Effect of meteorological conditions on interannual variability in timing and magnitude of the spring bloom in the Irminger Basin, North Atlantic

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Abstract

Interannual variability in the spring bloom in the Irminger Basin, northern North Atlantic, is investigated using SeaWiFS-derived chlorophyll-a (chl-a) concentration and satellite or model-derived meteorological data. Variability in the timing and magnitude of the spring bloom in the basin is evaluated. A method for estimating a time series of Sverdrup’s critical depth from satellite-derived data is introduced. Comparison with modelled mixed layer depth and chlorophyll concentration demonstrates that Sverdrup’s critical depth model is valid for the Irminger Basin spring bloom. The dependence of the timing and magnitude of the spring bloom on winter pre-conditioning is investigated. We find that in the Irminger Basin the start of the spring bloom can be estimated from the preceding winter’s mean wind speed and net heat flux. We also find that the maximum chl-a concentration during the bloom can be estimated from the frequency of winter storms. Increased storm activity results in a reduced bloom chlorophyll maximum by delaying the development of spring stratification, resulting in the bloom missing the ‘window of opportunity’ for optimum phytoplankton growth.

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1. Introduction

In many parts of the world’s oceans the annual cycle of phytoplankton growth is dominated by a rapid, intense population explosion—the spring bloom. The annual increase in phytoplankton in the North Atlantic has interested scientists since the earliest days of biological oceanography (see introduction of Lalli and Parsons, 1997). The earliest attempts to characterise the spring bloom relied on measurements taken from stationary weatherships (e.g., Sverdrup, 1953). With the development of the continuous plankton recorder, time series of data across large areas of the ocean could be developed (Colebrook and Robinson, 1965), and phytoplankton blooms have been the focus of several cruise programmes including the North Atlantic Bloom Experiment (Ducklow and Harris, 1993), British Ocean Flux Study (Savidge et al., 1992) and Joint Global Ocean Flux Study (Marra and Ho, 1993). Most recently, the advent of satellite data has

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allowed continuous monitoring of sea-surface conditions at high spatial and temporal resolution (Robinson and Mitchelson, 1983; Siegel et al., 2004). Well over 100 years of observations have resulted in the formulation of a classic cycle of phytoplankton abundance. The spring bloom is most pronounced in northern latitudes, where, during winter, phytoplankton growth is limited principally by low light levels. Strong winds and heat loss from the surface ocean drive convective overturning due to buoyancy loss. Although nutrient concentrations are uniformly high throughout the water column, the deep mixed layer (ML) ensures that phytoplankton are only in the sunlit surface layers for a short period of time before being mixed down into deeper, darker water. As spring approaches, the combined effects of longer, warmer days and reduced wind speeds lead to formation of a thermally stratified surface layer. This layer traps the phytoplankton, and the nutrients they require for growth, in the sunlit upper water column. Under these conditions phytoplankton grow very quickly, rapidly diminishing the stock of nutrients in the surface layer.

The timing of the spring bloom is important to the development of zooplankton (Rey et al., 1987), fish stocks (Platt et al., 2003) and the seasonality of the biological carbon pump (Eppley and Peterson, 1979). The classic quantitative description of bloom initiation was developed by Sverdrup (1953). In his critical depth model the relationship between the depth of water column mixing and the available light controls the timing of the onset of the bloom. Field studies have generally found Sverdrup’s critical depth model (SCDM) to be valid in temperate and sub-polar latitudes, where a distinct seasonal phytoplankton cycle occurs and where the winter ML is too deep to allow net growth. However, some studies have found that intermittent shallowing of the ML can result in increased chlorophyll concentrations (e.g., Ho and Marra, 1994) or that a spring bloom can occur in the absence of vertical stratification (Townsend et al., 1992; Durbin et al., 2003). SCDM is also less appropriate in the sub-tropical regions, where phytoplankton growth is hindered by nutrient, rather than light, limitation (e.g., Obata and Endoh, 1996; Dutkiewicz et al., 2001; Siegel et al., 2002).

