The onset of a bloom after deep winter convection in the northwestern Mediterranean sea: mesoscale process study with a primitive equation model

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Abstract

The importance of mesoscale processes for primary production predictions is examined in a process study concerning the onset of the spring bloom after deep winter convection in the northwestern Mediterranean sea. Winter deep convection brings nutrient to the enlightened surface layer, but inhibits photosynthesis; phytoplankton biomasses are very low. As soon as restratification occurs, vertical mixing is blocked and a strong bloom onsets. Coastal Zone Color Scanner images have emphasized a strong mesoscale signal in the sea surface chlorophyll during this period. Mesoscale heterogeneity of the mixed-layer depth, due to the baroclinic instabilities associated with the process of deep water formation, is indeed responsible for the mesoscale variability of primary production. To ascertain interactions between hydrological processes and primary production occurring at mesoscales, a primary production model with a parameterization of production inhibition in situations of deep mixing is embedded in a three-dimensional primitive equation model with explicit mixed-layer physics. The model is initialized with a circular chimney of dense water surrounded by a stratified ocean. Two experiments are performed using different treatments of lateral mixing. In the first experiment, the horizontal diffusion is set to a low level so that mesoscale activity can be explicitly resolved. Surface density meanders of 50 km wavelength develop at the periphery of the chimney. These meanders, and the associated vertical motions, induce the sinking and spreading of the chimney, and subsequent surface restratification. Upward motions are responsible for mesoscale mixed layer shallowing, leading to an enhancement of primary production. Maxima of productivity are obtained at the edge of the chimney, where mesoscale activity is the most intense, in agreement with in situ data. In the second experiment, the horizontal diffusion is set to a high level so that lateral mixing occurs primarily through those terms: explicit mesoscale activity is completely damped. The initial structure of the chimney progressively disappears due to the horizontal diffusion of density across the isopycnals instead of three-dimensional redistribution. Mixed-layer depth and productivity are homogeneous. It is shown that instantaneous primary production can be underestimated by a factor of 4 when mesoscale eddies are not explicitly solved. This finding questions the evolution of large-scale coarse resolution climatic models of the oceanic carbon cycle. © 1998 Elsevier Science B.V. All rights reserved.

Keywords: primary production; mesoscale; bloom; models

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1. Introduction

The quantification of primary production is a major issue in predicting the oceanic carbon cycle. Indeed, primary production is one of the main biological fluxes controlling the inorganic carbon content of the oceanic surface layer, as it involves fixation of $CO_2$ through photosynthesis in the surface enlightened layer. In middle and high latitude regions, primary production as well as oceanic surface $pCO_2$ are dominated by patchiness on scales of 10–100 km, corresponding to the scales of mesoscale eddies (Gower et al., 1980; Watson et al., 1991). Recent data bases with high horizontal resolution have enabled the correlation of observed patchiness in primary production to mesoscale physical processes, such as upwelling at fronts (Strass, 1992; Prieur and Sournia, 1994) or mesoscale eddies (Robinson et al., 1993). Process studies with quasi-geostrophic mesoscale physics (Flierl and Davis, 1993; Yoshimori and Kishi, 1994; McGillicuddy et al., 1995; Dadou et al., 1996) have shown that in situations of oligotrophy, mesoscale vertical velocities could be responsible for a local enhancement of primary production, as they could transport nutrients into the euphotic layer.

In wide regions of the ocean (part of the North Atlantic and of the Austral ocean), primary production cannot possibly be enhanced by mesoscale nutrient enrichment of the surface layer, as there is no nutrient stress. These regions are characterized by deep winter mixing, with very low phytoplankton biomasses within a nutrient rich euphotic layer. This paradox is explained by the dispersion of the phytoplankton cells within a mixed layer considerably deeper than the enlightened layer (Sverdrup, 1953; André, 1990; Yamazaki and Kamikyoski, 1991). The stabilization of the water column is the necessary condition for phytoplankton growth, and can arise from mesoscale eddies. In such a case, one can expect the heterogeneity in the mixed layer depth resulting from the mesoscale flow (Klein and Hua, 1988) to be responsible for primary production mesoscale variability.

A typical example of such a situation is the onset of the spring bloom after winter deep convection in the northwestern Mediterranean sea. Coastal Zone Color Scanner (CZCS) ocean color images have indeed shown strong mesoscale variability in the surface chlorophyll field in early spring (André, 1990), when nutrients have become fully available through convection. This region is one of the few in the world ocean where deep-water formation (DWF) occurs. Processes of DWF can be divided into three phases (Medoc group, 1970; Gascard, 1978): a preconditioning phase; a violent mixing phase; and a sinking and spreading phase. During the violent mixing phase, a neutrally stable column of dense water is formed (chimney), extending from the surface down to the ocean bottom. The final phase of the process is the breakup, or sinking and spreading of the chimney, with a rapid restratification of the surface waters. It involves mesoscale eddies, attributed to baroclinic instability processes and responsible for the redistribution of water masses (Kilworth, 1976; Gascard, 1978; Madec et al., 1991a). As expected this restratification causes the onset of the bloom (Jacques et al., 1973).

