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A 3-Dimensional Isopycnic Coordinate Model of the Seasonal Cycling of Carbon and Nitrogen in the Atlantic Ocean

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Abstract. Chemistry and pelagic ecosystem modules have been coupled to an isopycnic coordinate ocean general circulation model (OGCM) in order to model the seasonal cycling of carbon and nitrogen, and the associated organic and inorganic carbon and nitrogen fluxes, in the Atlantic Ocean. After an initial 20 years spin-up integration of the physical model, some of the results of a 5 years integration of the coupled model are presented. Comparison with in situ observations of nitrate, total dissolved inorganic carbon and surface water partial pressure of CO_2 indicate that the model reproduces the main features of the seasonal cycling of plant nutrients and carbon over large parts of the model domain. The model predicts that the Atlantic Ocean acts as a net sink of atmospheric CO_2 north of about 30°N (uptake of about 0.35 Pg-C yr^{-1}), whereas there is an outgassing of CO_2 in the tropical region. ©1997 Elsevier Science Ltd

1 Introduction

Atmospheric carbon dioxide has characteristic life times raging from decades to centuries, and about two-thirds of the man-generated greenhouse forcing is caused by increasing concentrations of atmospheric CO_2 (Houghton et al., 1996). It is therefore of paramount importance to improve our understanding of the past and present transport, mixing and cycling of carbon within and between the atmosphere, the terrestrial biosphere and the ocean carbon reservoirs if reliable predictions of the future global carbon cycle is to be made.

In this paper a coupled 3-dimensional (3-D) physical-biogeochemical model for one component of the carbon cycle is presented, namely for the cycling of carbon in the ocean. Whereas the majority of existing ocean carbon cycle models focus on the oceanic uptake of man-generated (mainly fossil fuel) CO_2 on decadal to centennial time scales, the Nansen Center Carbon Cycle Model (N3CM) has been developed to describe the natural cycling of carbon and nitrogen in the ocean on seasonal time scales. Through the modelled physical and biogeochemical variables, quantities like the surface water concentration of CO_2, the flux of CO_2 across the air sea-interface, the planktonic primary, regenerated, total and export productions (in carbon and nitrogen units), and the various sink and source terms for these quantities are derived.

N3CM represents in many ways an extension to the basin scale 3-D ecosystem model of Sarmiento et al. (1993). In N3CM, the ecosystem formulation includes both the cycling of carbon and nitrogen, the biogenic formation and subsequent dissolution of $CaCO_3$, the chemistry of the carbonic acid system in seawater, and a formulation of the exchange of CO_2 across the air-sea interface, wheres the model of Sarmiento et al. (1993) is a pure nitrogen model without carbon chemistry. In addition, N3CM is based on an isopycnic coordinate OGCM with explicit mixed layer (ML) dynamics, and for the transport of the biogeochemical compartments the non-diffusive numerical advection scheme of Smolarkiewicz and Grabowski (1990) is used; whereas Sarmiento et al. (1993) used a level model without explicit ML dynamics, and applied the diffusive upstream scheme for the transport of the ecosystem compartments.

The present version of N3CM covers the Atlantic Ocean between approx. 15°S and 80°N. The Atlantic version represents a first step towards a global biogeochemical model. The global model, which is now under development, will be used to assess the response on the oceanic cycling of carbon and plant nutrients on seasonal to decadal time scales to possible changes in the ocean circulation and thermodynamics caused by interannual variations in the climate system or to a modified atmosphere (for instance related to El Niño events and various global warming scenarios), and to anthropogenic generated input of nitrogen to the oceans (through river run-off and precipitation). In addition, the model will be used as a

