ÉTUDE DE LA VARIABILITÉ MULTI-ÉCHELLE DU PHYTOPLANCTON DANS L’OCÉAN ATLANTIQUE NORD-EST À PARTIR DE DONNÉES SATELLITE

MULTI SCALE PHYTOPLANKTON VARIABILITY IN THE NORTHEAST ATLANTIC OCEAN: A SATELLITE VIEW

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Acknowledgments

This thesis is the result of a collaboration between Université Pierre et Marie Curie (Paris 6), the Centre National d’Etudes spatiales (CNES) and the Tel Aviv University. It could not have been accomplished without the support and guidence of Marina Lévy and Eyal Heifetz to whom I am grateful.

I wish to thank Patrice Klein, Pierre-Yves Le Traon Alain Saliot, Eric Thouvenot and Laurent Mémery for accepting to take part in the thesis jury. Laurent Mémery is also thanked for his support during the first stages of the work.

Spacial thanks are to Francesco d’Ovidio for the close collaboration, the scientific guidence and the exchange of ideas. I am thankful to Jean-Michel André for his contribution to the analysis of the ocean color satellite images and to Laurent Dubroca for his help in the application of geostatistical tools. Carine Lesage is thanked for her help in the processing of the satellite data. I wish to thank Gilles Reverdin and Gurvan Medec for their helpful comments and for their willingness to share their knowledge on various aspects of oceanographic research.

I am grateful to Zvi Ben Avraham for having confidence in my abilities and for opening the way for me to conduct scientific research. Dina Prialnik, Pierre Soler and Eric Thouvenot are thanked for their help in obtaining the Ph.D. scholarship that was co-sponsored by the CNES and by Tel Aviv University.

Finally, I wish thank Cyril Lathuilière, Anne-Sophie Krémeur, Rym M’sadek, Julie Deshayes, Ariane Koch-Larrouy, Julie Leloup Kieth Rodgers, Cyril Guinet, Pierre Brochard and Andrew Lenton for their significant help in different stages of the work on this thesis: from the first computer programs to the proofreading of the manuscript. ...
Abstract

Spatial and temporal variations in phytoplankton biomass in regions of mode water subduction are of great importance for the oceanic carbon cycle: they drive changes in the biochemical properties of the surface waters before they are subducted to the ocean interior, where they are isolated from the influence of the atmosphere on decadal time scales.

The objective of this thesis is to characterize the patterns of phytoplankton variability and to identify the dynamical mechanisms driving them, in a region of mode water subduction in the northeast Atlantic. The patterns are examined over spatial scales ranging from a few to thousands of kilometers, and temporal scales ranging from a week to a few years. This is done by characterization of the apparent spatial and temporal variability of chlorophyll concentrations, a proxy for phytoplankton biomass, derived from Sea-viewing Wide Field-of-view Sensor (SeaWiFS) ocean color data, and comparison with dynamical fields derived from numerical models and from satellite measurements of sea level height (SSH) and sea surface temperature (SST).

The satellite-based analysis carried during this thesis is complementary to those of the Programme Océan Multidisciplinaire Méso Echelle (POMME) field experiment and related model studies. POMME included a series of interdisciplinary field experiments over an annual cycle between October 2000 and October 2001. Based on this analysis we first emphasize the mechanisms driving the large scale variability in the production system. This variability is shown to be almost entirely meridional, and is mostly driven by variations in the depth of the mixed layer. This analysis also allows us distinguish between three production regimes: the well known subpolar and subtropical regimes and the newly identified midlatitude regime. The identification of the midlatitude regime and its specific characteristics is of great importance as it overlaps the main area of mode water subduction and covers about half of the POMME area.

Then, our results show that the large scale variability is transferred into smaller scales through the process of direct horizontal cascade, associated with geostrophic stirring. The cascade mechanism leads to intrusions of chlorophyll rich structures (either eddies or filaments) from the north to the south. Another aspect of the cascade mechanism observed here, is the formation of chlorophyll anomalies within mesoscale eddies (both cyclonic and anti-cyclonic) through a process that involves water exchange between the interior of the eddy and its surrounding. In addition to that, our results suggest that phytoplankton variability is also driven by vertical injection of nutrients in regions of strong stirring.

Finally, we show that the mesoscale variability has an important impact on the large scale variability, through modulation of the annual cycle. This modulation results from mesoscale intrusion of nutrient rich waters.

We have developed three methods that are not commonly used in the field of ocean
color satellite imagery: i) interpretation of large-scale ocean color phytoplankton variability using ocean general circulation models; ii) analysis of transport properties of the velocity field and of their effect on the spatial distribution of chlorophyll with Lagrangian diagnostics of satellite derived geostrophic currents; iii) recovering of missing data in cloudy ocean color images using the kriging technique.
Résumé

Une meilleure connaissance des variations spatiales et temporelles de la biomasse phytoplanctonique dans les régions de subduction des eaux modales est fondamentale pour modéliser correctement le cycle océanique du carbone. En effet, ces variations sont à l’origine de modification des propriétés biogéochimiques des eaux de surface avant que celles-ci ne subductent vers l’océan intérieur, où elles sont alors isolées de l’influence de l’atmosphère à l’échelle de temps décennale.


Dans un premier temps nous décrivons les mécanismes à l’origine de la variabilité grande échelle du système de production. Nous montrons ainsi que cette variabilité est quasi-méridienne et résulte principalement des variations de la profondeur de la couche de mélange. Nous distinguons alors trois régimes de production : les régimes subtropicaux et subpolaires définis classiquement, et le régime des moyennes latitudes nouvellement introduit ici. L’identification de ce nouveau régime et de ses propriétés caractéristiques est fondamentale car il englobe la zone principale de subduction des eaux modales et inclut environ la moitié de la région POMME.

Par ailleurs, nos résultats mettent en évidence le transfert de variabilité des grandes échelles vers les petites échelles à travers des processus de cascade directe horizontale associés au mélange géostrophique. Le mécanisme de cascade entraîne un déplacement des structures riches en chlorophylle (tourbillons ou filaments) du nord vers le sud. Une autre propriété du processus de cascade observée ici est la formation d’anomalies de chlorophylle à l’intérieur des tourbillons mésoéchelles (cycloniques et anticycloniques), suite à un échange de matière (d’eau) entre l’intérieur du tourbillon et le milieu environnant. De plus, nos résultats suggèrent que la variabilité phytoplanctonique est contrôlée par l’injection verticale de nutriments dans les régions caractérisées par un mélange intense.

Enfin, nous montrons que la variabilité mésoéchelle a un impact important sur la variabilité grande échelle à travers la modulation du cycle saisonnier par l’intrusion de
structures mésoéchelles.

Nous avons ainsi développé trois méthodes originales, peu utilisées dans le domaine de
l’imagerie satellitale de la couleur de l’océan : i) L’interprétation de la variabilité grande
échelle de la couleur de l’océan et du phyto plankton à l’aide d’un modèle de circulation
générale de l’océan. ii) L’analyse des propriétés de transport du champ de vitesse et de
ses effets sur la distribution spatiale en chlorophylle, à l’aide de diagnostics lagrangiens
de courants géostrophiques dérivés de l’altimétrie satellitale. iii) Retrouver des données
manquantes dans des images "nuageuses" de couleur de l’océan en utilisant la technique
du "kriging".
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Introduction

In the northeast Atlantic, the spring phytoplankton bloom and the subduction of subpolar mode waters occur during the same period. Spatial and temporal variations in primary production associated with the bloom drive changes in the biochemical properties of the surface waters before they are subducted into the ocean interior, and are therefore of great importance for the oceanic carbon cycle.

Subduction mechanisms and processes responsible for the evolution of the biogeochemical properties of the mode waters were intensively investigated in the framework of the Programme Océan Multidisciplinaire Méso Échelle (POMME) research project [Mémery et al., 2005]. POMME included a series of interdisciplinary field experiments over an area that extends zonally from 16°W to 22°W and meridionally from 38°N to 45°N (Fig. 1). This area includes the area strong meridional gradient in winter mixed layer depth (MLD) between relatively deep winter mixed layers in the north and shallow mixed layers in the south, where subduction takes place. This area also corresponds to the transition zone between the productive region in the north and the oligotrophic region in the south.

In the framework of POMME four cruises were carried out (P0, P1, P2 and P3), covering different seasons over one annual cycle between October 2000 and October 2001. The two most extensive cruises were carried out from February through early May 2001 (P1 and P2), in order to study both late winter and early spring situations. Each of the cruises P1, P2 and P3 included two legs. POMME was conducted with strong emphasis on the role of mesoscale eddies, with a survey array at mesoscale resolution (typically 55 km) or higher. These mesoscale surveys, that included a simultaneous use of two vessels in several cruises within the same year, allowed the sampling of spatio-temporal scales ranging from several to several hundreds of kilometers and from days to seasons.

In terms of biological productivity, POMME observations showed a strong meridional gradient, with relatively high primary production rates in the north of the region and low production rates in the south. Primary production also showed a clear seasonal variability [Claustre et al., 2005]: there was a significant increase in primary production rates between the first leg of P1 (February) when phytoplankton growth was mainly limited by the availability of light [Blain et al., 2004] and the first leg of P2 (April) [Fernández et al., 2005b]. During P3 the nutrients were depleted from the ocean surface and there was a pronounced oligotrophic situation with low levels of biological activity.
Figure 1: Average SeaWiFS chlorophyll concentrations in the northeast Atlantic during the period 1998-2003. The frames mark the area of interest of this thesis (black lines) and of the POMME project (white lines).

Primary production was also influenced by the mesoscale dynamics associated with cyclonic and anticyclonic eddies [Claustre et al., 2005]. These rather stable [Assenbaum and Reverdin, 2005, Le Cann et al., 2005] mesoscale eddies were also found to have an important influence on the distribution of nitrate [Fernández et al., 2005a].

POMME observations have several limitations that stem from the difficulty of conducting field experiments in the ocean. Data gathered during POMME surveys are asynoptic [Lévy et al., 2005a]: it is difficult to distinguish between the spatial variability associated with the north-south gradient and temporal variability associated with the seasonal cycle. Another limitation in the non-continuity of the data in time: Although distinct variations in productivity from one season to the other are observed, accurate representation of production variability throughout the year is missing. Additionally, despite the relatively high spatial resolution of the POMME surveys, mesoscale and sub-mesoscale variabilities are only partially sampled [Lévy et al., 2005a]. Finally, the data are limited in their temporal and spatial extension. This inherently constrained the spatial and temporal scales of variability that can be studied and characterized.

The constraints described above limit the ability to characterize variations in the biogeochemical properties of the POMME region and to study the processes driving them. This rises the necessity for high resolution, synoptic and continuous observations that extend over a larger area (that includes important parts of the two contrasted regions) and over a longer period (that allows study of the interannual variabilities) than the area
and period covered by the POMME surveys. This is the objective of this thesis. More specifically we aim at:

i) Characterizing the phytoplankton variability at spatial and temporal scales that are either partially resolved or not resolved at all by the POMME observations.

ii) Identifying the dynamical mechanisms driving this variability.

The only feasible means for obtaining such extensive observations is to use satellite measurements of the ocean color, which provide a synoptic view of surface phytoplankton variations at spatio-temporal scales ranging from several to thousands of kilometers and from days to years [Yoder, 2000].

In this work we made use of such data to investigate phytoplankton variability over a period of several years, in a 20° latitudinal band that overlaps the area of the POMME experiment (Fig. 1). The characterization of spatial and temporal phytoplankton variabilities was done by analyzing ocean color data from the Sea-viewing Wide Field-of-view Sensor (SeaWiFS). The dynamical mechanisms driving these variabilities were investigated by comparison with dynamical fields derived from numerical models and from satellite measurements of sea level height (SSH) and sea surface temperature (SST).
This thesis covers a wide range of spatial and temporal scales (Fig. 2), thus providing a complementary data source to that obtained during the POMME surveys. We first investigated large scale variations by studying the phytoplankton bloom in terms of timing, intensity, intermittency, meridional extension and duration. These aspects are discussed with respect to spatial and temporal changes of the MLD and meteorological forcings. Secondly, we studied phytoplankton variabilities associated with the mesoscale dynamics, focusing on the impact of horizontal stirring on phytoplankton distribution at the time of the spring bloom. Finally, we investigated the coupling between the different scales of variability. This includes the formation of small scale patterns in phytoplankton distribution by cascade of the variability injected at the large scale and changes in the annual phytoplankton cycle that result from processes associated with the mesoscale dynamics.

The manuscript is organized as follows. In Chapter 1, we give a general overview on phytoplankton (Sec. 1.1) and on the northeast Atlantic (Sec. 1.2). In this chapter we also describe the general approach and the datasets that were used in this thesis (Sec. 1.3). The results are summarized in three articles that are the content of chapters 2-4. Each chapter starts with an extended abstract that includes a summary of the objectives, methods and main results. At the end of each chapter we conclude and propose some perspectives. In chapter 2, we give an overview on the physical and biogeochemical properties of the Northeast Atlantic from a climatological point of view. We then describe the large scale variations of the northeast Atlantic phytoplankton distribution. We identify spatial and temporal variations in the phytoplankton bloom and discuss how they are related to variations in the MLD. In chapter 3, we investigate the formation of small scale chlorophyll patterns by horizontal stirring associated with the mesoscale dynamics, by conducting a Lagrangian analysis of the geostrophic surface currents. In chapter 4, we apply the kriging interpolation method for reconstructing uninterrupted chlorophyll images that are then used for studying the impact of mesoscale dynamics on the annual chlorophyll cycle. In chapter 5 we conclude and propose some perspectives.
Chapter 1

Generalities

1.1 Phytoplankton

1.1.1 The role of phytoplankton in modulating the ocean’s biogeochemical properties

Phytoplankton are floating plants that inhabit the surface of the ocean. These autotrophic organisms produce organic compounds from $\text{CO}_2$ (as a source for carbon) and light (as a source of energy) through the process of photosynthesis. This primary production contributes roughly half of the world’s biological production and is the basis for the vast majority of marine life. Changes in patterns of phytoplankton distribution and abundance can have significant impact on the entire ecosystem.

The dependency on light restricts phytoplankton growth to the well lit layer close to the ocean’s surface. This layer is commonly referred to as the euphotic layer and is defined as the 1% incident light penetration depth. Aside from light, phytoplankton growth is crucially dependent on the availability of inorganic nutrients [Dugdale and Wilkerson, 1992] whose availability in the euphotic layer mostly depends on upwelling of deep, nutrient-rich waters [Williams and Follows, 2003]. Most of the phytoplankton are grazed locally and the nutrients comprising their biomass are regenerated at the surface and are available for another round of production. However, some fraction of the primary production finds its way to the deep sea, either through the settling of dead cells or through advection. Most of the organic carbon that finds its way to the deep sea is ultimately assimilated by bacteria, which regenerates it into inorganic forms. Phytoplankton growth, mortality and grazing by zooplankton are part of the so called biological pump [Longhurst, 1989], which is the sum of the biological processes that transport carbon from the ocean surface to its interior and therefore has an important role in the earth’s carbon cycle [Longhurst, 1991, Holligan, 1992]. Another important consequence of the biological pump is the creation of a sharp vertical gradient in the distribution of limiting nutrients, with minimum values near the ocean’s surface where they are consumed by phytoplankton and...
maximum values at depth where remineralization occur.

### 1.1.2 Phytoplankton variability

Phytoplankton abundance is mainly controlled by physical processes that determine the availability of light and nutrients. The different processes cover a wide range of temporal and spatial scales, from frontal processes that drive variations in primary production at length scales of several kilometers [Lévy, 2006] to climate changes that induce variabilities in the productivity of the entire world’s oceans [Behrenfeld et al., 2006]. The dynamical mechanisms and the consequent phytoplankton variabilities that were addressed in this thesis can be classified into three groups according to their typical time and length scales (Fig. 2): large scale, mesoscale and sub-mesoscale.

**Large scale variability**

We attribute the term ”Large scale” to seasonal and interannual variabilities with typical length scale of hundreds to thousands of kilometers. Variabilities at these scales are often driven by variations in the mixed layer depth (MLD) that determine the balance between the availability of the two limiting factors for phytoplankton growth: deep mixed layers lead to an extensive convective supply of nutrients to the euphotic layer but limit the amount of light experienced by the phytoplankton cells; Shallow mixed layers lead to an opposite scenario with the limiting factor being nutrient availability [Williams andFollows, 2003]. Large scale phytoplankton variabilities that are driven by changes in the MLD are the subject of chapter 2.

Production patterns at scales of thousands of kilometers are also strongly constrained by the wind driven circulation that consists of horizontal volume flux at the surface layer of the ocean, directed to the right of the wind stress in the northern hemisphere and to the left in the southern hemisphere. A horizontal divergence of this volume flux drives upwelling into the ocean’s surface layer whereas convergence drives downwelling. A general feature is that upwelling regions (e.g. subpolar gyres) are characterized by high productivity and regions of downwelling (e.g. subtropical gyres) are characterized by low productivity [Yoder et al., 1993].

**Mesoscale variability**

The term mesoscale refers to the scale of oceanic eddies. It covers the range of several tens to several hundreds of kilometers, and it is strongly correlated to the Rossby radius of deformation [Stammer, 1998] which is the length scale at which rotation effects become as important as buoyancy effects. The Rossby radius of deformation decreases from 300 km at the equator to a few kilometers at high latitudes, and ranges between 30 and 50 km at mid latitudes [Chelton et al., 1998].
Mesoscale circulation can influence productivity patterns because it is associated with vertical motions that transfer nutrients into the euphotic layer. In addition to this local effect of nutrient supply to the euphotic layer, eddies systematically transfer nutrients laterally. An example is the transport of nutrients by coherent eddies that are isolated from their surroundings, thus preventing inward and outward particle fluxes [Provenzale, 1999]. The displacement of such coherent eddies can transport tracers along hundreds of kilometers [Richardson, 1993]. An alternative mechanism that involves water exchange between the eddy’s interior and its surroundings due to the chaotic nature of the velocity field is proposed in chapter 3 of this manuscript.

Sub-mesoscale variability

Sub-mesoscale features include fronts and filaments between mesoscale eddies, and are characterized by spatial scales of few to several tens of kilometers. As shown by recent modeling studies, vertical motions within sub-mesoscale filaments surroundings the eddies can transport significant amount of nutrients into the euphotic layer [Lévy et al., 2001] thus contributing significantly to primary production. In this study, increase in model resolution to levels that allow representation of sub-mesoscale features leads to an increase of up to 50% in primary production.

Phytoplankton distribution at the sub-mesoscale is also modulated by lateral stirring and mixing [Martin, 2003]. Horizontal stirring leads to the production of smaller and smaller patterns. This mechanism of direct cascade was evidenced in a 2D numerical experiment conducted by Abraham [1998]. Observational evidence for the formation of sub-mesoscale patterns by horizontal stirring are given in chapter 3 of this manuscript.

1.1.3 Detecting phytoplankton from space

Phytoplankton cells contain chlorophyll, a green pigment that absorbs sunlight. Due to this green pigment, which is in fact the most important light-absorbing substance in the oceans, phytoplankton preferentially absorb the red and blue portions of the light spectrum and reflect green light (Fig. 1.1). Consequently, regions with high phytoplankton concentrations will appear as certain shades, from blue-green to green, depending upon the type and density of the phytoplankton population there.

A common and relatively simple approach for retrieving surface chlorophyll concentrations from water leaving radiance is based on the ratio between the radiance measured by the green and blue channels of the spectrometer, centered respectively at around 443nm and 550nm (the "blue-green" ratio). In open ocean waters, where optical properties are mainly controlled by phytoplankton (often referred to as Morel case 1 waters), this relation yields an estimate of near surface chlorophyll with an accuracy of about 30% [Gordon and Morel, 1981, 1983].
Figure 1.1: Spectral curves of surface reflectance for seawaters containing different quantities of chlorophyll. Increase in chlorophyll concentrations leads to a higher green to blue (centered at around 550 and 443 nm, respectively) reflectance ratio. The measurements were performed during June 2000 in Haifa Bay, Israel [Lehahn, 2002]. Note that the high chlorophyll concentrations presented in this figure are not typical to the area studied in this thesis.

Surface chlorophyll concentrations are used to estimate phytoplankton biomass and productivity throughout the euphotic zone [Morel and Berthon, 1989]. However, several factors limit the retrieval of these variables from surface chlorophyll measurements. An important factor is the vertical variability of phytoplankton biomass: since surface chlorophyll concentrations are representative of pigment concentrations within the mixed layer [André, 1992], they do not allow to detect production that takes place below this layer. A typical case for such deep production is the deep chlorophyll maximum that is a consistent oceanographic feature of tropical and subtropical oceans [Pérez et al., 2006]. Another limitation stems from the fact that the ratio between chlorophyll concentration and biological production varies between different phytoplankton species and is strongly dependent on the environmental conditions [Geider, 1987].