Although these examples give strong evidence that mesoscale physics strongly modify the necessary conditions for primary production (light and nutrient availability), global or basin scale models that can predict primary production do not explicitly resolve mesoscale processes (Sarmiento et al., 1993; Drange, 1994). Indeed, available computer power does not yet allow running biogeochemical eddy resolving models on such scales. Instead, mesoscale transfers are roughly parameterized by lateral mixing across the large-scale gradients. The aim of this paper is to emphasize the tremendous underestimation of primary production predictions when such a parameterization is used instead of explicit mesoscale resolution, focusing on the particular case study concerning the onset of a bloom during the sinking and spreading phase of DWF in the northwestern Mediterranean sea.

For this purpose, a primitive equation model is used, with an explicit mixed layer representation, and an embedded primary production model with a parameterization of production inhibition in situations of deep mixing. This case study approach follows that of Madec et al. (1991a) for the physics. The quantification of eddy-resolving versus coarse scale primary production predictions is done by comparing a simulation where mesoscale activity is explicitly solved, with a simulation where mesoscale
transfer processes are parameterized by a diffusion operator. After the description of the dynamical–bio-geochemical model components in Section 2, we present the main results of the two simulations in Section 3, and discuss their differences in Section 4.

2. Model description

2.1. The host physical model

In order to reproduce the main features of the collapse and spreading of a chimney of dense water, we use OPA, the multilevel primitive equation numerical model developed at LODYC (Madec et al., 1991a; Delecluse et al., 1993). The domain is a closed, flat square basin on a β-plane of 300 km × 300 km × 2500 m centered at 42°N, which accounts for the so-called ‘MEDOC area’ (Medoc group, 1970). The model has 20 vertical layers and the resolution varies from 10 to 25 m in the upper 130 m, and reaches 500 m near the ocean bottom. The horizontal grid spacing is uniformly set to 2.5 km on a C-grid and the time step is 8 min. This model has already been used in the same kind of configuration (Mediterranean convection area, eddy resolving horizontal grid) by Madec et al. (1991a,b, 1996).

Vertical eddy coefficients are computed from an embedded 1.5 turbulent closure model (Blanke and Delecluse, 1993) based on a prognostic equation for the turbulent kinetic energy, and a diagnostic computation of the mixing length. Cut-offs are applied on the diffusivity and viscosity coefficients to avoid numerical instabilities associated with too weak vertical diffusion (set to 10^{-5} m^2 s^{-1} and 10^{-6} m^2 s^{-1}, respectively). This closure allows us to deal with stably unstable density profiles, which lead to large values of diffusivity (up to 1 m^2 s^{-1}) and which restores the static stability of the water column. Hereafter, the mixed layer is defined as the surface layers where the vertical mixing coefficient is greater than 10^{-4} m^2 s^{-1}. As this coefficient exhibits a sharp vertical variation at the bottom of the mixed layer, this definition is quite insensitive to the minimum value chosen, as long as it is reasonable (between 5 × 10^{-4} and 5 × 10^{-3} m^2 s^{-1}).

Two experiments are performed that differ in the way mesoscale activity is treated. In the first eddy experiment, mesoscale eddies are explicitly resolved. Weak horizontal mixing of density and momentum is included through biharmonic friction terms which insure numerical stability by selectively dissipating the smallest horizontal scales of each field. The biharmonic coefficient was chosen as the same constant for all quantities, −0.32 × 10^9 m^4 s^{-4}. The second experiment is the non-eddy experiment. The horizontal resolution is kept at 2.5 km, but a diffusion operator is added, namely a standard horizontal Laplacian term. Such an operator is widely used in large-scale climatic ocean circulation models (see McWilliams (1996) for a review). It parameterizes the eddy transfer activity by horizontal diffusive fluxes across the large-scale gradients. The value of the viscosity and diffusivity coefficients, 500 m^2 s^{-1}, was chosen to get a significant dissipation of the mesoscale variations.