The start of the spring bloom has also been associated with variations in incident radiation and wind mixing (Townsend et al., 1994), air temperature (Nezlin and Li, 2003), sea-surface temperature (Thomas et al., 2003) and net heat flux (Azumaya et al., 2001). However, there have also been studies that found no, or complex, relationships between the timing of the bloom and wind speed or net heat flux (Anderson and Prieur, 2000; Waniek, 2003). The magnitude of the bloom is also subject to interannual variability, and this is usually attributed to changes in the depth of winter mixing, and hence concentration of nutrients available in spring (Koeve, 2001; Pätsch et al., 2002). In the Irminger Basin—the focal region of this study—Olafsson (2003), however, observed no correlation between winter ML depth and surface nutrient concentrations in the following spring. Determining the physical factors controlling the characteristics of the spring bloom can be complex. These processes vary for different regions, and often it is not one factor alone but rather the interaction between competing processes that impacts on the bloom.

In the northern North Atlantic, much of the annual export production occurs during the spring bloom (Falkowski et al., 2000). Knowledge of variability in export—the flux of biogenic material from surface waters to the ocean interior and to higher trophic levels—is essential to quantifying variability in the biological carbon pump. However, quantifying interannual variability in the spring bloom first requires knowledge of what is ‘normal’. To do this a long time series of measurements is ideally required. Such time series are rarely available in oceanography, particularly for biological measurements. In this study, 6 years of remote sensing data at high spatial and temporal resolution are analysed, ensuring that the variability in bloom characteristics is captured. By studying the inter-annual variability in the bloom and comparing it to variability in the meteorological forcing, the controls on the spring bloom are determined.

We focus on the Irminger Basin (IB), situated between Greenland and Iceland in the North Atlantic sub-polar gyre (see Fig. 1). The basin experiences severe winter meteorological conditions and may be a site of deep ocean convection (Pickart et al., 2003). Despite this, export production in the region is thought to be high, \(~100–150\text{ gC m}^{-2}\text{ yr}^{-1}\), on the basis of estimates from satellite-derived chlorophyll concentrations (Falkowski et al., 1998; Laws et al., 2000), although estimates of export based on in situ nutrient drawdown data are much lower at \(~35–60\text{ gC m}^{-2}\text{ yr}^{-1}\) (Henson et al., 2003; Sanders et al., 2005; Waniek et al., 2005; Henson et al., 2006). In this study, SeaWiFS chlorophyll-\(a\)
(chl-$a$) concentration data from 1998 to 2003 are used, along with the corresponding meteorological data, to elucidate factors affecting the timing and magnitude of the annual spring bloom.

2. Data

Daily Level 3 Standard Mapped Images of SeaWiFS chl-$a$ concentration, photosynthetically available radiation (PAR) and attenuation coefficient at 490 nm wavelength ($K_{490}$) at 9 km resolution for 1998–2003 were downloaded from http://www.disc.gsfc.nasa.gov/data/datapool/SEAWIFS/index.html. The data were then averaged to 0.5° resolution. For plotting purposes 7-day and 3-day means of the data were calculated from the daily files. A mean for a particular pixel was created only if at least 2 out of the 3 days or 4 out of the 7 days contained data. Prior to March and after October no data are available, because of low incident sun angle. $K_{490}$ was converted to $K$ (the attenuation coefficient in the PAR wavelengths) using the equation of Rochford et al. (2001): $K = 0.0085 + 1.6243 \times K_{490}$ (where $K_{490} < 1$). PAR was converted from units of Einsteins m$^{-2}$ day$^{-1}$ to W m$^{-2}$ as: Watts = (Einsteins $\times 6e^{23}$)/(86400 $\times 2.77e^{18}$).

All temperature profiles measured by Argo profiling floats in the study region from 1998 to 2003 were downloaded from http://www.ifremer.fr/coriolis/cdc. As only floats deployed after mid-2001 were equipped with salinity sensors it was not possible to calculate density profiles from temperature and salinity for the entire dataset. The ML
depth was therefore determined from Argo floats on the basis of temperature alone. The depth of the surface mixed layer is defined as the depth $z$ at which the temperature difference $\Delta T(z) = T(z) - T(z = 0)$ in the upper ocean exceeds a specified threshold value. If potential density measurements are available, a difference of $0.01 \text{ kg m}^{-3}$ is commonly used (Thomson and Fine, 2003). The equivalent temperature difference was determined by examination of CTD profiles taken during a series of cruises to the Irminger Basin in 2002 (UK Marine Productivity programme; http://www.nerc.ac.uk/marprod) and was found to be $\sim 0.07 \degree \text{C}$. The ML depth of all Argo temperature profiles was calculated in this way.