The optical properties of phytoplankton allow their detection by ocean color satellites that measure sunlight radiance backscattered out of the upper few meters of the ocean (water leaving radiance). Ocean color satellite imagery provides high quality images of the ocean biosphere on local to global scales and for time scales ranging from days to years [Yoder, 2000]. The use of ocean color satellites for the study of biological processes began in the late 1970s with the coastal zone color scanner (CZCS) aboard the Nimbus 7 satellite that acquired data from October 1978 to June 1986 [Evans and Gordon, 1994]. SeaWiFS, that was launched on August 1997 on board the OrbView-2, is the NASA successor to the CZCS [Siegel et al., 2004]. The SeaWiFS instrument became operational on September 1997 and routinely provides global coverage every 2 days [Hooker and McClain, 2000]. It is the second in a sequence of ocean color missions that began with the Japanese Advanced Earth Observation Satellite (ADEOS), which operated only from August 1996 to June 1997 and carried the Ocean Color and Temperature Scanner (OCTS) and the Polarization and Directionality of the Earth’s Reflectance (POLDER) instrument. Other ocean color satellite missions include the American Moderate Resolution Imaging Spectroradiometer
(MODIS), the European Medium Resolution Imaging Spectrometer (MERIS), the Indian Ocean Colour Sensor (OCM) and the Chinese Fenyeng 1c (FY-IC).

1.2 Regional context

1.2.1 Physical and biogeochemical characteristics of the northeast Atlantic

The northeast Atlantic is characterized by a sharp transition between an area of deep winter MLD in the north and relatively shallow MLD in the south [Arhan et al., 1994, Paillet and Arhan, 1996a, de Boyer-Montégut et al., 2004]. The strongest meridional gradient in winter MLD is approximately aligned with the line of zero annual net heat flux at around 43°N. Surface winds are generally directed eastward and rotate from a northward orientation at around 50°N to a southward orientation at 35°N. The resulting Ekman transport is southward and associated with upwelling in northern subpolar gyre and downwelling in the subtropical gyre [Williams and Follows, 1998, Williams et al., 2000].

These physical conditions drive a strong north-south contrast in the distribution of nutrients and divide the northeast Atlantic into two production regimes: a relatively productive regime in the north of the basin and an oligotrophic regime in the south (Fig. 1). In the north, deep winter mixed layers allow high rates of convective supply of nutrients to the euphotic layer [Williams et al., 2000]. This, together with the upwelling of nutrient rich water and the raising of the thermocline by the gyre circulation, enables an efficient phytoplankton growth. In the south, the opposite scenario of downwelling and a shallow mixed layer is found and the region is oligotrophic [Williams and Follows, 2003].

The study area is characterized by relatively low eddy kinetic energy (EKE) [Stammer et al., 2006]. Nevertheless, the large scale circulation is dominated by cyclonic and anti cyclonic mesoscale eddies [Assenbaum and Reverdin, 2005, Le Cann et al., 2005], with a typical size of several hundreds of kilometers. The mean surface circulation is directed southward [Paillet and Mercier, 1997].

1.2.2 The dual consequence of mixed layer shallowing: mode water subduction and phytoplankton bloom

The combination of the winter MLD gradient with the southward circulation leads to the subduction of 11°-12°C subpolar mode waters [Paillet and Arhan, 1996a,b]. Subduction, which involves the transfer of fluid from the mixed layer in to the stratified thermocline [Williams, 2001], is maximum over the area of maximum MLD gradient. It occurs at
the end of winter when the surface warming leads shallowing of the mixed mixed layer [Hazeleger and Drijfhout, 2000, Marshall et al., 1993, Valdivieso da Costa et al., 2005].

Another process associated with the seasonal restratification of the water column is the well known north Atlantic spring bloom [Siegel et al., 2002], which is characterized by a pronounced increase in phytoplankton biomass initiated when the mixed layer becomes shallower than Sverdrup’s critical depth [Sverdrup, 1953].

Because the phytoplankton bloom and subduction occur during the same period, spatial and temporal variabilities in primary production during this period play a major role in setting the biogeochemical properties of the water masses that will be isolated for a decade from the influence of the atmosphere, and therefore on the efficiency of the physical and biological pumps.

At large scales, the biogeochemical properties of subducted waters are strongly dependent on the bloom intensity and meridional extension and on the exact timing of the bloom with respect to the timing of subduction. In addition to that, the water masses properties are modified by mesoscale and sub-mesoscale dynamics, which are responsible for large variations in tracer distribution and production budgets [Oschlies and Garcon, 1998, McGillicuddy et al., 1998, Mahadevan and Archer, 2000, Lévy et al., 2001].

1.3 General approach and datasets

Our approach is to characterize the apparent spatial and temporal variability of satellite derived chlorophyll. We than relate the patterns of variability to those of physical variables derived from satellites and from numerical models. A summary of the datasets is given in table 1.1.

Chlorophyll

Estimations of chlorophyll concentrations were derived from SeaWiFS data obtained from the NASA Goddard Distribution Active Archive Center (DAAC). We used two data sets that differ in their spatial and temporal resolution and in their spatial coverage. For the study of the large scale variability (chapter 2) we used 5 years (1998-2002) of level 3 (9km). The images are composites of 8 consecutive days. The study of small scale variabilities (chapters 3 and 4) was done with high resolution (1km) daily images from 1998 to 2003. The retrieval of chlorophyll concentrations from the level 2 data was done by using NASA’s SeaWiFS Data Analysis System (SeaDAS) software package.

Geostrophic currents

Geostrophic currents (chapter 3) were calculated from satellite derived SSH, which is the sum of sea level anomalies (SLA) + mean dynamic topography. SLA were obtained from
Table 1.1: Datasets used in this thesis.

<table>
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<td>9 km</td>
<td>1998-2002</td>
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<td>1 km</td>
<td>1998-2003</td>
</tr>
<tr>
<td>Sea surface temperature</td>
<td>AVHRR</td>
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<td>1 km</td>
<td>October 2000-October 2001</td>
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<td>1/6°</td>
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<td>5</td>
<td>1/2°</td>
<td>1998-2002</td>
</tr>
<tr>
<td>Mixed layer depth</td>
<td>MERCATOR</td>
<td>1</td>
<td>1/3°</td>
<td>1998-2002</td>
</tr>
<tr>
<td>Wind stress</td>
<td>ECMWF</td>
<td>1</td>
<td>1/6°</td>
<td>1998-2002</td>
</tr>
<tr>
<td>Net heat flux</td>
<td>Clipper</td>
<td>5</td>
<td>1/6°</td>
<td>1998-2002</td>
</tr>
</tbody>
</table>

* from December 2001
** from March 2002

the AVISO database (http://www.aviso.cnes.fr). The distributed global product combines altimetric data from the Topex/Poseidon and ERS missions, from Jason-1 for data after December 2001 and from Envisat for data after March 2002. The spatial resolution is 1/3° with one data file every 7 days. The sea level anomalies were added to the Rio-05 mean dynamic topography [Rio and Hernández, 2004]. Currents were computed from the SSH measurements by solving the equation of geostrophic equilibrium.

**Sea surface temperature**

We used high resolution (1km) daily AVHRR data that were preprocessed towards SST by Meteo France in Lannion. These data were only available for the period of the POMME experiment (October 2000- October 2001).

**Mixed layer depth and net heat flux**

The depth of the mixed layer and net the heat fluxes for the years 1998-2002 (chapter 2) were derived from the ATL6 Clipper simulation of the North Atlantic [Tréguier et al., 2003]. The model was forced with daily ECMWF analysis. The heat forcing is corrected by applying a relaxation of SST towards Reynolds mixed satellite/in-situ product for bulk surface temperature, so that the net heat flux is equal to of the sum of the fluxes from the ECMWF analysis and the correction term. MLD is diagnosed as the depth at which the density differs from its surface value by 0.05 kg/m³. The spatial resolution is 1/6° and the outputs are 5 days averages. In order to evaluate the accuracy of the MLD estimated by the Clipper model, it was compared with MLDs from 3 other models: a high resolution (1/20°) regional model run from February to October 2001 [Lévy et al., 2005a]; the
1/2° resolution global configuration of OPA (http://locean-ipsl.upmc.fr/opa) and the 1/3° resolution operational model of the North Atlantic MERCATOR (http://www.mercator-ocean.fr). The results of this comparison are discussed in the text (chapter 2).

**Wind stress**

Daily wind stress data for the period 1998-2002 (chapter 2) were retrieved from the European Center for Medium range Weather Forecasting (ECMWF). The spatial resolution is 1/6°.
Chapter 2

Climatology and large scale variations in productivity

2.1 Extended abstract

Our analysis of the large scale phytoplankton variability is described in the Journal of Geophysical Research (JGR) paper: ”Production regimes in the northeast Atlantic: a study based on Sea-viewing Wide Field-of view Sensor (SeaWiFS) chlorophyll and ocean general circulation model mixed layer depth” [Lévy et al., 2005b, hereafter LLA05]. The main objective of this paper was to characterize several aspects of the large scale variability of the phytoplankton bloom and to study the way these aspects are related to changes in MLD. In addition to that, we give an overview on the physical and biogeochemical properties of the northeast Atlantic from a climatological point of view, and a detailed description of the POMME research project.

Phytoplankton variability patterns were characterized by using low resolution (9 km, 8 days composites) SeaWiFS chlorophyll images, between the years 1998 and 2002. Mechanisms driving these variabilities were studied by comparing the SeaWiFS data with MLD from the Clipper ocean circulation model of the North Atlantic. The impact of atmospheric forcings was explored by comparison with Clipper heat fluxes and with ECMWF wind data.

2.1.1 Main results

The large scale spatial variability is almost entirely meridional. Based on the meridional differences in the seasonal chlorophyll cycling and its relation to the seasonal cycling of MLD, we distinguish between three production regimes (LLA05, Fig. 3): the well known subpolar and subtropical regimes (in the northern and in the southern parts of the study area, respectively) and a newly identified midlatitude regime.

In the subpolar regime, the main features in the chlorophyll annual cycle are a major
bloom during spring and a secondary bloom in fall. Comparison between the evolution of chlorophyll and MLD shows that these blooms correspond respectively to mixed layer shallowing and deepening. During winter, production is inhibited by deep mixing that limits the availability of light. The springtime shallowing of the mixed layer (associated with a change in net heat flux from negative to positive) increases dramatically the amount of light experienced by the phytoplankton cells and leads to a rapid increase in phytoplankton biomass. During summer, the mixed layer is shallow and production is nutrient limited. The fall time mixed layer deepening increases nutrient availability and leads to the secondary fall bloom.

In the subtropical regime, the mixed layer is relatively shallow and phytoplankton growth is limited by the availability of nutrients. Productivity is maximal in winter and the chlorophyll signal is characterized by a single weak bloom that starts with the deepening the mixed layer during fall.

In the midlatitude regime, the seasonal production cycling and its relation to the evolution of the MLD differs remarkably from those of the subpolar and subtropical regimes. The chlorophyll signal is characterized by a single broad bloom, weaker than the subpolar spring bloom and stronger than the subtropical fall bloom. The midlatitude bloom begins in fall with the deepening of the mixed layer (as the subtropical bloom and subpolar secondary bloom) and keeps developing in winter, reaching its peak after the springtime restratification of the water column (as the subpolar spring bloom). The annual production cycling in the midlatitude regime is characterized by the merging of a fall bloom with a spring bloom and it is limited by the availability of both light and nutrients.

The identification of the midlatitude regime is an important finding of this work. Because it overlaps the area of maximum gradient in winter MLD, which is the main area of mode water subduction, the specific characteristics of the midlatitude regime are of importance to the ocean carbon cycle. Comparison between the three production regimes (LLA05, Fig. 9) shows that winter production is optimum in the midlatitude regime (since in the subpolar regime winter production is weak due to light limitation and in the subtropical regime it is weak due to nutrient limitation). This, together with the fact the midlatitude bloom starts prior to subduction and that production is active during winter is likely to reinforce the efficiency of the biological carbon pump during subduction.

In addition to that, the specific properties of the midlatitude regime may explain some unexpected characteristics of the POMME in situ observations. This is due to the fact that about half of the POMME area is included in this regime. An example for such a characteristic is the fact that in situ measurements of chlorophyll concentrations never exceed 1 mg/m$^3$, while higher concentrations were expected (for example during the North Atlantic Bloom Experiment (NABE) observed concentration during the bloom were 3-4 mg/m$^3$). Another unexpected result is the fact that winter is a period of significant
CHAPTER 2. CLIMATOLOGY AND LARGE SCALE VARIATIONS IN
PRODUCTIVITY

production, not much different than spring.

The latitudinal boundaries between the regimes were objectively identified for each year by applying objective criteria, based on the evolution of chlorophyll (for the boundary between the midlatitude and subpolar regimes) and on the timing of maximum MLD versus the timing of maximum chlorophyll (for the boundary between the midlatitude and subtropical regime). One of the main results is that the location of the boundaries between the production regimes shows interannual variability of several degrees (LLA05, Fig. 4). These variations are mainly attributed to interannual variations in wind stress that force meridional shifts in the locations of the winter MLD front and of the intergyre boundary between the subpolar gyre in the north and the subtropical gyre in the south (LLA05, Fig. 7).

Another result is that the strength of the subpolar spring bloom and of the subtropical bloom show interannual variations that we were able to relate to interannual variations in the MLD (LLA05, Fig. 6). In fact the two blooms are shown to be oppositely correlated with the depth of the mixed layer: whereas in the subpolar regime deep mixed layer (i.e. enhanced mixing) results in weaker bloom, in the subtropical regime deeper mixed layer intensifies the bloom. This opposite relationship between the bloom’s intensity and MLD results from the difference in the bloom’s limiting factor. In the subpolar regime the bloom is light limited so enhanced mixing decreases the amount of light experienced by the phytoplankton cells. In contrast, in the subtropical regime where the bloom is nutrient limited it is intensified by increasing nutrient input. The intensity of the midlatitude bloom shows no correlation with the depth of the mixed layer. This is explained by the fact that the production in the midlatitude regime is limited by the availability of both light and nutrient.

Our analysis shows that the timing of the bloom’s onset changes from one year to another according to timing of seasonal changes in the MLD. The subpolar restratification bloom starts when the mixed layer shallows to a depth less than 180m. This depth, that corresponds to twice the depth of the euphotic layer, has been determined empirically. The entrainment bloom (in the subtropical and mid-latitude regimes) starts when the mixed layer deepens below the depth of the nitracline (derived from Levitus nitrate climatology as the depth of maximum vertical gradient from the surface).

These mixed layer criteria for the triggering the blooms also explain the observed northward propagation of the subpolar spring bloom and southward propagation of the entrainment bloom (LLA05, Fig. 5). The propagation of the spring bloom is associated with the latitudinal change in the solar flux that determines the amount of light available to photosynthesis and partially forces mixed layer shoaling through its contribution to the net heat flux. The southward propagation of the entrainment bloom is mainly driven by the meridional gradient in the depth of the nitracline, which is shallower in the north than in the south.
Production regimes in the northeast Atlantic: A study based on Sea-viewing Wide Field-of-view Sensor (SeaWiFS) chlorophyll and ocean general circulation model mixed layer depth

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Received 15 October 2004; revised 31 January 2005; accepted 6 April 2005; published 20 July 2005.

[1] A 5 year time series of Sea-viewing Wide Field-of-View Sensor (SeaWiFS) ocean color images (SCHL) is compared with mixed layer depths (MLD) and atmospheric forcings from the Clipper model of the North Atlantic (1998–2002). This comparison is done over the region 16°–22°W, 30°–50°N, where subpolar mode waters are formed and which overlaps the region of the 2001 Programme Océan Multidisciplinaire Méso Echelle (POMME) experiment at sea. Three production regimes are identified on the basis of the seasonal cycling of SCHL and MLD: the well-known subpolar and subtropical regimes and a midlatitude regime. The midlatitude regime is characterized by a single broad bloom weaker than the subpolar spring bloom and stronger than the subtropical fall bloom, which starts in fall as an entrainment bloom and peaks in spring as a restratification bloom. This specific regime is found between 35°N and 40°N (±2°) in the northeast Atlantic. It corresponds to winter MLDs between Ze (the depth of the euphotic layer) and 2Ze, i.e., it lays between the region where the winter MLD is greater than Sverdrup’s critical depth (subpolar regime) and the region where the mixing is never deeper than the well-lit layer (subtropical regime). The very specific characteristics of the midlatitude regime strengthen the biological carbon pump since production is active in winter within the waters to be subducted. The midlatitude regime also may provide an explanation for the unexpectedly low f ratios sometimes observed during the bloom in the region (North Atlantic Bloom Experiment, POMME). A large interannual variability is observed for the three regimes in terms of the timing and the intensity of the blooms and of the geographical boundaries of the regimes. These variabilities appear to be mainly driven by the synoptic and the low-frequency atmospheric variabilities. It is also shown that in addition to the northward propagation of the subpolar spring bloom from 41°N (±1.3°) to 50°N, the (fall) entrainment bloom propagates southward over the whole latitudinal range (35°–50°N).


1. Introduction

[2] In the northeast Atlantic, the spring bloom and the formation of subpolar mode waters occur during the same period. Subpolar mode waters, characterized by a low potential vorticity and a temperature between 11° and 12°C, are subducted at the end of winter [Woods, 1985]. The intensity of the bloom and the exact timings of bloom and subduction play major roles in setting the biogeochemical characteristics of the water masses that will be isolated for a decade from the influence of the atmosphere, and therefore on the efficiency of the physical and biological carbon pumps. The Programme Océan Multidisciplinaire Méso Echelle (POMME) research project was designed to achieve a better understanding of the coupling between the dynamical processes involved in subduction and the biological processes involved in the oceanic carbon cycle [Mémery et al., 2005]. Field experiments were carried out over a seasonal cycle, from October 2000 to October 2001. The area of interest extends zonally from 16°W to 22°W and meridionally from 38°N to 45°N (Figure 1). This area

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0148-0227/05/2004JC002771S09.00
CHAPTER 2. CLIMATOLOGY AND LARGE SCALE VARIATIONS IN PRODUCTIVITY

Climatology

a) MLD (m) and line of zero annual heat flux

b) Wind Stress (W/m²)

c) Surface currents (cm/sec)

d) Cloud cover (%)

e) Surface Nitrate (mol/L)

f) SCHL (mg/m³)

Figure 1

2 of 16
corresponds to a transition zone between a productive region in the north, associated with quite deep late winter mixed layers (200–300 m), and an oligotrophic region in the south, associated with relatively shallow (100 m) winter mixed layers [Sathyendranath et al., 1995]. Both regions are characterized by a pronounced seasonal cycle, and an important year to year variability [Stramska et al., 1995; Michaels and Knapp, 1996; Williams et al., 2000; Dutkiewicz et al., 2001; Follows and Dutkiewicz, 2002; McClain et al., 2004; Dandonneau et al., 2004]. Estimates of surface ocean chlorophyll (SCHL) from space provide an opportunity to reveal this variability more clearly on a synoptic scale over several years.

This study focuses on SCHL as detected by SeaWiFS over the region 16°–22°W, 30°–50°N from 1998 to 2002, where subpolar mode waters are formed and which overlaps the region of the 2001 POMME experiment. It is aimed at explaining aspects of the seasonal cyclings and interannual changes in terms of mixed layer depth (MLD) behavior and meteorological forcing. This analysis is conducted with particular emphasis on the timing of the spring bloom and of subduction. Variability of the bloom in terms of timing, intensity, intermittency, duration and propagation is discussed with respect to three specific biogeochemical regimes, which will be identified in the data set: the well-known subpolar and subtropical regimes and a midlatitude regime. The study insists on the midlatitude regime (found in the Sargasso Sea) and the subpolar and subtropical regimes. Besides, the POMME region overlays the flanks of the subtropical gyre where Williams andFollows [1998] modeling study suggests that productivity is controlled by an Ekman injection of nitrogen from the north. However, by combining climatological nitrate profiles with MLD cycles and wind stress data, Williams et al. [2000] found that the interannual variability in nitrate supply is largely controlled by convection rather than by Ekman transfer and that the largest interannual variability in convective nitrate supplies occurs in the POMME area. Bograd et al. [2004] used SeaWiFS data to study the annual migration of the chlorophyll front in the transition zone of the north Pacific. They found significant interannual variations in the front location which they attributed to wind changes through a combination of two processes: changes in the wind stress curl which influences the redistribution of nutrients in the surface waters through Ekman horizontal divergence, and changes in the wind intensity which acts to mix nutrients up to the surface. The position of the boundary between the eutrophic regime in the north and the oligotrophic regime in the south of the POMME region is very likely to be affected by the same processes.