No-slip conditions and no heat flux are applied along solid boundaries. At the sea surface, we assume a zero wind stress and a zero net heat flux, held constant with time. The surface heat flux is split up into a penetrative part \( Q_{\text{sol}} = 150 \text{ W m}^{-2} \), exactly compensated by the non-penetrative part. This heat forcing is responsible for the generation of static instabilities (and is therefore an important source of vertical mixing) as the penetration of the heat flux causes heating of the first few layers, while only the very first layer is cooled. A linear equation of state is assumed, specific to the conditions of the northwestern Mediterranean sea: \( \rho = 1050.29 - 1.615 T \) (T in °C).

The initial density field is set with a homogeneous chimney of dense water in a patch with an approximate 20 km radius in the center of the domain (Fig. 1). Outside the patch, density is relaxed to a background stratification, within a horizontal gradient area of 80 km. The background density profile corresponds to the one observed in the area prior to the onset of localized deep convection. It is homogeneous over the first 80 m. Given the mean value of the Coriolis parameter over the domain, \( f = 10^{-4} \text{ s}^{-1} \), the first three internal Rossby radii of this stratification are 8.4, 3.8 and 2.5 km. The horizontal grid spacing thus equals the third Rossby radius of this stratification. The patch of dense water is perturbed with a sinusoid of mode 7 and 1 km in amplitude, corresponding to the single most unstable
mode of the system in the framework of the two-layer theory (Pedlosky, 1987). This approach has the advantage of providing a regular departure from axisymmetry. Hereafter, we distinguish three domains on the basis of the density field: the convective area is that covered by the chimney; the background area, where the background density is set; and the intermediate area corresponding to the area of horizontal density gradient. It will appear that mesoscale activity occurs in the intermediate area.

The initial horizontal velocity field is diagnosed from the discrete geostrophic equilibrium, assuming a zero barotropic component in agreement with the analytical calculations of Crépon et al. (1989). Such an initialization has the advantage of strongly reducing the generation of transient superinertial waves generated during the gravitational adjustment period (Hermann and Owens, 1992).

2.2. The primary production model

The primary production model is a simplified version of the biogeochemical model developed in a one-dimensional vertical study by Lévy et al. (1998) for the northwestern Mediterranean seasonal cycle of primary production and export fluxes. Indeed, the particular period of interest (the onset of the spring bloom) and the characteristics of the trophic web during this period (new production, no regeneration) allow a reduction in the number of trophic variables while insuring the same predictions of stocks and fluxes (Lévy, 1996). The resulting model has four prognostic variables (instead of ten): nutrients (N); phytoplankton (P); zooplankton (Z); and detritus (D). They are expressed in terms of their nitrogen content. The governing equations are given Table 1 and the associated parameters are given in Table 2. Biogeochemical interactions between the compartments are primary production, grazing of phytoplankton and detritus (see Table 3 for formulation), phytoplankton mortality, zooplankton mortality and excretion (these two processes are inhibited when zooplankton concentration is below a given thresh-

Table 1

<table>
<thead>
<tr>
<th>Source/sink budget due to biogeochemical processes for the four compartments of the NPZD model in the first 200 m³</th>
</tr>
</thead>
<tbody>
<tr>
<td>( S(N) = - \mu_P L_N P + \mu Z + \mu_D D ) (1.1)</td>
</tr>
<tr>
<td>( S(P) = \mu_P L_N P - G_P - m_P P ) (1.2)</td>
</tr>
<tr>
<td>( S(Z) = a_P G_P + a_P G_P - \mu_P Z - m_P Z ) (1.3)</td>
</tr>
<tr>
<td>( S(D) = (1 - a_P) G_D + (1 - a_P) G_D + m_P P ) (1.4)</td>
</tr>
<tr>
<td>( + m_P Z - \mu_D D - G_D )</td>
</tr>
</tbody>
</table>

*See text and Tables 2–5 for coefficient definitions.*
ous studies of Sverdrup 1953, Woods and Onken (1980) in agreement with the previous studies of Sverdrup (1953), Woods and Onken

Grazing formulation

\[ \frac{d}{dt} G_{p} = g_{p} Z_{p} \left( \frac{P}{P + D} \right) K_{Z} + \frac{P^{2}}{D^{2}} \]  

(3.1)

Grazed detritus

\[ \frac{d}{dt} G_{d} = g_{d} Z_{d} \left( \frac{P}{P + D} \right) K_{Z} + \frac{P^{2}}{D^{2}} \]  

(3.2)

