Auxiliary Material

The GYRE-LOBSTER configuration: Model set up and analysis of the mean state

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Auxiliary material for this article contains the set up of the GYRE-LOBSTER configuration that has been used in the paper. It is followed by a complementary analysis of the basic features of the GYRE-LOBSTER configuration. Data are from the World Ocean Atlas [Locarnini et al., 2006; Antonov et al., 2006].

1. The model set up

This study uses a coupled physical and biological model. The NEMO primitive-equation ocean circulation model [Madec, 2008, available at http://www.locean-ipsl.upmc.fr/NEMO] is used. The configuration consists of an idealized double gyre on the θ-plane (2,120 * 3,180 km) that mimics the North Atlantic circulation with its western boundary current. It is adapted from the rectangular configuration developed by Hazeleger [1998] to study the subtropical mode water formation and its variability [Hazeleger and Drijfhout, 1998, 1999]. The basin is rotated with an angle of 45° in order to capture the western boundary current with maximum length. The domain corresponds to a reduced version of the North Atlantic: It extends from 15°N to 50°N and from 85°W to 55°W with a grid spacing of 35 km in both directions. This eddy-permitting configuration will be referred to as the ”1/3° configuration”. The vertical grid has 31 nonuniform vertical levels whose thickness vary from approximately 5m at the surface to 250 m near the bottom. A flat bottom is set at the last vertical level (roughly at 4 km depth) and the lateral boundaries are straight and closed. We use free slip conditions and a non-linear friction is applied at the bottom. For simplicity, a bilinear equation of state is used where the density $\rho$ (in kg m$^{-3}$) is a function of temperature $T$ (in °C) and salinity $S$ (in psu):

$$\rho = \rho_0(1 - (\alpha T - \beta S)).$$

Here $\beta = 7.7 \times 10^{-4}$ kg m$^{-3}$ psu$^{-1}$ and $\alpha = 2 \times 10^{-4}$ kg
\( m^{-3} K^{-1} \) are the expansion coefficients for salinity and temperature respectively and \( \rho_0 = 1020 \text{ kg m}^{-3} \). The model uses a level-coordinate system and has an implicit free surface treatment of the barotropic equations. The month and the year are fixed to 30 and 360 days, respectively.

1.1. Forcings of the physical model

The model is forced analytically by a penetrative solar radiation, a newtonian-type net heat flux, a fresh water flux and wind stress. All forcings vary zonally and seasonally, as depicted in fig. 1. The forcings were taken and adapted from Hazeleger and Drijfhout [1998] and a solar radiation flux was explicitly implemented. This flux is needed for the ecosystem model. A newtonian-type net heat flux (Eq. (1)) simulates the retroaction of the sea surface temperature (SST) on atmospheric fluxes with a restoring coefficient \( \gamma \) equal to 40 W m\(^{-2}\)K\(^{-1}\). This net heat flux is applied to a zonally averaged apparent temperature (Eq. (2), fig. 1a). The formulation for the penetrative solar radiation flux (Eq. (3), fig. 1b) has been established from mean ECMWF fluxes (averaged between 1979 and 1993). The maximum and minimum insolations are set to 225 W m\(^{-2}\) and 190 W m\(^{-2}\) in the south and 175 W m\(^{-2}\) and 80 W m\(^{-2}\) in the north, respectively. The freshwater flux (evaporation minus precipitation budget, Eq. (4) and (5), fig. 1c) is negative in the north (-1.2 m year\(^{-1}\)) and positive in the south (0.5 m year\(^{-1}\)). In the subtropical gyre the evaporation is less than observed in order that the basin integrated freshwater flux is nul. The wind intensity (Eq. (6), fig. 1d) has a maximum at 36°N and a minimum at 21°N to represent the westerlies and easterlies, respectively. At 36°N, the wind stress curl becomes negative to allow the subpolar gyre formation. The forcings vary sinusoidally between the 21th of December (\( t_{\text{dec}} \)) and the 21th of June (\( t_{\text{jun}} \)) for all fluxes except for
the restoring apparent temperature which varies between the 21th of January ($t_{jan}$) and the 21th of July ($t_{jul}$). This one-month time lag between the incident solar forcing and the restoring temperature is necessary to correctly simulate the restratification of the water column. Hereafter the equations used for the analytical forcings are presented with $Q$ the total surface heat flux (in W m$^{-2}$), $T^*$ the restoring apparent temperature (in °C), $qs_r$ the solar radiation flux (in W m$^{-2}$), $EmP$ the freshwater flux (in m year$^{-1}$), $\tau$ the wind stress (in N m$^{-2}$), and $y$ the latitude:

\[ Q = \gamma(T^* - SST) \]  

\[ T^* = 28.3 \left( 1 + \frac{1}{50} \cos(\pi \frac{t - t_{jul}}{t_{jul} - t_{jan}}) \right) \times \cos \left( \frac{\pi}{2 \times 53.5} \left( \frac{y - 5}{1 + \frac{11}{53.5} \cos(\pi \frac{t - t_{jul}}{t_{jul} - t_{jan}})} \right) \right) \]  

\[ qsr = 230 \cos \left( \frac{\pi}{0.9 - 180} \left( \frac{y - 23.5 \times \cos(\pi \frac{t - t_{jul}}{t_{jul} - t_{dec}})}{0.9 - 180} \right) \right) \]  

If $y \in [15 ; 37.2]$ then

\[ EmP = 0.7 \sin \left( \frac{\pi}{2} \left( \frac{y - 37.2}{24.6 - 37.2} \right) \right) \times \left( 1 - \frac{0.1}{0.7} \cos(\pi \frac{t - t_{jul}}{t_{jul} - t_{dec}}) \right) \]  

If $y \in [37.2 ; 50]$ then

\[ EmP = -0.8 \sin \left( \frac{\pi}{2} \left( \frac{y - 37.2}{46.8 - 37.2} \right) \right) \times \left( 1 - \frac{0.1}{0.8} \cos(\pi \frac{t - t_{jul}}{t_{jul} - t_{dec}}) \right) \]
\[ \tau = \sin \left( \frac{\pi y - 15}{29 - 15} \right) \times \left( \frac{0.105}{\sqrt{2}} - 0.015 \cos \left( \pi \frac{t - t_{\text{jun}}}{t_{\text{jun}} - t_{\text{dec}}} \right) \right) \] (6)

1.2. Model physics

Mixing of temperature and salinity is included through a Laplacian operator and is done along isopycnal surfaces. Mixing of momentum is included through a bi-Laplacian horizontal operator. The lateral viscosity and diffusivity coefficients are \(-10 \times 10^{11}\) and 300 m² s⁻¹, respectively. The vertical viscosity and diffusivity are computed using the TKE mixing parameterization [Blanke and Delecluse, 1993] with background values of \(1.2 \times 10^{-4}\) and \(1.2 \times 10^{-5}\) m² s⁻¹, respectively. For advection of temperature and salinity, we use the positive, low-diffusive, total variation diminishing (TVD) scheme with corrected flux [Lévy et al., 2001a]. The time step is 1 hour.

