantic phytoplankton blooms are one of the largest manifestations of this phenomenon in the global ocean [e.g., Esaias et al., 1986]. Although the general pattern of seasonal phytoplankton blooms has been known for a long time, the reasons for interannual variability of their timing and intensity, particularly in the polar regions, remain undocumented. Several studies have indicated in the past that synoptic and interannual to decadal variability of phytoplankton concentration in the North Atlantic can be influenced by meteorological forcing [e.g., Dutkiewicz et al., 2001; Follows and Dutkiewicz, 2002; Reid et al., 1998; Stramska and Dickey, 1994; Stramska et al., 1995; Taylor and Stephens, 1980]. However, those studies

did not examine polar regions of the North Atlantic, which are the primary focus of the present paper. The geographical region of interest covers waters of the north polar Atlantic between 70°N and 80°N within the meridional zone between 11°W and 11°E. This region includes the Norwegian Atlantic and West Spitsbergen Currents, Greenland Sea Gyre, and East Greenland Current. Our objective is to examine the relationships between the interannual meteorological variability and phytoplankton dynamics in this region.

2. Basic Relationships

[5] The links between the penetration of light, vertical mixing, and biological productivity have been the central themes in oceanography for many years. The early phytoplankton bloom models have been based on simple relationships [*Riley et al.*, 1949; *Sverdrup*, 1953]. It has been assumed that when nutrients are abundant, phytoplankton production rate, P(z), is linearly dependent on irradiance E,

$$P(\mathbf{z}) = \alpha E,\tag{1}$$

where α is the slope of the light-productivity relationship and *E* is the scalar irradiance of photosynthetically available radiation (PAR). The average PAR available to phytoplankton in a well-mixed water column can be estimated from

$$E = E_{\rm o} e^{-K \,\rm MLD} \left(e^{K \,\rm MLD/2} - e^{K \,\rm MLD/2} \right) / K \,\rm MLD, \qquad (2)$$

where E_o is E just below the water surface, MLD is the mixed layer depth, and K is the diffuse attenuation coefficient for PAR averaged over the euphotic zone. K can be estimated from chlorophyll concentration (Chl) according to the relationship given by *Morel* [1988],

$$K = 0.121 \text{ Chl}^{0.428}.$$
 (3)

The two quantities central in Sverdrup's model are the compensation depth, Z_c , and the critical depth, Z_{CR} . The compensation depth is the water depth where during a 24-hour time period P(z) is equal to all community loss processes, R(z). The community loss processes include autotrophic and heterotrophic respiration, grazing, and vertical export by sinking particles. Alternatively, compensation irradiance E_c is defined as the irradiance at which photosynthesis equals losses R. The critical depth, Z_{CR} , is defined as the depth for which 24-hour vertically integrated water column productivity is equal to water column integrated losses. *Sverdrup* [1953] postulated that a phytoplankton bloom develops when MLD becomes less than the critical depth Z_{CR} .

[6] The assessment of Z_{CR} can be made from a simple relationship,

$$Z_{\rm CR} = E_{\rm o}/K E_{\rm c},\tag{4}$$

if values for E_0 , K, and E_c are available and if Z_{CR} is large compared with 1/K. Unfortunately, the critical depth model appears to have more value as a general concept in pointing

to the importance of the relationship between vertical mixing and light for phytoplankton growth than it has for practical use [see also Smetacek and Passow, 1990]. The major reason for this is our limited knowledge about the community loss rates, which may vary spatially and with season, and may depend on many factors such as variation in respiratory demands in different species and states of physiological adaptation, selective grazing, and phytoplankton seeding strategies. Recent estimates of the compensation irradiance vary by an order of magnitude [see Siegel et al., 2002; Marra, 2004]; therefore it is not practical to apply the critical depth concept to predict phytoplankton blooms. Another shortcoming of the critical depth model is the assumption of active mixing. As pointed out by Sverdrup [1953] himself, if turbulence is moderate or weak, phytoplankton population may increase independently of the relation between MLD and Z_{CR} .

[7] Note that the Sverdrup's critical depth concept refers only to the possibility of phytoplankton growth. In the real ocean, factors not included in his conceptual model may limit such growth. One of the most important factors limiting phytoplankton growth, particularly in later phases of phytoplankton blooms, can be a low supply of nutrients, with nitrogen being considered the dominant element limiting phytoplankton growth in the oceans. The basic theory for the control of primary production by nitrogen and the circulation of this element throughout the euphotic zone ecosystem has been developed by Dugdale [1967] and Dugdale and Goering [1967]. The ideas from the Sverdrup's critical depth model and Dugdale and Goering's nitrogen cycling scheme form the basis for the many recent models used to simulate phytoplankton cycles in the ocean [e.g., Doney et al., 1996; Fasham et al., 1990; Gregg, 2002; Marra and Ho, 1993; Stramska and Dickey, 1994]. These models include less or more complicated approaches which try to account for issues such as diversity of planktonic organisms, nonlinearity of growth and predation processes, and nonpassive plankton behavior. It is, however, very striking that even models with relatively simple theoretical treatment of biological processes are able to reproduce many important aspects of phytoplankton dynamics in the ocean [e.g., Tett and Edwards, 1984].