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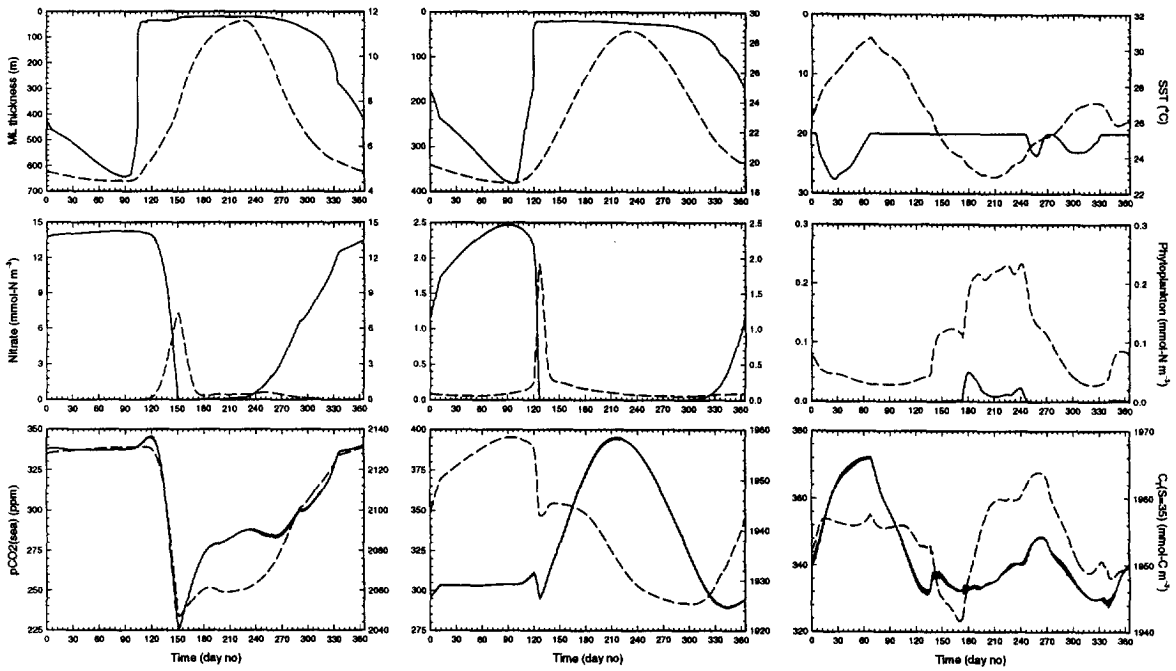

Fig. 1. Modelled seasonal evolution of ML layer thickness and temperature (upper panels), ML layer nitrate and phytoplankton concentrations (mid panels), and surface water pCO_2 and ML total dissolved inorganic carbon content normalized to a constant salinity of 35 per mil (lower panels) for a station south-west of Iceland (left row), Bermuda station 'S' (mid row) and a station off the coast of Angola (right row). The solid lines are plotted against the left, and the dashed lines against the right, y-axes. Note the different scaling of the y-axes.

platform for modelling other climatically active components (as dimenthyl sulfide and nitrous oxide), and to examine the potential effect lack of micronutrients and increased ultra violet radiation may have on the marine biota.

A description of an early version of the coupled model is given by Drange (1994), whereas the version of the model presented here is currently being documented. In the following sections, the coupled physical-biogeochemical model is briefly described, and some of the model results are presented.

2 The coupled physical-biogeochemical model

2.1 Physical model

The coupled model is driven by the Miami Isopycnic Coordinate Ocean Model MICOM (Bleck et al., 1992), but in the implementation of the James Rennell Centre for Ocean Circulation, Southampton (New et al., 1995). The main differences between the present model and the model discussed by New et al. are that we use the Gaspar (1988), and not the Kraus and Turner (1967), parameterization for the dissipation of the turbulent kinetic energy in the surface mixed layer (ML), and that the prescribed densities of the isopycnals are somewhat different in order to improve the Equatorial dynamics (A. New, 1994, pers. comm.).

On the top of the model ocean the ML formulation incorporates the integral effect of wind stirring and buoyancy fluxes at the surface. The prognostic variables for the ML are the horizontal velocity vector, the layer thickness, temperature T and salinity S. The density of the mixed layer is thus allowed to evolve in time and space according to the surface forcing and mixing between different water masses. Below the mixed layer there are 19 layers of prescribed (and constant) density with the horizontal velocity vector, the layer thickness and salinity as prognostic variables. The potential densities of the isopycnic layers in the dimensionless σ_0-units (reference pressure at surface) are 24.70, 25.28, 25.77, 26.18, 26.52, 26.80, 27.03, 27.22, 27.38, 27.52, 27.64, 27.74, 27.82, 27.88, 27.92, 28.00, 28.06, 28.09 and 28.12. Unstable stratification is removed by mixing the actual water masses (and all of their physical and biogeochemical properties) uniformly in the vertical.