1.3. The biogeochemical model

1.3.1. Description of the biogeochemical model

The LOBSTER biogeochemical model [Lévy et al., 2001b, 2005a] is a simple nitrogen-based pelagic model with six compartments: nitrate (\(NO_3\)), ammonium (\(NH_4\)), phytoplankton (\(P\)), zooplankton (\(Z\)), detritus (\(D\)) and semi-labile dissolved organic matter (\(DOM\)). Each of these tracers \(C\) is freely advected (\(-\nabla (uC)\)) and diffused by the flow (vertical diffusion (\(\frac{\partial}{\partial z} (k_z \frac{\partial C}{\partial z})\)) and lateral diffusion (\(D_{\text{lat}}\)) and is submitted to a specific source/sink term \(SMS(C)\) which describes the biogeochemical interactions among the tracers. In Lévy et al. [2001b, 2005a], the use of simplified equations below the euphotic layer was justified by the short duration of the experiments. Here, since we focus on longer
time scales, the same set of equations is applied throughout the whole water column to
avoid an artificial gradient between the two layers. The biogeochemical model is described
by the equations in Table 1. Fast sinking detritus which leave the euphotic layer \((-v_d \partial_z D)\)
are a source of nitrate in the aphotic layer which decreases till the bottom \((+v_d \partial_z D)\). A
fraction of these detritus reaches the bottom and is trapped in a two dimensional \((x,y)\)
model sediment trap that releases nitrogen in the water column regularly. A fraction of the
zooplankton mortality flux feeds instantaneously the aphotic layer \((\partial_z f (z))\) and is then
remineralized in nitrate in the aphotic layer. Remineralization is assumed to occur below
the euphotic layer with the downwards vertical fluxes of detritus and dead zooplankton
decreasing exponentially with depth [similar to the scheme of Najjar et al., 1992]. The
parameter values used in the simulations are listed in Table 2. They were tuned from the
POMME parameters [Lévy et al., 2005a]. Published papers provide some guidance for the
remaining parameters, but not for all. With these caveats in mind, the unknown model
parameters were tuned within reasonable ranges in order to best represent data in the
North Atlantic. For further details on the optical model, the reader is referred to Lévy
et al. [2005a]. The time step and the tracer advection and diffusion schemes are the same
as for temperature and salinity.

1.3.2. Choice of the DOM representation

The choice of the DOM representation was subjected to parametrization tests. The
DOM pool is a continuum usually separated into labile, semi-labile and refractory DOM
pools which make up 1%, 10% and 90% of the total pool respectively. The labile and
semi-labile pools are reported to be biologically reactive on time-scales of hours to days
[Bronk et al., 1994; Kiel and Kirchman, 1999] and months to year [Vidal. et al., 1999;
Carlson, 2002], respectively. The model includes the labile component of the DOM via the 
NH4 compartment. The fraction of refractory DOM pool in the ocean which may escape 
degradation over many ocean circulation times [Loh et al., 2004] is not simulated in the 
model. Refractory DOM is photochemically degraded in the upper layers on timescales 
of 6-10 years [Mopper et al., 1991], and is not taken into consideration in this study. The 
semi-labile pool is explicitly represented by DOM, and also implicitly represented in 
the model by the phytoplankton exsudation reaction from which a large fraction (75%) 
is quickly transformed in NO3. The breakdown rate of DOM is set in the model to 1/(3 
years), which is lower than the typical degradation range but is consistent with the relative 
small size of the basin, and allows the representation of a semi-labile to semi-refractory 
DOM pool. Simulated DOM at the surface exhibits a strong contrast in the opposite way 
of nitrate with mean values of 9 mmol m\(^{-3}\) in the south and 7 mmol m\(^{-3}\) in the north. 
Observed DOM at the surface shows high concentrations of nearly 9.5 mmol m\(^{-3}\) at 15\(^{\circ}\)N 
[Vidal. et al., 1999], and low concentrations of 4.4 - 7.4 mmol m\(^{-3}\) between 35\(^{\circ}\)N and 
50\(^{\circ}\)N at 20\(^{\circ}\)W [Kahler and Koeve, 2001]. This DOM representation is in adequation with 
data.

1.3.3. Estimate of chlorophyll from the pronostic PHY variable

The surface chlorophyll (CHL) is not a pronostic variable in the model but can be 
diagnosed from the model data. The CHL is computed from phytoplankton averaged 
over the upper 5 m (Eq. 7), using a ratio which depends both on the ratio between 
nitrogen and chlorophyll, and on nutrients and light limitations. The Chl:N ratio (\(R_{\text{Chl:N}}\), 
Eq. 8) is equal to 1.32 g chlorophyll per mol nitrogen which corresponds to a chlorophyll 
to carbon mass ratio of 1:60 and a C:N Redfield ratio of 6.6 used throughout the paper.
And as CHL depends on nutrients and light availability, limitations by nutrients \( (L_{NO_3}) \) and light \( (L_I) \) were added to the calculation.

\[
\text{Chl} = R_{\text{Chl}:N} \times P \\
R_{\text{Chl}:N} = \frac{1}{1.32L_IL_{NO_3}}
\]

1.4. Initializations and spin up

To speed up the spin up, it is started at a lower resolution (106 km). This coarse configuration will be referred to as the "1° configuration".

