1. Introduction

Coastal environments play an important role in the global biogeochemical cycles of carbon and macronutrients (DIN, Si(OH)₄ and PO₄) (RSB). Despite their relatively small volume, they contribute disproportionately to global primary production (Rousseau et al., 2002). In coastal transition systems, tightly coupled geological, hydrodynamic, geochemical and biological processes are modulated by a wide array of forcing mechanisms, which include, light, temperature, wind stress, waves, tides, freshwater discharge, as well as continental and oceanic nutrient inputs. These forcings operate on very different temporal and spatial scales and, together, set strong constraints on the biogeochemical functioning of the coastal ocean. They determine the spatial and temporal distributions of dissolved and particulate species and markedly influence the magnitude of biogeochemical reaction rates. Achieving a detailed and mechanistic understanding of the processes that control the biogeochemical dynamics along the river–ocean continuum is thus particularly challenging, yet, essential for predicting the response of the coastal ecosystem to the continuously evolving anthropogenic pressure.

Over the past decades, human activity has significantly accelerated the supply of land-derived nutrients to the coastal ocean and, as a result, nearshore areas bordering highly populated regions have experienced increased eutrophication (Neal et al., 2000; Cloern, 2001; Turner et al., 2003). In north-western Europe, massive developments of Phaeocystis colonies have regularly been observed along the coastline of the Eastern Channel and the Southern Bight of the North Sea (Lancelot et al., 1987; Riegman et al., 1992; Mills et al., 1994; Rousseau et al., 2002). A characteristic and recurrent diatom–Phaeocystis–diatom succession develops in response to the seasonal availability of nutrients (Rousseau et al., 2002). In this region, nutrient
levels are controlled by the mixing of Atlantic waters with nutrient-rich freshwater from the Scheldt, the Rhine/Meuse and the Seine estuary (Pingree and Maddock, 1977; Salomon and Breton, 1993; Lacroix et al., 2004, 2007b). The present study focuses on the Belgian coastal zone where phytoplankton and nutrient concentrations have been intensively monitored at one sampling station (St. 330; N51°26.05; E2°48.50) between 1988 and 2000, at time intervals ranging from twice a week to twice a month (Breton et al., 2006). The long-term time series thus provide key information on the variable nutrient supply from the Scheldt estuary in this highly dynamic area. Nonetheless, high-resolution data are only available at a single sampling station and can thus not easily be extrapolated at the scale of the entire coastal ecosystem. To date, observations on nutrient and phytoplankton distributions in the Belgian coastal zone are available but they are limited in time span and spatial coverage and merely resolve the monthly timescale (Borges and Frankignoulle, 1999, 2002; Muylaert et al., 2006; Van der Zee and Chou, 2005).