[8] The major phases of phytoplankton blooms described by these models can be summarized as follows. In early spring, there is a good supply of nutrients and there may be at the top of the water column adequate illumination, but phytoplankton do not remain long enough near the surface to make significant growth. With the progression of seasonal stratification in spring and early summer, there are enough nutrients and light to support phytoplankton growth and initiation of phytoplankton bloom. Zooplankton grazing becomes considerable. In addition, during this time of the year, periodic events of increased mixing due to storm passage result in the removal of significant portion of the biomass out of the surface waters and replacement of nutrients [Marra et al., 1990; Stramska et al., 1995]. Finally, later in the season, there is enough light but not enough nutrients to support net growth, and phytoplankton biomass decreases. In the fall, intermittent mixing due to storms and seasonal decrease in water stratification may restock nutrients and bring about secondary phytoplankton bloom. This phytoplankton bloom scenario briefly described here



Figure 1. Mixed layer depth (MLD) climatology within the study region. MLD was estimated from temperature profiles using 0.2°C temperature difference criterion [*de Boyer Montégut et al.*, 2004]. Description of the data can be found at http://www.lodyc.jussieu.fr/~cdblod/mld.html. See color version of this figure at back of this issue.

demonstrates that the progression of seasonal stratification of oceanic surface waters has a crucial influence on phytoplankton dynamics.

[9] The issue remains, however, whether phytoplankton biomass in the real ocean averaged over large areas depends to the first approximation on the intensity of vertical mixing of water column and whether atmospheric forcing can be linked to interannual variability of blooms. To address this question we use meteorological data to quantify the intensity of vertical mixing during spring and summer in the north polar Atlantic through the bulk mixed layer theories [Kraus and Turner, 1967; Niiler and Kraus, 1977] and test if there exists a significant correlation between the mixing intensity and Chl. According to the bulk mixed layer theories [Kraus and Turner, 1967; Niiler and Kraus, 1977], vertical mixing of water properties in the vertically homogenous oceanic mixed layer is related to the rate of generation of turbulent kinetic energy (TKE_{RT}). The TKE_{RT} can be quantified in terms of wind stirring and buoyancy forcing,

$$\text{TKE}_{\text{RT}} = \int_{h}^{0} \frac{d(\text{TKE})}{dt} = m_1 u_*^3 + m_2 \frac{\alpha g}{\rho c_p} \frac{\text{MLD}}{2} (-H_0), \quad (5)$$

where TKE is the turbulent kinetic energy, MLD is the mixed layer depth, u_* is the wind-induced friction velocity, ρ is the water density, c_p is the specific heat, g is the gravitational acceleration, α is the coefficient of logarithmic

expansion of ρ as a function of water temperature, and H_{ρ} is the net heat flux. The coefficients m_1 and m_2 are difficult to quantify; however, several authors [e.g., Kraus et al., 1988] have assumed $m_1 = 1.25$ and $m_2 = 1$ for negative buoyancy forcing (heat loss from the surface ocean) and $m_2 = 0.2$ for positive buoyancy forcing (surface ocean gains heat). We used these values for the m_1 and m_2 coefficients. Note that equation (5) estimates the TKE_{RT} with the assumption that effects due to internal waves, energy dissipation, and variable vertical distribution of penetrative radiation due to water clarity are small and can be neglected. Similar parameterization of the mixing intensity has been used recently by Follows and Dutkiewicz [2002] to study the meteorological influences on phytoplankton blooms in subtropical and subpolar Atlantic. The first term on the right-hand side of equation (5) $(m_1 u^3)$ indicating the rate of work by the wind will be referred to as TKE_w. The second term on the right-hand side of equation (5) $(m_2 \frac{\alpha g}{\alpha c_n} \frac{MLD}{2})$ $(-H_0)$) representing the rate of potential energy change produced by heat fluxes across the ocean surface will be denoted by TKE_H. In our notation, H_0 is negative, and TKE_{H} is positive in the case of the heat loss by the ocean to the atmosphere. Note that the wind action always increases TKE_{RT} in the oceanic boundary layer. The buoyancy forcing can either increase TKE_{RT} when the water column is cooled from above ($H_0 < 0$), or reduce TKE_{RT} when the water column is stratifying owing to heat input to the surface ocean from the atmosphere $(H_0 > 0)$.

[10] Our approach of investigating the correlation between the atmospheric forcing and Chl is simple, and



Figure 2. Schematic diagram of the Greenland Sea showing (top) the surface circulation and (bottom) water mass structure at a section across the central gyre. (From *Arctic Monitoring and Assessment Programme* [1998]). See color version of this figure at back of this issue.

we recognize its deficiencies. However, the important advantage of such an approach is that we avoid the many assumptions inherent in phytoplankton bloom models. We expect to identify certain time periods when atmospheric forcing is particularly important for phytoplankton dynamics and when correlation between Chl and atmospheric parameters is high. At some other time periods, processes such as zooplankton grazing, phytoplankton mortality, or advection may dominate the variability of phytoplankton biomass in a given location, and the correlation between atmospheric parameters and Chl may be low or insignificant.