1.4.1. Initializations

Initializations are made for the 1° configuration. Interpolation of the spun-up state of the 1° configuration is then used to initialize the 1/3° configuration. Vertical profiles of temperature and of salinity are constructed from the World Ocean Atlas climatology \cite{Levitus and Boyer, 2005} by averaging over 25°N-30°N and 80°W-0°W, and are applied to the whole domain. Nitrate is initialized in winter as a function of the winter density of the 500-years spun-up state of the 1° model. Above the isopycnal \( \sigma = 24.5 \) (equivalent to 1,024.5 kg m\(^{-3}\)), nitrate is uniformly set to 2 mmol m\(^{-3}\). Below the isopycnal \( \sigma = 24.5 \), nitrate (in mmol m\(^{-3}\)) is defined through a linear relationship with the potential density:

\[
NO_3 = (15.55 \times \sigma) - 380.11.
\]

This relation has been established from the density and nitrate data in the subtropical gyre \cite{Conkright et al., 2002}. All remaining biogeochemical compartments are initialized with a uniform value: 0.1 mmol m\(^{-3}\) for P, Z, DET and NH4, and 1 mmol m\(^{-3}\) for DOM.

1.4.2. Spin up
The 1° physical model is first spun up alone for 500 years. Then, the biogeochemical model is inserted and the integration continues online for another 350 years. A surface salinity relaxation with a timescale of one year is added during this second part of the spin up to dampen low-frequency oscillations. This relaxation is not maintained after the spin up. After 850 years, the 1° coupled physical-biogeochemical model has reached a mean state. The spatially-averaged drift at 450 m for the last 10 years of the spin up is 0.0004°C/10 years for temperature, 0.00021 psu/10 years for salinity and 0.023 mmol m⁻³/10 years for total nitrogen. Then the 1/3° coupled physical-biogeochemical model is initialized from the spun-up state of the 1° simulation and run for another 50 years to adjust the basin with the new resolution. After that, 1/3° simulated temperature, salinity and biogeochemical properties in the thermocline are found to be quite stable: The spatially-averaged drift at 450m for the last 10 years of the 1/3° configuration spin up is 0.008°C/10 years for temperature, 0.0038 psu/10 years for salinity and 0.023 mmol m⁻³/10 years for the total nitrogen. The drifts are slightly higher for the 1/3° than for the 1° configuration but somewhat moderate. The model presents residual interannual variability despite its climatological forcing. The sources for this interannual variability are both mesoscale eddies (which are only marginally resolved) and internal modes of variability. This interannual variability has been identified by Hazeleger [1998] and Hazeleger and Drijfhout [2000]. He founds two significant modes: one with a 8-year period and associated with a large spatial scale, and another with a 4.5-year period and associated with a smaller spatial scale. The amplitude of this variability is weak in our model, less than 4% for the subtropical mode water mean thickness for example. As the purpose of this work is...
to focus on the mean seasonal cycle, most of our analysis is based on a climatology built from a 10-yrs run conducted after the spin up with outputs saved every three days.

2. Analysis and validation of the mean state

2.1. Barotropic stream function

From the forcings applied (fig. 1) and the basin geometry, the model simulates an annual-mean barotropic stream function (BSF, fig. 2) with a large anticyclonic subtropical gyre and a weaker cyclonic subpolar gyre. The subtropical western boundary current (WBC hereafter, the model equivalent of the Gulf Stream) is concentrated at the coastline between 25°N and 36°N, and heads offshore approximately at 36°N. This WBC recirculates through the large subtropical gyre with transports up to 30 Sv. The simulated transport is consistent with that estimated through the Florida strait [Larsen, 1992; Baringer and Larsen, 2001].