Table 2
Set of parameters for the NPZD model

<table>
<thead>
<tr>
<th>Nutrients</th>
<th>Half-saturation constant ( K_{N} )</th>
<th>mmol/m³</th>
</tr>
</thead>
<tbody>
<tr>
<td>Phytoplankton</td>
<td>Maximal growth rate ( \mu_{p} )</td>
<td>2 day⁻¹</td>
</tr>
<tr>
<td>Carbon/chlorophyll ratio ( R_{C/Chl} )</td>
<td>55 g C/g Chl</td>
<td></td>
</tr>
<tr>
<td>Carbon/nitrogen ratio ( R_{C/N} )</td>
<td>6.6 mmol/mmol</td>
<td></td>
</tr>
<tr>
<td>Half-saturation constant for light ( K_{PAR} )</td>
<td>33.33 W/m²</td>
<td></td>
</tr>
<tr>
<td>Zooplankton</td>
<td>Maximal grazing rate ( g_{z} )</td>
<td>0.75 day⁻¹</td>
</tr>
<tr>
<td>Half-saturation constant ( K_{z} ) for grazing</td>
<td>1 mmol/m³</td>
<td></td>
</tr>
<tr>
<td>Assimilated fraction of phytoplankton ( a_{p} )</td>
<td>0.7</td>
<td></td>
</tr>
<tr>
<td>Assimilated fraction of detritus ( a_{d} )</td>
<td>0.5</td>
<td></td>
</tr>
<tr>
<td>Excretion rate ( m_{e} )</td>
<td>0.1 day⁻¹</td>
<td></td>
</tr>
<tr>
<td>Mortality rate ( m_{m} )</td>
<td>0.03 day⁻¹</td>
<td></td>
</tr>
<tr>
<td>Concentration threshold ( Z_{\text{max}} )</td>
<td>0.015 mmol/m³</td>
<td></td>
</tr>
<tr>
<td>Detritus</td>
<td>Sedimentation speed ( V_{d} )</td>
<td>5 m day⁻¹</td>
</tr>
<tr>
<td>Remineralization rate ( \mu_{R} )</td>
<td>0.09 day⁻¹</td>
<td></td>
</tr>
</tbody>
</table>

Light absorption model

\[ L_{S} = \frac{N}{N + K_{N}} \]  

(2)

Light limitation is of the Webb et al. (1974) type, modulated by a parameterization of lagarian production inhibition in situations of deep mixing through a specific coefficient \( \gamma_{m} \). This parameterization, proposed by André (1990) in agreement with the previous studies of Sverdrup (1953), Woods and Onken (1982), Yamazaki and Kamikowski (1991), yielded encouraging results in Lévy et al. (1998):

\[ L_{i} = \gamma_{m} \left( 1 - e^{-\frac{PAR}{K_{PAR}}} \right) \]  

(3)

The photosynthetic available radiation (PAR) is derived from a light absorption model (Tables 4 and 5). The euphotic layer depth is then computed as the 1% incident light depth. Light limitation depends on the mixed layer and the euphotic layer depths through \( \gamma_{m} \) as described in Table 6. The parameter \( \gamma_{m} \) equals 1 (no limitation) when the mixed layer is shallower than the euphotic layer. In such a case, cells within the mixed layer are assumed to experience a mean PAR over the mixed layer. When the mixed layer becomes deeper than the euphotic layer, \( \gamma_{m} \) decreases accordingly down to a threshold value of 0.1. Production is inhibited because only \( \gamma_{m} \) percent of the cells remains long enough (≈ 1 day) within the euphotic layer to grow. These cells experience a mean PAR over the euphotic layer.

Table 4
Light absorption model

<table>
<thead>
<tr>
<th>( \lambda_{A} )</th>
<th>( \lambda_{B} )</th>
<th>( \lambda_{C} )</th>
<th>( \lambda_{D} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \lambda_{A} = \lambda_{A0} + \lambda_{A1} Chl^{1/2} )</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \lambda_{B} = \lambda_{B0} + \lambda_{B1} Chl^{1/2} )</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \lambda_{C} = \lambda_{C0} + \lambda_{C1} Chl^{1/2} )</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \lambda_{D} = \lambda_{D0} + \lambda_{D1} Chl^{1/2} )</td>
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</tbody>
</table>

Table 5
Parameters of the light absorption model

<table>
<thead>
<tr>
<th>( \lambda_{A} )</th>
<th>( \lambda_{B} )</th>
<th>( \lambda_{C} )</th>
<th>( \lambda_{D} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \lambda_{A} = \lambda_{A0} + \lambda_{A1} Chl^{1/2} )</td>
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<tr>
<td>( \lambda_{B} = \lambda_{B0} + \lambda_{B1} Chl^{1/2} )</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>( \lambda_{C} = \lambda_{C0} + \lambda_{C1} Chl^{1/2} )</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \lambda_{D} = \lambda_{D0} + \lambda_{D1} Chl^{1/2} )</td>
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</table>