3. Data Sources

[11] This study is based on historical data of satellitederived chlorophyll concentration and meteorological and hydrographic data for the north polar Atlantic. The primary data set, ocean surface chlorophyll *a* concentration (Chl, mg m⁻³), covers seven successive years (1998–2004) of data obtained from the SeaWiFS instrument [e.g., *Hooker and McClain*, 2000]. The SeaWiFS mission provides global coverage of water leaving-radiance at eight spectral bands in the visible and near-infrared spectral region approximately every 2 days. Chlorophyll concentrations are esti-

mated from satellite-derived water-leaving radiances using the empirical algorithm OC4v4 [O'Reilly et al., 1998, 2000]. The standard data processing procedures involve atmospheric correction and removal of pixels with land, ice, clouds, or heavy aerosol load prior to calculation of Chl. Our analysis is based on Level 3 standard mapped images (SMI), which are projections of the Global-Area-Coverage data onto a global, equal-angle grid with a nominal 9 km \times 9 km resolution (recent reprocessing version 4). These data were obtained from the Goddard Earth Sciences Data and Information Services Center, DAAC (http://daac.gsfc.nasa. gov/data/data set/SEAWIFS/). The 8-day and monthly composites of Chl were selected for our analysis because of the reduced loss of data due to cloud cover in comparison to the daily composites. Since we are interested in large-scale patterns, the Chl data were binned to $2^{\circ} \times 2^{\circ}$ grid to filter out the smaller scale variability. A potential source of uncertainty in our results is the choice of a global algorithm for Chl rather than regionally based one. There is evidence that Chl in the north polar Atlantic can be better described with regional algorithms [e.g., Stramska et al., 2003], but because regional algorithms are not available for the entire region of interest to this study, we assume that global algorithms are sufficiently accurate to address spatial and interannual variability in Chl.



Figure 3. The 7-year averaged monthly means of turbulent energy generation rate due to heat fluxes across the ocean surface (TKE_H) for the months of March through August.

[12] To evaluate atmospheric forcing, we compared *Chl* estimates with the meteorological data from the NOAA-CIRES Climate Diagnostic Center NCEP/NCAR (National Centers for Environmental Prediction and National Center for Atmospheric Research) Reanalysis Project, which uses state-of-the-art analysis/forecast system to assimilate global meteorological data from various available sources from 1948 to the present. Data relevant to our study region have been extracted and reformatted from netCDF to plain ASCII format by Climatic Research Unit (CRU), University of East Anglia (http://www.cru.uea.ac.uk/cru/data/ncep/). In particular, we used the net latent and sensible heat flux, along with the net longwave and shortwave radiation estimates, to calculate the net heat flux H_0 at the sea surface. We also used the wind speed and wind stress data and estimated wind friction velocity [Liu et al., 1979].

[13] Another environmental parameter used in our study is the oceanic mixed layer depth (MLD). The most recent available climatological MLD data were obtained from the Laboratoire d'Océanographie Dynamique et de Climatologie, Université Pierre et Marie Curie in Paris [*de Boyer Montégut et al.*, 2004]. In comparison with older MLD estimates [*Kara et al.*, 2003; *Monterey and Levitus*, 1997], this new climatology represents somewhat lower MLD values in northern North Atlantic. For example, the new winter climatology for the Greenland Sea yields maximum MLD of ~750 m, while the older estimate was ~1000 m. Several reasons may contribute to these differences (see discussion by *de Boyer Montégut et al.* [2004]. The new climatology is based on larger hydrographic data sets and on 0.2°C temperature criterion applied to the individual unsmoothed profiles, in comparison to the 0.5°C or 0.8°C threshold values applied to already averaged profiles in earlier MLD estimates [*Kara et al.*, 2003; *Monterey and Levitus*, 1997]. It seems that the larger temperature criteria tend to represent changes of the main thermocline rather than changes in mixing of the top water column, and this is the reason why we have decided to use the more recent MLD data.

[14] The uncertainty in the MLD estimates is a potential source of error in our TKE_{RT} estimates. To evaluate this problem, we repeated all our computations twice. In the first case we used the MLD estimates as shown in Figure 1, with MLD greater than ~100 m in March and about 20–30 m in late summer in most of the region. In the second case we assumed that the MLD was on the order of 200 m in early spring (March–April) and 80 m during the rest of the investigated time period. Such MLD estimate corresponds



Figure 4. As in Figure 3 but for the turbulent energy generation rate due to wind action (TKE_W).

more closely to the earlier oceanographic climatologies for the region [*Kara et al.*, 2003; *Monterey and Levitus*, 1997]. Our calculations indicated that although the differences in the assumed MLD have some effect on the values of turbulent kinetic energy generation rates (TKE_{RT}), the correlations between meteorological parameters and Chl are not sensitive to these differences.