The model equations are formulated on a 126 by 104 points Mercator grid mesh, yielding a horizontal resolution of 1° by 1° along 40°W and Equator, and 0.5° by 0.5° in the Gulf of Mexico and Gulf of Guinea. The Nordic Seas are covered by 12 by 13 grid cells, which means that this region should be considered as a boundary zone for the Atlantic Ocean.

A sea-ice model was not available with the present version of MICOM, but the presence of ice was mimicked by setting the surface wind stirring, the heat fluxes and the CO_2 gas exchange coefficient to zero where the modelled ML temperature went below -1.8°C. Furthermore, the northern and southern model boundaries are solid walls, and the isopy-

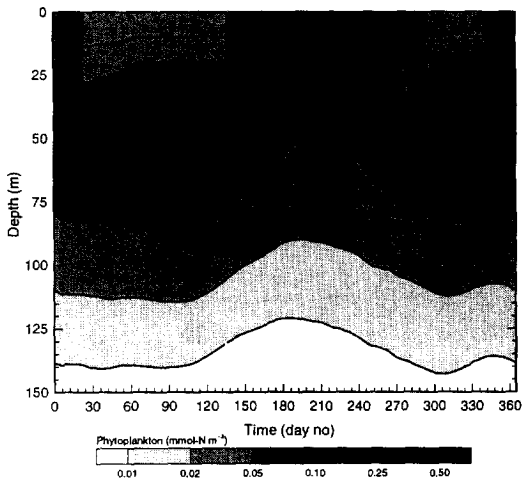

Fig. 2. The modelled vertical distribution of phytoplankton (mmol-N m^{-3}) throughout the year at the Angola station. The solid lines show the layer interfaces (the second upper-most line represents the ML depth).

cnal diffusivity is increased linearly by an order of magnitude, from the interior value of about 1×10^3 m^2 s^{-1} to 1×10^4 m^2 s^{-1}, over the ten grid cells closest to the walls. Diapycnal mixing is included by assuming that the diapycnal diffusivity is given by the stability dependent expression $1 \times 10^{-7}/N$ m^2 s^{-2} (Gargett, 1984), where N (s^{-1}) is the local buoyancy frequency. A no slip lateral boundary condition was used for the velocity fields.

2.2 Biochemical model

The ecosystem formulation adopted for the euphotic zone is built on the seven-compartment nitrogen-based model of Fasham et al. (1990) (hereafter FDM). The FDM-model was originally developed to study the annual cycle of nitrogen at Station 'S' near Bermuda (32° 10'N, 64° 30'W), and has later been slightly adjusted and tested against observations from both Station 'S' and the Ocean Weather Station 'India' (59°N, 19°W) (Fasham, 1993). It is this latter version of the FDM-model that has been used in this study. The original version of the FDM-model has been successfully coupled to the Princeton general circulation model for the North Atlantic (Sarmiento et al., 1993), and may be taken as a state-of-the-art formulation of the cycling of nitrogen in the ocean.

The FDM-model has phytoplankton P, zooplankton Z and bacteria B as the living biota; the nutrients are ammonium NH_4^+ and nitrate NO_3^-; in addition there is one pool of dissolved organic nitrogen DON and one pool of particulate organic nitrogen PON.