Reactive-transport models (RTMs) are particularly suitable to develop a quantitative understanding of biogeochemical transformations and fluxes in coastal environments. In particular, they complement field measurements, because they provide a mechanistic description of process interactions over a spectrum of scales that cannot be resolved by field observations. More than two decades ago, Nihoul and Hecq (1984) identified the dominant patterns in the ecosystem structure of the Belgian coastal zone using hydrodynamic simulations to support the interpretation of ten year-averaged field observations of nutrients, phytoplankton, zooplankton and suspended sediments. Since then, valuable insights on the biogeochemical dynamics of the Belgian coastal zone have been gained through further modeling efforts. In particular, the MIRO model (Lancelot et al., 2005) was designed to describe diatom and Phaeocystis colony blooms and the related carbon and nutrient (N, P, and Si) cycles in the eastern channel and southern North Sea with a focus on the Belgian coastal zone. It was first implemented in a multi-box framework to study the link between anthropogenic nutrient loads by the Seine and Scheldt rivers and the magnitude and extent of diatom/Phaeocystis blooms (Lancelot et al., 2005), as well as its evolution over the last decade (Gypens et al., 2007) and the last 50 years (Lancelot et al., 2007). The carbon dynamics and CO2 air/water exchange which result from this coupling has also been addressed (Gypens et al., 2004, 2009). Recently, MIRO has been coupled to the COHERENS-3D hydrodynamic model (Lacroix et al., 2007b). The coupled model has been used to simulate and assess the relative impact of river (Seine, Somme, Scheldt, and Rhine/Meuse) and Atlantic waters on the contemporary (1991–2003) spatio-temporal distribution of diatom/Phaeocystis blooms and the related carbon and nutrient cycling in the Channel and Southern Bight of the North Sea (Lacroix et al., 2007a). At the larger scale of the North Sea, three-dimensional box-averaged biogeochemical models forced by spatially-resolved transport models (Baretta et al., 1995; Patsch and Radach, 1997), two-dimensional simplified biogeochemical models (Hydes et al., 1996) and three-dimensional coupled transport and biogeochemical models (Skogen et al., 1995; Delhez, 1998; Moll, 1998) have also been used to simulate primary production (Moll, 1998; Skogen and Soland, 1998; Luyten et al., 1999; Skogen and Moll, 2005), toxic algae blooms (Cugier et al., 2005), nutrient cycling (Allen et al., 2001; Vichi et al., 2004) and long-term eutrophication (Patsch and Radach, 1997) in the entire North Sea. In parallel to the development of North Sea models, a suite of RTMs that couple the physics and biogeochemistry of estuaries have been developed over the last two decades. The physical extension of these estuarine models was generally restricted to the area between the estuarine mouth and the upper limit of the salt intrusion (Soetaert and Herman, 1995a, b; Regnier et al., 1997, Regnier and Steefel, 1999; Vanderborght et al., 2002; Tappin et al., 2003; Arndt et al., 2005). For instance, Regnier and Steefel (1999) used the 1D model CONTRASTE to calculate the nitrogen export fluxes to the coastal zone in an estuarine system subject to fully-transient conditions. However, the control of these fluxes on coastal biogeochemistry was not explicitly simulated. Although, integrated representations of the morphologically complex river–ocean continuum are increasingly used (Proctor et al., 2000; Tappin et al., 2003; Cugier et al., 2005; Lancelot et al., 2007; Vanderborght et al., 2007) they generally compromise on the spatial, temporal and dynamical resolution. In particular, no study has addressed so far the seasonal evolution in ecosystem structure and functioning of the Scheldt estuary and adjacent coastal zone using a continuum approach, that is, none has considered the influence of the estuarine domain on the coastal environment and vice-versa. Nutrient supplies are either specified by measured riverine fluxes discharging directly into the coastal zone or, at best, transiting through a poorly resolved estuarine system. Therefore, these approaches do not provide a realistic description of the retention and transformation of land-derived nutrient inputs within the estuary, which may influence both the magnitude and timing of export fluxes to the coastal zone. In addition, the feedbacks of coastal processes on the estuarine environment are currently not accounted for. Estuarine models generally assume that the biogeochemistry of the estuary is exclusively influenced by the riverine inputs from the watershed. Therefore, combined observational and modeling efforts do not provide a comprehensive, synoptic view of the biogeochemical dynamics along the entire mixing-zone between river and marine waters and the quantitative significance of the estuary for the functioning of the adjacent coastal zone remains poorly evaluated. As a result, the mechanisms that lead to the emergence and short-term temporal variability in ecosystem structure along such an estuary–coastal zone continuum are still largely unknown.

In this study, a fully transient two-dimensional, nested-grid hydrodynamic model of the Scheldt (B/NL) and adjacent nearshore coastal zone continuum is coupled to the biogeochemical models MIRO and CONTRASTE for the coastal zone and the estuary, respectively. Transient model simulations are performed with a high spatial (80–750 m) and temporal (30 min) resolution over a period of one year (January–December 1995). The system scale simulations provide time series of nutrient transformations and fluxes along the entire estuary–coastal zone continuum, as well as weekly-resolved nutrient inventories for the estuarine and the coastal zone sub-domains. These results are then used to investigate the respective importance of the physical and biogeochemical controls on the spatio-temporal structure of an entire estuary–coastal zone transition continuum.