Table 6
Parameters of the light absorption model

<table>
<thead>
<tr>
<th>( \lambda_{A} )</th>
<th>( \lambda_{B} )</th>
<th>( \lambda_{C} )</th>
<th>( \lambda_{D} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \lambda_{A} = \lambda_{A0} + \lambda_{A1} Chl^{1/2} )</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \lambda_{B} = \lambda_{B0} + \lambda_{B1} Chl^{1/2} )</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>( \lambda_{C} = \lambda_{C0} + \lambda_{C1} Chl^{1/2} )</td>
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<tr>
<td>( \lambda_{D} = \lambda_{D0} + \lambda_{D1} Chl^{1/2} )</td>
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</tbody>
</table>

Table 7
Parameters of the light absorption model

<table>
<thead>
<tr>
<th>( \lambda_{A} )</th>
<th>( \lambda_{B} )</th>
<th>( \lambda_{C} )</th>
<th>( \lambda_{D} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \lambda_{A} = \lambda_{A0} + \lambda_{A1} Chl^{1/2} )</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \lambda_{B} = \lambda_{B0} + \lambda_{B1} Chl^{1/2} )</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( \lambda_{C} = \lambda_{C0} + \lambda_{C1} Chl^{1/2} )</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>( \lambda_{D} = \lambda_{D0} + \lambda_{D1} Chl^{1/2} )</td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

Only a fraction (43%) of the light field in the water can be used for photosynthesis. Two different light wave lengths are considered. The absorption coefficients depend on the local phytoplankton concentrations.

old), fecal pellet production, detritus sedimentation and detritus remineralization. Phytoplankton growth \( \mu_{p} L_{N} L_{P} P \) is nutrient and light limited. Nutrient limitation formulation is the widespread Michaelis–Menten kinetics:

\[ L_{N} = \frac{N}{N + K_{N}} \]  

(2)

The photosynthetic available radiation (PAR) is derived from a light absorption model (Tables 4 and 5). The euphotic layer depth is then computed as the 1% incident light depth. Light limitation depends on the mixed layer and the euphotic layer depths through PAR and \( \gamma_{m} \) as described in Table 6. The parameter \( \gamma_{m} \) equals 1 (no limitation) when the mixed layer is shallower than the euphotic layer. In such a case, cells within the mixed layer are assumed to experience a mean PAR over the mixed layer. When the mixed layer becomes deeper than the euphotic layer, \( \gamma_{m} \) decreases accordingly down to a threshold value of 0.1. Production is inhibited because only \( \gamma_{m} \) percent of the cells remains long enough (~ 1 day) within the euphotic layer to grow. These cells experience a mean PAR over the euphotic layer.
Table 6

Expressions of $\text{PAR}$ and $\gamma_M$ as a function of the mixed layer depth $z_M$ and the euphotic layer depth $z_E$ ($z$ is the local depth)

Case 1: the mixed layer is shallower than the euphotic layer

$$z < z_M, \quad \text{PAR} = \frac{1}{z_M} \int_0^{z_M} \text{PAR}(z) \, dz$$

Case 2: the mixed layer is deeper than the euphotic layer

$$z > z_E, \quad \text{PAR} = \text{PAR}(z)$$

$$\gamma_M = 1$$

$$\gamma_M = \text{max}(0.1, 1 - 0.9 \left( \frac{z_M}{z_E} - 1 \right))$$

The equations for the biogeochemical model are only solved for the uppermost 12 model levels ($\sim 200$ m). Below, the three biogenic compartments decay to nutrients, with a decaying rate varying from 1 to 20 days.

2.3. The on-line time integration procedure

The Eulerian time evolution of any of the four state variable concentrations $C$ is controlled by biogeochemical processes, advection and vertical diffusion:

$$\frac{\partial C}{\partial t} = S(C) + \nabla \cdot (U C) + \frac{\partial}{\partial z} \left( k \frac{\partial C}{\partial z} \right)$$

where $U$ and $k$ are, respectively, the velocity vector and the diffusion coefficient computed by the physical model. As in the host physical model, the governing equations are solved on a C-grid. An absolute requirement is that the tracer concentrations remain positive. Therefore, the centered finite difference scheme used for the advection of density in the host physical model cannot be used, even when the diffusive biharmonic term is added. Instead, a positive definite advection transport scheme (Smolarkiewicz and Clark, 1986) is used. It is slightly diffusive, therefore, no additional horizontal diffusion is needed to insure the numerical stability of the system in the eddy experiment. In the non-eddy experiment, a Laplacian horizontal mixing term is added to the right-hand side of Eq. (4), with the same coefficient used for the physics. An explicit time integration is used, with time-centered velocities, except for vertical mixing which is implicit. These discretizations combine to insure total nitrogen conservation. The time step is 32 min, four times that of the physical model.