Our approach is guided by previous studies on SCHL variability in the region. Phytoplankton growth involves a balance between nutrient supply and light forcing. The deepening of the oceanic mixed layer (ML) supplies nutrients while decreasing the averaged light level experienced by the phytoplankton population. The interplay between these two competing effects has been rationalized by Dutkiewicz et al. [2001]. They identify two regimes based on the ratio of the spring critical depth to the ML at the end of winter: anomalously high spring mixing decreases SCHL in the subpolar regime, and enhances SCHL in the subtropical regime. This classification is based on Sverdrup’s [1953] critical depth theory and on Menzel and Ryther [1961] observations that in the Sargasso Sea the chlorophyll generally occurs when the ML is at its deepest. By applying Dutkiewicz et al. [2001] concepts to SeaWiFS time series observations binned into 5° × 5° regions in conjunction with reanalyzed NCEP meteorological data, Follows and Dutkiewicz [2002] show that the expected SCHL-to-mixing relationships emerge in both regimes. In this low-resolution study, no clear trend emerges in the POMME region which appears as a transition between the two regimes. Besides, the POMME region overlays the flanks of the subtropical gyre where Williams and Follows [1998] modeling study suggests that productivity is controlled by an Ekman injection of nitrogen from the north. However, by combining climatological nitrate profiles with MLD cycles and wind stress data, Williams et al. [2000] found that the interannual variability in nitrate supply is largely controlled by convection rather than by Ekman transfer and that the largest interannual variability in convective nitrate supplies occurs in the POMME area. Bograd et al. [2004] used SeaWiFS data to study the annual migration of the chlorophyll front in the transition zone of the north Pacific. They found significant interannual variations in the front location which they attributed to wind changes through a combination of two processes: changes in the wind stress curl which influences the redistribution of nutrients in the surface waters through Ekman horizontal divergence, and changes in the wind intensity which acts to mix nutrients up to the surface. The position of the boundary between the eutrophic regime in the north and the oligotrophic regime in the south of the POMME region is very likely to be affected by the same processes.

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The analysis of the spatio-temporal variability is based on time series of satellite images and model outputs that are now described.

2. Satellite-Derived Chlorophyll

SCHL estimates are derived from SeaWiFS. They are representative of the phytoplankton pigment concentration within the mixed layer [e.g., André, 1992]. We used the Level 3 (9 km) weekly composites from the Distributed Active Archive Center of the Goddard Earth Sciences (GES-DAAC) at NASA after reprocessing #4. The data set corresponds to the period from January 1998 to December 2002. The accuracy of satellite retrieval over the POMME region was estimated by comparing satellite estimates of SCHL with coincident (in time and space) in

Figure 1. A climatological view of the northeast Atlantic. (a) Winter mixed layer depth (MLD) (after de Boyer Montégut et al. [2004]) and the line of zero annual net heat flux in Clipper between the years 1998 and 2002. (b) Wind stress (European Centre for Medium-Range Weather Forecasts (ECMWF)) and line of zero wind stress curl. (c) Surface currents (after Paillet and Mercier [1997]). (d) Cloud cover as percent of flag occurrence over 1998–2002 in Sea-viewing Wide Field-of-view Sensor (SeaWiFS) weekly chlorophyll products (see text) and the line of 50% cloud cover. (e) Surface nitrate [Levitus, 1982]. (f) Surface chlorophyll (average of SeaWiFS ocean color images (SCHL) over 1998–2002). The solid frame marks the Programme Ocean Multidisciplinaire Méso Echelle (POMME) region, and the dashed frame marks the region examined in the present study. See color version of this figure at back of this issue.
situ measurements during POMME. The estimated ratio of satellite-to-in-situ SCHL is 1.06 ± 0.28%, i.e., no significant bias and a small standard deviation (more details will be available in the work of H. Llois et al. (manuscript in preparation, 2005)). Level 3 weekly composites are significantly cloudy for the north Atlantic in winter. An average cloud cover was computed as the percentage of masking occurrence for each pixel over the 5 years. Very high SCHL were often observed in the vicinity of clouds. After examination of in situ regional measurements, values greater than 5 mg/m³ were considered as unrealistic and were filtered out.

2.2. Model Products

The evolutions of the MLD and of the net heat flux are derived from the ATL6 Clipper simulation of the North Atlantic (described in detail by Treguier et al. [2003]). This simulation was performed with the primitive equation model OPA [Madec et al., 1999] in a domain that covers the Atlantic Ocean from 75°S to 70°N with a horizontal resolution of 1/6°. Vertical resolution is 12 m at the surface and decreases to 200 m below 1550 m. The model is forced with the daily ECMWF analysis from 1979 to 2002. The heat forcing is corrected by applying a relaxation of SST toward Reynolds satellite data. In this paper, the net heat flux is defined as the sum of the ECMWF analysis heat flux and of this correction term. Wind stress is from ECMWF analysis. Clipper outputs are 5 day averages.

MLD is diagnosed as the depth at which the density differs from its surface value by 0.05 kg/m³. In order to evaluate the quality of the Clipper MLDs, they have been compared to the MLDs predicted by three other models and with POMME data. The results of this comparison will be discussed in section 4.

2.3. Analysis of the Variability

Our analysis of the SCHL variability relies on comparisons with the MLD. We examine three scales of variability. We start with an overview of the main large-scale climatological patterns over the northeast Atlantic (5°–35°W, 30°–50°N). This meridional extension covers significant parts of the subpolar oligotrophic region and of the northern productive region, as well as the transition region in between. In a second stage, we examine the seasonal cycling of SCHL and MLD from September 2000 to October 2001 (the period of the POMME experiment) and restrict our analysis to the zonal extent of the SCHL region (16°–22°W). When computing zonal averages, this range is large enough to counterbalance the relatively large cloud cover. In a third stage, we analyze the interannual variations of the coupled cycles between 1998 and 2002.

In order to derive the bloom onset and the bloom duration, a low-pass time filter is applied to the satellite time series. This allows to retain only the seasonal signal and to remove sources of variability due to higher frequencies. The bloom onset is an important parameter whose determination has been designed according to the bloom dynamics. In particular, two dynamics of bloom onset were observed (see later) which called for specific criteria. For sharp SCHL rising following fast ML retreats (spring blooms), the onset is defined as the time of SCHL maximum derivative (maximum growth). For mild SCHL increases driven by slow ML deepenings (entrainment blooms), the onset is defined as the time of SCHL zero derivative (beginning of growth). In both cases, the end of the bloom is defined as the time when the SCHL comes back to its onset value. The propagation speed of the bloom onset is computed as the time derivative of its location, averaged over the latitude range of the propagation. Finally, the bloom strength is defined as the average SCHL over the duration of the bloom.

3. Results

3.1. Climatological Patterns

A central feature of the northeast Atlantic is the sharp transition between the areas of deep winter MLDs in the north, and shallower MLDs in the south [Arhan et al., 1994; Paillet and Arhan, 1996a]. Figure 1a shows how clearly this emerges in the recent climatology of de Boyer Montégut et al. [2004]; in February, MLDs reach 220m in the north and 100m in the south. The strongest winter MLD gradient (located around 43°N) is aligned with the line of zero annual net heat flux. This gradient is mostly meridional, although the front is slightly inclined from southwest to northeast due to the presence of the North Atlantic Current (NAC) in the northwest (Figure 1c). Besides the NAC, the geostrophic circulation at the surface, as reconstructed from an inverse model by Paillet and Mercier [1997], reveals a strong eastward current farther south (around 33°N), known as the Azores Current (AC). Two branches of southward recirculation appear around 30°W and 22°W. These branches are connected with the AC in the south. Surface winds are generally directed eastward (Figure 1b), and rotate from northeastward at 50°N to southeastward at 35°N. The zero wind stress curl is located at around 45°N and slightly inclined (Figure 1b). The resulting Ekman transport is directed southward and associated with upwelling in the northern subpolar gyre and downwelling in the southern subtropical gyre [e.g., Williams and Follows, 1998; Williams et al., 2000]. It is the combination of the winter meridional MLD gradient with the weak southward circulation which leads to the subduction of subpolar mode water [Paillet and Arhan, 1996a, 1996b]. Subduction occurs when mixed layer retreats and it is maximum over the area of maximum MLD gradient [Marshall et al., 1993; Hazeleger and Drijfhout, 2000; Valdivieso da Costa et al., 2005]. In the SeaWiFS climatology (Figure 1f), the two well-known productivity regimes of the north Atlantic appear. The north is productive with high mean values of SCHL (>0.4 mg/m³), and the south is oligotrophic with low mean SCHL values (<0.1 mg/m³). This spatial structure is the result of the meridional variations in the distribution of nutrients (Figure 1e) resulting itself from the physical forcings structure (see above). In the north, deep winter mixed layers allow high rates of convective supply of nutrients to the euphotic layer [Williams et al., 2000]. This, together with the upwelling of nutrient rich water and the raising of the thermocline by the gyre circulation enables an efficient phytoplankton growth and the region is relatively productive. In the south, the opposite scenario of downwelling from a shallow mixed layer is found and the region is oligotrophic [Williams and Follows, 2003]. Cloud cover increases with latitude, from 10% south of 35°N to more...
than 60% north of 50°N (Figure 1d). In the region of the POMME experiment, the cloud cover ranges between 20% and 40%.

[14] To sum up, the POMME region is characterized by relatively weak currents, small zonal variability and strong meridional gradients. The area is split by the lines of zero annual net heat flux, zero wind stress curl, maximum MLD and SCHL gradients.

3.2. Seasonal Cycling (September 2000–October 2001, the Year of the POMME Experiment)

[15] The period from September 2000 to October 2001 is used to identify the main features of the seasonal cycling. Besides their seasonal variations, all parameters exhibit high-frequency signals. The timing and intensity of the high-frequency events vary from one year to another. They will be specifically described for the POMME year.

3.2.1. Production Regimes

[16] The zonal character of most features in the climatologies (Figure 1) drives to focus on meridional and temporal variations. Zonal averages, as presented in Figure 2, reveal the expected strong seasonality of all fields.

[17] The net heat flux (Qnet) switches from negative to positive values around March, and becomes negative again during September (Figure 2a). The timing of the zero Qnet varies with latitude: warming occurs earlier in the south (in spring) and cooling occurs earlier in the north (in fall), both by about 1 month between 30°N and 50°N. Qnet is also strongly modulated by the occurrence of wind bursts, which are frequent in autumn and winter, particularly north of 40°N (Figure 2b). In the northeast Atlantic, the ML variability is essentially driven by Qnet. When Qnet is positive, the ocean warms up, the water column stratifies and the ML shallows. When Qnet is negative, static instabilities are generated and convection deepens the ML. Over the whole latitude range examined, the ML is thus forced to deepen in fall (earlier in the north) and to shoal in spring (earlier in the south (Figure 2c)). Vertical mixing generated by strong wind events drive short-term ML variability that can prevail over a stabilizing effect of Qnet; an example is the ML deepening event at the end of P2L2 (strong wind and positive Qnet at the beginning of May).

[18] The seasonality of SCHL is driven by the seasonality of the atmospheric forcing, in particular through the solar flux and the MLD variations. The seasonal cycling of SCHL changes with latitude (Figure 2d). This meridional change results from much deeper winter MLs in the north than in the south (300 m versus 120 m, as predicted by the Clipper model).

[19] In the north (north of 40°N on the 2001 picture), the main features are a major bloom in spring ("subpolar spring bloom") and a relatively small bloom in autumn (thereafter "subpolar fall bloom"). In summer. During the northern winter, production is inhibited by too deep a mixing and the subpolar spring bloom is triggered by the restratification of the water column. As it appears in Figure 2, the period of highest biomass closely follows the period of deepest ML.

[20] In the south (south of 35°N on the 2001 picture), the SCHL signal is characterized by a single weak bloom that starts in the fall and reaches its peak in February. As the fall bloom of the subpolar regime, this "subtropical (fall) bloom" is an entrainment bloom [Michaels et al., 1996], initiated by a deepening of the mixed layer. In contrast to what happens in the northern winter, the southern ML does not deepen beyond a critical threshold; the subpolar bloom ends with the exhaustion of nutrients [Fernández et al., 2005], and not because the ML becomes too deep. It appears in Figure 2 that the period of highest biomass matches rather closely the period of deepest ML.

[21] At midlatitudes (between 35°N and 40°N on the 2001 picture), SCHL appears to follow a seasonal cycling in relation to the MLD cycling which differs markedly from the typical subpolar and subtropical cycling. The SCHL signal is characterized by a single broad bloom of intermediate amplitude, weaker than the subpolar spring bloom and stronger than the subpolar and subtropical fall blooms. This single bloom begins at fall with the deepening of the ML, which is characteristic of an entrainment bloom (as in both the subpolar and subpolar regimes) and keeps developing in winter. It reaches its peak in spring, after restratification, which is characteristic of a spring bloom (as in the subpolar regime). Figure 2d might give the impression of a progressive evolution from the subtropical regime to the subpolar regime. However, we believe that a third specific regime emerges from this analysis, a "midlatitude regime." This midlatitude regime is characterized by the merging of a fall bloom with a spring bloom. The midlatitude bloom is both light and nutrient limited, contrary to the spring bloom which is essentially light limited, and to the entrainment blooms which are essentially nutrient limited. Specific criterion can be drawn to identify the midlatitude regime. First, the absence/presence of a fall bloom enables to distinguish between the midlatitude and the subpolar regimes. Second, the midlatitude and subtropical regimes differ in the dynamics of their single bloom. In the subtropical regime, nutrient limitation prevails; when nutrients are provided by the deepening of the ML, the bloom proceeds at high efficiency. This results in the peak of the bloom occurring simultaneously or slightly before the maximum MLD. In the midlatitude regime, phytoplankton is both light and nutrient limited; when the ML deepens, nutrient limitation decreases but light limitation increases. The bloom starts at a slow rate. Only when the ML shallows again does the bloom reach its maximum efficiency. This is characterized by the peak of the bloom occurring after the maximum MLD. The midlatitude regime has specific biogeochemical implications, which will be discussed farther.

[22] To make the picture clearer, the 2001 SCHL and MLD seasonal cycles have been averaged for the 3 specific regimes, i.e., for this particular year, over the 30°–35°N, 35°–40°N and 40°–45°N latitude bands. After filtering out high-frequency variability, they are plotted in Figure 3. As shown by the paired curves, the "midlatitude bloom" starts as a fall bloom and switches to a spring bloom. It lasts...
longer than the other blooms. Its amplitude is intermediate between that of the subpolar spring bloom and that of the subtropical fall blooms. The time shift between maximum MLD and maximum SCHL is also intermediate between that for the subtropical bloom and that for the subpolar spring bloom. Section 3.3 will provide further assessment of the 3 regimes.

3.2.2. Focus on the POMME Cruises

From September 2000 to October 2001, four POMME cruises took place. The first cruise (P0) was conducted in September 2000. The most complete coverage occurred during P1 (February–March 2001) and P2 (March–May 2001). The last cruise P3 occurred in late August–early October, 1 year after P0 (vertical white lines on Figure 2 indicate the first and last date of each cruise). The cruises were usually divided into two legs. The first legs (L1 on Figure 2) were dedicated to a large-scale survey of the area [Mémery et al., 2005]. The main study area (39°–44.5°N, 16°–21°W) was covered with 7 latitudinal transects, with CTD stations approximately 50 km apart and basic JGOFS type of measurements (nutrients, oxygen, inorganic and organic carbon, phytoplankton, primary pro-

Figure 2. Temporal and meridional variations of zonal averages (16°–22°W) from September 2000 to October 2001 of (a) net heat flux (Clipper), (b) wind stress (ECMWF), (c) MLD (Clipper), and (d) SCHL (SeaWiFS). The vertical lines mark the period of each POMME survey (P0 to P3). The numbers mark the location and time of the time series stations. Legs 1 (L1) and 2 (L2) of each survey are also indicated. See color version of this figure at back of this issue.
This coverage lasted around 3 weeks. During the second legs (L2), four time series stations (numbered 1 to 4 on Figure 2) were carried out to investigate 1D biogeochemical processes. They generally lasted 2 days. A complete set of physical, chemical and biological parameters were measured in specific regions of the mesoscale field to give information on the mesoscale dynamics, on the mixed layer evolution, on the biogeochemical stocks and fluxes and on the ecosystem structure [Mémery et al., 2005]. The aim of P1 was to describe the winter conditions in order to estimate the maximum MLD and set the initial prebloom conditions. The spring cruise P2 immediately followed, with the objective of characterizing the mode waters and the bloom evolution. The aim of P3 was to capture oligotrophic conditions. POMME in situ data were used to validate the SCHL and MLD used in this work (see sections 2.1 and 4.1).

A drawback of POMME observational strategy is the strong asynopticity in the data [Lévy et al., 2005]. Particularly during the bloom, the duration of the first leg is long compared to the characteristic timescales of the weather, of the MLD and of the ecosystem. During the second legs, it is again difficult to determine if the biogeochemical differences observed from one station to another are due to their specific location, to the progression along the seasonal cycle or to a response to different local weather conditions.

SeaWiFS data provide a synoptic view that enables us to situate these four POMME cruises within the seasonal cycle and the series of wind events. Figure 2d reveals very clearly that P0 occurred during the 2000 fall bloom. P1 was initially planned to represent winter preconditioning conditions. SeaWiFS data reveal that this winter regime was actually reached prior to P1, between P0 and P1. P1 appears as a period of slow growth: the ML is at its maximum depth. Between 35°C176N and 40°C176N, P1 covers the beginning of the midlatitude bloom, i.e., the period during which it has the characteristics of an entrainment bloom. Growth is accelerated during two short periods of fine weather, one during P1-L1, and the other during P1-L2 (stations 3 and 4). Maximum SCHL values are reached during P2; the second part of the midlatitude bloom (spring bloom type) occurs during P2. The wind event in May occurs during the end of P2, and causes dilution of phytoplankton at P2 time series stations 3 and 4. P3 appears as a typical summer oligotrophic regime, close to equilibrium, and ends with the 2001 fall bloom.

This 2001 example shows that the seasonal cycle is perturbed by synoptic atmospheric events. These can either accelerate (as in mid-February) or decelerate (as in the beginning of May) phytoplankton growth. It is noteworthy that these high-frequency events spotted in totally independent data sets (ECMWF and SeaWiFS) are remarkably in phase.

3.3. Variability Over the 5 Years

In order to examine how the seasonal cycling changes with latitude throughout the 5 years, SCHL and MLD were averaged zonally between 16°C176W to 22°C176W and meridionally in 1° bands from 30°N to 46°N. The resulting seasonal cycles are presented in Figure 4. For each situation (year and 1° latitude band), the paired SCHL and MLD curves were compared to the three “models” of production regime proposed in Figure 3. Each situation was determined to belong to one of the three specific regimes on the basis of objective criteria. More precisely, the boundary between the subpolar and the midlatitude regime was defined as the latitude at which the fall bloom merges with the spring bloom. A smoothing was applied to the SCHL curves to...
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PRODUCTIVITY

Figure 4

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remove high-frequency variability. Smoothed SCHL curves where a local minimum (in summer) was found in between two local maxima (in autumn and spring) were assigned to the subpolar regime. The boundary between the midlatitude and the subtropical regime was determined on a criterion involving the timing of the maximum MLD versus the timing of the maximum SCHL peak. Situations where the SCHL peak followed the MLD peak were assigned to the midlatitude regime, and inversely situations where the SCHL peak preceded or was simultaneous to the MLD peak were assigned to the subtropical regime. According to these criteria, seasonal cycles of subpolar type were marked in Figure 4 with a dark shaded background, midlatitude ones with white and subtropical ones with light shading.

3.3.1. General Features

[28] On the basis of Figure 4, we now describe the features of the three regimes that are consistent throughout the 5 years and the features that vary.

[29] In the subpolar regime, the maximum MLD is always greater than 250 m. The spring bloom is characterized by an abrupt increase in SCHL in response to a rapid shoaling of the ML. The bloom develops at the end of the stratification transition and its intensity is always higher than 1 mg/m³. Beside the spring bloom, a fall bloom occurs in all of the situations in Figure 4. Driven by the ML deepening after the summer minimum, the fall bloom reaches its peak (of the order of 0.5 mg/m³) well before the end of the winter deepening.

[30] In the subtropical regime, the maximum MLD is always less than 120m. There is always a single and relatively weak fall bloom, driven by the ML deepening of the mixed layer. The biomass peaks at less than 0.4 mg/m³ at about the same time as the MLD reaches its maximum.

[31] In the midlatitudes regime, the maximum MLD comprises between 120 m and 250 m. There is always a single bloom. Its intensity ranges between the maximum value encountered for the subtropical fall bloom (0.4 mg/m³) and the minimum value observed for the subpolar spring bloom (1 mg/m³). The midlatitude bloom peaks during the shoaling phases of the mixed layer and lasts longer than the blooms in the two other regimes.

[32] As regards the timing of the blooms, Figure 4 shows significant variability within the latitude range, as well as from one year to the next. For the subpolar spring bloom, the onset occurs earlier in the south and varies with an overall amplitude of almost 6 months. For the subpolar fall bloom, the onset (generally in August–September) occurs earlier in the north and varies with an overall amplitude of 2 months. It is exceptionally late in 1998 (late October). The midlatitude (fall) bloom onset (most generally in September–October) also occurs earlier in the north and the overall amplitude of its meridional/interannual change is of the order of 2 months. It is exceptionally late in 1998 (early November). The timing of the subtropical bloom (onset always in November) also occurs earlier in the north but the overall variation of the onset is of the order of only 1 month.