4. Regional Conditions

4.1. Hydrography

[15] A brief description of regional hydrographic conditions, atmospheric forcing, and phytoplankton distribution is given next. The region of north polar Atlantic located between $70^{\circ}N-80^{\circ}N$ and $11^{\circ}E-11^{\circ}W$ (Figure 2) covers the northward flowing Atlantic Water within the Norwegian and West Spitsbergen Currents, and the southward cold East Greenland Current [e.g., *Orvik and Niiler*, 2002; *Cisewski et al.*, 2003; *Flatau et al.*, 2003; *Jakobsen et al.*, 2003]. The central Greenland Sea behaves like a large gyre, which is relatively isolated from the surrounding waters. Toward the east, the Arctic front marks the transition to the northward moving Atlantic Water. The hydrography of the central region in late winter is characterized by several hundred meters deep mixed layer. Near the ice edge on the Green-

land shelf the surface waters in early spring are less salty and more stratified than in other areas, owing to influence of melting ice [e.g., Feruvik et al., 2002]. Low-salinity Polar Waters are also present near Spitsbergen. During summer the hydrographic environment is characterized by increased surface water temperature and stratification with mixed layer depth of the order of tens of meters. The influence of the meltwater during summer can be noted in the west part of the Greenland Sea and near Spitsbergen [e.g., Tomczak and Godfrey, 1994; Saloranta and Svendsen, 2001]. The biomass of phytoplankton decreases with time during late summer within the Greenland Sea, mainly owing to nutrient limitation [e.g., Rey et al., 2000]. Nutrients are periodically supplied to the surface waters through the storm passage. Nutrient limitation is less pronounced in the East Greenland Current [Rey et al., 2000]. The hydrography of this area, with deep presence of meltwater as indicated by relatively low salinities (34.5) at 125 m, shows that the region is dynamic in terms of vertical mixing and potential transport of nutrients into the euphotic zone [see also Smith et al., 1985; Johannessen et al., 1983].

4.2. Atmospheric Forcing

[16] The atmospheric parameters described below are based on the analysis of meteorological data from the



Figure 5. As in Figure 3 but for the total turbulent energy generation rate (TKE_{RT}).

NOAA-CIRES Climate Diagnostic Center NCEP/NCAR Reanalysis Project. The results are presented in Figures 3-5 as the 7-year averages (1998-2004) of the mean monthly estimates of the three terms from equation (5): TKE_W , TKE_H, and TKE_{RT}. These estimates are shown for the months of March-August, i.e., for the part of the calendar year for which phytoplankton bloom dynamics is discussed in this paper. The estimates of TKE_H in March and April (Figure 3) are consistently positive over the entire region, indicating intensive loss of heat by the ocean. The highest values of TKE_H are observed in the region of the cold East Greenland Current in March. In May the 7-year averaged monthly means of TKE_H can be negative or positive, with the absolute values more than an order of magnitude smaller than the March values. In June, July, and August the 7-year averaged monthly means of TKE_H are negative in the entire region, indicating the net heat gain by the ocean from the atmosphere. During this time period, $\ensuremath{\text{TKE}_{\text{H}}}$ reduces the TKE_{RT} generated by the wind action. The seasonal minimum of TKE_H (maximum H_0) is generally observed in June, except in the southwestern part of the region where it is observed in May.

[17] The generation rate of turbulent kinetic energy due to the wind action is also higher in March and April than in other months (Figure 4), with highest values of 7-year averaged monthly mean TKE_W observed in March. Note, however, that during this time of the year TKE_H contributes more to TKE_{RT} than does TKE_W . In May the 7-year averaged monthly mean TKE_W decreases to about half of TKE_W observed in April, but its contribution to TKE_{RT} is now greater than the contribution from TKE_H . In June, July, and August, TKE_W is about half of the May TKE_W , with somewhat higher values in the southeastern part of the region.

[18] The seasonal changes of the 7-year averaged monthly mean TKE_{RT} shown in Figure 5 are in agreement with the patterns in TKE_H and TKE_W. The maximum values of TKE_{RT} are observed in March and April, with March values being the highest in most of the region. In addition, the spatial variability of TKE_{RT} available in March is characterized by the relatively high values in the northwest part of the region. The lowest 7-year averaged monthly means of TKE_{RT} in March-April are present in the central and southern parts of the study area. From the comparison of Figures 3, 4, and 5 it is evident that the TKE_H dominates the TKE_{RT} in March and April. Interannual variability (not shown) of TKE_{RT} during this time period is usually correlated with both TKE_H and TKE_W. Interannual variability of TKE_{RT} in spring is the strongest in the region of East Greenland Current and the weakest in



Figure 6. SeaWiFS monthly composites of surface chlorophyll concentration in May of 1998 through 2004 for the north polar Atlantic region, which indicate significant interannual variability of phytoplankton biomass. See color version of this figure at back of this issue.

the central Greenland Sea and near the Spitsbergen. In May the 7-year averaged monthly mean TKE_{RT} decreases fivefold to tenfold from its April values, but it is still about 2–4 times higher than the 7-year averaged monthly means in June–August time period. In contrast to March and April, it is the TKE_W which dominates the TKE_{RT} in May. Through the rest of the season (June–August) the interannual changes of TKE_{RT} are well correlated with TKE_W (not shown) and TKE_H acts to decrease the TKE_{RT}. The lowest seasonal values of TKE_{RT} are observed during the June–July time period.