Initialization of the biogeochemical variables is done after an 8 day spin-up of the host physical model. The physical adjustment of the system is therefore fully established at the beginning of the biogeochemical simulation. As suggested by observations (Coste et al., 1972; Jacques et al., 1973), initial nitrate and phytoplankton fields are derived from the density field at day 8. High densities are correlated with high nitrate and low phytoplankton concentrations. The four-variable system is close to one-dimensional equilibrium in the background conditions. The background profiles of zooplankton and detritus are extended to the entire domain.

3. Model results

The initial geostrophic balance combines baroclinic currents with the horizontal density gradient surrounding the neutrally stable column. In the eddy experiment, a cyclonic gyre forms in the upper 500 m whose core velocity is $\sim 20$ cm s$^{-1}$, while a weaker countercurrent ($\sim 2$ cm s$^{-1}$) appears in the deep layers. Due to the important slope of the isopycnals at the edge of the chimney, and to the relevant shears in the velocity field, this situation is unstable (Pedlosky, 1987). After a few days, mixed barotropic–baroclinic instabilities start to develop. Their signatures on the surface density field are the development of meanders along the front that defines the neutrally stable column. A time series of the surface density field (Fig. 2) shows the development of 14 small meanders at the outer edge of the convective region (day 12), with a large growth rate, rapidly caught up by 7 slower (but stronger in magnitude) meanders of $\sim 50$ km wavelength (day 16). This wavelength corresponds to the first baroclinic mode of oscillation, while the first smaller wavelength ($\sim 20$ km) was that of the second mode. Meanders grow from days 12 to 24, when they break into 7 eddies. As soon as the meanders appear, they start propagating in the anticyclonic direction with a
Fig. 2. Time series of the surface isopycnals for the eddy experiment. Contour interval is 0.05 in units of $\sigma_z$. Shaded areas indicate the location where the mean vertical velocities over the first 100 m are greater than 10 m d$^{-1}$ (upward velocity cells), and correspond to the contoured areas of Fig. 3. Note that only a zoom of the total domain is represented (from 50 to 250 km).

speed of 1.8 cm s$^{-1}$. This phase speed is about one-tenth of the mean surface velocity, in agreement with the analytical solution proposed by Gascard (1978). Such meanders have been observed by in situ measurements (Gascard, 1978), and their role in the sinking and spreading of the initial chimney was emphasized by the modeling study of Madec et al. (1991a).

Cells of upward velocity appear on the downstream side of the surface density crests of the meanders (shaded area superimposed to the surface density fields in Fig. 2), while cells of downward velocity appear on the upstream side of the density crests. These cells extend throughout the water column without vertical phase lag, as already mentioned in Madec et al. (1991a). Maximum vertical velocities
are found between 300 and 1000 m, where typical values are \( \pm 50 \, \text{m d}^{-1} \), but can sometimes exceed \( \pm 200 \, \text{m d}^{-1} \). Such a distribution of vertical velocities tends to sink and spread the initial chimney, as the denser waters sink out of the convective area while the lighter peripheral water masses are advected up towards the center.

A time series of the mixed-layer depth superimposed on the horizontal signature of the upward velocity cells is shown in Fig. 3. As there is no wind,

Fig. 3. Time series of the mixed-layer depth in meters, for the eddy experiment. Contoured areas show the location where the mean vertical velocities over the first 100 m are greater than 10 m d\(^{-1}\) (upward velocity cells), and correspond to the shaded areas of Fig. 2.
the main source of vertical turbulence at the surface is the penetration of radiative heat flux; the mixed layer is heat-driven. It is very deep (greater than 500 m) within the convective area, due to the absence of stratification. In the background area, it is bounded at 80 m by the top of the thermocline. As expected from the quasi-geostrophic study of Klein and Hua (1988), the mesoscale flow causes heterogeneity of the mixed layer; in the intermediate area, its depth varies between 20 and 80 m in response to vertical

Fig. 4. Time series of surface primary production for the eddy experiment.
motions associated with the meanders. When meanders are fully developed (days 16–20), minima of mixed-layer depth appear in-between two surface density crests, in the downstream side of upward velocity cells. Relative maxima appear at density crests, in downward velocity cells. This mixed-layer response to vertical motions is asymmetrical, as the mixed layer can be locally shallowed, but can never become deeper than its background depth. Such an asymmetrical response has also been described in the quasi-geostrophic-surface boundary layer study of McGillicuddy et al. (1995). One possible explanation for this is that upward motions drive lighter waters on the top of the thermocline, thus shallowing the gradient, while downward vertical velocities causes detrainment of mixed-layer fluid along the isopycnes, thus keeping the gradient as it is.