[33] The SCHL maximum is highly variable from one year to the next. In the subpolar regime, it does not exceed much 1 mg/m³ in 1998 and 1999 but reaches 1.5 to 1.8 mg/m³ in some of the 2000, 2001 and 2002 situations. It is worth noting that, for each of the 5 years, the bloom is not a smooth SCHL variation; most generally it is briefly interrupted by a sudden redeepening of the ML, to resume with the definite seasonal restratification. As regards the subpolar fall bloom, its peak ranges from a minimum value of about 0.5 mg/m³ in 2002 to a maximum of about 0.7 mg/m³ in 1998. In the midlatitude regime, the bloom peak ranges from a minimum value of around 0.5 mg/m³ (encountered over the whole region in 1998 and in its southern part for all years) to a maximum of around 1 mg/m³ (encountered in the northern part of the region in 1999, 2000 and 2002). The midlatitude bloom is most generally characterized by a single peak. Exceptionally, in 2000, an event of redeepening during the ML shoaling period leads to a secondary SCHL peak. In the subtropical regime, the (fall) bloom shows a lesser variability in its intensity, with an annual peak generally between 0.21 mg/m³ and 0.27 mg/m³. An exception is the bloom of 1999–2000, with two strong peaks at about 0.4 mg/m³. It is also worth noting that both the MLD and SCHL signals exhibit a rather moderate variability at high frequency, compared with the two other regimes.

[34] The location of the regimes boundaries changes significantly from one year to another. The southern boundary of the subpolar regime is at 42°–43°N at the beginning of the time series in 1998–1999 and shifts toward 40°N at the end in 2002. For the northern boundary of the subtropical regime, the (fall) bloom shows a lesser variability in its intensity, with an annual peak generally between 0.21 mg/m³ and 0.27 mg/m³. An exception is the bloom of 1999–2000, with two strong peaks at about 0.4 mg/m³. It is also worth noting that both the MLD and SCHL signals exhibit a rather moderate variability at high frequency, compared with the two other regimes.

3.3.2. Physical Interpretation

3.3.2.1. Bloom Onset and Propagation

[35] The information about the variations of the bloom onset has been synthesized in Figure 5. The well-known northward “propagation” of the subpolar spring bloom is shown in Figure 5a. Interestingly, a southward propagation is also evidenced. This southward propagation concerns the entrainment bloom, i.e., the succession from north to south of the subpolar fall bloom, the midlatitude bloom and the subtropical bloom. When looking at the y-t SCHL representation (Figure 2d), the impression is of a northward propagation over the whole latitudinal range, i.e., from 30°N to 50°N. This northward propagation (Figure 2d) is that of the SCHL peak; it is reached in March in the subpolar regions, in April at midlatitudes and in May in the north. The peak marks the beginning of the decline of the bloom, and this northward propagation (already observed by, e.g., Yoder et al. [1993], Siegel et al. [2002], Follows and Dutkiewicz [2002], and Dandonneau et al. [2004]), likely reflects the meridional gradient in nutrient limitation rather
than the northward propagation in irradiance: nutrients being more abundant in the north, their exhaustion last longer. In our analysis, we are dealing with the propagation of the onset of the bloom. It appears that the northward propagation (of the bloom onset) only concerns the subpolar spring bloom, which only exists north of approximately 40°N. The propagation speed of this bloom strongly varies with latitude and from one year to the next. It ranges between an average of 7 km/d in 2000 and 50 km/d in 2001 (the value of 20 km/d was published by Siegel et al. [2002]). The propagation speed of the fall bloom is also variable, around a mean of approximately 20 km/d.

Figure 5b presents the times at which the ML reaches bloom-triggering values. The time when the ML shoals to a depth less than 180 m (i.e., twice the typical euphotic depth; see the discussion section 4.2) has been empirically determined to explain the restratification bloom (subpolar spring bloom). The time when the ML deepens below the nitracline depth was retained for the entrainment blooms. The nitracline depth was computed from the fall Levitus nitrate climatology, as the depth of maximum vertical gradient from the surface. As shown, the propagation of the subpolar spring bloom and its variations are rather well explained by the ML criterion. A first reason is that mixing depth is the most important limiting factor for production in this situation. A second reason is that the latitudinal propagation of solar light, which forces photosynthesis, is parallel to the propagation of the ML shoaling, and as a matter of fact, partially forces this propagation (through its contribution to Qnet).

As for the entrainment bloom, the time when the ML deepens to the nitracline depth also appears to be an adequate criterion to explain the southward propagation of the bloom. Further analysis (not shown) reveals that this propagation is mainly driven by the nitracline meridional gradient (the nitracline being shallower in the north than it is in the south), and not so much by the southward propagation of the mixed layer deepening (which is much more rapid). This also suggests that interannual variability in the entrainment bloom onset and propagation is related to interannual variations in the nitrate distribution, which is absent from this analysis (Levitus climatological data had to be used for nitrate).

3.3.2.2. Bloom Strength

The purpose of Figure 6 is to evaluate to what extent the bloom strength is explained by the vertical mixing intensity. For that, the mean biomass during the bloom, as a measure of its “strength,” is plotted against the mean MLD during the same period, as an index of vertical mixing intensity. As shown, the strength of the subpolar spring bloom (black symbols, upper line in Figure 6) and the strength of the subtropical bloom (shaded symbols with black frame, lower line) are actually correlated to the depth of the mixing, whereas no coherence emerges from a linear regression analysis for the midlatitude bloom (white symbols) and for the subpolar fall bloom (frameless shaded symbols in the bottom part of Figure 6). As expected, the restratification bloom and the entrainment bloom exhibit opposite responses. In the subpolar regime, a deeper ML (i.e., enhanced mixing) results in a weaker bloom. Indeed, the intermittent mixing events which keep perturbing the bloom might either lead to light limitation (a deeper ML further decreases the averaged light experienced by the phytoplankton cells) or simply lead to dilution. In contrast, in the subtropical regime the bloom is nutrient limited, and a deeper mean ML intensifies the bloom by increasing the nutrients input. As for the midlatitude regime, the concomitance of light and nutrient limitation forbids the emergence of a definite relationship. This competition between the two limiting factors is somewhat reflected in the distribution of...
the midlatitude points in Figure 6: they tend to join one or the other of the organized groups. As for the subpolar fall bloom, the point distribution looks rather parallel to the subtropical line. If no clear correlation emerges, it is likely because the bloom does not end due to the exhaustion of a nutrient pool determined by the vertical mixing during the bloom period, as it is the case for the subtropical bloom, but due to light limitation.

3.3.2.3. Regime Boundaries

[39] The interannual variations in latitude of the regime boundaries were described previously (3.3.1). In order to analyze the shift in terms of the physical forcings, the meridional variations of annual SCHL, maximum winter MLD, winter wind stress intensity and annual wind curl are plotted in Figure 7 for 1998 (black line) and 2001 (shaded line). These years were selected as extreme cases of the meridional variability in the 1998–2002 series. The meridional extension of the midlatitude regime was reported from Figure 4 for the 2 selected years (black frame for 1998, shading for 2001). For the 2 years the northern boundary of the box corresponds approximately to the same value of 0.2 mg/m^3 for the annual mean SCHL (Figure 7a) and to the same value of 250 m for the maximum winter MLD (Figure 7b) while the southern boundary corresponds approximately to 0.1 mg/m^3 and 110 m (large dots on Figure 7). This suggests that the three regimes can also be distinguished according to thresholds in annual mean biomass or maximum winter MLD.

[40] In 2001 the winter wind stress is stronger than in 1998 from 35 to 45°N (Figure 7d) and the line of zero wind stress curl (located at about 50°N) is found farther south than in 1998 (at around 55°N, Figure 7c). This is an indication of a southward shift in the northeast Atlantic wind system. Both the intergyre boundary and the ML front (Figure 7b) follow and thus the production regime boundaries. As noted before, the southward displacement of the southern boundary exceeds that of the northern limit. Indeed, an increase in wind is associated with larger heat losses, which have potentially a greater impact along the southern side where the ML is shallower than on the northern side.

4. Discussion

4.1. Methodology

[41] In order to analyze the variability of SCHL over the northeast Atlantic, 5 years of SCHL data from SeaWiFS were combined with MLDs from the Clipper ocean circulation model. This comparison revealed remarkable phase relationships in these totally independent data sets, at synoptic, seasonal and interannual timescales. A posteriori, the overall success of the MLD/SCHL comparison provides a validation of the phase of the MLD variability in Clipper. Besides, the phytoplankton bloom is influenced by other factors than the MLD, such as lateral advection [Williams and Follows, 1998], mesoscale activity [Levy et al., 1998; McGillicuddy and Robinson, 1997; Oschlies and Garçon, 1998], nitrogen fixation [Michaels et al., 1994] and grazing [Popova et al., 2002]. The relationships that nevertheless emerged between the variations in MLD and SCHL emphasize that the MLD variability is the main factor influencing the SCHL variability in that region.

[42] In order to mitigate our results, errors associated with satellite and model products are now discussed. The relatively high cloud cover in the region forces the use of 8 day composites of SeaWiFS images, in which there is a higher amount of cloud-free pixels. This filters out events of higher frequency. It also makes more difficult the comparison with model MLDs, which are an average over 5 days. Another consequence is that the precision of our timing estimates cannot be better than a week. Besides, many of our results are based on the analysis of spatial means. Owing to the nonhomogeneous character of a rather high cloud cover (twice higher in the northern part of the region than in the south), the values of meridional gradients in bloom properties have to be taken with some caution.

[43] Another uncertainty concerns the MLD absolute value estimated by the Clipper model. In order to evaluate its accuracy, a comparison was made with MLDs from three other models (Figure 8): a high-resolution (1/20°) regional model run from February to October 2001 (extension of the model run presented by Levy et al. [2005]); the 1° resolution global configuration of OPA (http://www.lodyc.jussieu.fr/opa) and the 1/3° resolution operational model of the North Atlantic MERCATOR (http://www.mercator-ocean.fr), for the period 1998–2002. The four models are based on the same numerical code (OPA), with the same vertical mixing scheme, but with different resolutions, lateral physics, initial and boundary conditions, and forcings. The POMME model is used as a reference since its MLDs have been validated against in situ observations taken during P1 and P2 [Levy et al., 2005; Paci et al., 2005]. The timing of the MLD deepenings and shallows are the same for all models, emphasizing again the correct
representation of the MLD phase (this is also true for the 4 other years; not shown). However, the MLDs amplitudes appear overestimated in Clipper, by up to 100m in March. This overestimate is due to the ECMWF heat forcing, which is too cold and not fully compensated by the SST restoring term (the best estimate annual net heat flux over POMME is 22 W/m² [Caniaux et al., 2005], it is −10 W/m² in ECMWF and 6 W/m² in Clipper with the restoring term).

In view of this overestimate, maximum MLDs from Clipper need to be corrected before attempting to derive critical values for the MLDs.

[44] Finally, the determination of regime types for each situation (Figure 4) is somewhat dependant on the data filtering prior to the application of the two criteria (section 3.3). This is however not likely to lead to worse than a 1° error in latitude when locating the boundaries.
Besides these technical limitations (time resolution, cloud cover, model MLDs), our approach has also inherent limitations. It only enables to address that part of the variability which is driven by vertical mixing and the biogeochemical variability is interpreted on the basis of surface (or rather ML) chlorophyll only. On the whole, comparison of fields from ocean models with satellite retrieved chlorophyll has however proven to be an interesting alternative to coupled physical-biogeochemical models for examination of the interannual variability of the production regimes, with the benefit that conclusions are not subordinated to the uncertain choice of parameterizations and parameters in the biological model.

4.2. Production Regimes

Three production regimes were identified on the basis of the seasonal cycling of SCHL versus MLD. The well-known subpolar regime and subtropical regime have naturally emerged from our analysis and the data were treated with sufficient meridional resolution to emphasize a third regime, the midlatitude regime, with very specific characteristics.

The characterization of the regimes presented in this work is consistent with the regime identification of Dutkiewicz et al. [2001], in spite of the difference in the approach. Dutkiewicz et al. [2001] define the subpolar and subtropical regions as the regions where enhanced mixing decreases or increases the bloom intensity, respectively. They found an intergyre region, where none of these relationships emerge. Using ocean time series observations and remote SCHL estimates from SeaWiFS, Follows and Dutkiewicz [2002] examine the interannual variability of the bloom intensity, in the conceptual framework of these regional regimes. Here, the regimes are defined in terms of coupled seasonal cyclings. This provides a detailed description of the characteristics of three distinct regimes. The regionalization of the regimes (Figure 4), and the relationship between bloom intensity and mixing (Figure 6), are two characteristics that emerge from the analysis of the 5 years SCHL time series.

To complement the description of the regimes given for the year 2001 in section 3.2.1 and the documentation of meridional and interannual variations of their parameters given in section 3.3, a synthesis of the mean characteristics of each regime is now presented in Table 1. Basically, the regimes are organized along meridional regions. The subpolar regime is found approximately north of 41°N (±1.3°), the midlatitude regime between 36°N and 41°N and the subtropical regime south of 36°N (±2.7°). These regions correspond to different typical magnitudes of winter MLD (300 m, 150 m, and 100 m respectively). The deepest MLDs are reached around the end of February (±2 weeks), with a time shift of only 1 week between 35°N and 50°N. The annual mean SCHL in the midlatitude regime is approximately intermediate between the subpolar and subtropical regimes. The largest annual SCHL are found where winter MLDs are the deepest, and therefore where the nutrient

![Figure 8. Time evolution of the MLD averaged over the POMME region (16°–22°W, 38°–45°N) predicted by different models from October 2000 to September 2001.](image-url)

Table 1. Mean and Standard Deviation Characteristics of the Subpolar, Midlatitude, and Subtropical Regime Throughout the Period 1998–2002*

<table>
<thead>
<tr>
<th>Regime</th>
<th>Subpolar</th>
<th>Midlatitude</th>
<th>Subtropical</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latitude range, deg</td>
<td>north of 41.2N ± 1.3</td>
<td>36.2N ± 2.7–41.2N ± 1.3</td>
<td>south of 36.2N ± 2.7</td>
</tr>
<tr>
<td>Max annual MLD, m</td>
<td>310 ± 51</td>
<td>155 ± 32</td>
<td>110 ± 16</td>
</tr>
<tr>
<td>Day of maximum annual MLD</td>
<td>3 Mar ± 16</td>
<td>28 Feb ± 16</td>
<td>22 Feb ± 14</td>
</tr>
<tr>
<td>Mean annual SCHL, mg/m³</td>
<td>0.4 ± 0.04</td>
<td>0.25 ± 0.07</td>
<td>0.15 ± 0.04</td>
</tr>
<tr>
<td>Peak of the bloom, mg/m³</td>
<td>0.81 ± 0.36</td>
<td>0.38 ± 0.16</td>
<td></td>
</tr>
<tr>
<td>Fall bloom</td>
<td>0.44 ± 0.14</td>
<td>0.84 ± 0.34</td>
<td></td>
</tr>
<tr>
<td>Spring bloom</td>
<td>1.42 ± 0.32</td>
<td>1.54 ± 0.32</td>
<td></td>
</tr>
<tr>
<td>Day of bloom onset</td>
<td>7 Oct ± 18</td>
<td>29 Oct ± 26</td>
<td></td>
</tr>
<tr>
<td>Fall bloom</td>
<td>18 Sep ± 17</td>
<td>4 Apr ± 25</td>
<td></td>
</tr>
<tr>
<td>Spring bloom</td>
<td>1 Apr ± 22</td>
<td>4 Mar ± 14</td>
<td></td>
</tr>
<tr>
<td>Day of SCHL peak</td>
<td>10 Nov ± 14</td>
<td>7 May ± 24</td>
<td>204 ± 68</td>
</tr>
<tr>
<td>Bloom duration, days</td>
<td>117 ± 18</td>
<td>259 ± 15</td>
<td></td>
</tr>
<tr>
<td>Fall bloom</td>
<td>73 ± 22</td>
<td>204 ± 68</td>
<td></td>
</tr>
</tbody>
</table>

*Abbreviations are as follows: MLD, mixed layer depth; SCHL, SeaWiFS ocean color images.

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Figure 9. SCHL at the time of maximum MLD as a function of latitude for each of the 5 years.

supplies are the largest. First, this suggests that the production budget between 30°N and 50°N is primarily driven by the convective supply of nutrients, in agreement with the quantitative analysis of Williams et al. [2000].

This suggests secondly that the seasonal cycling is strongly modulated by light limitation. At all latitudes, an entrainment bloom is initiated in fall by the deepening of the ML (from mid-September in the subpolar regime to the end of October in the subtropical regime). This bloom is more intense at midlatitudes than farther south due to the larger availability of nutrients. Traditionally, the subpolar cycling is described as the succession of an entrainment bloom in fall and an intense bloom after vernal stratification. An alternate picture of the subpolar seasonal cycling is that production is interrupted in winter (from mid-January to beginning of April) and resumes in spring. The subpolar regime is characterized by the existence of this specific winter regime, during which phytoplankton growth is inhibited by deep mixing. The subpolar winter regime does not occur in the midlatitudes and subtropics, because the ML does not get deep enough to inhibit growth. Consequently, the midlatitude bloom is the longer-lasting bloom (Table 1).

Throughout the 5 years, the midlatitude regime meridional extent corresponds to relatively constant range of winter MLD (Figure 7b, from 250 m in the north to 110–140 m in the south). Although we have low confidence in these absolute values, the fact that they are consistent throughout the 5 years implies that there might be “threshold values” in winter MLD that determine the regime of a region. Following the concepts first introduced by Sverdrup [1953], we can derive critical values of the ratio of winter mixed layer to euphotic layer depth (Ze). Ze estimated during P1 [Claustre et al., 2005] is approximately 90 m. On the basis of Figure 9, we apply a ~30% correction to the Clipper winter MLDS. Coming back to Figure 7, the midlatitude regime appears now as the regime where the winter maximum MLD comprises between Ze and 2Ze. These winter conditions allow slow phytoplankton growth. In the subpolar regime, winter MLD exceeds 2Ze. This is too deep to allow the growth of phytoplankton. In the subtropical regime, the MLD does not get deeper than Ze. It is worth noting that these MLD/Ze bounds were chosen by Lévy et al. [1998] to parameterize the inhibition of production in situations of deep and intermediate mixing in the northwestern Mediterranean Sea.

The midlatitude regime covers about half of the POMME area (the other half pertaining to the subpolar regime). This may explain some unexpected characteristics of the POMME in situ observations. The first characteristic is that in situ SCHL never exceeded 1 mg Chl/m³, even during the bloom, while higher SCHL concentrations were expected (3–4 mg Chl/m³ were observed during the North Atlantic Bloom Experiment (NABE) [McGillicuddy et al., 1995]). As seen in Table 1, this value is consistent with the order of magnitude of the peak of the midlatitude bloom (0.81 ± 0.36 mg Chl/m³). Figure 4 also shows that the peak SCHL value rapidly increases with latitude above 43°N, which is consistent with the much higher values observed during NABE at 47°N. The second characteristic is that averaged primary production profiles measured during POMME (second legs of P1 and P2) clearly revealed that winter is a period of significant production, not so much different from the bloom period (37% higher primary production during spring than during winter [Claustre et al., 2005]). This result was unexpected, but is less surprising in the light of the midlatitude regime concept; indeed, as mentioned before, the bloom onsets in fall in the midlatitude regime and lasts longer. The third characteristic is that the rather small f ratios sometimes observed during the bloom in the region (during NABE [Garside and Garside, 1998] and during POMME (S. Lhelguen, personal communication, 2004)) may also be explained by the fact that the bloom has actually started in winter, thus allowing the regeneration network to be fully functional by the beginning of spring.

4.3. Production in Mode Waters

The production level within mode waters is of significant biogeochemical importance due to the long-term storage of surface properties subducted into deep waters. These waters are formed in winter [Woods, 1985] when the ML begins to retreat [Marshall et al., 1993; Hazeleger and Driftouw, 2000; Valdivieso da Costa et al., 2005] in regions of large meridional MLD gradient [Paillet and Arhan, 196a, 1996b]. For this reason, we plot in Figure 9 SCHL at the time of maximum MLD (hereafter winter SCHL) as a function of latitude. Winter SCHL is representative of the SCHL content of the water to be subducted. It is an index of the initial biogeochemical properties of the subducted waters and thus of the efficiency of the biological pump of carbon.

Figure 9 shows a bowl shape structure of winter SCHL against latitude: a general feature that emerges is that winter SCHL is maximum at midlatitudes (between 35° ± 2°N and 40° ± 2°N), and minimum at high and subtropical latitudes. This singular structure ensues from the particular
properties of the midlatitude regime. Indeed, in the subpolar regime, production is weak in winter due to light limitation, and in the subtropical regime, production is always weak due to the small vertical excursion of the ML; winter production is actually optimum in the midlatitude regime. Intriguingly, the region of the midlatitude regime is the main area of subduction since it overlaps the area of maximum gradient in winter MLD (Figure 7b). This analysis therefore suggests that winter production is the highest in the region of subduction, thus reinforcing the efficiency of the biological carbon pump during subduction. Note that this result is not intuitive and relies on the particularity of the midlatitude regime. The winter MLD gradient also overlays the region of the subpolar regime (Figure 7b), but in that case winter production is weak and the pathway for carbon sequestration is thus less efficient than in the midlatitude regime.

Figure 9 also suggests that the biogeochemical properties of the waters to be subducted show a large variability from one year to the next. However, this variability cannot be fully assessed without better data sets. Indeed, the difference in temporal resolution of the SCHL and MLD time series forces to interpolate the value of SCHL at the time of maximum MLD. This interpolation is uncertain, as it involves fast varying biological parameters, and may be responsible for the jagged aspect of the curves.