In this study, organic and inorganic carbon and total alkalinity have been coupled to the flows of nitrogen and ammonium in order to simulate the cycling of carbon in the ocean. Both simple and relatively complex marine carbon cycle and ecosystem models (for instance Peng et al., 1987; Aksnes and Lie, 1990; Maier-Reimer, 1993), and primary production estimates (Eppley and Peterson, 1979), are generally based on the assumption that the stoichiometric composition of the planktonic organic material is on average constant and equals (or is close to) the chemical composition of seawater. In this study it is assumed that the carbon to nitrogen ratios for phytoplankton, zooplankton and bacteria are 7, 5.5, and 5, respectively (Redfield et al., 1963; Fasham et al., 1990; Anderson, 1992). This means that the atomic carbon to nitrogen ratio for dissolved and particulate organic matter varies according to the sources and sinks of these compartments. Total alkalinity A_T and the carbon reservoirs total dissolved inorganic carbon C_T, dissolved and particulate organic carbon (DOC and POC, respectively), have therefore been added to the FDM-model, yielding a total of 11 biogeochemical compartmets.

The formation and subsequent dissolution of biogenic $CaCO_3$ is not explicitly included in the ecosystem model. However, the formation of $CaCO_3$ is parameterized by reducing C_T by one concentration unit and increasing A_T by two charge units for each concentration unit of $CaCO_3$ formed. It has been assumed that the flux of $CaCO_3$ that sinks out of the euphotic zone is proportional (on a C-mole basis) to that of organic matter, and the proportionallity factor is set to 0.2 (Broecker and Peng, 1982, p. 11 and 269) in waters with a temperature of 20°C or more, but is reduced in cold waters (Maier-Reimer, 1993).

In addition, C_T in the uppermost biogeochemical layer is influenced by the exchange of CO_2 across the air-water interface according to the flux expression (Wanninkhof, 1992)

$$F = 0.35\, U_{10}^2\, K_0\, (\mathrm{Sc}/660)^{-0.5}\, (p\mathrm{CO_2}^{\mathrm{atm}} - p\mathrm{CO_2}^{\mathrm{sea}}). \quad (1)$$

Here U_{10} (m s^{-1}) is the friction velocity, K_0 (mol-C m^{-3} atm^{-1}) is the solubility of CO_2 in seawater, Sc (dimensionless) is the Schmidt number for CO_2 in sea water, and $p\mathrm{CO_2}^{\mathrm{atm}}$ and $p\mathrm{CO_2}^{\mathrm{sea}}$ are the concentration of CO_2 in air and sea, respectively, both evaluated at the water saturation pressure. The atmospheric concentration of CO_2 is prescribed and is based on observed $p\mathrm{CO_2}$ values from the 1980's (Boden et al., 1991), with an increasing trend of 1.6 ppm yr^{-1} and seasonal amplitudes ranging from 0.6 ppm at 20°S to 15 ppm poleward of 75°N. Furthermore, A_T and C_T have been adjusted for variations in the water balance according to the evaporation and precipitation fields (see Subsec. 3.1).

Below the euphotic zone we follow the parameterization used by Sarmiento et al. (1993), and let the biogenic compartments P, Z, B and DON decay to NH_4^+, and then to NO_3^-, with constant decay rate of 0.1 day^{-1}. In a similar way, DOC decays to C_T with a similar decay rate. Particulate organic material falling out of the euphotic zone is remineralized according to the expression derived by Martin et al. (1987), whereas biogenic $CaCO_3$ is distributed in the vertical with an e-folding length of 4000 m, and is then dissolved into seawater. All biogenic matter that ends up on the ocean floor is remineralized in the bottom layer.

Given C_T, A_T, T and S, the partial pressure of CO_2 in seawater has been computed based on the chemical model of Peng et al. (1987) (see Subsec. 3.1).