The meander dynamics have a remarkable impact on primary production, mainly through the variations of the mixed-layer depth, indirectly related to the vertical velocities, and also through the advection of phytoplankton cells. Indeed, mesoscale nutrient enhancement has a very low impact during this period as nutrients are still plentiful and zooplankton has not yet developed, as expected during this period. Therefore, phytoplankton growth is mainly controlled by the mixing depth and by its own distribution. Production is the strongest in regions where the mixed layer is shallower than the euphotic layer, and in regions where biomasses reach the highest concentrations. Primary production predictions at the surface are shown in Fig. 4. Their spatial structures are very similar to that of the mixed-layer depth (Fig. 3). In the convective area, primary production is strongly limited by deep convection and is close to zero. In the background area, where the mixed layer is about twice as deep as the euphotic layer (80 m), it hardly reaches 0.1 mmol N m⁻³ d⁻¹. Maximum values of primary production are found in the trough of the meanders, where the mixed layer is shallower then the euphotic layer (15–30 m), and can reach up to 0.7 mmol N m⁻³ d⁻¹. The mesoscale variability of the phytoplankton field, resulting from the competition between areas of productivity and export of the cells from these areas (Lévy, 1996) will be presented in a subsequent publication.

In the non-eddy experiment, the primary effect of horizontal diffusion is the dissipation of mesoscale structures. The problem is axially symmetrical, no meander development is observed (not shown). The only lateral mixing that occurs is achieved by the horizontal diffusion operator that works against the large-scale gradients. The isopycnsals are, therefore, horizontally stretched out, which weakens the horizontal density gradients and consequently weakens the cyclonic gyre at the surface; after 12 days, the mean velocity is about half that of the eddy experiment at the same time, 10 cm s⁻¹. Heat is brought towards the convective area by lateral diffusion and no variability in the mixed-layer depth field is observed. The mixed layer has a constant depth of 80 m in the whole basin, except for the very central zone, where it can reach 100 m due to a weaker stratification. It is always deeper than the euphotic layer (40 m) and, therefore, the André (1990) parameterization induces strong inhibition of production. Consequently, productivity is low and homogeneous within the intermediate and the background areas, and is minimum in the convective area.

4. Discussion

As we focus our attention on the large-scale differences between the two simulations, the axial symmetry of the problem has led us to compare the radially averaged properties across the frontal zone for the two simulations (time series, Figs. 5 and 6). In the non-eddy experiment, the interior zone appears to be covered by lighter water within a few days, the whole area being restratified by day 12 (Fig. 5). The whole process is horizontal. Horizontal gradients stretch out, while in the eddy experiment, eddies not only widen the frontal zone, but also locally strengthen the mean gradients (R = 30 and 70 km).

Horizontal diffusion has consequently a tremendous impact on the prediction of the mixed layer. Radial averages of the mixed-layer depths for the two experiments (Fig. 6, dotted curves) are very different in the intermediate and convective area, but are identical in the background area. In the eddy experiment, the averaged mixed-layer depth shows a clear maximum in the convective area, while this
maximum narrows with time. It is minimum in the intermediate area, and tends to ‘squeeze’ toward the center as meanders restratify the water column. In the non-eddy experiment, the mixed layer is homogeneously set at its background level (80 m) outside the convective area, where its depth is therefore always deeper than in the eddy experiment. Inside the convective area, because of quicker restratification, the mixed layer is not as deep as in the eddy experiment; it does not exceed 100 m.

These differences in the mixed layer have important consequences on the radially averaged primary production (Fig. 6). In the eddy experiment the primary production is maximum in the intermediate area, where the mixed layer is the shallowest, and its maximum value grows with time, in response to phytoplankton development as well as restratification. Such a maximum of productivity at the border of the convective area has been observed by in situ measurements by Nival et al. (1972). The radial maximum of primary production shifts towards the center, in response to increasing stratification. Production observations carried out in the MEDOC area during the two legs of the Mediprod 1 cruise (Jacques et al., 1973), in March and in April, show a similar shift. In the non-eddy experiment, the expected maximum of production in the intermediate area is not simulated at all. Instead, production is more homogeneous, as expected from the mixed-layer field, with lower values than in the eddy experiment in the intermediate area, higher values in the convective area, and similar values in the background area.

The main process responsible for the differences between the two experiments is production inhibition by deep vertical mixing. Indeed, primary production is either controlled by light or by nutrient availabil-
Fig. 6. Radially averaged vertical section of primary production. The 0 km radius corresponds to the center of the domain (x = 150 km, y = 150 km). Contour interval is 0.025 mmol N m\(^{-3}\) d\(^{-1}\). The dotted curve is the radially averaged mixed-layer depth. Top: eddy experiment. Bottom: non-eddy experiment.