The net upward flux of long-wave radiation, latent heat flux, sensible heat flux and flux of short-wave solar radiation are available at 2$^\circ$-resolution from the NCEP/NCAR reanalysis project at http://www.cdc.noaa.gov/cdc/data.ncep.reanalysis.html. The net heat flux was calculated as the sum of the components, where a positive flux represents heat entering the ocean. All daily data from 1998 to 2003 were downloaded and the data linearly interpolated onto a 0.5$^\circ$-resolution grid.

QuikScat scatterometer wind speed data are available from the Remote Sensing Systems website http://www.remss.com/. The data were provided at 0.25$^\circ$-resolution as daily files for 2000–2003 and were then averaged to 0.5$^\circ$ resolution. Note that pre-2000 satellite-derived wind speed was available only as a weekly product from ERS-2. The storms examined in this paper are expected to last only 3–4 days and so are not adequately represented in the weekly ERS-2 product. For this reason, only daily QuikScat data are used, limiting the storm frequency analysis to the years 2000–2003. Note that for the seasonal mean wind speed calculations the weekly ERS-2 data for 1998 and 1999 are included in the analysis.

A 1-D vertical mixing model which uses a Kraus and Turner (1967) mixing scheme is also used to estimate ML depth. The fine details of the model can be found in Rahmstorf (1991) and Waniek (2003). The model was initialised using temperature and salinity profiles from CTD stations carried out in 2002 during the NERC Marine Productivity cruise programme (http://www.nerc.ac.uk/marprod). The model is forced using QuikScat wind speed and NCEP reanalysis air temperature, humidity, solar radiation and cloudiness products. The modelled ML depth was estimated from temperature output using the same criteria as for the Argo float profiles.

In order to quantify the spatial variability in the basin, empirical orthogonal function (EOF) analysis was used to determine biogeographical zones in the Irminger Basin on the basis of SeaWiFS chl-$a$. The zonation seeks to reduce a large dataset to a few zones whose characteristics are representative of the area they cover, and yet are distinct from the other zones. The EOF method allows the identification of the principal modes of variability in the dataset and returns maps of the spatial patterns. Details of the methodology can be found in Preisendorfer (1988). Prior to the analysis the seasonal mean and interannual trend were removed from the data. The first four modes of the EOF analysis were found to be non-degenerate on the basis of North’s ‘rule of thumb’ (North et al., 1982) and explain a total of 44% of the variability. A varimax rotation was found to facilitate interpretation of the results (Richman, 1986) and was performed on the first four modes. The spatial patterns of variability are displayed as homogenous correlation maps—the correlation between the time series of the EOF mode and the time series of the original chl-$a$ data at each pixel.

3. Results and discussion

3.1. Biogeographical zones

The spatial patterns of the first four modes of the EOF analysis are displayed in Fig. 2. These maps highlight the ‘centre of action’ of each mode within which the characteristics of the chl-$a$ signal will be similar. The chl-$a$ characteristics of each mode will also be distinct from the other zones. Mode 1 is located to the southwest of Iceland (Iceland Shelf: IS), mode 2 is on the east coast of Greenland (East Greenland: EG), mode 3 is centred over the Reykjanes Ridge (RR), and mode 4 is in the North Central Basin (NCB). The zones are spatially distinct from each other and highly correlated with the original chl-$a$ data.

These results are consistent with a zonation of the Irminger Basin carried out by Holliday et al. (2006) and Waniek and Holliday (2006). They used CTD data to divide the IB into zones on the basis of the physical properties of the water column (temperature, salinity and known positions of currents and fronts). The authors report that the division of their two East Greenland regions (Polar and Atlantic) is
arbitrary in terms of surface water characteristics. The analysis of surface chl-a presented in this paper results in only one zone on the east coast of Greenland (EG zone) and hence supports their conclusion. Additionally Holliday et al. (2006) state that their northern Irminger Current and Central Irminger Sea zones are barely distinguishable in terms of chl-a biomass. Again the results in this paper confirm that, in terms of chlorophyll characteristics, the northern and central basin are part of the same zone (NCB). The RR zone defined in this study coincides with the Holliday et al. (2006) Reykjanes Ridge province, and their ‘Iceland basin water type’ is equivalent to the IS zone in this manuscript.