2.2. Mixed-layer depth

The maximum mixed-layer depth (MLD, fig. 3a) is 250 m just south of the WBC (marked by the zero-contour of the barotropic stream function reported from fig. 2), which is in agreement with observations (fig. 3b, WOA, 1994). Mixed-layers deeper than 1,000 m are found in the north in the subpolar gyre, also consistent with observations (fig. 3a, b). The mixed-layer shows a strong gradient around 30°N as in the data. This gradient is crucial for the formation of mode water [Worthington, 1959; McCartney, 1982]. During summer, the MLD is less than 50 m in the entire subtropical gyre (not shown).
2.3. Surface chlorophyll

Chlorophyll (CHL) is not a prognostic variable of the model. It is diagnosed from the concentration of phytoplankton using a variable N:Chl ratio (Eq. (7)). The diagnosed annual-averaged surface chlorophyll pattern is displayed in fig. 3c and the corresponding annual-averaged chlorophyll data (SeaWIFS, 1997-2006) in fig. 3d. The model reproduces the observed large-scale contrasts of SeaWIFS climatology: a sharp latitudinal gradient between the low-CHL subtropical gyre and the high-CHL subpolar gyre. The model does not properly capture the observed longitudinal gradient of CHL with rich waters in the east and poor ones in the west. In the subtropical gyre, modeled CHL concentrations (lower than 0.2 mgChl m$^{-3}$) are slightly above the satellite values (0.1 mgChl m$^{-3}$). These two differences may be due to the relatively small size of our basin: the center of the gyre is closer to the boundary with the subpolar gyre and less isolated from intergyre Ekman transport [Williams and Follows, 1998]. In the subpolar gyre, modeled CHL concentrations (0.9 - 1 mgChl m$^{-3}$) are consistent with SeaWIFS values (1 mgChl m$^{-3}$). Note the high CHL concentrations in the south-east of the basin due to upwelling along the boundary (the model equivalent of the Northern African upwelling).

2.4. Seasonal cycle at BATS and NABE

Here we compare the model’s seasonal cycle with observations from two JGOFS stations located in the subtropical and subpolar gyres, respectively (fig. 4): the Bermuda Atlantic Time serie Study [BATS, 31°N, 64°W, Michaels and Knapp, 1996] and the North Atlantic Bloom Experiment [NABE, 47°N, 20°W, Weeks et al., 1993]. BATS is located near the northwestern edge of the North Atlantic subtropical gyre in a region of weak Gulf Stream recirculation with a mean flow toward the southwest. NABE is situated near
the southeastern edge of the North Atlantic subpolar gyre with deeper winter MLD than at BATS. It is characterized by an intense spring bloom, as shown by the data collected during summer (fig. 4 b,d). The location of the corresponding sites $29^\circ 5N, 75^\circ W$ and $40^\circ 5N, 59^\circ W$ in our idealized basin were chosen in similar regions with respect to the general circulation patterns (fig. 3 a,b). The simulated seasonal cycle of the MLD at each site is also coherent with the one in the data (fig. 4).

Modeled surface nitrate concentrations at BATS are fairly close to climatology (fig. 4a). The maximum values in winter are of the order of 0.4 mmol m$^{-3}$. Nitrate is exhausted in summer both in the observations and in the model. PP in the model also has the correct magnitude and timing (fig. 4b), with a peak in winter around 1 mmolC m$^{-3}$ day$^{-1}$ and lower values (0.5 mmolC m$^{-3}$ day$^{-1}$) in summer. Note that summer surface PP is sustained at very low ambient nitrate concentrations (0.01 mmol m$^{-3}$). In the model, summer surface PP is sustained by the recycling of the suspended dissolved inorganic matter and is a bit overestimated (0.6 instead of 0.4 mmolC m$^{-3}$ day$^{-1}$). The model reproduces well the summer PP which is difficult with simpler ecosystem models [Fasham, 1993; Follows and Dutkiewicz, 2001].

The modeled nitrate values at NABE (6 mmol m$^{-3}$) are lower than observed (8 mmol m$^{-3}$) during winter, and are depleted after the bloom as in the data (less than 0.5 mmol m$^{-3}$, fig. 4c). The simulated amplitude of PP (5.3 mmolC m$^{-3}$ day$^{-1}$) is very similar to the data (5 mmolC m$^{-3}$ day$^{-1}$, fig. 4d). In our modeled time series, the spring bloom at NABE occurs a month later than at BATS but a month earlier than one would expect from NABE data (fig. 4d). Again, this difference between the model and the
observations can be explained by the small size of the domain, which does not allow to capture the total amplitude of the bloom time lag with latitude.