In both experiments there is no nutrient stress and the light limitation is different through the \(\gamma_m\) coefficient (which strongly depends on the depth of the mixed layer versus the euphotic layer). To illustrate this, both the eddy and non-eddy experiment have been rerun without this inhibiting process. In such a case, the settling of the bloom is no longer controlled by the depth of the mixed layer. In both experiments, the primary production was quite similar and predicted phytoplankton biomasses became huge completely unrealistic values of one order of magnitude larger than in the standard eddy experiment.

In order to more globally quantify the differences in primary production predictions between the two experiments, primary productions were averaged over the 100-m-depth 100-km-radius domain (Fig. 7). This average excludes the background area, away from mesoscale influence, where predictions are similar. Until day 12, global production is higher in the non-eddy experiment, as restratification has started earlier, but only slightly, as the convective area has much lesser extent when compared to that of the intermediate area (1250 and 30,000 km\(^2\), respectively). From days 12–24, global primary production decreases with oscillations in the non-eddy experiment, before reaching a constant value. The oscillations are due to the limitation of production by self-shading of the cells (not shown). During the same period, primary production greatly increases in the eddy experiment until day 22 when the eddy dynamics and their shallowing effect on the mixed layer begins to weakens. The difference between the predictions of the two experiments varies with time.
The ratio of the two predictions is very close to 1 until day 12, and almost reaches the value of 4 at day 22. As a whole and for the whole duration of the simulation, primary production is underestimated by a factor of 2 when the mesoscale structures are not resolved (4.5 against 2.5 mmol N m$^{-2}$ d$^{-1}$ for the eddy and the non-eddy experiment, respectively).

5. Summary and conclusions

A primary production model has been embedded in a primitive equation ocean model with mixed-layer physics and has been used to reproduce the onset of a bloom during the sinking and spreading phase of a convective chimney in the MEDOC area. In this situation, the main factor controlling photosynthesis is not nutrient availability, as the environment is not nutrient stressed at this time of the year, but vertical mixing. Therefore, unlike previous mesoscale process studies, mesoscale transport of nutrients to the euphotic layer does not enhance primary production. However, an eddy-resolving experiment (eddy experiment) has clearly indicated a strong correlation between mesoscale activity and the enhancement of primary production, in agreement with CZCS observations. The reason for such an interaction concerns another controlling factor of primary production: light. Indeed, when the mixed layer is deeper than the euphotic layer, intense vertical mixing strongly reduces the average time over which phytoplankton is exposed to light. The mixed-layer depth becomes a key parameter of the physical environment controlling primary production. Mixed-layer depth minima are due to the mesoscale activity associated with the sinking and spreading phase of DWF. One can now understand how oceanic mesoscale variability strongly interacts with primary production in the absence of nutrient stress. The model has succeeded in predicting the highest productivity at the edge of the convective chimney and the lowest productivity inside, as previously measured (Nival et al., 1972). The chimney edge is, indeed, the region where mesoscale activity is the strongest, and causes strong shallowing of the mixed layer.

When the mesoscale processes contributing to the sinking of the convective area are not explicitly solved, but are crudely parameterized by a widely used second-order diffusion operator (non-eddy experiment), the mixed layer field remains homogeneous and deeper than the euphotic layer all along the simulation. Consequently, primary production remains homogeneous and low. Because of the strong non-linear response of primary production to the environment, the averaged primary production fluxes show great differences between the eddy and the non-eddy experiments. In our case, we have shown that primary production is underestimated by up to a factor of 4 when mesoscale eddies are not explicitly solved. This indicates that, even in a no nutrient stress condition, primary production is greatly and globally boosted by continuous enhancement of production in restricted mesoscale areas through mixed-layer depth and euphotic layer depth ‘interactions’.
Mesoscale influence on primary production can no longer be rejected in global oceanic carbon cycle models. There seems to be two alternatives. The first is to generalize such an eddy-resolving model to larger time and space scales, even though it is very expensive. The second is to parameterize mesoscale interactions between dynamical and biogeochemical processes. The horizontal mixing parameterization we have used in our non-eddy experiment is very crude. A more sophisticated parameterization (Gent and McWilliams, 1990; Visbeck et al., 1997) might lead to improved results. The Gent and McWilliams (1990) scheme indeed produces a similar lateral transport of water masses as the eddying model, in contrast to a diffusive scheme. This second option could be considered when primary production is only controlled by the physics, but this approach still seems quite unrealistic when different types of mesoscale interactions start interfering, such as nutrient or zooplankton stress.

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