The 6-year mean annual cycle of chl-a in each of the four zones defined by the EOF analysis is shown in Fig. 3. Each zone is represented by the mean of a 1° × 1° box centred on 63°N, 22°W for the IS zone, 62°N, 40°W for the EG, 57°N, 30°W for RR zone and 64°N, 32°W for the NCB. In the IS zone a sudden increase in chl-a occurs in mid-April, and it reaches a peak of ∼2 mg m⁻³ at the end of May. Chl-a remains high at ∼1.2 mg m⁻³ during summer, with a gradual return to winter concentrations beginning in mid-September. In the EG, RR and NCB regions winter chl-a is ∼0.2 mg m⁻³ until mid-April. Throughout April chl-a increases slowly and is of similar magnitude in these three regions. In the EG region in mid-May chl-a increases suddenly to a maximum of ∼1.6 mg m⁻³, but by mid-June levels have returned to ∼0.4 mg m⁻³. From mid-July to early September chl-a remains at ∼0.7 mg m⁻³. In the NCB chl-a increases throughout April, with a sharper increase at the end of May, reaching a maximum of ∼1 mg m⁻³. In the NCB chl-a drops again in mid-June and remains at ∼0.7 mg m⁻³ during the summer. In the RR chl-a begins to increase in early April but never reaches a sharp peak. Throughout spring and summer chl-a remains at ∼0.7 mg m⁻³. From early September the NCB, RR and EG zones have similar signatures, with chl-a dropping rapidly back to winter levels. The EOF analysis therefore successfully identified four regions in the Irminger Basin, which have distinct chl-a signals. These four zones will be used in the
following sections to characterise the spatial variability of chlorophyll distribution in the Irminger Basin.

3.2. Timing of the spring bloom

In Fig. 4, the 6-year time series of SeaWiFS chl-a, averaged over the 1° boxes representative of each of the four zones (as defined in the previous paragraph), are plotted. Variability in the timing, duration and magnitude of the bloom clearly occurs in all regions. In the Iceland Shelf region the highest magnitude bloom occurred in 2000, with strong blooms also occurring in 2001 and 2003. In the East Greenland zone 2000, 2001 and 2003 also experienced strong blooms, whilst in 2002 the bloom was much reduced. In the Reykjanes Ridge region 1998 had a strong spring bloom, whilst 1999 and 2002 experienced elevated chlorophyll concentrations throughout autumn. In the NCB the 2002 spring bloom was weaker than in other years.

A key indicator of interannual variability in spring bloom characteristics is the timing of the start of the bloom. The start of the bloom can be defined as the time when chlorophyll concentration rises above a certain fixed threshold (e.g., in Fleming and Kaitala (2005) the threshold is set at 5 mg m$^{-3}$). However, this method is not suited to a spatially variable region, such as the Irminger Basin, where maximum chlorophyll concentrations vary between coastal and open-ocean areas by a factor of.
two (Fig. 3). The start of the spring bloom in each year was therefore estimated as the day of year when the chlorophyll concentration first rises 5% above that year’s annual median for each zone (a method proposed by Siegel et al., 2002). In addition, the chlorophyll concentration was required to remain at this elevated level for at least 3 days in order to eliminate temporary increases in chl-\(a\) resulting from transient events, such as spring storms (Henson et al., 2006). This method is objective and defines a threshold chl-\(a\) level, which importantly, is different for each geographical zone. The start of the spring bloom was estimated for each year in each province. The results are presented in Table 1, and the start of the bloom is marked on Fig. 4. The method successfully detects the initiation of the spring bloom in all provinces and years (Fig. 4). The start date of the bloom in the Irminger Basin ranges from mid-April to late May, with the Iceland and Greenland shelf regions blooming earliest, followed by the Reykjanes Ridge region and finally the Northern Central Basin. Within each province the start date of the bloom can vary by up to 30 days interannually. On average the bloom starts earliest in 1998 and 1999 and latest in 2002 (Table 1).

3.3. Estimating Sverdrup’s critical depth

The classic model for the necessary conditions for bloom initiation was introduced by Sverdrup (1953). As the ML shoals in spring, the bloom can start when the ML depth is shallow enough that, on average, the phytoplankton population receive sufficient light for net growth. During spring not only does the ML shallow, but the surface incident irradiance also increases. The irradiance received by any one phytoplankter therefore increases, not only because the surface solar irradiance increases during spring, but, due to the shallowing of the ML, the depth-averaged irradiance is also increasing. This balance point was defined in Sverdrup’s (1953)
critical depth model. He concluded that for a spring bloom, i.e., a net increase in phytoplankton biomass, to occur the ML depth must be shallower than a certain critical depth. Sverdrup (1953) defined this as the depth at which integrated production is equal to integrated respiration, and net community growth can occur. He also defined a compensation depth, and corresponding compensation irradiance, $I_C$, at which daily production is equal to daily respiration. The compensation irradiance, $I_C$, is defined as