2. The model

2.1. Hydrodynamics

The estuarine and coastal hydrodynamics is described using the nested hydrodynamic model MIKE 21 NHDC® (DHI). The model simulates unsteady two-dimensional flows in a vertically homogenous water column. It solves simultaneously the set of depth-averaged Saint-Venant equations for barotropic flow:

$$\frac{\partial \xi}{\partial t} + \frac{\partial p}{\partial x} + \frac{\partial q}{\partial y} = 0$$  \hspace{1cm} (1)

$$\frac{\partial p}{\partial t} + \frac{\partial}{\partial x} \left( \frac{p^2}{\eta} \right) + \frac{\partial}{\partial y} \left( \frac{pq}{\eta} \right) + gh \frac{\partial \xi}{\partial x} + \frac{gq}{\eta} \frac{\partial^2 p}{\partial y^2} + \frac{g}{\eta} \frac{\rho_c}{\rho} \frac{\partial^2 q}{\partial y^2} + \frac{\rho_c}{\rho} \frac{\partial^2 \xi}{\partial y^2} = -\Omega_y q - \frac{2}{\eta} \left( \frac{\partial^2 p}{\partial x^2} + \frac{\partial^2 q}{\partial y^2} \right) - \frac{v_w}{\rho} \frac{\partial w_x}{\partial x} + \frac{h}{\rho} \frac{\partial}{\partial y} \frac{p_n}{\rho} = 0$$  \hspace{1cm} (2)

$$\frac{\partial q}{\partial t} + \frac{\partial}{\partial x} \left( \frac{q^2}{\eta} \right) + \frac{\partial}{\partial y} \left( \frac{pq}{\eta} \right) + gh \frac{\partial \eta}{\partial y} + \frac{gq}{\eta} \frac{\partial^2 q}{\partial y^2} + \frac{g}{\eta} \frac{\rho_c}{\rho} \frac{\partial^2 p}{\partial y^2} + \frac{\rho_c}{\rho} \frac{\partial^2 q}{\partial y^2} = \Omega_y p - \frac{2}{\eta} \left( \frac{\partial^2 p}{\partial x^2} + \frac{\partial^2 q}{\partial y^2} \right) - \frac{v_w}{\rho} \frac{\partial w_y}{\partial y} + \frac{h}{\rho} \frac{\partial}{\partial x} \frac{p_n}{\rho} = 0$$  \hspace{1cm} (3)
where \( \xi \) is the water elevation \((m)\); \( p = v_x, h, q = v_y, h \) are the flux densities \((m^2/s)\); \( h \) is the water depth \((m)\); \( g \) is the acceleration due to gravity \((m/s^2)\); \( Ch \) is the Chézy coefficient \((m^{1/2}/s)\); \( \Omega_x, \Omega_y \) are the Coriolis coefficients \((1/s)\); \( E \) is the eddy viscosity \((m^2/s)\); \( f \) is the wind friction factor; \( w \) is the wind speed; \( u_x, u_y \) are the wind components in \( x \)- and \( y \)-directions \((m/s)\); \( \rho_w \) is the water density \((kg/m^3)\) and \( \rho_0 \) the atmospheric pressure \((kg/m^2/s^2)\). Eqs. (1)-(3) are spatially discretised on dynamically nested, rectangular, staggered grids and integrated using a ‘fractioned-step’ technique combined with an Alternating Direction Implicit (ADI) algorithm \((\text{Leendertse, 1967})\). The ADI scheme leads to a tri-diagonal system which is solved by applying a double-sweep algorithm. Three dynamically nested grids which cover the brackish and saline zones of the Scheldt estuary, as well as the well-mixed part of the Belgian coastal zone form the model support. The grid size is \(83.33 \times 83.33\) \(m\) for the channel up to the Belgian/Dutch border \((\text{grid 3, Fig. 1})\), \(250 \times 250\) \(m\) for the lower estuary \((\text{grid 2, Fig. 1})\) and \(750 \times 750\) \(m\) for the coastal zone \((\text{grid 1, Fig. 1})\). The total amount of grid points is larger than \(60,000\) for a total surface area of \(8700\) \(km^2\). The temporal resolution of the hydrodynamic simulations is limited by the Courant criterium and set to \(30\) \(s\). Simulations are performed from June 1994 to December 1995 and results are processed for the period between January 1995 to December 1995, after the model has reached a hydrodynamic regime.