To sum up, the midlatitude regime overlays the area of subpolar mode water subduction of the northeast Atlantic. Since subducted waters originate from the ML when the MLD is largest, the fact that the bloom starts prior to that date very likely enhances the efficiency of the biological carbon pump. Production during that period enables to convert inorganic carbon into organic form, and to sequester this carbon after the water mass is isolated from the influence of the atmosphere.

5. Conclusion

A simple method was applied to study the variations of SCHL in relation to variations in MLD. Synoptic observations from space were found to be a valuable tool to assess the driving mechanisms of the SCHL variability over the northeast Atlantic and during POMME. This study has emphasized the large interannual variability of the production system over the northeast Atlantic, in terms of bloom timing, intensity and of the position of the geographical frontier between the various regimes. Comparison with MLD and atmospheric fields has revealed that a large part of this variability can be attributed to interannual variability in the atmosphere. At synoptic scales, the passage of storms varies from one year to another in timing and intensity. This explains why the MLD cycle is not smooth, but is punctuated by mixing events. The consequence is the different timing and intensity of the bloom, and the high frequency in SCHL. At lower frequencies, the westerlies shift from a more southerly to a more northerly track. This shift forces the shift of the winter MLD gradient (through a change in wind stress), and the location of the intergyre boundary (through a change in wind stress curl), and hence the boundaries between the biogeochemical regimes. The dominant mode of atmospheric low-frequency variability is associated with the North Atlantic Oscillation (NAO) [Hurrell, 1995]. Williams et al. [2000] have shown that the variability in nitrate supply is significantly correlated with the NAO over parts of the central and western Atlantic, but not over the eastern Atlantic. This suggests that another mode of variability may prevail in the POMME region, that still has to be elucidated, thanks to a long enough time series.

After confronting our results to method limitations, they proved robust enough to support a novel interpretation of the dynamics of the North Atlantic blooms. Indeed, the midlatitude regime which emerged from our analysis fills the gap between the singular Sverdrup case and the prevalent situation of nutrient limited production. The midlatitude regime is characterized by winter MLDS in the range Ze to 2 Ze and by a bloom which evolves from autumn to spring. Besides, it covers parts of the mode water formation region. This combination likely has crucial positive impacts on the long-term carbon sequestration.

The midlatitude regime needs to be further documented. Other biogeochemical parameters can be retrieved from space that should provide further insight, such as dissolved organic components (H. Loisel et al., manuscript in preparation, 2005) or phytoplankton community composition (J. Uitz et al., From surface chlorophyll a to phytoplankton community composition in oceanic waters, submitted to Global Biogeochemical Cycles, 2004). More sophisticated tools are needed, such as interannual coupled biological-dynamical models calibrated with the POMME data set. Their development is in the ongoing POMME project. The relationships between SCHL and MLD that emerged from this work should guide the analysis of such models.

Acknowledgments. Financial support for this research was provided by CNES, MERCATOR, CNRS, IRD, and the PROOF program. The work of Y.L. was supported by CNES and by Tel Aviv University. We are thankful to A.-M. Tréguet, B. Barnier, J.-M. Molines, and S. Theetten for providing the outputs from the Clipper model. C. Elbe and G. Madec are acknowledged for providing the outputs from the ORCA model. We wish to thank M. Gavart for the extraction of the MERCATOR products. We thank C. de Boyer Montégut and J. Paillet for allowing the use of their data sets. M. Raizada is acknowledged for his assistance in the treatment of the SeaWIFS data. Special thanks are to the Goddard Space Flight Center (GSP/CNASA) for providing the SeaWIFS images used in this work.

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CHAPTER 2. CLIMATOLOGY AND LARGE SCALE VARIATIONS IN PRODUCTIVITY

Climatology

a) MLD (m) and line of zero annual heat flux

b) Wind Stress (W/m²)

c) Surface currents (cm/sec)

d) Cloud cover (%)

e) Surface Nitrate (mol/L)

f) SCHL (mg/m³)

Figure 1

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Figure 1. A climatological view of the northeast Atlantic. (a) Winter mixed layer depth (MLD) (after de Boyer Montegut et al. [2004]) and the line of zero annual net heat flux in Clipper between the years 1998 and 2002. (b) Wind stress (European Centre for Medium-Range Weather Forecasts (ECMWF)) and line of zero wind stress curl. (c) Surface currents (after Paillet and Mercier [1997]). (d) Cloud cover as percent of flag occurrence over 1998–2002 in Sea-viewing Wide Field-of-view Sensor (SeaWiFS) weekly chlorophyll products (see text) and the line of 50% cloud cover. (e) Surface nitrate [Levitus, 1982]. (f) Surface chlorophyll (average of SeaWiFS ocean color images (SCHL) over 1998–2002). The solid frame marks the Programme Océan Multidisciplinaire Méso Echelle (POMME) region, and the dashed frame marks the region examined in the present study.
Figure 2. Temporal and meridional variations of zonal averages ($16^\circ$–$22^\circ$W) from September 2000 to October 2001 of (a) net heat flux (Clipper), (b) wind stress (ECMWF), (c) MLD (Clipper), and (d) SCHL (SeaWiFS). The vertical lines mark the period of each POMME survey (P0 to P3). The numbers mark the location and time of the time series stations. Legs 1 (L1) and 2 (L2) of each survey are also indicated.
CHAPTER 2. CLIMATOLOGY AND LARGE SCALE VARIATIONS IN PRODUCTIVITY

2.2 Conclusions and perspectives

In this chapter we characterized large scale spatial (from several hundreds to several thousands of kilometers) and temporal (seasonal to interannual) variability patterns in the northeast Atlantic production system. Analysis of the mechanisms driving these variabilities was done by applying a simple method of comparison between the apparent changes in SeaWiFS chlorophyll data and variations in the MLD derived from the Clipper model. The MLD data was also used to identify the timing and location of subduction, thus allowing estimation of the relative importance of biological productivity in the waters that are subducted.

Based on our results we conclude that the POMME region is actually divided into two different production regimes: the subpolar regime associated with the well known spring bloom in the north and the midlatitude regime in the south. In 2001 (the year of the main POMME cruises) the boundary between the regimes is around 41°N. The newly identified regime is of great importance as it overlaps the main region of subduction (associated with strong winter MLD gradient) and can explain the rather high production rates observed during the end of winter POMME cruises (P1). The biogeochemical properties of the midlatitude regime should be further investigated. Products derived from ocean color satellite data such as classification of phytoplankton groups [Alvain et al., 2005] and estimations of particulate organic carbon [Stramski et al., 1999, Loisel et al., 2002] provide appropriate tools for such investigation.

Comparison with dynamical fields derived from ocean circulation models is proven to be an adequate tool for studying mechanism driving large scale variability patterns in ocean color satellite images. This method has the drawback of not taking into account the impact of mesoscale dynamics, that can have an important role in transferring tracers along relatively large distances, thus influencing the large scale production variability [Lee et al., 1997, Lee and Williams, 2000]. The main aspect of the mesoscale dynamics that should be further studied with respect to the large scale variability is the transfer of nutrients from the subpolar regime to the midlatitude regime. This process can have an important contribution to the biological production in the subducted waters.

The five year (1998-2002) dataset used in this work is long enough to emphasize the large interannual variability of the production system. On the other hand, the temporal coverage of the dataset prevents us relating the observed variabilities to climate changes. The length of the SeaWiFS record now available (nearly a decade) opens the way for studying the relationship between changes in the production system over the northeast Atlantic and changes in the world’s climate. This follows the work of Behrenfeld et al. [2006], which shows that the ocean’s productivity is strongly influenced by climate changes.
Chapter 3

Formation of spatial variability patterns by horizontal stirring

3.1 Extended abstract

In the paper "Stirring of the northeast Atlantic bloom: a Lagrangian analysis based on multi satellite data", accepted for publication in JGR [Lehahn et al., in press, hereafter LOL07], a systematic analysis of the effect of geostrophic stirring on phytoplankton variability at the time of the spring bloom is presented. Chlorophyll patterns were analyzed in 23 daily high resolution (1 km) SeaWiFS images obtained between 1998 and 2003. Geostrophic velocities were calculated from sea surface height, generated by combining SLA from the AVISO database of satellite derived altimetry and the Rio-05 mean dynamic topography. Daily high resolution (1 km) images of sea surface temperature from AVHRR were also used.

3.1.1 Lagrangian analysis of the velocity field

The effect of horizontal stirring on chlorophyll distribution was studied using Lagrangian diagnostics that are based on the detection of properties of the velocity field along particle trajectories. While being a standard tool for the study of transport properties in the atmosphere, in the field of oceanography they have been introduced more recently and their use is still limited. The transport properties of the geostrophic surface currents are studied by computing Lyapunov exponents using the finite size Lyapunov exponent (FSLE) method. The Lyapunov exponents are obtained by measuring the exponential divergence of trajectories that are initiated at nearby positions. As the initial distance between the trajectories (that also defines the resolution of the Lyapunov map) we used a value $0.01^\circ$ and for the final separation we choose $0.8^\circ$.

The transport properties are described in terms of hyperbolic points and manifolds. Hyperbolic points, that are located between coherent eddies at the intersection of a di-
verging and converging region, are of special relevance to transport: a passive tracer initialized nearby is effected at the same time by contraction along the converging direction and stretching along the diverging direction. The two directions are respectively called the stable and unstable manifolds of the hyperbolic point. Due to the converging component of the velocity field, a tracer can not cross an unstable manifold and it is forced to stick to it. In the case of a time dependent velocity field unstable manifolds evolve in lobular structures that allow a tracer to intrude other regions. For these reasons, unstable manifolds can be regarded as "sticking" transport barriers for the 2D advection of passive tracers. The unstable manifolds are identified as local maxima (ridges) of finite size Lyapunov exponent values. Since the intensity of the Lyapunov exponents measures the time scale at which a tracer relaxes over the manifold (i.e. the intensity of the convergent field), stronger manifold are expected to have more effect on tracers.

3.1.2 Main results

Following this diagnostics, horizontal transport by the geostrophic currents was shown to form three types of chlorophyll distribution patterns (LOL07, Fig. 5): (i) lobular fronts in correspondence to strong unstable manifolds; (ii) sub-mesoscale patterns inside eddies, characterized by weakly permeable cores (chlorophyll spots) connected to the eddies periphery by spiraling lobes of unstable manifolds; (iii) filaments aligned to the diverging direction for chlorophyll that is initialized over hyperbolic points.

The efficiency of unstable manifolds as transport barriers is expressed by a strong spatial correlation between unstable manifolds and chlorophyll fronts (LOL07, Fig. 7). In order to quantify the overall agreement between the unstable manifolds and chlorophyll distribution, we calculate the angle $\beta$ between the manifolds and chlorophyll iso-lines so that a perfect agreement corresponds to $\beta=0^\circ$ and an absolute mismatch corresponds to $\beta=90^\circ$ (LOL07, Fig. 8). After smoothing the chlorophyll images and taking into account all pixels that include the 10% strongest manifolds and highest values of chlorophyll gradients, the angle $\beta$ is smaller than $20^\circ$ in almost half of the pixels (43%) and larger than $70^\circ$ in only 9% of the pixels.

Complementary to the diagnostics based on hyperbolic points and unstable manifolds, to isolate the effect of horizontal transport, chlorophyll patterns were also compared with distribution patterns of synthetic passive particles that are advected by the satellite derived velocities.

In most cases observed here, the source of initial heterogeneity, which is essential for the formation of small scale structures in tracer’s distribution by horizontal stirring, is the large scale meridional chlorophyll gradient associated with the northward propagation of the subpolar bloom. In these cases the formation of small scale patterns result from the direct cascade of the chlorophyll variance injected at large scale. While this
process has been simulated in model studies, to our knowledge this work provides the first observational evidence of this cascade process. The cascade consists of a modulation of the large scale chlorophyll front and of intrusions of chlorophyll rich structures (either eddies or filaments) from the north to the south (or inversely, intrusions of chlorophyll poor structures from the south to the north). These intrusions reach up to 3° in latitude and act as an efficient diffusive mechanism for the large scale.

Our results indicate that the 2D cascade of the initial large-scale gradient can create chlorophyll anomalies within mesoscale eddies (both cyclonic and anti-cyclonic, LOL07, Fig. 10). The proposed mechanism involves water exchange between the interior of the eddy and its surroundings, and is made possible by the chaotic (i.e. time dependent) nature of the geostrophic velocity field. This is in contrast with the commonly referred “eddy pumping” mechanism that explains positive chlorophyll anomalies within eddies by vertical transport of nutrients through the doming of isopycnals. The time-dependency of the geostrophic velocity field, that generates inward and outward flows toward and from the eddy, also explains the formation of spirals within eddies. These sub-mesoscale features are a purely dynamical phenomenon that is controlled by the time variability of the velocity field and not by its spatial scale. The inner core of the eddy is protected against mixing between the interior and the exterior, which is consistent with floats and subsurface water masses analysis from POMME field measurements. In addition to the direct effect of horizontal transport on chlorophyll pattern formation, horizontal hyperbolic regions are also shown to support local production of chlorophyll anomalies, through vertical advection of nutrients along unstable manifolds (LOL07, Fig. 11). The proposed mechanism acts in phase with the tracer filamentation aligned to the diverging direction producing chlorophyll filaments along unstable manifolds. Based on the comparison with chlorophyll images, we deduce that application of the FSLE method on satellite altimetry data is adequate for the study of mesoscale patterns and in some cases even for sub-mesoscale transport processes. The FSLE method is shown to be a useful tool to identify transport barriers and hence to affect chlorophyll pattern formation.
Stirring of the northeast Atlantic spring bloom: a Lagrangian analysis based on multi-satellite data

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Abstract. The effect of the geostrophic stirring on phytoplankton variability during the Northeast Atlantic spring bloom is studied by analyzing satellite derived surface chlorophyll, sea surface temperature and sea surface height. The calculation of unstable manifolds is used as a diagnostic of the transport properties of the geostrophic velocity field (calculated from the sea surface height). We identify two mechanisms by which the geostrophic velocity field acts on chlorophyll patterns. The first mechanism is a direct effect of the horizontal transport on already-formed chlorophyll. By acting as "sticking" transport barriers, the unstable manifolds are shown to (i) modulate the fronts of already-formed phytoplankton in lobular structures; (ii) create spiralling chlorophyll anomalies within eddies; (iii) produce chlorophyll filaments. The second mechanism is an indirect effect on in-situ chlorophyll production mediated by nutrient upwelling. Supported by a recent study on the vertical velocities of the Northeast Atlantic [Legal et al., 2006], we argue that the horizontal unstable manifolds also shape the filamentary, vertical velocity cells, and hence the patterns of in-situ produced chlorophyll through sub-mesoscale vertical nutrient injection.

1. Introduction

The North Atlantic spring bloom is a pronounced increase in phytoplankton biomass, triggered by the shallowing of the mixed layer (ML) below Sverdrup’s critical depth [Sverdrup, 1953]. As a result of the northward propagation of the water column restratification, the bloom propagates northward throughout spring. This propagation is followed by a northward propagation of the large scale surface chlorophyll front between the productive waters in the north and the oligotrophic waters in the south [Lévy et al., 2005b].

The North Atlantic spring bloom is a basin scale phenomenon [Esajas et al., 1986; Follos and Dutkiewicz, 2002] with great importance for ocean productivity and has been the focus of several field experiments, including the JGOFS North Atlantic Bloom Experiment (NABE) and the Programme Océan Multidisciplinaire Méso Échelle (POMME). In situ observations made in the framework of these projects have shown that even though the bloom is a large scale phenomenon, mesoscale variability in the biological fields is also observed [Robinson et al., 1993; Memery et al., 2005].

The aim of this paper is to elucidate some of the mechanisms by which the geostrophic velocity field acts on the pattern formation of mesoscale and sub-mesoscale chlorophyll patches. This is done by focusing on the spring bloom in the Northeast Atlantic and analyzing altimetry data (Topex/Poseidon, ERS, Jason and Envisat), sea color images from the Sea-viewing Wide Field-of-view Sensor (SeaWiFS) and when relevant sea surface temperature (SST) images from the Advanced Very High Resolution Radiometer (AVHRR).

The analysis is based on the derivation of Lagrangian diagnostics from the time-dependent geostrophic velocity field, and more precisely transport barriers and trajectories of synthetic passive tracers. As we shall see, these Lagrangian tools are very effective in reconstructing the specific effect of horizontal stirring on
individual patterns. They provide a complementary approach to previous techniques derived from turbulence, like calculation of turbulent diffusion and spectral analysis, that are instead very powerful in detecting statistical properties (see for instance [Pasquero, 2005] and the review in [Martin, 2003]). Lagrangian techniques are based, in general, on the identification of the velocity field characteristics along particle trajectories. They are very well suited for diagnosing properties of tracers like chlorophyll, since they allow to quantify the dynamical properties experienced by a parcel of water during its motion. In spite of the fact that Lagrangian tools are extensively employed for atmospheric tracers, their use for chlorophyll has been confined to a small number of isolated cases. Abraham et al., [2000] studied a local phytoplankton bloom initiated by the iron fertilization of a patch of water in the Southern Ocean and highlighted the efficiency of the deformation field associated with mesoscale circulation to produce long and thin horizontal chlorophyll filaments in a short period; Toner et al., [2003] used a Lagrangian method to describe the evolution of two chlorophyll plums in the Gulf of Mexico.

In this work we present a systematic analysis of the impact of horizontal stirring in modulating chlorophyll variability, covering the period of the spring bloom in six consecutive years (1998-2003). The main Lagrangian concept that we use is the notion of unstable manifold. Unstable manifolds are material lines with a converging, transverse dynamics. Due to the converging dynamics transverse to them, unstable manifolds act on any advected tracer as "sticking" transport barriers, creating fronts and filaments. The concept of unstable manifold suggests some typical scenarios induced by the advection of post-bloom chlorophyll and by in-situ phytoplankton production. The suggested scenarios are verified by comparing chlorophyll fronts and filaments with reconstructed unstable manifolds. In order to isolate the effect of the 2D transport, chlorophyll patterns are compared with distribution patterns of synthetic passives particles that are advected by the satellite derived geostrophic velocities. When relevant, SST structures are also being compared.

The good agreement between the expected and the observed chlorophyll patterns indicates the central role of horizontal stirring in shaping the spatial phytoplankton variability in the Northeast Atlantic for post-bloom and locally produced chlorophyll. Remarkably, the unstable manifolds allow to predict some structures that are below the altimetric resolution. These small-scale patterns result by the time-dependent (chaotic) evolution of mesoscale coherent structures of the velocity field and can be captured by the Lagrangian detection of transport barriers.

The paper is organized as follows. Section 2 gives an overview of the regional context. Section 3 details the data and methods, with an emphasis on the concept of unstable manifolds and their interaction with post-bloom chlorophyll and in situ production. In Sec. 4 we compare chlorophyll patterns and passive tracer evolution with unstable manifolds, focusing on three typical structures. The results are discussed and concluded in Sec. 5.

![Figure 1.](image_url)

**Figure 1.** a) Average SeaWiFS chlorophyll concentrations in the North Atlantic during the period 1998-2003. The black frame marks the location of the region examined in this study. The white frame marks the location of the POMME experiment. b) Time series of zonally averaged (13°-23°W) chlorophyll. The six vertical black lines mark the time and meridional extension of the six images shown in Fig. 7.
2. Regional context

The study is focused on a 20° latitudinal band in the Northeast Atlantic (13°–23°W/30°–50°N, Fig. 1a). This area overlaps the region of the POMME experiment that was conducted in order to study the role of mesoscale eddies in the formation and subduction of 11° and 13° mode waters in the Northeast Atlantic. POMME included a series of inter-disciplinary field experiments over a seasonal cycle from October 2000 to October 2001, with more intense observations during the period of the spring bloom [Mémery et al., 2005].

This domain covers part of the subpolar gyre and part of the subtropical gyre. It is characterized by a sharp transition between an area of deep winter mixed layer depth (MLD) in the north and an area of shallower MLD in the south [Arhan et al., 1994; Paillet and Arhan, 1996a]. The strong meridional gradient in winter MLD is clearly seen in the recent climatology of [de Boyer-Montegut et al., 2004], with values reaching 220m in the north and 100m in the south. Although the area is of relatively low eddy kinetic energy (EKE) [Stammer et al., 2006], the large scale circulation is dominated by cyclonic and anti cyclonic mesoscale eddies [Assenbaum and Reverdin, 2005; Le Cann et al., 2005], with a typical eddy size of a several hundreds of km (Fig. 2a). The mean Ekman transport is directed southward. The combination of the winter meridional MLD gradient with the mean southward circulation leads to subduction of subpolar mode water [Paillet and Arhan, 1996a, b].

In terms of chlorophyll, the area is characterized by a strong meridional gradient and relatively weak zonal variations (Fig. 1a). The north is productive with high values of chlorophyll (annual mean > 0.4 mg/m³), and the south is oligotrophic with low chlorophyll values (annual mean < 0.1 mg/m³). The large scale meridional chlorophyll gradient results essentially from the meridional variations in winter convective supply of nutrients. In the north, deeper winter ML combined with a shallower nutricline allows higher supply of nutrients to the euphotic layer. In the south, the winter source of nutrients is much smaller and the region is oligotrophic [Williams et al., 2000; Williams and Follows, 2003].