$$I_C = I_0 e^{-kZ_C},$$

where $I_0$ is the incident surface PAR (photosynthetically available radiation), $k$ is the attenuation coefficient of PAR and $Z_C$ is the compensation depth. In deriving the critical depth Sverdrup (1953) then assumed that production at depth $z$ is proportional to the available light so that in time interval $dt$, the change in production is $dp = a I_z dt$ and for respiration $dr = b dt$. Factors $a$ and $b$ represent community ‘growth’ and ‘loss’, respectively. By definition $dp = dr$ at the compensation depth, where $I_z = I_C$, and therefore $I_C = b/a$. Integrating over the whole time interval (1 day) and water column depth, Sverdrup derived an equation for the critical depth, $Z_{CR}$:

$$\frac{1}{kZ_{CR}} \left(1 - e^{-kZ_{CR}}\right) = \frac{I_C}{I_0}. \quad (2)$$

Of the terms in Eq. (2), $I_0$ and $k$ can be estimated from SeaWiFS data, but $I_C$, the irradiance at which respiration is equal to production, is more difficult to quantify. Losses to the phytoplankton community occur not only through phytoplankton respiration, but also through grazing, excretion and sinking (Smetacek and Passow, 1990; Platt et al., 1991). How can these community-wide losses be accounted for in Sverdrup’s formulation? By considering an increase in biomass then, implicitly, community gains are overcoming total community losses. This is not necessarily true for an increase in production.

Following Sverdrup’s hypothesis, at the time of spring bloom initiation community gains are outstripping losses for the first time, and the depth of the ML must be equal to the critical depth, $Z_{CR}$ (Siegel et al., 2002). In order to estimate the ML depth at the start of the bloom, the SeaWiFS chl-a coincident (within 1° of latitude/longitude and ± 1 day) with each Argo profile was determined. An attempt was made to estimate the ML depth separately for each province and each year. However, because the Argo floats drift with the prevailing currents, there can be long periods of time when no floats are present within a particular province. Additionally, Argo floats take profiles only every 10 days, and often no estimate of ML depth is available near the start date of the bloom. Instead, each occasion when a ML depth estimate is coincident with a SeaWiFS chl-a measurement is plotted for all provinces and years in Fig. 5. The sharp increase in biomass indicative of the start of the spring bloom occurred only when the ML depth was shallower than ~100 m, irrespective of year or location within the basin.

The SeaWiFS attenuation coefficient and PAR at the start of the bloom in each zone was determined (see Section 2). Eq. (2) was then applied to calculate the community compensation irradiance, $I_C$, in each zone (results in Table 2). $I_C$ varies between years and provinces, with a mean of ~2.5 (±0.5) mol photons m$^{-2}$ day$^{-1}$, which is equivalent to a compensation depth of ~20 m (from Eq. (1)). This estimate is consistent with published values of community compensation irradiance (~1–3.5 mol photons m$^{-2}$ day$^{-1}$; Riley, 1957; Siegel et al., 2002). It is also far greater than phytoplankton-only $I_C$ as estimated in laboratory experiments (~0.1–0.8 mol photons m$^{-2}$ day$^{-1}$; Langdon, 1988), giving confidence

### Table 1
Day of the year on which the spring bloom begins in 1998–2003 for each of four provinces

<table>
<thead>
<tr>
<th>Year</th>
<th>Iceland Shelf</th>
<th>East Greenland</th>
<th>Reykjanes Ridge</th>
<th>North Central Basin</th>
<th>Mean</th>
</tr>
</thead>
<tbody>
<tr>
<td>1998</td>
<td>120</td>
<td>128</td>
<td>110</td>
<td>133</td>
<td>123</td>
</tr>
<tr>
<td>1999</td>
<td>140</td>
<td>104</td>
<td>116</td>
<td>133</td>
<td>123</td>
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<tr>
<td>2000</td>
<td>130</td>
<td>112</td>
<td>121</td>
<td>137</td>
<td>125</td>
</tr>
<tr>
<td>2001</td>
<td>126</td>
<td>120</td>
<td>130</td>
<td>124</td>
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</tr>
<tr>
<td>2002</td>
<td>138</td>
<td>135</td>
<td>140</td>
<td>154</td>
<td>142</td>
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<tr>
<td>2003</td>
<td>134</td>
<td>132</td>
<td>136</td>
<td>144</td>
<td>137</td>
</tr>
</tbody>
</table>

The final column shows the mean start date for each year.
that this method accounts for all community losses, not just phytoplankton respiration.