**References**


<table>
<thead>
<tr>
<th>Description</th>
<th>Equation</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Tracer transport</strong></td>
<td>$\frac{\partial C}{\partial t} = SMS(C) - \nabla \cdot (uC) + \frac{\partial}{\partial z} \left( k_z \frac{\partial C}{\partial z} \right) + D_{lat}$</td>
</tr>
<tr>
<td><strong>Nitrate source/sink</strong></td>
<td>$SMS(NO_3) = -\mu_p L_I L_{NO_3} P + \mu_n NH_4 (+\partial_z f(z))$</td>
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<tr>
<td><strong>Ammonium source/sink</strong></td>
<td>$SMS(NH_4) = \alpha_p \gamma \mu_p L_I (L_{NO_3} + L_{NH_4}) P - \mu_p L_I L_{NH_4} P - \mu_n NH_4 + \alpha_z \mu_z Z + \alpha_d \mu_d D + \mu_{dom} DOM$</td>
</tr>
<tr>
<td><strong>Phytoplankton source/sink</strong></td>
<td>$SMS(P) = (1 - \gamma) \mu_p L_I (L_{NO_3} + L_{NH_4}) P - G_p - m_p P$</td>
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<tr>
<td><strong>Zooplankton source/sink</strong></td>
<td>$SMS(Z) = a_z (G_p + G_d) - m_z Z^2 - \mu_z Z$</td>
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<tr>
<td><strong>Detritus source/sink</strong></td>
<td>$SMS(D) = (1 - a_z) (G_p + G_d) + m_p P - G_d + f_z m_z Z^2 - \mu_d D - \nu_d \partial_z D$</td>
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<tr>
<td><strong>DOM source/sink</strong></td>
<td>$SMS(DOM) = (1 - \alpha_p) \gamma \mu_p L_I (L_{NO_3} + L_{NH_4}) P + (1 - \alpha_z) \mu_z Z + (1 - \alpha_d) \mu_d D - \mu_{dom} DOM$</td>
</tr>
<tr>
<td><strong>Light limitation</strong></td>
<td>$L_I = 1 - e^{-\frac{P_{PAR}}{k_{par}}} \frac{m_{NH_4}}{m_{CHL:N}}$</td>
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<td><strong>Nitrate limitation</strong></td>
<td>$L_{NO_3} = \frac{NO_3}{NO_3 + K_{no}} e^{-\psi_{NH_4}}$</td>
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<tr>
<td><strong>Ammonium limitation</strong></td>
<td>$L_{NH_4} = \frac{NH_4}{NH_4 + K_{nh}}$</td>
</tr>
<tr>
<td><strong>Grazing of phytoplankton</strong></td>
<td>$G_p = g_z \frac{p_P}{K_z + p_P + (1 - p) D} Z$</td>
</tr>
<tr>
<td><strong>Grazing of detritus</strong></td>
<td>$G_d = g_z \frac{(1 - p) D}{K_z + p_P + (1 - p) D} Z$</td>
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<tr>
<td><strong>Preference for phytoplankton</strong></td>
<td>$p = \frac{p_P}{p_P + (1 - p) D}$</td>
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<tr>
<td><strong>Remineralization flux</strong></td>
<td>$f(z) = f(z_{bio}) \left( \frac{z}{z_{bio}} \right)^{-r}$</td>
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<tr>
<td><strong>Instantaneous export</strong></td>
<td>$f(z_{bio}) = \int_{z_0}^{z_{bio}} (1 - f_z) m_z Z^2 dz$</td>
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**Table 1.** Biogeochemical equations
<table>
<thead>
<tr>
<th>Parameter name</th>
<th>Symbol</th>
<th>Value</th>
<th>Unit</th>
</tr>
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<tbody>
<tr>
<td>Nitrate limitation half-saturation value</td>
<td>( K_{\text{no}_3} )</td>
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<td>Ammonium limitation half-saturation value</td>
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<td>Inhibition of nitrate uptake by ammonium</td>
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<td>Light limitation half-saturation value</td>
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<td>( \mu_p )</td>
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<td>day(^{-1})</td>
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<td>Phytoplankton exsudation fraction</td>
<td>( \gamma )</td>
<td>0.05</td>
<td></td>
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<tr>
<td>Phytoplankton mortality rate</td>
<td>( m_p )</td>
<td>0.050</td>
<td>day(^{-1})</td>
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<tr>
<td>NH4 fraction of phytoplankton exsudation</td>
<td>( \alpha_p )</td>
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<td>Grazing half-saturation value</td>
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<td>mmol m(^{-3})</td>
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<td>Zooplankton maximal grazing rate</td>
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<td>Assimilated food fraction by zooplankton</td>
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<td>Preference for phytoplankton</td>
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<tr>
<td>Zooplankton excretion rate</td>
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<td>day(^{-1})</td>
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<tr>
<td>NH4 fraction of zooplankton excretion</td>
<td>( \alpha_z )</td>
<td>0.5</td>
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<tr>
<td>Zooplankton mortality rate</td>
<td>( m_z )</td>
<td>0.120</td>
<td>day(^{-1}) mmol(^{-1}) m(^{-3})</td>
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<tr>
<td>Fraction of slow sinking mortality</td>
<td>( f_z )</td>
<td>0.5</td>
<td></td>
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<tr>
<td>Nitrification rate</td>
<td>( \mu_n )</td>
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<td>day(^{-1})</td>
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<tr>
<td>DOM breakdown rate</td>
<td>( \mu_{\text{dom}} )</td>
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<td>day(^{-1})</td>
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<tr>
<td>NH4 fraction of detritus dissolution</td>
<td>( \alpha_d )</td>
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</tr>
<tr>
<td>Detritus sedimentation speed</td>
<td>( v_d )</td>
<td>3</td>
<td>m day(^{-1})</td>
</tr>
<tr>
<td>Detritus remineralization rate</td>
<td>( \mu_d )</td>
<td>0.050</td>
<td>day(^{-1})</td>
</tr>
<tr>
<td>C/N Redfield Ratio</td>
<td>( R_{C:N} )</td>
<td>6.6</td>
<td>mol C mol N(^{-1})</td>
</tr>
<tr>
<td>Depth of the euphotic layer</td>
<td>( z_{\text{bio}} )</td>
<td>150</td>
<td>m</td>
</tr>
</tbody>
</table>