Chlorophyll seasonality is mainly driven by variations in the MLD. The latter result from changes in the solar flux and atmospheric forcings. The study is focused on the period of the subpolar spring bloom (crudely between the months of March and July), when the meridional chlorophyll gradient is the strongest (Fig. 1b) and the water column is stratified.

3. Data and method

3.1. Satellite data

Daily images of chlorophyll concentrations were processed from high resolution (1 km)
Level 2 SeaWiFS data, obtained from the NASA Goddard Distribution Active Archive Center (DAAC). Data were processed using NASA’s SeaWiFS Data Analysis System (SeaDAS software package). Over 700 chlorophyll images during the spring bloom and for the period 1998-2003 were examined. We focused on images overlapping the large-scale meridional chlorophyll gradient (c.f. Fig 1b). We selected an image size of 7° x 7° in order to encompass several eddies. Because of high cloud cover, only 23 such images could be exploited (Table 1). In these images, the cloud coverage is smaller than 30%.

Sea level anomalies (SLA) were obtained from the AVISO database (http://www-aviso.cnes.fr). The distributed global product combines altimetric data from the Topex/Poseidon and ERS missions, from Jason-1 for data after December 2001 and from Envisat for data after March 2002. The SLA is gridded on a 1/3° x 1/3° Mercator grid, with one data file every seven days. The Rio-05 mean dynamic topography (MDT) [Rio and Hernández, 2004] was added to the SLA to get the sea surface height (SSH). Geostrophic velocities were calculated from this SSH, by solving the equation for geostrophic equilibrium [Pedlosky, 1987] with centred, finite differences.

Daily AVHRR data with a resolution of 1 km were processed towards SST images by Meteo France in Lannion. The processing only covers the period of the POMME experiment (from October 2000 to October 2001).

3.2. Lagrangian trajectories and Finite-size Lyapunov exponent calculation

The trajectories used for the passive tracer experiments and for the Lyapunov exponent calculation have been computed by integrating the geostrophic velocities with a Runge-Kutta scheme of the fourth order with a fixed time step of 6 hours. The velocities derived from the altimetric data has been interpolated in space and time with a multilinear algorithm. Transport properties are studied by computing Lyapunov exponent on the sea surface [Abraham and Bowen, 2002; d’Ovidio et al., 2004; Waugh et al., 2006]. More precisely we use the finite-size method and parameters as in [d’Ovidio et al., 2004]. Lyapunov exponents are obtained by measuring the exponential divergence of nearby trajectories. For the final deformation we choose 0.8° while for the initial separation we used a value of 0.01°, that also defines the resolution of the Lyapunov map. By equating the initial separation to the resolution of the map, we guarantee that all the space is sampled at least once, and no more than once. In order to avoid the dependence on the orientation of the pairs, Lyapunov exponents are obtained by diagonalizing the linear transformation of a square whose diagonals are formed by two pairs [Ott, 1993]. As we shall see in the next section, in this paper we are interested in unstable manifolds of hyperbolic points (see the appendix for a detailed discussion of the method). Such structures can be identified as local maxima (ridges) of Finite Size Lyapunov Exponents (FSLE) values (see Appendix). Then, for clarity, we define the strong manifolds as the manifolds associated with the largest 10% FSLE values (Fig. 4b). The intensity of the Lyapunov exponent measures the timescale at which a tracer relaxes over the manifold, i.e., the intensity of the convergent field. The strong manifolds are therefore the manifolds that most affect tracers.

3.3. Lagrangian analysis of transport barriers

The transport properties of the currents are described in terms of hyperbolic points and manifolds embedded in the velocity field. For the reader not familiar with these concepts, we now provide some brief overview, and refer to more specific works [Haller and Yuan, 2000; Boffetta et al., 2001; Koh and Legras, 2002; Mancho et al., 2004; Wiggins, 2005].

Figure 3. a) Geostrophic velocity field (small arrows) and pathways (thick arrows) of convergence and divergence towards and from a hyperbolic point. b) An illustration of the stretching of a passive tracer that is initialized near a hyperbolic point.
Let us start by a region dominated by mesoscale eddies, like the one depicted in Fig. 3a, and suppose that the velocity field is stationary. Without calculation, one can detect a point at the intersection of a diverging and converging region (black dot). Such hyperbolic points are of special relevance for transport, as shown in Fig. 3b. A passive tracer initialized nearby the hyperbolic point is affected at the same time by contraction along the converging direction (called stable manifold of the hyperbolic point) and stretching along the diverging direction (unstable manifold). The tracer approaches the hyperbolic point as a thinner and thinner filament and eventually aligns along the unstable manifold. Note that due to the converging transverse component of the velocity field, the tracer cannot cross such a line. For this reason, unstable manifolds of hyperbolic points act as transport barriers and control the formation of fronts. An example of this is shown in Fig. 4a where two patches initialized in the vicinity of the hyperbolic point are advected backward (blue) and forward (green) in time and approach respectively the stable and unstable manifold as thin filaments after a few days.

The above description of hyperbolic structures is rigorous for stationary velocity fields, but can still be applied to the time-dependent case, provided that the evolution of the velocity field is on a slower timescale than tracer advection. If this is the case, a tracer can relax over a manifold (the timescale being the inverse of the value of the Lyapunov exponent, as explained later), and then evolve together with the manifold. For the ocean, there is a clear timescale separation, because the propagation of the eddies (and thus of the unstable manifolds) is much slower than the mesoscale velocity field. Mid-latitudes mesoscale eddies have typical lifetimes of several months and a speed of the order of 10 km/week while mesoscale velocities are one order of magnitude larger. While a tracer cannot cross an unstable manifold, it is forced by the transverse, converging velocity to stick to it. As the manifold moves, the tracer front is thus shaped and transported. In particular, in the case of a time-dependent field, unstable moving manifolds evolve in convoluted and lobular structures that allow a tracer to intrude other regions. Therefore moving unstable manifold does not only act as transport barriers, but also control exchange and mixing. Figure 4c-f compares the cases of a stationary (i.e. frozen in time) and time-dependent velocities. In the first case, synthetic tracer trajectories are forced to follow altimetric iso-lines. Indeed, in geostrophic balance, altimetric isolines are the streamlines and therefore for a stationary velocity field the altimetric iso-lines coincide with the trajectories. A consequence of this stationarity is that the eddies are characterized by concentric closed circles ("tori", as they are properly referred to) that perfectly isolate from the surrounding (Fig. 4c,e). On the other hand, for the case of time-dependent velocities, the identity between trajectories and streamlines does not hold anymore. In particular, a tracer released inside an eddy does not

![Figure 4](image-url)
follow a closed path and after one revolution does not come back exactly to its initial position. In this way, the impermeable barriers formed by the concentric iso-lines are replaced by spirals, as can be easily seen by detecting transport barriers with the Lyapunov exponent calculation (Fig. 4f), or simply by releasing a tracer (Fig. 4d). The spirals are tighter and closely resemble impermeable tori where the streamlines are not strongly affected by the time dependency, that is, at the eddy cores. In these regions, passive tracers are trapped for a long time and can escape the eddy only after several revolutions. The picture of an isolated eddy core is indeed in agreement with observed trajectories of released floaters in the POMME region [G. Reverdin, personal communication] and, as we shall see, has a strong effect on the shaping of submesoscale chlorophyll patterns.

On the other hand, tori break in looser spirals at the periphery, allowing a relatively stronger exchange. The formation of spirals is a purely dynamical phenomenon, that is controlled by the time variability of the velocity field and not by the spatial scale of the velocity field. For this reason, a passively advected tracer can be distributed with spatial structures that are below the resolution of the velocity field itself. Note however, that spirals can also result from non-geostrophic stationary velocity fields, such as the well-known Ekman spirals.

Probing a velocity field by advecting a passive tracer as in Fig. 4a is a way of detecting unstable manifolds. The Finite-size Lyapunov exponent technique [Aurell et al., 1997; d’Ovidio et al., 2004] that we use in this work is a systematic way of doing the tracer experiment of Fig. 4a. By paving the domain and repeating this process, we measure the dynamical (Lagrangian) stretching rates, that is known to provide good candidates for points along unstable manifolds. The global result is the full complexity of these manifolds (Fig. 2b,c). For clarity, we define the strong manifolds as the manifolds associated with the largest 10% FSLE values (Fig. 4b).

### 3.4. Geostrophic transport barriers and chlorophyll patterns

The main idea of the previous section that will guide us in the analysis of geostrophic velocity field and chlorophyll images, is that unstable manifolds can be regarded as “sticking” transport barriers, that move and shape the fronts of any advected tracer. The use of Lagrangian diagnostics for the characterization of tracer distributions is extensively made for the atmosphere (see for instance [Koh and Legras, 2002] and references therein), and much less for the ocean. For the case of the Northeast Atlantic and phytoplankton patterns, we will focus on two mechanisms that point on a strong spatial correlation between unstable manifolds and chlorophyll fronts. The first one is a direct effect of transport on post-bloom chlorophyll. The second one is an indirect effect of the horizontal hyperbolic regions to support local production of chlorophyll, mediated by vertical advection of nutrients upon unstable manifolds. Let us now elaborate on the two mechanisms and speculate over the features that we expect to find in the satellite observations. The North Atlantic spring bloom is primarily induced by a shallowing of the mixed layer, and can be seen as a large scale, northward propagating chlorophyll front. Over a short period...
(1-2 weeks) a large reservoir of phytoplankton is created. This chlorophyll pool is affected by the horizontal transport. Chlorophyll relaxes onto the unstable manifolds and is then trapped and advected by lobes, and ultimately mixed with poorer water. Recalling the discussion of Sec. 3.2, we therefore expect three types of patterns as the direct effect of stirring of the post-bloom chlorophyll pool: (i) lobular fronts in correspondence of strong manifolds; (ii) submesoscale patterns inside eddies, characterized by weakly permeable cores (chlorophyll spots) connected to the periphery by spiraling lobes of unstable manifolds; (iii) filaments aligned to the diverging direction for chlorophyll initialized over hyperbolic points. Figure 5a-c presents in a schematic view the genesis of these three patterns.

In spite of the fact that the velocities do not contain vertical component, a relationship between unstable manifolds and in situ chlorophyll production through upwelling of nutrients might be expected [Lapeyre and Klein, 2006]. The vertical velocity field in the POMME region was described in [Legal et al., 2006] on the basis of SEASOR data combined with the analysis of altimetry data. Their analysis revealed a vertical velocity field $w$, involving elongated thin structures, in agreement with previous estimations from numerical simulations in the POMME region [Paci et al., 2005]. The 3D dynamics involved is the restoration of the thermal wind balance within small-scale density filaments that are elongated by horizontal stirring processes [Hakim et al., 2002]. It induces a strong correlation between density anomalies and $w$, with $w$ positive in lighter filaments and negative in denser filaments within regions of high strain. In Fig. 6 we apply this mechanism to an SST front shaped by an unstable manifold and sketch the expected effect on plankton formation. Note that the phytoplankton is not expected to be directly in phase with the upwelling cell because of the strong ageostrophic transport toward the downwelling cell. Instead, it is expected to be shifted toward the unstable manifold. This mechanism acts in phase with

![Figure 6. Chlorophyll pattern formation by local nutrient upwelling. Upwelling and downwelling occur respectively on the negative (warm) and positive (cold) anomaly. By continuity, an ageostrophic circulation connects the warm branch to the cold branch [Hakim et al., 2001; Legal et al., 2006; Lapeyre and Klein 2006b]. Nutrients are advected to the photic layer along the warm branch and then conveyed toward the cold branch by the ageostrophic circulation. This generates a local bloom on the top of the unstable manifold.](image-url)

![Figure 7. Superposition of strong unstable manifolds (black lines) and of the geostrophic velocities (white arrows) over SeaWiFS chlorophyll concentrations (colors). The timing and meridional extension of the images with respect to the large scale chlorophyll variability are marked by black lines in Fig. 1b. Throughout the paper, chlorophyll images are plotted using the same color-scales as in Fig 7f.](image-url)
the previous one (Fig. 5a-c), producing chlorophyll filaments along unstable manifolds.

4. Results

4.1. Chlorophyll fronts

Figure 7 shows six examples of SeaWiFS chlorophyll images overlaid with the geostrophic velocities (white arrows), and with the unstable manifolds (black lines). In order to facilitate the comparison between the fields, only the strong manifolds are plotted. The time-space location of these six images with respect to the evolution and northward propagation of the bloom is indicated by the black lines on Fig. 1b. Note that all images cover the large-scale chlorophyll front associated with the bloom propagation. As mentioned before, this large-scale chlorophyll front mainly results from the larger winter convective supply of nutrients in the north. Figure 7 reveals that the large-scale chlorophyll front is shaped by the mesoscale flow field. Furthermore, there are indications of intrusions, associated as well with the flow field, of rich water to the south and of poor water to the north (c.f. for instance the patch of relatively rich water at around 21°W/43°N in Fig. 7a, to which we will come back in the next sub-section).

The relationship between the chlorophyll distribution and the flow field is evidenced by the alignment of the chlorophyll front (and notably of chlorophyll intrusions) with the unstable manifolds and supports the scenario described in Fig. 5a. Note that this relationship is much less distinct when using the instantaneous velocity field instead of the manifolds. Similarly, the orientation of the intrusion filaments (of either chlorophyll-rich or poor water) is in much better agreement with the unstable manifolds than with the instantaneous velocity field. This seemingly counterintuitive observation comes from the fact that individual water parcels follow the velocity field, while fronts are shaped by the time evolution of the velocity field, as depicted in Fig. 5a. The overall agreement between the manifolds and the chlorophyll distribution is quantified by the histogram in Fig. 8. This histogram describes the

![Figure 8](image)

**Figure 8.** Distribution of the angle $\beta$, between the direction of the chlorophyll iso-lines and the direction of the unstable manifolds. Only pixels with strong manifolds and strong chlorophyll gradients are accounted for. A perfect agreement corresponds to $\beta = 0^\circ$ (see text).

![Figure 9](image)

**Figure 9.** a,b) SeaWiFS chlorophyll (colors, same scale as in Fig. 7f); c,d) AVHRR SST (colors); e,f) time evolution of synthetic passive particles (green dots) advected by the geostrophic velocities. The particles are initiated on March 26, 2001 as a meridional front at 38.5°N. The snapshots are from consecutive days 2 weeks apart. The black lines correspond to the strong unstable manifolds.
distribution of the angle $\beta$ between the direction of the unstable manifold and of the chlorophyll iso-lines, so that a perfect agreement corresponds to $\beta = 0^o$ and an absolute miss-match corresponds to $\beta = 90^o$. In order to filter out the highest frequencies (that are not detected by the manifolds, see further in the discussion), the chlorophyll images are first smoothed with a running window of 30 km. For the overall 23 images, all pixels including both strong manifolds and 10% of the strongest values of chlorophyll gradient are examined. The angle $\beta$ is smaller than 20$^o$ in almost half (43%) of the pixels and is larger than 70$^o$ in only 9% of them.

This alignment of the chlorophyll front along unstable manifolds confirms that the meandering of this front primarily results from transport barriers stemming from the geostrophic advection. In Fig. 9, two snapshots of the same chlorophyll front 15 days apart (corresponding to the front around 19$^o$W/38$^o$N in Fig. 7c) provide an opportunity for exploring the time evolution of the chlorophyll front versus the evolution of the unstable manifolds. Figure 9a,b shows that, during the 15 days, the chlorophyll front and the unstable manifolds have both moved toward the south. A similar coherency is found with the SST front (Fig. 9c,d). Incidentally, one can also note that smaller scale structures can be seen along the fronts in the chlorophyll and SST images and are not captured by the manifolds.

In order to further examine the role of 2D geostrophic advection in shaping the chlorophyll front, we eliminate other processes (such as phytoplankton growth and decay, and vertical advection) by simulating the evolution of synthetic, purely passive particles. These passive particles are initially organized as a large-scale meridional front (not shown). This front resembles crudely the chlorophyll distribution at the large scale (c.f. Fig. 1b) in the initial date of the simulation (March 26, 2001). The passive particles are solely advected by the horizontal geostrophic velocities, derived from the satellite altimetry data. The evolution of the particles (Fig. 9e,f) shows a clear correspondence with the fronts of chlorophyll and SST at the mesoscale.

### 4.2. Chlorophyll spirals

Figure 7 shows several examples of cyclonic and anti-cyclonic eddies with chlorophyll levels higher than their surroundings: two cyclones on Fig. 7a located at 21$^o$W/43$^o$N and 22$^o$W/47$^o$N, one anticyclone on Fig. 7d at 18$^o$W/47.5$^o$N, and another anticyclone on Fig. 7e at 20.5$^o$W/45.5$^o$N. The chlorophyll distribution within these eddies is not homogeneous. It often has a spiral like shape, and occasionally a spot at the center, particularly remarkable in Fig. 7d.

As depicted in Fig. 5b, the time-dependency of the velocity field is enough to generate inward and outward flows toward and from geostrophic eddies, which have the shape of spirals. This spiral flow can either introduce relatively rich waters into the eddy, thus creating a positive anomaly within the eddy, or vice-versa.

In order to test whether this mechanism of horizontal enrichment of eddies could apply to the positive anomaly located at 21$^o$W/43$^o$N in Fig. 7a, Fig. 10 zooms over this eddy and shows the evolution of synthetic, passive particles initially organized as a front north of the eddy (Fig. 10a). This integration shows that, within the time scale of a month, the particles...
are entrained within the eddy in a spiral manner (Fig. 10c) that resembles the chlorophyll spiral seen in the data (Fig. 10d). Note that the shape of the spiral is also controlled by the location of the unstable manifolds (Fig. 7a and 10c).

This example highlights how a large-scale front of chlorophyll can cascade to smaller scales by filamentation into spirals that eventually enrich eddies. In this particular example, the eddy is located to the south of the large scale chlorophyll front and the unstable manifolds associated with the eddy are crossing it. This allows cross-front exchange and penetration of chlorophyll to the south of the front.

Small spots at the core of some eddies (either as a positive anomaly as in Fig. 7d or as a negative anomaly as in Fig. 7a) are shielded against mixing with the exterior by the spiral flow. Indeed, the discrepancy between trajectories and streamlines in the case of a non-stationary velocity field is more pronounced at the eddy periphery than at its core, since the inner core is less affected by the time-dependency of the velocity field. This leads to the formation of an impermeable kernel, as discussed in Sec. 3.2.

4.3. Chlorophyll filaments

Lastly, we examine the particular case of a chlorophyll filament located over a hyperbolic point (Fig. 11). In this location, the flow field exhibits a large strain that stretches passive tracers along the unstable manifold and contracts it along the stable manifold (Fig. 11a,b). Consequently, the chlorophyll distribution forms a filament along the unstable manifold (Fig. 11c). This raises the question of the source of the initial positive chlorophyll anomaly that is being stretched. At first glance, from the geostrophic velocities (Fig. 11d), it seems that this filament is not fed by the large chlorophyll values found in the northwest corner, because the velocities are directed toward the northwest. The filament is not inside a lobe of the unstable manifold, that is a mechanism by which an intrusion against the instantaneous velocity field is possible (see Fig. 5a). We are thus left with two explanations. The first is intermittent Ekman transport acting against the geostrophic flow. Ekman transport can displace a patch of chlorophyll over the unstable manifold, generating a filament by the mechanism demonstrated in Fig. 5c. Indeed, this scenario is supported by ECMWF wind stress analysis (Fig. 11e), that shows strong wind activity in the days before the formation of the filament. The second explanation for the filament formation is the mechanism discussed in Sec. 3.3, i.e. frontogenesis and ageostrophic circulation in the region of an hyperbolic point, with consequent local formation of chlorophyll by sub-mesoscale upwelling of nutrients along the unstable manifold. This scenario is supported by SST data (Fig. 11f), that show strong temperature gradients in the same region of the filament. An equation for estimating the vertical

Figure 11. a,b) Time evolution of synthetic passive particles (green dots) advected by the geostrophic velocity field. The particles are initiated on June 24, 2001 as a patch around the hyperbolic point at 20°W/45°N; c) chlorophyll (colors, same scale as in Fig. 7f); d) Geostrophic velocity field (arrows); e) wind stress (arrows, the frame marks the boundaries of the other images in this figure); f) superposition of chlorophyll (contours) and SST (colors). The data are from June 30, 2001. The red and black lines correspond respectively to the strong stable and unstable manifolds.
velocity \( w \) at a depth of 200m in the Northeast Atlantic from the density anomalies is proposed in [Legal et al., 2006]: \( w = -250\Delta \rho \text{ m/day} \) where \( \Delta \rho \) is the density difference over a typical lengthscale of 10 km. Figure 11f shows a thermal gradient of about 0.5 °C. Assuming typical thermal expansion coefficients of the order of 0.1-0.2/°C, we get to an estimation of \( w \) in the range of 25 m/day, corresponding to an energetic vertical cell capable of nutrient upwelling. We conclude that both Ekman transport and frontogenesis contributed to the filament formation. This is a typical case in which both mechanisms suggested in Sec. 3.3 act together for the formation of the same pattern.

5. Discussion

In this study, satellite data were used to investigate how the large scale phytoplankton spring bloom in the Northeast Atlantic is modulated by the mesoscale dynamics. Despite the region being an area of low EKE, mesoscale stirring by the geostrophic currents is found to generate mesoscale and sub-mesoscale chlorophyll structures.