Although we can estimate critical depth at the start of the bloom, it is the changes in $Z_{CR}$ during the initiation of the spring bloom that are of most interest. A time series of critical depth can be estimated using Eq. (2) iteratively using daily SeaWiFS PAR and $K_{490}$ data and corresponding $I_C$ in each biogeographical zone. In Fig. 6, the time series of critical depth in 2003 is plotted along with modelled ML depth and SeaWiFS chlorophyll concentration in each zone. (For brevity, only results from 2003 are presented. All years show similar results). Modelled ML depth is used here because the Argo float data was too sparse to satisfactorily compare to the daily critical depth estimates. The timing of the start of the bloom, as calculated in the preceding section, is also marked on Fig. 6. The critical depth varied between 0 and 50 m during the winter, whilst the ML depth was $\sim$350–400 m. In spring the ML shallowed very rapidly and during summer was fairly constant at $\sim$30–50 m. At the same time the critical depth deepened to a maximum of $\sim$150–250 m, shallowing again in late summer and autumn. After the bloom has begun, and as long as the ML depth is shallower than the critical depth, variations in $Z_{CR}$ do not impact the phytoplankton population. The variability in chlorophyll concentrations in late spring and summer are likely to stem instead from changes in nutrient availability and zooplankton grazing. The critical depth covaries to some extent with the chl-a concentration because of the influence of $k$ in the calculation of $Z_{CR}$. In the Irminger Basin there is little other material, such as gelbstoff or suspended particulate matter, which could contribute to $k$, so attenuation is almost entirely due to chlorophyll in the water column.

The spring bloom never starts whilst the ML is deeper than the critical depth. There is, however, a delay in the RR, IS and NCB regions of $\sim$10 days between the ML becoming shallower than the critical depth and the bloom starting. Whilst Sverdrup’s model determines whether growth is possible, it provides no information on how quickly the biomass might increase; i.e., the critical depth criterion is ‘necessary but not sufficient’ for a spring bloom to start (Platt et al., 1991). Nevertheless, Sverdrup’s model is valid for the Irminger Basin spring bloom.

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**Table 2**

<table>
<thead>
<tr>
<th>Year</th>
<th>Iceland Shelf</th>
<th>East Greenland</th>
<th>Reykjanes Ridge</th>
<th>North Central Basin</th>
</tr>
</thead>
<tbody>
<tr>
<td>1998</td>
<td>2.02</td>
<td>1.88</td>
<td>2.57</td>
<td>2.13</td>
</tr>
<tr>
<td>1999</td>
<td>3.14</td>
<td>2.38</td>
<td>2.77</td>
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<td>1.72</td>
<td>1.65</td>
<td>1.64</td>
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<td>2.86</td>
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<td>2003</td>
<td>2.18</td>
<td>2.44</td>
<td>1.93</td>
<td>2.99</td>
</tr>
</tbody>
</table>
In the method outlined here Sverdrup’s original assumptions regarding the critical depth model remain. Principally these are a linear photosynthesis–light relationship and use of a daily average of surface irradiance. An exact, analytical solution for the daily rate of production, which accounted for a non-linear photosynthesis–light curve, was formulated by Platt et al. (1991) and resulted in only a ~10% difference from Sverdrup’s original estimate. Another potential source of error lies in the treatment of respiration in this method. Smetacek and Passow (1990) raised concerns that field-based studies of critical depth had not accounted for losses to the phytoplankton community, other than through phytoplankton respiration. This method accounts for these other losses, as there is no requirement for knowledge of the details of the total production or respiration occurring within the water column—if a bloom has started, biomass is increasing and phytoplankton community gains are therefore greater than community losses.