4.3. Chlorophyll

[19] The results presented in this section are based on the analysis of SeaWiFS monthly Chl composites. The terms such as "SeaWiFS monthly Chl composite for the month of May" will be interchangeably referred to as "monthly Chl" or "May Chl." The 5-month April–August average of Chl will be referred to as "seasonal Chl." The high-latitude North Atlantic region is characterized by a very wide seasonal range of Chl, with a prominent spring/summer phytoplankton bloom and with significant interannual variability in local timing and magnitude of the bloom. To demonstrate the interannual variability, we show in Figure 6 the SeaWiFS monthly composites of Chl in the northern

North Atlantic for the months of May in years 1998–2004. In addition, the seasonal changes of Chl distribution and its interannual variability during the 5 months of the productive season from April through August are shown as scatterplots in Figures 7-11.

[20] In April, during the initial period of bloom development, Chl is still relatively low, with lowest values observed in the region of the central Greenland Sea (Figure 7). Nevertheless, it is clear that considerable growth is occurring already in April in many locations, where Chl values are significantly higher than the end-of-March concentrations (not shown here). The 7-year averaged April Chl is highest in the northeast (near Spitsbergen) and southwest parts of the region.

[21] In May and June an increased surface irradiance and mixed layer shoaling coincident with lower TKE_{RT} , as well as a large supply of nutrients from winter and early spring entrainment, are conducive to very large growth of phytoplankton. The highest 7-year averaged May Chl was observed in the western and northern parts of the region, while lower Chl was observed in the Greenland Gyre (Figure 8). Chl in the western areas showed the largest interannual variation. In June, the 7-year averaged Chl was also quite high (Figure 9) and we did not observe as strong regional differentiation as in May. The regionally averaged



longitude [degrees]

Figure 7. Interannual variability of Chl in the month of April.

Chl was highest either in May (in 1998 and 2003) or in June.

[22] The 7-year average of monthly Chl in July is on the order of 1 mg m⁻³ in most of the region (Figure 10), while for August it is usually somewhat lower (Figure 11). Nutrient limitation has been observed in late summer in the region of Greenland Gyre, while it is not considered to be as significant a problem within the East Greenland Current waters [e.g., *Rey et al.*, 2000]. The August decline of surface Chl is coincident with the seasonal decrease of daily net shortwave radiation flux and increase of the TKE_{RT} in comparison to June and July values.

[23] The geographical distribution of the 7-year averaged seasonal Chl is shown in Figure 12. The highest 7-year seasonal Chl is observed in the region of East Greenland Current and near Spitsbergen. The seasonal Chl averaged over the entire region of interest was highest in 1998 and lowest in 2003 and 2004. The correlation analysis indicates that seasonal Chl is strongly correlated with May/June Chl averages (not shown). Note also that seasonal Chl in the north polar Atlantic is quite high in comparison to many other open ocean regions. Such high Chl values are possible only as long as nutrients are replenished to the surface waters during the productive season. It is expected that mechanisms enhancing vertical mixing in the water column play a crucial role in replenishing these nutrients and supporting phytoplankton blooms in the region. Other factors limiting Chl in North Atlantic are various loss rates due to biological processes such as metabolism and zooplankton grazing, but remote sensing data analyzed in this paper do not allow us to estimate such losses.

5. Relationship Between Atmospheric Forcing and Chlorophyll

[24] We will now describe the links between atmospheric forcing and interannual chlorophyll variability. For the sake of clarity of our discussion we somewhat arbitrarily distinguish four phases in the bloom cycle, namely pre-bloom (March), bloom initiation (April), late spring/early summer (May and June), and mid-to late summer (July and August) phases. We do not include in our analysis the fall and winter seasons because of the limited availability of satellite ocean color data (or no data at all) in the region owing to cloudy skies and polar night.



Figure 8. As in Figure 7 but for the month of May.



Figure 9. As in Figure 7 but for the month of June.

5.1. Pre-Bloom Phase

[25] Because the end-of-winter phytoplankton stock serves as an inoculum for spring phytoplankton bloom, we will now illustrate the interannual variability in the pre-bloom conditions. The early spring SeaWiFS data of Chl are missing for 2003 and 2004 owing to overcast skies over most of the region, so our analysis of the end-of-winter and bloom initiation phases is limited to years 1998-2002. Composites of Chl obtained for the 8-day time period starting on day 89 each year (March 29 in years 1998, 1999, 2001, and 2002, and March 28 in 2000) and the monthly means of TKE_{RT} in March were used to calculate the correlation coefficients between the two parameters. The results indicate a significant relationship between Chl and TKE_{RT} as summarized in Figure 13. The correlation coefficient is negative in most of the region. The positive correlation coefficients were obtained for the spatial bins located in the area influenced by the cold water currents. Chl averaged for the entire region was highest in 1999 (0.41 mg m^{-3}) and lowest in 2001 (0.23 mg m^{-3}) .