Here, we used two independent satellite data sets: chlorophyll and SLA. From the latter, the geostrophic velocities are derived. From the velocity filed, we use the FSLE technique to locate the unstable manifolds of the Lagrangian chaotic flow. These manifolds are expected to act as strong sticking barriers for the 2D advection of passive tracers, and indeed, we found good agreement between the manifolds and the chlorophyll structures, indicating that 2D stirring is responsible to a large extent for the formation of mesoscale chlorophyll patterns.

The formation of small scale structures in tracer’s distribution by horizontal stirring requires an initial source of spatial heterogeneity. In most cases observed here, the source of heterogeneity is the large scale meridional chlorophyll gradient associated with the northward propagation of the Northeast Atlantic bloom. In these cases, the formation of mesoscale and sub-mesoscale patterns result from the direct cascade of the chlorophyll variance injected at large scale. While this process has been simulated with an idealized 2D-turbulence model including the evolution of biological tracers [Abraham, 1998], this work provides, to the best of our knowledge, the first observational evidence of this cascade process.

The cascade appears as a modulation of the large scale chlorophyll front and in intrusions of chlorophyll rich structures (either eddies or filaments) from the north to the south (or inversely, intrusions of chlorophyll poor structures from the south to the north). These intrusions reach up to 3° in latitude and act as an efficient diffusive mechanism for the large scale. The importance of this diffusion mechanism mediated by mesoscale eddies has been brought up for idealized tracers and nutrients in eddy-resolving experiments [Lee et al., 1997; Lee and Williams, 2000].

An interesting finding is that the 2D cascade of the initial large-scale gradient can create chlorophyll anomalies within mesoscale eddies (both cyclonic and anti-cyclonic). The proposed mechanism involves water exchange between the interior of the eddy and its surroundings, and is made possible by the chaotic stirring and mixing, stemmed by the time dependent nature of the. This mechanism also explains the formation of spirals within eddies. The inner core of the eddy is protected against mixing between the interior and the exterior. Note that this is also consistent with floats and subsurface water masses analysis from POMME field measurements. The protection of the inner core explains the small spots of high chlorophyll observed on some of the images: they are not mixed with the surrounding poorer waters. The formation of positive chlorophyll anomalies within eddies through 2D cascade were shown in recent model studies [Lévy and Klein, 2004; Lévy, 2003]. Another commonly referred mechanism to explain positive chlorophyll anomalies within eddies is the "eddy pumping" which involves a vertical transport of nutrients through the doming of isopycnals [McGillicuddy et al., 1998].

Here, comparing the data with the simulation of purely 2D passive tracers evolving from a large-scale front, we show that the observed positive anomalies can result from the 2D cascade. This is in agreement with Uz and Yoder [2004] who, based on the relationship between mesoscale anomalies of satellite derived chlorophyll and SST, found that the most important
influence of mesoscale motion on the distribution of chlorophyll is advection of the existing gradients.

Our results suggest that chlorophyll variance is also injected at small scales within specific areas of the flow associated with hyperbolic points as local events of upwelling. This mechanism is supported by recent numerical studies, which show that vertical motions within stirring regions can have a significant impact on the vertical injection of any tracer characterized by a strong vertical gradient [Mahadevan and Archer, 2000; Lévy et al., 2001; Martin et al., 2002; Lévy and Klein, 2004].

Note that another potential importance of these vertical motions is their effect on the re-stratification of the upper layers of the ocean [Lapeyre et al., 2006]. The effect can be dominant in pre-bloom conditions as shown by the numerical study of [Lévy et al., 2005a] for the POMME area, but is negligible in bloom and post-bloom conditions.

Based on the comparison with chlorophyll images, we deduce that application of the FSLE method on satellite altimetry data is adequate for the study of mesoscale patterns and in some cases even for sub-mesoscale transport processes. The FSLE method is shown to be a useful tool to identify transport barriers and hence to affect chlorophyll pattern formation. The fact that the area is characterized by rather low EKE values, suggests that the method would be even more efficient in a region of high EKE.

There are three main limitations to the approach described here. The first limitation is the relatively low resolution (in space and in time) of the altimetric data from which we derive the geostrophic velocities. While the Lagrangian analysis allows us to detect sub-mesoscale filaments coming from the chaotic stirring, some other small scale features are clearly not resolved (see for instance the small meandering of the chlorophyll and SST fronts in Fig. 9). In some cases the low time resolution in altimetry poses a difficulty in accurately localizing even mesoscale structures when the geostrophic field evolves faster than the one-week sampling rate. The second limitation is that, when estimating the surface velocity from altimetry we did not take into account quantitatively ageostrophic components, such as the secondary circulation described qualitatively in Fig. 6 or the Ekman transport associated with the wind. Finally, in this study we do not take into account the impact of biological activity that is known to have an important role in modulating plankton distribution at the mesoscale [López et al., 2001; Martin et al., 2002; Srokosz et al., 2003]. These limitations explain some of the mismatch that sometimes appears between fronts and manifolds.

The fact that both horizontal transport and vertical upwelling act in phase, strengthening chlorophyll fronts over unstable manifolds, is an advantage for our analysis, supporting the correlation between chlorophyll fronts and manifolds for both post-bloom and locally produced chlorophyll. On the other hand, the same argument shows that passive horizontal advection and vertical production are intrinsically entangled, and that it is not possible to resolve the effect of the two by an analysis of chlorophyll and geostrophic velocities images alone. A natural extension of our results is therefore the integration of a Lagrangian analysis with biogeochemical models and with vertical velocity estimation by density anomalies and wind stress. A similar combined approach should be able to attempt a quantitative prediction of chlorophyll gradients and a parameterization of transport on plankton bloom.

Appendix A: Manifold detection by Lyapunov exponent calculation

Let us come back to Fig. 4a, that is the typical situation one has in mind when probing a field with Lyapunov exponent calculation. A tracer initialized in the vicinity of the stable manifold (green dots) is advected toward the equilibrium point where it is subjected to the stretching effect of the unstable manifold. Conversely, for a tracer initialized in a region dominated by re-circulation, the stretching effect is smaller. The simplest way for quantifying the stretching is to consider the maximum growth rate among two points in the tracer volume. For the case of the green points, the growth is given by the exponential expansion due to the unstable manifold. Calling the initial separation \( \delta_0 \), the separation after time \( \tau \), and
the positive eigenvalue of the equilibrium $\lambda$, the growth is expressed by:

$$\delta_r = \delta_0 e^{\lambda \tau} \quad (A1)$$

By inverting Eq. A1, one can obtain $\lambda$ from the tracer experiment of Fig. 4a:

$$\lambda = \frac{1}{\tau} \log(\delta_r / \delta_0). \quad (A2)$$

Equation A2 in fact defines a way for probing the space and looking for the stable manifolds of strong equilibrium points. At any point (longitude, latitude) and at any time $t$, we can initialize a volume (that is a surface for a 2D case), advect it, and by using Eq. A2 we can measure a growth rate $\lambda$. All volumes initialized over the stable manifold go toward the equilibrium point where experience the same stretching and thus (except for a small transient) provide the same value for $\lambda$. The stable manifolds appear in the $\lambda$ field as lines of local maxima. The unstable manifolds can be also detected with the same algorithm, by integrating the velocity field backward in time. Equation A2 corresponds to the Lyapunov exponent in the limit field backward in time. Equation A2 is obtained from Eq. A1, setting the time of integration $\tau$.

One possibility (finite-time Lyapunov exponent) is to prescribe $\tau$ a priori, using the same value for all the points; another possibility (finite-size Lyapunov exponent) is to prescribe $\tau$ implicitly, setting the value of $\delta_r$, and finding for each initial condition the time $\tau$ when points initially separated by $\delta_0$ reach a final distance $\delta_r$. From a theoretical viewpoint finite-size and finite-time Lyapunov exponents are quite different. Moreover, by scanning a in $\delta_0$, the finite-size Lyapunov exponents provides a powerful tool for probing a turbulent field at different scale (see e.g. [Artale et al., 1997]). However for the use that we make here (the detection of the manifolds) they are similar. The FSLEs are slightly better for our application, in the sense that they allow to tune the time of integration depending on the local properties of the field. In fact there is a trade-off for the choice of $\tau$: a large time allows to obtain a more precise value for $\lambda$ (by weighting more the stretching in the vicinity of the equilibrium point) but can also be more affected by the time variation of equilibrium and manifolds. The FSLEs allow to use a small integration time whenever a strong hyperbolic structure is encountered, since in this case the final separation $\delta_r$ is reached faster.

Acknowledgments. This study was supported by the Centre National d’Études Spatiales (CNES) and MERCATOR. F. d’O. is a Marie-Curie fellow, grant 024717-DEMETRA. Carine Lesage is thanked for processing of SeaWiFS data and Youssuf Amar for processing of AVHRR data. Eric Greiner and Marie-Hélène Rio are acknowledged for providing the RIO-5 MSLH MDT. E. H. and Y. L. are grateful to the Porter School of Environmental Studies at Tel-Aviv University (grant 0603413101). Special thanks are to the Goddard Space Flight Center (GSFC/NASA) for providing the SeaWiFS images used in this work. The altimeter products were produced by Ssalto/Duacs and distributed by Aviso, with support from CNes.

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3.2 Conclusions and perspectives

Horizontal stirring by geostrophic surface currents was shown to have an important effect on chlorophyll distribution during the period of the spring bloom. This effect is responsible for the formation of spatial patterns over a wide range of scales, including small scale chlorophyll spirals and filaments and large scale chlorophyll fronts.

The transport properties of the geostrophic velocity field were characterized by applying Lagrangian analysis tools that included calculation of unstable manifolds and advection of synthetic passive tracers. These tools were found to be highly efficient in studying the role of the surface currents in controlling the distribution of passive tracers. Nevertheless, several limitations should be taken into consideration. These limitations explain the mismatches between the orientations of the strong unstable manifolds and high chlorophyll gradients (LOL07, Fig. 8). The first limitation is the relatively low resolution (in time and in space) of the altimetry data from which we derive the geostrophic velocities that prevents from detecting some of the sub-mesoscale and even mesoscale patterns. A second limitation is that, when estimating the surface velocities from altimetry we did not take into account ageostrophic components. Finally, in this study we did not take into account the impact of biological activity that is known to have an important role in modulating plankton distribution at the mesoscale [Srokosz et al., 2003, Martin et al., 2002, López et al., 2001]. This limitation can be overcome by integration of the Lagrangian methods applied here with biochemical models.

The importance of horizontal transport in modulating chlorophyll distribution implies that it also has an important impact on the properties of the subducted mode waters. The main aspect that should be further investigated is the advection of nutrients towards the area of subduction, associated with the area of strong winter MLD gradient. This could be done by combining the methods described here with MLD data from numerical models (see chapter 2 for a detailed description).

Another mechanism that may have an important contribution to primary production and should be more thoroughly investigated is the nutrient injection by vertical motions within sub-mesoscale filaments [Lévy et al., 2001]. Such local events of vertical motions are often found in specific areas of the flow that can be identified by Lagrangian analysis of the velocity field. Since the efficiency of the mechanism also depends on the MLD, combining information from ocean color images (for identification of local chlorophyll anomalies), Lagrangian analysis of the velocity field (for identification the relevant areas of flow) and MLD from models provide an adequate framework for such investigation. Furthermore, such a combined approach would allow identification of the contribution of these local upwelling events to the productivity of the subducted waters.
Chapter 4

Recovering and analysis of small scale chlorophyll variabilities

4.1 Extended abstract

As shown in the previous chapters, chlorophyll images derived from satellite measurements of ocean color provide a powerful tool for the study of phytoplankton variability. However, the use of this satellite data is severely limited by the presence of clouds that often mask large parts of the images, thus reducing the amount of available data and preventing continuity in time and in space. The problem of missing data mainly hinders the study of small scale processes and prevents a systematic analysis of small scale variabilities.

A standard approach used for limiting the effect of clouds is the construction of image composites. This method has the drawbacks of degrading both the spatial and temporal resolution of the satellite data proportionally to the number of images needed to remove clouds, filtering out small-scale features in an uncontrolled way. Another approach is to recover missing data by interpolating over the gaps in the remotely sensed data.

4.1.1 Geostatistics and kriging

In the paper "Recovering small scale variabilities in cloudy ocean color images using kriging" [Lehahn et al., in preparation, hereafter LOD07], which is in its final stages of preparation, interpolated high resolution (1km) chlorophyll images from SeaWiFS are used in a systematic analysis of the impact of mesoscale variabilities on the annual chlorophyll cycle. The interpolation is done by applying the kriging method. Although kriging is commonly used in many geophysical fields, to our knowledge it was never used for the reconstruction of chlorophyll patterns in ocean color satellite images.

Kriging is one tool of geostatistics, a branch of applied statistics that focuses on the modeling and estimation of spatial patterns. Geostatistics is associated with a class of techniques used to analyze and predict values of variables distributed in space. Such
values are implicitly assumed to be correlated with each other, with the expectation that, statistically, samples close to each other are more similar than those that are further apart. Before using any geostatistical estimation method, this spatial autocorrelation must be inferred using spatial continuity tools, that measure the strength of correlation among the samples or their similarity or dissimilarity with distance. A geostatistical tool that summarizes the spatial continuity for all possible data pairings for a given range of distances between point is the variogram [Rossi et al., 1994].

The kriging method relies on quantification of the spatial variability to provide estimates for unsampled locations by taking a weighted linear average of available samples. With its internal quantification of spatial variability through the variogram, kriging is able to produce maps of predictions and the associated prediction errors from incomplete and noisy spatial data. The reproduction of the variance is not part of the criteria searched for in kriging. It is indeed well known that kriging "smoothes" the signal, so that the histogram of the values estimated will display more values around the mean and less extreme values than the histogram of the observations [Biau et al., 1999].

Here we first test the efficiency and reliability of kriging to recover missing data in high resolution ocean color images. This is done using 20 relatively cloud free chlorophyll images on which we impose cloud masks that are randomly taken from the dataset of 6 years (1998-2003) SeaWiFS images of the study area. Altogether 1000 combinations of chlorophyll images and cloud masks were used. The artificially masked images are used for testing the interpolation error, by comparing the difference between chlorophyll values in the masked images after they are interpolated with the values in the original cloud free images.

4.1.2 Main results

In general kriging gives remarkable results in recovering missing chlorophyll values. Furthermore, despite its characteristic of smoothing the spatial signal, the kriging method allows a successful reconstruction of mesoscale chlorophyll patterns even in cases of extensive cloud coverage (LOD07, Fig. 1e). An exception was in the case of large compact clouds that significantly reduce the number of nearby points used for the interpolation, therefore degrading strongly the performance of kriging and prevailing a reliable estimation of the interpolation error (LOD07, Fig. 1f). The problem of compact clouds was approached by applying a criteria for patchiness, using a structure function provided by the interpolation procedure itself and discarding all reconstructed images that did not match the criteria. This solution ruled out 44% of the interpolated images.

In the remaining images interpolated values in the masked images were very close to the values in the cloud free images. The interpolation error of log(chlorophyll) was smaller than 0.18 (corresponding to calibration and validation objective for SeaWiFS data of pro-
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ducing chlorophyll concentrations within ±35% over the range of 0.05-50mg/m³) in 80% of the interpolated pixels. Furthermore, in these images the error estimation embedded within kriging is correlated with the real error (the difference between chlorophyll values in the interpolated image and the cloud free images), therefore providing a reliable evaluation of the interpolation error for a given pixel (LOD07, Fig. 2).

After evaluating the efficiency and reliability of the method in recovering chlorophyll patterns in high resolution ocean color images, we applied the kriging method for the construction of a dataset of uninterrupted daily chlorophyll images. This dataset was used for studying small scale changes in chlorophyll evolution throughout the year. More precisely we investigate the impact of horizontal advection by the mesoscale dynamics (that was described in details in chapter 3) on the annual chlorophyll cycle. This is done by extracting time series of chlorophyll concentrations in two boxes of 3 x 3 km, 150 km apart (LOD07, Fig. 3). Here we present time series of chlorophyll evolution throughout 1999 from two locations at the vicinity of the boundary between the midlatitude and subpolar regimes. The two time series include values from 70 images. Although they are located only 150 km apart they show significant differences in the annual chlorophyll cycle. The main difference is in the duration of the bloom that is about two months longer, extending into summer, in the northern time series. This difference is attributed to an intrusion of rich waters from the productive region in the north to the oligotrophic region in the south that reaches to about 42°N, north of the southern station. The intrusion is associated with the mesoscale eddy centered around 21°W/43°N, through the mechanism of horizontal transport that was discussed in chapter 3.

It is important to note that this analysis would not have been possible without using the interpolated data since the time series was too sparse to have robust conclusions concerning small scale variations in the annual chlorophyll cycle.

These results emphasize the important role of the mesoscale dynamics (through horizontal transport) on large scale production patterns, mainly in areas that are adjacent to the boundaries between contrasted biogeochemical regimes. In the example presented here, in addition to the direct impact on the annual chlorophyll, the mesoscale dynamics leads to a secondary consequence of shifting the boundary between the midlatitude and subpolar regimes.
Recovering small scale variabilities in cloudy ocean color images using kriging

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The possibility of using satellite-based sea color images at the nominal space and time resolution is often hindered by clouds, that force to limit the analyses to hand picked, cloud free cases, or to construct composites at degraded resolution. Focusing on the NE Atlantic, here we test the reliability of the kriging technique (also known as optimal interpolation) for reconstructing contiguous, high resolution datasets from SeaWiFS observations. By extracting cloud masks and applying them to cloud-free images, we construct a large set of cases where we assess the the reliability of kriging in reconstructing the pixel values and to estimate the error for different cloud coverage and patchiness. As an example of application, a case of small-scale phytoplankton variability induced by mesoscale eddies is studied by extracting a time series of chlorophyll concentrations over an annual cycle at two 3x3km regions 150km apart.

1. Introduction

Recent studies show with growing evidence that phytoplankton variability at the mesoscale and sub-mesoscale contributes significantly to global primary production budgets (see [Lévy, 2006] for a review). High resolution modeling studies suggest that incorrect representation of mesoscale and sub-meso variability in ocean global circulation models (OGCM) can result in errors of near 30% [Oschlies and Garcon, 1998; Mahadevan and Archer, 2000; McGillicuddy et al., 2003] in primary production estimations. This rises the necessity of synoptic observations that characterize the mesoscale and sub-mesoscale phytoplankton variability and hence allow parameterization and validation of the OGCM.

The only feasible means for obtaining such detailed observations is provided by high resolution ocean color sensors as the Sea-viewing Wide Field-of-view Sensor (SeaWiFS) and the Moderate Resolution Imaging Spectroradiometer (MODIS), that allow a global characterization of small scale chlorophyll variability patterns [Yoder, 2000]. However, the exploration of these data is strongly limited by the presence of clouds that mask part of the ocean’s surface and prevent space and time contiguity. The lack of contiguity has various and severe effects on the analysis of space and time variability. A first example is provided by spectral analysis [Denman and Platt, 1976]. Spectral analysis (and similar methods) have been shown able to extract important information of space and time plankton variability, but require uninterrupted datasets. The presence of cloud in this case often limit a systematic analysis, since a careful human-based choice of cloud-free space and time regions (see e.g. [Denman and Abbott, 1994]) is usually required.

A second important example of a problem affected by cloud coverage is provided by the integration of satellite data with in-situ measurements. The comparison of sea color images with other in-situ parameters is a fundamental step for the calibration of the satellite data and for our understanding of the biogeochemical cycles. However, in-situ measurements have typically a very sparse and localized coverage, strongly constraining the space and time coordinates of the satellite observations. Their
integration with satellite data is therefore extremely vulnerable to the presence of clouds.

A standard approach used for limiting the effect of clouds is the construction composite images. By aggregating over time (e.g., two or more consecutive days), the data are less sparse. This method has the drawbacks of degrading both the spatial and temporal resolution of the satellite instrument proportionally to the number of images needed to remove clouds, filtering in an uncontrolled way small-scale features.

In this paper we explore the possibility of recovering missing data in cloudy ocean color images using the kriging technique, which is also known as optimal interpolation [Emery and Thomson, 1998]. Historically, kriging is a generic name adopted by geostatisticians for a family of generalized least-square regressions algorithms that share the objective of minimizing the estimation or error variance under the constraints of unbiasedness of the estimator [Wackernagel, 1998]. Kriging is commonly used in the field of satellite image analysis Cressie et al. [2006] and for the reconstruction of uninterrupted space and time datasets in different geophysical fields as oceanography [de Boyer-Montégut et al., 2004], ecology [Hückstädt and Christina, 2004] and geology [Goovaerts, 1997].

Quite surprisingly, and although geostatistical tools have already been used for characterizing variability patterns in SeaWiFS data [Doney et al., 2003; Fuentes et al., 2000], to our knowledge kriging has never been used before for recovering missing data in ocean color images.

The study is focused on a 20° latitudinal band in the northeast Atlantic (13° – 23°W/30° – 50°N). That has been the subject of many model and campaign studies [Méromy et al., 2005]. The domain is characterized by a sharp transition between areas of deep winter mixed layer depth (MLD) in the north and shallower MLD in the south [de Boyer-Montégut et al., 2004]. This leads to variations in the convective supply of nutrients and produces a large scale gradient in productivity, with relatively productive waters in the north and oligotrophic waters in the south [Lévy et al., 2005]. The difference in productivity is expressed by a meridional chlorophyll front that interacts with mesoscale eddies giving rise to a strong mesoscale and sub-mesoscale phytoplankton variability [Lehahn et al., 2007].