3.4. Role of winter pre-conditioning

If the timing of the bloom adheres to the critical depth theory then it will be determined by the timing of the shallowing of the ML in spring. This in turn depends on the depth of the winter ML and the meteorological conditions, in particular the wind speed and net heat flux into the ocean. The winter/early spring (January–March) mean wind speed is plotted against the start date of the bloom in Fig. 7(a). The data for each of the four zones from 1998–2003 are plotted together. The trend is for the bloom to start later with increasing mean winter wind speed. A linear regression between the start date of the bloom and the mean winter wind speed (U) returned (r = 0.59, p < 0.01,
However, high wind speeds do not occur in isolation—they are expected to work in combination with net heat loss to deepen the ML, resulting in a delayed start to the bloom. In Fig. 7(b), the mean winter/early spring (January–March) net heat flux is plotted against the start date of the bloom. A negative net heat flux indicates loss of heat from the ocean. The data are highly scattered, although there may be a tendency for the bloom to start later with increased loss of heat from the ocean during winter. Nevertheless, including net heat flux in the estimation of the start of the bloom improves the regression statistics considerably ($r = 0.69$, $p < 0.001$, rmse = 8.3, $n = 24$). The equation describing the multiple linear regression of mean winter wind speed ($U$) and net heat flux ($Q$) against the start date of the bloom is

\[
\text{Start} = 30 + 6.6U - 0.09Q.
\]

In Fig. 7(c) the start date of the bloom as predicted from Eq. (2) is plotted against the observed start date (from Table 1). So a combination of increased winter wind speeds and heat loss from the ocean result in a deeper winter ML, and hence a later bloom start. The timing of the start of the bloom may thus be estimated from knowledge of the preceding winter’s meteorological conditions. Therefore, not only can short time scale weather events, such as a windy day, impact on phytoplankton populations (e.g., Nezlin and Li, 2003; Waniek, 2006).
but meteorological conditions can act over seasonal timescales to affect spring bloom timing.

3.5. Magnitude of the spring bloom

Considerable interannual variability in the timing of the bloom occurs, but is this reflected in the magnitude of the bloom? Does a late start result in a reduced magnitude bloom and vice versa? In Table 3 the maximum chl-\(a\) concentration observed in each year and each province is presented, along with the mean maximum chl-\(a\) for each year. 2002 experienced the lowest mean maximum chl-\(a\), with the highest mean maximum chl-\(a\) in 2000. In 2002, the spring bloom started later (Table 1) and had lower chl-\(a\) concentrations than in other years. This might suggest that a bloom that begins late in the year has missed the ‘window of opportunity’ for optimum bloom conditions; i.e., if a bloom starts late there may be a shorter period during which the light and mixing conditions have the potential to result in phytoplankton growth, resulting in a reduced bloom. Alternatively, a late start to the bloom may allow more time for zooplankton to reproduce. By the time the bloom starts the grazers may be able to keep pace with phytoplankton growth, preventing an explosive bloom.

The shallowing of the ML in spring is not a smooth transition and is likely to be interrupted by mixing events, such as storms. The frequency of storms during February and March was determined from satellite-derived wind speed data for each of the four zones for 2000–2003 (no satellite-derived daily wind speed data are available for 1998 and 1999—see Section 2). A stormy day was defined as one on which the wind speed was greater than 17 m s\(^{-1}\), i.e., a gale force wind (as in Moore, 2003). The percentage of stormy days is plotted against the maximum bloom chlorophyll concentration in Fig. 8 for all four regions. The data are split into two groups: the coastal areas (IS and EG) and the open ocean regions (RR and NCB). The coastal areas experienced fewer winter storms than the open ocean regions, but nevertheless the trend is the same for the two regions.

In the Irminger Basin, there is a reduction in maximum chl-\(a\) with increasing frequency of mixing events (Fig. 8). This suggest that storms result in a delay in spring stratification, with the consequence that the bloom misses the prime ‘window of opportunity’. These intermittent mixing events lead to a fine balance between increased nutrient entrainment and reduction of depth-averaged irradiance. In regions that have relatively shallow ML depths or are at low latitudes, and so are unlikely to be light limited, nutrient concentration will be the key factor controlling the magnitude of the bloom. Storm events will thus entrain nutrients and may lead to an enhanced bloom. In high latitude regions,