5.2. Bloom Initiation

[26] For the purpose of our analysis we define the timing of the bloom onset in each grid element as the day of the year when 8-day Chl composite data binned to $2^{\circ} \times 2^{\circ}$ grid reached for the first time at least 0.5 mg m⁻³. We recognize that this Chl threshold value has been chosen somewhat arbitrarily, but the exact value of selected threshold is not so important as long as it is significantly higher than the average winter Chl and lower than Chl when phytoplankton bloom is fully developed.

[27] The dates of the bloom onset in the north polar Atlantic in different years as defined by the 0.5 mg m⁻³ Chl threshold are shown in Figure 14. The timing of the bloom onset exhibits significant interannual variability. In general, the most consistent pattern is observed in waters west off Spitsbergen coast where the bloom is always starting at the beginning of April. Importantly, our analysis shows a positive correlation between the date when Chl for the first time reached or exceeded 0.5 mg m⁻³ and TKE_{RT} integrated for the month of March (Figure 15). In most cases the correlation coefficient is >0.5. This indicates that the timing of the bloom onset in a given area is delayed in years when high TKE_{RT} occurs in early spring.

[28] It is interesting to note that the increase of Chl to the 0.5 mg m⁻³ value in the northeast (west off Spitsbergen) and the western part of the region (off Greenland) occurs during the time period when the 8-day average TKE_{RT} is consistently positive and supports mixing in the surface







Figure 11. As in Figure 10 but for the month of August.

waters (Figure 16). In the central part of the study area this phase of the bloom development is also associated with positive TKE_{RT} averaged over the 8-day time period, although short (1-2 days) intermittent periods of heat gain were occasionally present and possibly sometimes a weak restratification of surface waters occurred. The observation that the increase of Chl in the open ocean starts when TKE_{RT} is still positive indicates that the bloom initiation actually precedes the onset of seasonal surface water stratification. Although this result may seem somewhat surprising in the light of classical phytoplankton bloom models, it is in agreement with Sverdrup's [1953] comments on phytoplankton blooms in conditions of moderate turbulence. Similar results based on in situ experiments and models that resolve the vertical turbulent structure in the surface layer have been reported by others [e.g., Townsend et al., 1992; Stramska and Dickey, 1994; Huisman et al., 1999]. It is, however, remarkable that our results presented here indicate that the blooms preceding the onset of seasonal stratification are more of a rule than an unusual situation in the north polar Atlantic. It is possible that the bloom near the Greenland and Spitsbergen can be accelerated by the presence of meltwater stabilizing the water column.

[29] To gain more insight into the early stage of phytoplankton bloom, we next examined for the month of April the relationships between Chl and (1) total turbulent kinetic energy generation rate (TKE_{RT}), (2) turbulent kinetic energy generation rate due to the wind action (TKE_W), and (3) turbulent kinetic energy generation rate due to net heat flux (TKE_H). The results are summarized in Figure 17. The most noteworthy finding is the significant negative correlation between Chl and TKE_{RT} in most parts of the study area. The net heat flux (included in the second term of equation (5)) dominates the changes in TKE_{RT} during this time of the year and represents the conditions of active heat loss from the ocean to the atmosphere (compare Figures 3 and 5). Wind stirring (Figure 4) is also active and reinforces the effects of heat loss, but most of the TKE_{RT} variability is due to the variability in H_0 . The correlation between Chl and TKE_w is not significant in April in most parts of the region (with the exception of the northeastern part). The correlation coefficient between Chl and TKE_H is similar to that for Chl and TKE_{RT} (Figure 17). The relationship between Chl and TKE_{RT} is also significant if we consider Chl and TKE_{RT} averaged over the entire region of interest (Figure 18). This



Figure 12. The 7-year averaged seasonal (April–August) means of Chl.



Figure 13. Correlation coefficient (*r*) between the turbulent energy generation rate (TKE_{RT}) in March and surface chlorophyll *a* concentration (Chl) estimated from 8-day SeaWiFS composites starting on day 89 of the year. The results are based on data from 1998 through 2002. Horizontal lines indicate $r = \pm 0.5$.

means that in years with higher TKE_{RT} , the region-averaged surface Chl in April is lower. In summary, our results for prebloom and bloom initiation phases indicate a crucial role of atmospheric forcing for interannual Chl variability during this time period. This outcome is what one would expect for the early phase of bloom based on phytoplankton models. The rationale for this is that during this time of the year, atmospheric forcing is very energetic and has dramatic effects on redistribution of water properties in weakly stratified ocean. Increased mixing within the water column in April decreases phytoplankton growth, because average light energy available to phytoplankton cells becomes lower (equations (1) and (2)), and increases phytoplankton losses because cells are removed from mixed layer during mixing/ restratification events.

5.3. May and June

[30] We will describe now the remaining phases of the bloom: the late spring to early summer period (May and June) and mid-summer to late summer period (July and August). Interestingly, our analysis indicates that phytoplankton stock displayed different responses to increased mixing in the late spring–early summer period compared to mid-to-late summer period, and also regionally. We will first illustrate the responses typical for the May–June period.