The eddy viscosity coefficient, \( E \), in Eqs. (2)-(3) is calculated according to the Smagorinsky formula, which relates \( E \) to the local flow velocities using a Smagorinsky constant of \(0.5\) \((\text{Smagorinsky, 1963})\). Bed friction exerted on the moving water is characterized by means of a roughness formulation with a Chézy coefficient, \( Ch \). This coefficient is the key parameter for the hydrodynamic model calibration, performed by comparing observed and simulated tidal amplitudes and phases. The best calibration is achieved with \( Ch = 90\) \(m^{1/2}/s\) for the coastal zone and \( Ch = 75\) \(m^{1/2}/s\) for the Scheldt estuary. The wind friction factor \( f \) is set to \(0.006\).

2.2. Transport

Coupled reaction and mass transport is described by the two-dimensional continuity equation for scalar quantities:

\[
\frac{\partial h C_i}{\partial t} + \frac{\partial}{\partial x}(v_x h C_i) + \frac{\partial}{\partial y}(v_y h C_i) = \frac{\partial}{\partial x} \left( h \cdot D_{xx} \frac{\partial C_i}{\partial x} \right) + \frac{\partial}{\partial y} \left( h \cdot D_{yy} \frac{\partial C_i}{\partial y} \right) + h \sum R_{n,i}
\]

(4)

\[
\frac{\partial h C_i}{\partial t} + \frac{\partial}{\partial y}(v_x h C_i) = \frac{\partial}{\partial x} \left( h \cdot D_{xy} \frac{\partial C_i}{\partial y} \right) + h \sum R_{n,i}
\]

(5)

where \( C \) is the concentration of species \( i \); \( v_x, v_y \) are the horizontal velocity vector components; \( D_{xx, D_{yy, D_{xy}}} \) are the horizontal dispersion coefficients and \( R_i = \sum R_{n,i} \) denotes the sum of all transformation processes, \( n \), affecting species \( i \). Eqs. (4) and (5) are integrated numerically using an operator splitting approach. The advection-dispersion terms are solved with the third order finite difference scheme QUICKEST \((\text{Ekebjaerg and Justesen, 1991})\), which avoids instabilities associated with central differencing and numerical dissipation. The source/sink term in the continuity equation is integrated using the Euler method \((\text{e.g. Press et al., 1992})\) with a time step of \(1800\) \(s\), and is discussed in further detail in the following section. The dispersion coefficients account mainly for the non-resolved sub-grid scale processes and are calibrated using longitudinal salinity profiles as an indicator of the mixing of water masses. A good agreement with observations is achieved with dispersion coefficients of \(40\) \(m^2/s\), \(100\) \(m^2/s\) and \(150\) \(m^2/s\) for the upper estuary, the lower estuary and the coastal zone, respectively.

2.3. Biogeochemistry

The biogeochemical reaction-network is coupled to the hydrodynamic and transport processes using the ECOLab©environmental modeling tool \((\text{DHI})\). The reaction network is based on a combination of the biogeochemical MIRO model \((\text{Lancelot et al., 2005})\), which is implemented for the coastal zone \((\text{grid 1, Fig. 1})\) and the biogeochemical reaction network of the CONTRASTE model \((\text{Regnier and Steefel, 1999; Vanderborght et al., 2007})\), which simulates the biogeochemical transformations along the estuarine gradient \((\text{grids 2 and 3, Fig. 1})\).