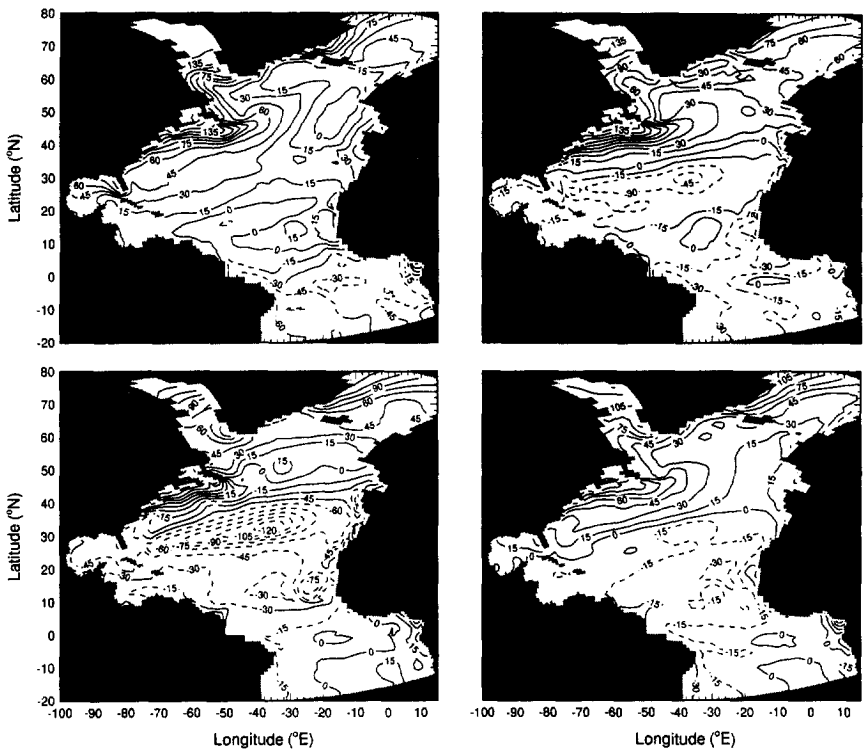

Fig. 3. Modelled $pCO_2{}^{atm} - pCO_2{}^{sea}$ (ppm) for Jan-Mar (upper left panel), Apr-Jun (upper right), Jul-Sep (lower left), and Oct-Dec.

3 Model simulation

3.1 Initialization and spin-up

The physical model was initialized by the September temperature and salinity fields of the Levitus (1982) data set, and spun up from rest. The forcing functions used were the Hellerman and Rosenstein (1983) winds, the Esbensen and Kushnir (1981) heat fluxes, and the Jaeger (1976) precipitation minus the Esbensen and Kushnir (1981) evaporation fields. The ML temperature and salinity were relaxed towards the monthly mean Levitus' temperature and salinity climatology. The relaxation coefficient for the heat fluxes was 35 W m^{-2} K^{-1}, and a corresponding coefficient was used for the salinity flux, yielding approximately a 4-month relaxation time scale for a 100 m thick ML.

After an integration time of 20 years, the basin-averaged kinetic energy and the basin-averaged layer thicknesses had reached a quasi steady-state annual cycle. It should here be mentioned that a true annual steady-state circulation or an exact annual thermodynamic balance between the surface water and the surface forcing will not occur with the present model set-up since the surface forcing is realistic and there is no exchange of water through the northern and the southern model boundaries.

The ecosystem and chemical modules were run fully coupled to the physical model for 5 years, starting on January 1 of year 20 of the physical model run. A quasi-annual cycle was obtained after 2–3 years. The biogeochemical model was initialized with the annual mean nitrate field compiled by J. L. Reid and A. Mantyla at Scripps Institution of Oceanography, University of California, and presented by Levitus et al. (1993). For the C_T and A_T fields, data from the following cruises and programmes were merged and interpolated onto the isopycnic model coordinate system: GEOSECS, TTO/NAS, AJAX, SAVE, NABE, and the Hudson-82 and Mosby-93 cruises. The data sets from the first five of these programmes were supplied by T. Takahashi and co-workers at Lamont-Doherty Earth Observatory of Columbia Univ., New York. For these data sets the quantities pCO_2 and C_T have been measured, and A_T is computed from the chemical model of Peng et al. (1987). To ensure consistency between the carbonic data and the model, the chemical module of Peng et al. (1987) has been adopted for this study. The Hudson-82 and Mosby-93 data sets have been slightly adjusted in order to make them consistent with the TTO/NAS data set (Lundberg, 1995).

For the initial concentration of the remaining biogeochemical compartments, we follow the values given by Sarmiento et al. (1993), and set $P = 0.14$ mmol m^{-3}, $Z = B = 0.014$ mmol m^{-3}, and $NH_4^+ = DON = PON = 0.1$ mmol m^{-3} in the upper biogeochemical layer, with exponentially decreasing concentrations with depth with a scale length of 100 m.