[31] Our analysis suggests that for both May and June, there is no strong relationship between interannual variability of Chl and TKE_{RT} if considered on a monthly basis, although we did observe some negative correlation between Chl and TKE_{RT} in May (not shown). The cause for such weak relationship during this time period can be explained as follows. Atmospheric forcing expressed by the magnitude of TKE_{RT}, TKE_W, and TKE_H is much weaker in May/ June time period than it is in March/April. In addition, surface waters become more stratified later in season and more energy is needed for significant deepening of MLD. As a result, the mixing/restratification events are less intense in May/June than in April. Note that during this time period, increased mixing can have counteracting effects on phytoplankton: Replenished nutrients allow for greater phytoplankton growth, while decreased average light energy received by phytoplankton cells decreases the growth rates. In addition, mixing/restratification events remove the biomass to deep waters, which increases phytoplankton loss rates. The main difference between April and May/June phases of the bloom is that in April there is no benefit to



Figure 14. Timing of the onset of phytoplankton bloom (day of the year when Chl reached or exceeded 0.5 mg m^{-3}). The results are based on 8-day Chl composites from 1998 through 2002.



Figure 15. Correlation coefficient (r) between turbulent energy generation rate (TKE_{RT}) in March and the timing of the phytoplankton bloom onset displayed in Figure 14.

phytoplankton from mixing due to replenishing of nutrients because nutrients are plentiful anyways, and that mixing events are more dramatic which means stronger impact on phytoplankton growth and losses than in the May/June phase. As a result, we observe a strong negative correlation between TKE_{RT} and Chl in April, while the effects of atmospheric forcing on May/June Chl are not as evident.

[32] Interestingly, the interannual variability of Chl averaged over the 2-month period of May and June showed significant correlation with the early spring values of TKE_{RT}, TKE_W, and TKE_H (i.e., when TKE_{RT}, TKE_W, and TKE_H are averaged over a 2-month period of March and April). Recall that on average, TKE_H is always significantly larger than the TKE_W in March and April. This is why the spatial patterns of the correlation coefficient for Chl versus TKE_{RT} shown in Figure 19 most closely resemble the patterns for Chl versus TKE_H. These results indicate that the correlation coefficients between Chl and TKE_{RT} as well as between Chl and TKE_H in the central part of the study area are positive. In the remaining parts of the region the correlation coefficients are negative. Without access to the full information about the water column density structure, we can only speculate about the possible reasons for the different patterns in the correlations. Most likely the observed relationships are related to the regional differences in the atmospheric forcing and local hydrography. Recall

that on average the spatial variability of TKE_{RT} available in March and April is characterized by the relatively high and low values of TKE_{RT} in the northwestern and central parts of the region, respectively. It is possible that some increase in the relatively low TKE_{RT} values in early spring in the central part of the Greenland Sea can delay the onset of seasonal stratification and eventual nutrient depletion. Note that nutrient limitation is usually not as big a problem in the East Greenland Current region as it is within the Greenland Sea Gyre [Rey et al., 2000], which means that the benefits from larger nutrient supply are not as essential for phytoplankton growth in this region. Instead, further increase of the already high TKE_{RT} values in the East Greenland Current region can have inhibiting effect on the development of the bloom through deepening of the mixed layer and decreasing of the light levels.

5.4. July and August

[33] The interannual variability of Chl averaged over a 2-month period of July and August is not well correlated with the spring TKE_{RT} (not shown), but is more sensitive to atmospheric forcing in summer. There is a significant



Figure 16. Turbulent energy generation rate (TKE_{RT}) averaged over the 8-day time periods corresponding to the data shown in Figure 14.



Figure 17. Correlation coefficient (r) between surface chlorophyll (Chl) and turbulent energy generation rate due to heat fluxes (TKE_H), wind (TKE_W), and total turbulent energy generation rate (TKE_{RT}) in April.

positive correlation between Chl and TKE_H and negative correlation between Chl and TKE_W in the northern and western parts of the region (Figure 20). Because the absolute value of TKE_H is relatively low, the TKE_{RT} variability is relatively more affected by TKE_W in July and August than it was in early spring (compare Figures 3–5). In particular, the effect of wind energy on TKE_{RT} is stronger in the southeastern part of the region. The increase of mixing (or weaker stratification) either through TKE_W or TKE_H is expected to have beneficial effects for sustaining phytoplankton population, as nutrients are most likely one of the factors strongly limiting phytoplankton growth during this time period.

5.5. Seasonally Averaged Chl (April Through August)

[34] We now briefly summarize the interannual variability of Chl averaged over the 5-month time period (April through August). This 5-month average of Chl will be referred to as "seasonal Chl." The most important observation is that the seasonal Chl in the north polar Atlantic displays a very similar dependence on early spring atmo-



Figure 18. Relationship between regionally averaged surface chlorophyll (Chl) and turbulent energy generation rate (TKE_{RT}) in April.

spheric forcing conditions as does Chl averaged over May and June (Figure 21). This is mainly attributed to the fact that the highest values of Chl are observed in the May–June time period. Thus, in the Greenland Gyre, higher seasonal



Figure 19. Correlation coefficient (*r*) between surface chlorophyll (Chl) averaged over the 2-month period of May and June, and the turbulent energy generation rate (TKE_{RT}, TKE_H, and TKE_W), all averaged over the 2-month period of March and April.