Table 2. Biogeochemical parameters for the North Atlantic
Figure 1. Analytical forcings as a function of latitude used to force the model: (a) apparent temperature in °C, (b) penetrative solar radiation flux in W m\(^{-2}\), (c) fresh water flux in m year\(^{-1}\) and (d) the wind stress in N m\(^{-2}\), for winter (solid thin line) and summer (solid thick line).
Figure 2. Annual-mean barotropic stream function (BSF, in Sv).
Figure 3. Model (left panels) and data (right panels): (a,b) maximum MLD reached at each point (in m) and (c,d) annual-mean surface chlorophyll (in \(\text{log} \,(\text{mg m}^{-3})\)). The subtropical gyre is bounded by the zero-contour of the barotropic stream function (dashed black line, fig. 2) on the model solutions (a,c), and by the chlorophyll contour (black line, from d) on data (b). Hydrographic data are taken from the World Ocean Atlas [Locarnini et al., 2006; Antonov et al., 2006]. The location of BATS and NABE in the model and in the real ocean are indicated in the MLD pattern (a,b). Observed chlorophyll is taken from the SeaWIFS data (averaged from January 1997 to December 2006, available at http://oceancolor.gsfc.nasa.gov/).
Figure 4. Seasonal cycle for the model (black line) and data (open black circles) of surface nitrate in mmol m$^{-3}$ (upper panel) and surface primary production in mmolC m$^{-3}$ day$^{-1}$ (lower panel) at two JGOFS stations: BATS (left) and NABE (right). The locations of BATS (31°N, 64°W) and NABE (47°N, 20°W) are indicated on fig. 3 in the mixed-layer pattern. The mixed-layer seasonal cycle is also represented for the model (grey line) and data (grey crosses). All values shown are from the climatology constructed from the 10-yrs run after the spin up for the model and from climatological data for BATS [Michaels and Knapp, 1996], and data are restricted to the spring bloom and early summer periods at NABE [Weeks et al., 1993].