We first test the performance of the interpolation and error estimation on a library of 20 cloud-free SeaWiFS images to which we overimpose real cloud masks (a total of 1000 masks). As an example of application, we then apply kriging over a full year of daily images and show the capability of the technique to allow the extraction of time series of chlorophyll evolution with apparent small-scale variabilities. In particular, we show the dramatic effect of a mesoscale eddy on

2. Data and method

Daily images of chlorophyll concentrations were processed from high resolution (1 km) Level 2 SeaWiFS data, obtained from the NASA Goddard Distribution Active Archive Center (DAAC). Data were processed using NASA’s SeaWiFS Data Analysis System (SeaDAS software package). Because chlorophyll concentrations display a log-normal distribution [Campbell, 1995], the values were log-transformed.

We used 500x500 km high resolution (1 km) ocean color images from SeaWiFS between the years 1998-2003 in the NE Atlantic window 13° – 23°W/30° – 50°N. For studying the reliability of kriging we selected a set of 20 500x500 km relatively cloud-free (1-30% cloud coverage). The images were masked by clouds that were randomly taken from SeaWiFS images in the same area, sampling uniformly in space and in time. Altogether 1000 clouds masks with cloud coverage ranging from 1% to 97% were applied. The artificially masked images are used for testing the interpolation error, which is defined as the difference between chlorophyll values in the interpolated image and the cloud free images.

The missing information in the chlorophyll data is recovered by ordinary kriging [e.g. Wackernagel, 1998]. The kriging method relies on the calculation of a semivariogram that measures the local spatial variation of a random field by describing how sample data are related
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with distance and direction [Journal and Huijbrechts, 1978]. By using the semivariogram approach for characterizing chlorophyll variability patterns we follow [Doney et al., 2003; Fuents et al., 2000] who also give a detailed description of the method. Based on this quantification of the spatial variability, kriging provides estimates for unsampled locations by taking a weighted linear average of available samples. With its internal quantification of spatial variability through the semivariogram, kriging methodology is able to produce maps of optimal predictions and associated prediction of standard errors from incomplete and noisy spatial data.

The computation was performed in the R Free Software environment. R is an integrated suite of software facilities for data manipulation, calculation and graphical display [R Development Core Team, 2006]. The interpolation was done by using the gstat program library [Pebesma, 2004]. In order to speed up computation, the variogram calculation was modified using the algorithm proposed by [Marcotte, 1996].

For the exemplificative use of kriging, a dataset of consecutive daily images has been constructed for the year 1999.

3. Results

Figure 1 shows an example of reconstruction for an originally cloud-free image artificially covered with two different cloud masks. The chlorophyll pattern in the cloud-free image (Fig. 1a) shows characteristic mesoscale and submesoscale structures of the northeast Atlantic spring bloom. The pattern is originated by the interaction of a mesoscale eddy (centered around 21°W, 43°N) with the large-scale chlorophyll front along the 45° meridian. Two cloud masks that cover 77% and 93% of the image (Fig. 1b and 1c respectively) are artificially applied. The two masks have been chosen in order to show the effect of cloud patchiness on the ability of quality of interpolation. They represent contrasted cloud distribution patterns: i) a patchy distribution where the clouds are clustered into a large number of relatively small patches (Fig. 1b) and ii) a homogeneous distribution where the clouds are clustered into a small number of large patches (Fig. 1c).

As can be seen in Fig. 1e, kriging works remarkably well in recovering data that is masked by clouds organized as relatively small patches. Although most of the image area is masked, there is enough available information for recovering the large scale, mesoscale and even part of the sub-mesoscale chlorophyll patterns. The reconstruction is relatively less effective for the case of the more homogeneous clouds. Nevertheless, the mesoscale structure of the chlorophyll intrusion is correctly reproduced with correct quantitative values (Fig. 1f). Notice that for the homogeneous mask a small number of cloud-free pixels exist, that allow the mesoscale reconstruction. Such cloud-free pixels can be

![Figure 1](image-url)

Figure 1. a) Original and b,c) masked (77% and 93% of the pixels respectively) SeaWiFS chlorophyll image from July 7 1999. d) Number of pairs of cloud-free pixels as a function of distance between points for the images in Fig. 1a,b,c (black, blue and red points respectively) divided by the mean value of all distances. e,f) The resulted images after recovering the missing data from the image in Fig. 1b,c, respectively. The two black points in Fig. 1a mark the locations of the time series in Fig. 3.
essentially recognized in the reconstructed image as well, forming the vertices of the triangular chlorophyll tiles reconstructed by kriging.

The reconstructed values have two sources of errors. The first one is the variability among the values used for the interpolation. The second one is the presence of large patches that may drastically reduce the number of nearby points used for the interpolation. Kriging directly takes into account the first source, by producing a per pixel error estimation based on a weighted standard deviation. The second error component dominates in the case of homogeneous cloud patches (as shown in Fig. 1d) but cannot be computed directly, since it depends on the values of the unknown nearby pixels. While there may be a seasonal correlation between total coverage and homogeneous cloud patch, total coverage cannot be used as an indicator of this second error source, since a large patch may occupy a relatively small part of the image. An indirect estimation of the error induced by patchiness can be done in the following way. When computing the weights for interpolating an image, kriging calculates the number of pairs $P(d)$ of cloud-free pixels as a function of distance $d$ between points (Fig. 1d). By defining a mask $M(x)$ for the cloud-free pixels ($M(x) = 1$ for cloud free pixels and $M(x) = 0$ in the presence of a cloud), it becomes apparent that the function $P(d)$ is in fact a discrete autocorrelation function for the cloud mask: $P(d) = \sum_{|x_1-x_2|=d} M(x_1)M(x_2)$.

For this reason, the scales at which the cloud mask affects the reconstruction by kriging can be analyzed looking at $P(d)$ (see Fig. 1e for the comparison of $P(d)$ for the three cloud masks). Several diagnostics of cloud patchiness can be defined over $P(d)$. Leaving a detailed analysis to a future publication, we consider here the position $d_M$ of the maximum as an indication of the cutoff induced by presence of very large cloud patches. Based on a careful examination of the images we choose a value of $d_M = 240$ km as a threshold below which a masked image is considered as non-recoverable. Application of this threshold rules out 44% of the images. The exclusion of interpolated images that are characterized by $d_M$ smaller than 240 improves significantly the reliability of the error estimation embedded within the kriging technique.

This is emphasized in Fig. 2 where we plot the interpolation error (defined as the difference between chlorophyll values in the interpolated image and in the cloud free image) vs. the error estimated by kriging for all the masked pixels. When taking into account only recoverable images (corresponding to $d_M \geq 240$, circles) the kriging estimated error is proportional and slightly smaller than the real error, thus providing a reliable measure to it. On the other hand when using all images (crosses), pixels that are badly reconstructed (i.e. pixels located in compact clouds) are associated to low values of estimated error. This result from the fact that the large cloud patches reduces significantly the number of pixels used for the calculation of the estimated error.

In order to evaluate the quality of the interpolation we count the number of pixels in which the error in estimating log chlorophyll concentrations is smaller than 0.18. This value corresponds to the calibration and validation objective for SeaWiFS data of producing chlorophyll concentration within ±35% over the range of 0.05-50mg/m³ (Hooker and McClain, 2000). After excluding the non-recoverable images the interpolation error of log chlorophyll is smaller than 0.18 in 80% of the pixels.

The information that is made available by recovering the missing data with kriging is used for studying small scale variabilities in chlorophyll distribution and evolution. This is done by extracting time series of surface chlorophyll evolution during 1999 at two locations (hereafter referred to as northern and southern stations) 150 km apart (Fig. 3). The values are taken as the average over a box of 3X3 km, which is small enough to prevent variabilities within the averaged area. The two time series include values from 70 images characterized by $d_M$ larger or equal to 240 (circles). In order to show the importance of having the additional information from the recovered data, we also show time series that are extracted from the non-interpolated images (crosses). The signal from the reconstructed data is much less sparse and reveals small scale patterns that can not be identified in the time series extracted.
4. Discussion

In this paper, we analyzed a dataset of high resolution SeaWiFS images for the study of small scale chlorophyll variabilities in the northeast Atlantic. The exploitation of this data set is severely limited by the presence of clouds that mask a large part of the images area and prevent a systematic analysis of small-scale features. This problem is addressed by recovering missing data using the kriging technique. Such technique, currently used in many other geophysical fields, is able to reconstruct a contiguous dataset from sparse observations, also providing error estimations. The technique and the error estimation embedded within it performed remarkably well for patchy clouds even in the case of extreme coverage, since the presence of nearby isolated cloud-free regions are sufficient for reconstructing the gradient in between. On the other hand, the performance of kriging is strongly degraded in the presence of large, homogeneous clouds, that appeared as the main limitation of kriging. Since the algorithm interpolates without forcing small-scale structures, regions covered by large and compact clouds are recovered as smooth gradients. In such regions, pixels may have values possibly close to the real ones, but obviously the small-scale spatial variability is lost. A second problem related to large compact clouds, is that such regions are reconstructed by interpolating sparse points. For this reason, the kriging error estimation is in this case unreliable. We approached the problem of compact clouds only from the original (non interpolated) images, which consist of only 22 (northern station) and 23 (southern station) time steps.

Both in the northern station (Fig. 3, upper panel) and southern station (lower panel) the evolution of surface chlorophyll concentrations is characterized by a steady increase during autumn and winter, reaching maximum values at the end of April. These changes in chlorophyll concentrations throughout the year are driven by seasonal variation in the depth of the mixed layer [Lévy et al., 2005]. Following the annual chlorophyll peak the two time series differ significantly (notice especially the period between the two vertical lines): Whereas in the southern station chlorophyll concentrations continuously decrease and the months of June and July are characterized by relatively low and constant values, in the northern station a secondary increase at the beginning of June leads to a period of relatively high chlorophyll concentrations with a local maximum at the beginning of July. This difference results from an intrusion of chlorophyll rich waters that reaches to about 41.5°N north of the southern station (see Fig. 1a).
by defining a computational effective criterium for patchiness, based on a correlation function computed during the interpolation process and discarding all reconstructed images that did not match the criterium. Other options may be available, like the use of wavelet or inverse Fourier transform for forcing small-scales spatial structures.

After validating the reliability of error estimation, the interpolated data were used for characterizing small scale chlorophyll variability patterns. Comparison between time series of surface chlorophyll concentrations extracted from a series of interpolated images shows the local impact of the mesoscale dynamics on the annual chlorophyll cycle. The effect of horizontal stirring on chlorophyll distribution in this area was studied by [Lehahn et al., 2007] on isolated bloom events. The extraction of time series from interpolated images that we presented here provides the first opportunity to observe the local impact of these transport processes on the entire annual cycle.

Acknowledgments. This study was supported by the Centre National d’Etudes Spatiales (CNES) and MERCATOR. F. d’O. is a Marie-Curie fellow, grant 024717-DEMETRA. Special thanks to the Goddard Space Flight Center (GSFC/NASA) for providing the SeaWiFS images used in this work. Carine Lesage is thanked for processing of the SeaWiFS data. Discussion with J. Deshayes and C. Lathuilière. Discussion with G. Lapeyre and J. Iern-Foontanet. R routines for optimal interpolation.

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4.2 Conclusions and perspectives

In this part of the work we showed, to our knowledge for the first time, the efficiency and reliability of the kriging method to recover missing data in high resolution ocean color satellite images. Application of this method opens the way for constructing datasets of uninterrupted high resolution chlorophyll images and increases significantly the possibility to characterize small scale variabilities in the ocean biosphere.

As a primary example we used a set of interpolated ocean color data for studying the impact of the mesoscale dynamics on the annual chlorophyll cycle in the midlatitude regime. Comparison between time series extracted from this dataset shows that the mesoscale dynamics has a crucial impact on the chlorophyll evolution throughout the year. Horizontal advection of chlorophyll-rich waters associated with a mesoscale eddy leads to an increase in chlorophyll concentrations to levels that are of the same order as the annual chlorophyll peak. Furthermore, the high chlorophyll concentrations associated with the mesoscale circulation extend over a period of two months during summer, when the midlatitude regime is highly oligotrophic.

The analysis of the chlorophyll time series presented emphasizes the need for a systematic characterization of small scale variability patterns in primary production, and to study the impact of small scale processes on large-scale seasonal and interannual variabilities. This can be done by extracting and analyzing similar time series from different locations, covering the three northeast Atlantic production regimes and extending over a period of several years. In addition to that, the availability of uninterrupted chlorophyll images opens the way for a systematic analysis of the spectral properties characterizing chlorophyll distribution. This will allow identification of the relative importance of the different scales of phytoplankton variability, and to study the way it changes in time and in space.
Conclusions and perspectives

In this thesis we used SeaWiFS ocean color data to characterize phytoplankton variability in a region of subpolar mode water subduction in the northeast Atlantic. The dynamical mechanisms driving the phytoplankton variability were investigated by comparison with physical fields derived from numerical models and from other satellite measurements. The results from this thesis cover spatio-temporal scales ranging from several to thousands of kilometers and from days to years. They are intended to provide complementary information to that gathered in the framework of the POMME research project that included a series of interdisciplinary field experiments between October 2000 and October 2001.

This thesis has emphasized the large scale variability of the production system over the northeast Atlantic. The large scale variability is almost entirely meridional, and is mostly driven by variations in the depth of the mixed layer that result from changes in the atmospheric forcings. Based on the meridional differences in the seasonal chlorophyll cycling and its relation to the seasonal cycling of MLD, we distinguish between three production regimes: the well known subpolar and subtropical regimes and the newly identified midlatitude regime. In the northern subpolar regime the main features in the chlorophyll signal are a major bloom during spring and a secondary bloom in fall. These blooms correspond respectively to mixed layer shallowing and deepening. In the southern subtropical regime the chlorophyll signal is characterized by a single weak bloom that starts with the deepening of the mixed layer during fall, and reaches its peak in winter when the mixed layer is at its maximum depth. The midlatitude regime is characterized by a single broad bloom, weaker than the subpolar spring bloom and stronger than the subtropical fall bloom, which starts in fall with deepening of the mixed layer and peaks in spring with the mixed layer restratification.

The identification of the midlatitude regime and its specific characteristics is a meaningful result that have important implications in our understanding of the ocean carbon cycle. This importance stems from the fact that the midlatitude regime overlaps the area of maximum gradient in winter MLD, which is the main area of mode water subduction. The fact that winter production is optimum in the midlatitude regime (since in the subpolar regime, winter production is weak due to light limitation, and in the subtropical regime, production is weak due to nutrient limitation), that the midlatitude bloom starts
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prior to subduction and that production is active during winter are likely to reinforce the
efficiency of the biological carbon pump during subduction.

The specific properties of the midlatitude regime, that covers about half of the POMME
area, may also explain some unexpected characteristics of the POMME in situ observa-
tions. An example is the fact that in situ measurements of chlorophyll concentrations
never exceeds 1 mg/m$^3$, while higher concentrations were expected (for example during
the North Atlantic Bloom Experiment (NABE) observed concentration during the bloom
were 3-4 mg/m$^3$). Another unexpected characteristics is the fact that winter is found to
be a period of significant production, not much different than the period of the spring
bloom.

In the subpolar and subtropical regimes production is limited respectively by the
availability of light and nutrients. Consequently production levels in these two regions
are strongly and inversely dependent on the depth of the mixed layer that balance between
the availability of these two limiting factors: whereas in the subpolar regime deeper mixed
layers lead to weaker blooms, in the subtropical regime, deeper mixed layer intensifies the
bloom. This result coincides with the recent work of Behrenfeld et al. [2006] that shows
that due to the dependency on vertical mixing, global ocean productivity is strongly
influenced by climate changes. In the permanently stratified regions of the ocean (such as
the subtropical regime) climatic conditions that lead to surface warming (and consequently
shallower mixed layers) are shown to be accompanied by reduction in productivity.

Our results also show that phytoplankton variability is strongly effected by the process
of direct horizontal cascade of the large scale variability associated with the large scale
meridional chlorophyll gradient. To our knowledge this thesis provides the first observa-
tional evidence of this cascade mechanism that consists of a modulation of the large scale
chlorophyll front and intrusions of chlorophyll rich structures (either eddies or filaments)
from the north to the south. These intrusions reach up to 3° in latitude and act as an effi-
cient diffusive mechanism for the large scale. Indeed, analysis of time series of chlorophyll
concentrations shows that intrusion of chlorophyll rich waters from the subpolar region
has an important impact on the annual chlorophyll cycle in the midlatitude region.

Another aspect of the horizontal cascade, observed here for the first time, is the forma-
tion of chlorophyll anomalies within mesoscale eddies (both cyclonic and anti-cyclonic).
The proposed mechanism involves water exchange between the interior of the eddy and
its surroundings, and is made possible by the chaotic (i.e. time dependent) nature of the
geostrophic velocity field. This horizontal mechanism is an alternative to the commonly
referred "eddy pumping" mechanism that explains positive chlorophyll anomalies within
eddies by vertical transport of nutrients through the doming of isopycnals.

Further analysis of the relationship between the mesoscale dynamics and the large
scale phytoplankton variability would provide important insight regarding the biogeo-
chemical properties of the subducted mode water. Such analysis should first include the
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evaluation of the amount of nutrient rich waters that are transported from the subpolar region to the midlatitude region. Calculation of unstable manifolds and comparison with distribution patterns of chlorophyll and MLD provide an adequate template for such analysis. Additional diagnostics that could contribute to characterization of subducted water properties is a systematic comparison, over several years, between chlorophyll cycles at different locations within the midlatitude regime. Such diagnostics requires the construction of datasets of contiguous chlorophyll images, which is made possible by recovering missing data with the kriging technique.

Another aspect of the mesoscale dynamics that can have an important impact on production in the subducted waters is the input of nutrients through local events of upwelling. This mechanism can be of great importance at the midlatitude regime where a major factor limiting phytoplankton bloom is availability of nutrients. Analysis of the importance of this process for the region's productivity can be done by combining Lagrangian analysis of the geostrophic flow, that would allow identification of specific area where nutrients are likely to be injected, with high resolution images of chlorophyll and SST.

In this thesis we applied three methods that are not commonly used in the field of ocean color satellite imagery. These methods are shown to be very efficient for the study of phytoplankton variability patterns and of the mechanisms driving them, and provide a very useful framework for similar studies in other regions of the ocean.

Our results show that ocean general circulation models have enough precision to interpret spatial and temporal variabilities observed in ocean color satellite images. Since ocean color data is now available on daily and global coverage, this provides a powerful tool for the study of biological and physical interactions in the world’s oceans. Indeed, the success of the method presented here has motivated similar works in the North Atlantic (Lathuilière et al., Seasonal and intra seasonal surface chlorophyll-a variability along the northwest African coast, manuscript in preparation) and in the Indian Ocean [Lévy et al., submitted], in the frame of the CNES project MULTICOLOR..

Lagrangian diagnostics of satellite derived geostrophic currents are shown to be very useful for analyzing transport properties of the velocity field and particularly their effect on the spatial distribution of tracers as chlorophyll. While being a standard tool for the study of transport properties in the atmosphere, in the field of oceanography they have been introduced more recently and their use is still rather limited. Based on comparison with unstable manifolds that were calculated by applying the FSLE technique, we deduced that a large part of the small scale chlorophyll patterns observed in ocean color images of the northeast Atlantic result from horizontal stirring by the time varying velocity field. The unstable manifolds were shown to i) modulate the shape of already existing fronts; ii) create spiraling chlorophyll anomalies within eddies; iii) produce chlorophyll filaments. Our results also suggest that Lagrangian analysis can be used for identification of specific
areas of the flow where chlorophyll variance is likely to be injected at small scales by local events of vertical motion. These areas are associated with strong stirring and are characterized by strong manifold adjacent to hyperbolic points. A natural extension of this work is the integration of Lagrangian analysis with biogeochemical models.

The kriging technique, one tool of geostatistics, is found to be an efficient and reliable tool for recovering missing data in cloudy ocean color images. Although it is commonly used in many geophysical fields to our knowledge it was never applied for the reconstruction of ocean color satellite data. Application of this method opens the way to the formation of datasets of uninterrupted high resolution chlorophyll images and provides the means for systematic analysis of phytoplankton variabilities at scales ranging from the sub-mesoscale to the global scale. In addition to that, the ability of the kriging method that relies on the quantification of the spatial variability, to successfully reconstruct mesoscale chlorophyll structures ensures the reliability of geostatistical techniques in characterizing mesoscale variability in ocean color satellite images. The use of geostatistical techniques for analyzing spatial variabilities in ocean color satellite images, as was done by Doney et al. [2003], is still not common.

The methods proposed in this thesis open the way for more systematic analysis of ocean color variability and of its relationship with dynamical mechanisms, over a wide range of spatial and temporal scales. There is no doubt that our understanding of the marine biosphere and the way it is related to physical processes will continue to benefit from the large amount of high resolution ocean color data that are now available.
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