3.2 Results and discussion

The ecosystem dynamics at high and mid latitudes is highly dependent on the seasonal evolution of the ML depth. Compared with ML depth estimates based on hydrography (Levitus, 1982; McCartney and Talley, 1982; Lamb, 1984), it is found that the modelled summer ML follows the observational data closely over the entire Atlantic Ocean. There is also a general agreement between the observational and modelled ML depth for the winter situation, but the ML in the southern part of the sub-tropical gyre is too deep and extends too far to the west. One candidate for this failure is that the modelled Gulf Stream is too diffusive in the model and does not separate from the coast at Cape Hatteras as observed. This is a common problem for OGCM's with horizontal resolution of about 1° by 1° (Holland and Bryan, 1994; New et al., 1995). The mis-placed Gulf Stream affects the dynamics and thermodynamics of parts of the south-western sub-tropical gyre, leading to a too cold, and therefore to a too deep, winter ML here.

The seasonal evolution of some of the modelled physical and biogeochemical fields from locations south-west of Iceland (64°N, 26°W), near Bermuda St. 'S' (32°N, 65°W), and off the coast of Angola (2°S, 5°W) are plotted in Fig. 1. These, and the other model results presented in this section, have been taken after the coupled model has been run for 5 years.

3.2.1 High latitude dynamics

At high latitudes during winter (represented by the Iceland station), cooling of the surface water leads to ML depths of 500 m or more, and high ML NO_3^- and C_T concentrations, in qualitative agreement with hydrographic and nutrient observations (Levitus, 1982; Takahashi et al., 1993). In spring, when the air temperature and the solar irradiance increase and the wind mixing is relatively low, heating of the upper parts of the water column leads to rapid shallowing of the ML. Relatively high irradiance, together with strong stratification of the upper parts of the water column and low concentrations of zooplankton, leads to rapid growth of phytoplankton. The phytoplankton spring bloom lasts until the ML nitrate concentration is depleted around day 150. The formation of organic matter lowers the C_T content (normalized to a constant salinity of 35 per mil) by ~80 μmol-C kg^{-1}, and leads to a 120 ppm drop in the surface water pCO_2, both quantities in agreement with observations (Takahashi et al., 1993). After the spring bloom, planktonic release of ammonium and dissolved inorganic carbon leads to a more rapid build-up of C_T than of NO_3^-. The modelled pCO_2 field is closely correlated with the C_T field except for increased pCO_2 values prior to and after the phytoplankton bloom. This mis-match is mainly governed by heating (cooling) of the surface water as pCO_2 increases (decreases) by about 4% for each degree increase (decrease) in temperature (Takahashi et al., 1993).

3.2.2 Mid latitude dynamics

The modelled seasonal evolution of the ML depth at the Bermuda station is similar, but exhibits a weaker amplitude, to that of the Iceland station. Hydrography from the Bermuda station shows a very strong interannual variability in the maximum winter ML depth, ranging from 75–375 m over the years 1954–1987, and with a mean value of about 200 m (Steele and Henderson, 1993). It is therefore hard to compare the modelled fields with observations since the model represents (at best) the climatological, or mean, situation. However, the modelled ML depth is significantly deeper than the mean depth of about 200 m, indicating that the modelled ML is too deep. Observations of the surface water NO_3^- concentration show considerable interannual variability (Conkright et al., 1994), but consistent with the modelled ML depth, the winter surface water NO_3^- concentration is probably too high. It is seen in the lower panel in Fig. 1 for the Bermuda station that the effect of the spring 'bloom' on C_T and pCO_2 is relatively weak. In fact, at the Bermuda station the surface water pCO_2 field is mainly governed by the seasonal evolution of the SST field. According to the surface ocean CO_2 measurements of Keeling (1993) from the years 1983–1991, it follows that the modelled C_T and pCO_2 fields are in reasonable agreement with observations. For instance, the observed C_T field (normalized to a constant salinity of 35 pro mil, as in Fig. 1) vary between 1930 and 1965 μmol-C kg^{-1}, and the pCO_2 field between 300 and 380 ppm.