Table 3

<table>
<thead>
<tr>
<th>Year</th>
<th>Iceland Shelf</th>
<th>East Greenland</th>
<th>Reykjanes Ridge</th>
<th>North Central Basin</th>
<th>Mean</th>
</tr>
</thead>
<tbody>
<tr>
<td>1998</td>
<td>1.82</td>
<td>1.86</td>
<td>1.86</td>
<td>1.50</td>
<td>1.76</td>
</tr>
<tr>
<td>1999</td>
<td>1.73</td>
<td>1.59</td>
<td>1.88</td>
<td>1.75</td>
<td>1.74</td>
</tr>
<tr>
<td>2000</td>
<td>4.42</td>
<td>2.93</td>
<td>0.92</td>
<td>1.20</td>
<td>2.37</td>
</tr>
<tr>
<td>2001</td>
<td>2.76</td>
<td>2.27</td>
<td>1.13</td>
<td>1.77</td>
<td>1.98</td>
</tr>
<tr>
<td>2002</td>
<td>1.84</td>
<td>1.23</td>
<td>1.73</td>
<td>1.06</td>
<td>1.47</td>
</tr>
<tr>
<td>2003</td>
<td>2.94</td>
<td>2.94</td>
<td>0.84</td>
<td>1.52</td>
<td>2.06</td>
</tr>
</tbody>
</table>

The mean maximum chl-\(a\) concentration for each year is shown in the final column.
which typically have deep ML and low irradiance, but high nutrient concentrations. Changes in the underwater light field will likely have the most significant impact on the bloom magnitude. In the Irminger Basin any additional nutrients that may be entrained by the deepening of the ML are not reflected in the peak chlorophyll concentration. In situ measurements taken during the Marine Productivity cruise programme to the region showed that the nutricline is at \( \sim 150\,\text{m} \) depth (Sanders et al., 2005; their Fig. 3). The maximum winter ML depth always exceeds this (data not shown), so that there is no gradient in the nutrient profile prior to the spring bloom. If a storm event occurs before the bloom, no additional nutrients will be entrained, as the surface pool is at the same concentration as the sub-ML pool. Therefore, no advantage is gained from a mixing event, which in the Irminger Basin, only serves to interrupt the development of the spring stratification resulting in a delayed, low magnitude bloom.

4. Conclusion

Considerable interannual variability in spring bloom timing and magnitude occurs in the Irminger Basin. Quantifying the variability and its roots in variability in physical forcing has only been possible through the long time series of high spatial and temporal resolution data provided by satellite-based measurements.

A novel method for estimating Sverdrup’s critical depth from satellite data was developed and the critical depth criteria found to be valid for the Irminger Basin. The meteorological conditions in winter are fundamental to the timing and magnitude of the region’s spring bloom. We found that the start of the bloom can be predicted from a combination of the mean winter wind speed and net heat flux. Additionally, the magnitude of the bloom may be estimated from the frequency of winter storms.

The results presented here compare favourably with previous studies conducted in the northern North Atlantic. Time series of data constructed from Continuous Plankton Recorders suggest that increased mixing results in a later bloom, with shorter duration (Colebrook and Robinson, 1965; Robinson, 1970; Colebrook, 1979). A similar result arose from interannually resolved data collected during the Marine Light-Mixed Layers programme, which demonstrated that deeper mixing in winter led to a later, weaker bloom (Stramska et al., 2004 and other papers in that special issue). Satellite data or model results permit interannual variability to be more easily addressed, and several studies have investigated the timing of the bloom. By comparing turbulent kinetic energy to SeaWiFS chl-\(\alpha\) both Follows and Dutkiewicz (2002) and Stramska (2005) concluded that greater mixing results in reduced intensity blooms in sub-polar regions. Follows and Dutkiewicz (2002) also note that the relationship between turbulent kinetic energy and bloom intensity shows more variation between sub-regions than within them (compare to our Fig. 8). By comparing EOFs of satellite chlorophyll and wind speed, Ueyama and Monger (2005) also concluded that increased wind-induced mixing delayed the onset of the bloom and resulted in reduced bloom magnitude over the subpolar Atlantic. The importance of the timing of the phytoplankton spring bloom to zooplankton populations has been confirmed in both modelling and in situ studies. A later bloom tends to result in increased grazing pressure, as zooplankton have had longer to reproduce, thus limiting the magnitude of the spring bloom (Colebrook, 1979; Marra et al., 1995; Gifford et al., 1995).

Our results confirm the importance of sufficient light to high latitude blooms, where nutrients are abundant at the start of the growth season. The critical depth model—essentially a requirement that each phytoplankter receives, on average, a minimum critical light level—is valid. Storms and the associated high wind speeds and loss of heat from the ocean result in a delay to the bloom start and a reduction in its magnitude—the consequence of insufficient depth-averaged irradiance for phytoplankton to flourish.

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References