Figure 20. Correlation coefficient (*r*) between surface chlorophyll (Chl) and the turbulent energy generation rate (TKE_{RT}, TKE_H, and TKE_W), all averaged over the 2-month period of July and August.

Chl is favored by stronger spring mixing of surface waters, while the opposite is true for the areas outside the gyre.

6. Summary and Discussion

[35] We have examined the interannual variability of phytoplankton blooms in the region of north polar Atlantic in response to meteorological variability. Our results identify subregions and times of the year of contrasting responses of Chl to regional atmospheric forcing. In early spring (April), the chlorophyll concentration and timing of the bloom onset are significantly correlated with TKE_{RT}. In years with low TKE_{RT} the timing of the significant increase of Chl in comparison to winter concentrations is earlier and regionally averaged Chl in April is higher than in years with elevated TKE_{RT}. This response of Chl to TKE_{RT} is in agreement with phytoplankton models which predict development of blooms when mixing in the surface waters decreases significantly in comparison to winter.

[36] With time of the year the relationship between Chl and TKE_{RT} changes. During the May-June period when regional TKE_{RT} is significantly lower than in April, the responses to meteorological conditions are not as explicit. This is associated with opposite effects of enhanced mixing on phytoplankton population in the situation of moderate stratification of the water column. In this case, increase of mixing can increase phytoplankton growth when nutrients are needed, decrease growth if light energy becomes limiting, and intensify loss rates through removal of biomass from surface to deep waters. On the other hand, our data suggest that Chl averaged over May and June is influenced by the early spring (March-April) TKE_{RT}. Chl within the Greenland Gyre reaches higher values in years when March-April TKE_{RT} is elevated than in years with low early spring TKE_{RT}. The opposite is true for the areas outside the gyre. In these areas in years with weatherenhanced oceanic boundary layer mixing in early spring, the May-June Chl does not reach as high values as in years with low TKE_{RT}. These regional differences in responses to early spring atmospheric forcing may be related to differences in hydrographic and meteorological conditions which



Figure 21. As in Figure 20 but for surface chlorophyll (Chl) averaged over the April–August period.

regulate local input of nutrients through vertical mixing processes.

[37] In the July-August period, nutrients are usually depleted; consequently, meteorological conditions which favor lower water stratification support higher Chl. In the western part of the study region, which is influenced by the cold East Greenland Current, the net heat flux tends to have stronger impact on Chl than the wind. This may be associated with the fact that lower wind speed is observed here than in other areas. Thus in the vicinity of the East Greenland Current in years when the heat gain by surface water is low the July-August Chl tends to have higher values than when the heat gain is high. In the southeastern part of the study region, higher Chl corresponds to stronger winds. However, it is the interannual variability of early spring TKE_{RT} that controls the seasonally averaged (April-August) Chl variability (primarily through influences on Chl in the May-June period when Chl reaches seasonal maximum).

[38] Our methods of data analysis were similar to those used by Follows and Dutkiewicz [2002]. They postulated that phytoplankton responses to interannual differences in atmospheric forcing can be divided into two regimes defined by the ratio of the spring critical layer depth (h_c) and the winter mixed layer depth (MLD_w). They hypothesized that regions with large h_c/MLD_w are characterized by an enhanced bloom in response to augmented mixing, while in the low h_c/MLD_w regime the blooms are less intense in years with strong mixing. Follows and Dutkiewicz [2002] supported their hypothesis with the relationships derived for the subtropics (large h_c/MLD_w), but subpolar data representing low h_c/MLD_w regime did not show a clear pattern to back up their hypothesis. Note that our data from the polar region of North Atlantic do not support their hypothesis either. Instead, our results suggest that in the regions where nutrients limit the seasonal phytoplankton productivity, an increased mixing tends to favor stronger blooms, while in the regions where nutrients are in a larger supply, this is not the case.

[39] As a main factor influencing the interannual variability of phytoplankton blooms in the open ocean, we have considered the turbulent energy generation rate. Note that in a hypothetical situation when Chl depended only on TKE_{RT} the absolute value of the correlation coefficient between Chl and TKE_{RT} would be unity. Our results indicate that this value is less than unity, which is a consequence of the fact that factors not included in our analysis also influence phytoplankton blooms. Such factors include various biological processes such as grazing, mortality, and differences in composition of phytoplankton population.

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Figure 1. Mixed layer depth (MLD) climatology within the study region. MLD was estimated from temperature profiles using 0.2°C temperature difference criterion [*de Boyer Montégut et al.*, 2004]. Description of the data can be found at http://www.lodyc.jussieu.fr/~cdblod/mld.html.



Figure 2. Schematic diagram of the Greenland Sea showing (top) the surface circulation and (bottom) water mass structure at a section across the central gyre. (From *Arctic Monitoring and Assessment Programme* [1998]).



Figure 6. SeaWiFS monthly composites of surface chlorophyll concentration in May of 1998 through 2004 for the north polar Atlantic region, which indicate significant interannual variability of phytoplankton biomass.