3.2.3 Equatorial dynamics

An illustration of the Equatorial biogeochemical dynamics is shown in the right row in Fig. 1. Here the ML is shallow and almost constant throughout the year (the modelled ML thickness is not allowed to go below 20 m), and the evolution of the modelled NO_3^- and P concentrations are very different from those at the Iceland and Bermuda stations. The increase in NO_3^-, P and C_T in mid to late summer are caused by divergence in the surface velocity field, and thus by upwelling of subsurface waters, as seen in Fig. 2. Some correlation between the pCO_2 and temperature fields is seen, especially for the first 3 months of the year.

3.2.4 Seasonal evolution of ΔpCO_2

The seasonal evolution of the modelled ΔpCO_2 is shown in Fig. 3. In the first quarter of the year, there is an oceanic uptake of carbon poleward of about 10°N. The large ΔpCO_2 south of the coast of New Foundland for this and the following quarters are caused by cooling of the warm Gulf Stream water flowing northwards along the coast of North America. This effect is real, but the position is shifted and the magnitude is too large compared to analyzed ΔpCO_2 maps based on observations from the region (Lefèvre, 1995; Takahashi et al., 1995). Heating of the surface water leads in general to increased pCO_2^{sea} in the second and third quarter, although the increase in SST is partly balanced and even counteracted

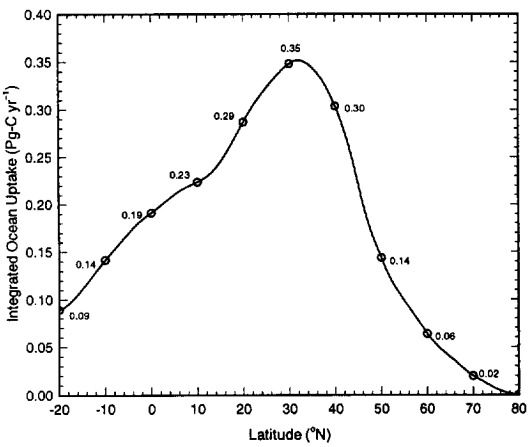

Fig. 4. The net ocean uptake of carbon in Pg-C yr^{-1} between the indicated latitudes and 80°N.

at high and mid latitudes. In the central subtropical gyre in late summer the surface water pCO_2 is up to 120 ppm higher than pCO_2^{atm}. Such a feature is probably real (the data coverage is sparse), but strong stratification of the upper part of the water column and climatological forcing without variability prevents mixing of subsurface and ML waters, leading to a warm and nutrient depleted ML, and probably to a overestimated pCO_2^{sea} field.

3.2.5 Air-Sea exchange of CO_2

The modelled ocean uptake of carbon, integrated from the northern boundary and southwards, is shown in Fig. 4. There is a net ocean uptake of about 0.35 Gt-C yr^{-1} north of about 30°N, and a net outgassing of carbon at lower latitudes. Northward of Equator, the uptake is 0.19 Gt-C yr^{-1}, and at 20°S, the uptake is 0.09 Gt-C yr^{-1}. Since the model domain is closed, the latter result shows that there is an accumulation of carbon in the model. This is partly due to the steady increase in pCO_2^{air} during the model run (pCO_2^{air} increases by 1.6 ppm yr^{-1}), and partly due to the fact that neither the physical model (see Subsec. 3.1) nor the biogeochemical model are in an exact seasonal steady-state mode.

3.3 Conclusions

This paper presents, to the best of our knowledge, the first attempt to couple ecosystem dynamics and carbon chemistry to an isopycnic coordinate OGCM. The present version of the model covers the the North Atlantic region. Several multi year simulations have been carried out with the coupled model, and some of the key results from a 5 years simulation are presented. Comparison with in situ observations of nitrate, total dissolved inorganic carbon and surface water partial pressure of CO_2 indicate that the model reproduces the main features of the seasonal cycling of plant nutrients and carbon over large parts of the model domain. The model predicts that the Atlantic Ocean acts as a net sink of atmospheric CO_2 north of about 30°N (uptake of about 0.35 Pg-C yr^{-1}), whereas there is an outgassing of CO_2 in the tropical region.

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