2.3.1. The MIRO model

MIRO is a mechanistic biogeochemical model describing carbon, nitrogen, phosphorous and silica cycling through aggregated components of the planktonic and benthic realms of \(\text{Phaeocystis}\) dominated ecosystem \((\text{Lancelot et al., 2005})\). Its structure includes thirty-eight state variables assembled in four modules describing the dynamics of phytoplankton \((\text{diatoms, nanoflagellates and \text{Phaeocystis} colonies})\) in a nutrient \((\text{DIN, DIP, Si(OH)}_{4})\)- and light-limited environment, zooplankton \((\text{copepods and microzooplankton})\), dissolved and particulate organic matter \((\text{each with two classes of biodegradability})\) degradation and nutrients \((\text{NO}_3, \text{NH}_4, \text{PO}_4\) and \(\text{Si(OH)}_{4})\) regeneration by bacteria in the water column and the sediment. Equations and parameters were formulated based on current knowledge on the kinetics and the factors controlling the main auto- and heterotrophic processes involved in the functioning of the coastal marine ecosystem.

![Fig. 1. Bathymetric map of the Scheldt (Belgium/The Netherlands) estuary (grid 1 and grid 2) and the adjacent Dutch/Belgian coastal zone (grid 3). The limits of the three nested grids are indicated by black boxes.](image-url)
These are fully documented in Lancelot et al. (2005) and www.int-res.com/journals/suppl/appendix_lancelot.pdf. Compared to the original MIRO model, the two-dimensional implementation described in this work does not account for diagenetic recycling processes, since at the intra-annual timescale the benthic regeneration of nutrients is generally negligible compared to the pelagic recycling processes and the estuarine nutrient fluxes in the nearshore coastal zone (Lancelot et al., 2005). In addition, the description of photosynthesis was modified to account for the coupling between primary production, light availability and SPM (see Section 2.3.3). Furthermore, the MIRO model was extended by an explicit description of oxygen to allow for a realistic description of O₂ dynamics in the estuary. Oxygen is affected by the following source/sink terms:

\[ \sum R_{O_2} = R_{\text{air,O}_2} + R_{\text{NPP,O}_2} - 2R_{\text{min,O}_2} - R_{\text{aer,O}_2} \] (6)

In Eq. (6), \( R_{\text{air,O}_2} \) denotes the rate of \( O_2 \)-transfer through the air/water interface, \( R_{\text{NPP,O}_2} \) is the oxygen production by net primary production, while \( R_{\text{min,O}_2} \) and \( R_{\text{aer,O}_2} \) are the rates of pelagic oxygen consumption through nitrification and aerobic respiration, respectively.

2.3.2. The CONTRASTE model

The description of biogeochemical processes in the estuary with the exception of primary production is based on the latest version of the 1D-CONTRASTE model (Regnier et al., 1997; Regnier and Steefel, 1999; Vanderborght et al., 2002, 2007; Arndt et al., 2007, 2009) and includes organic matter, nutrients (ammonium, nitrate, phosphate and silica) and oxygen as state variables. Processes included are heterotrophic respiration, nitrification, denitrification and gas transfer at the water/air interface. Further details about their respective formulation can be found in Regnier and Steefel (1999) and Vanderborght et al. (2002, 2007). The CONTRASTE model is extended by a simple description of phosphate dynamics, where phosphate is taken up by phytoplankton and released during aerobic respiration and denitrification. The description of estuarine phytoplankton dynamics is adapted from the MIRO approach. Based on field observations (Koeman et al., 1992; Kromkamp and Peepe, 1995; Muylaert and Sabbe, 1999; Muylaert et al., 2009), diatoms are considered as an explicit state variable. Phaeocystis and nanoflagellates are assumed to lyse and remineralize within the estuarine domain. Studies of phytoplankton speciation along the Scheldt estuarine gradient show a complex distribution from typical fresh-water/estuarine species (e.g. Cyclotella spp.) in the freshwater tidal reaches to marine species (e.g. Chaetoceros spp., Rhizosolenia spp., Guinardia delicatula) in the vicinity of the estuarine mouth (Rijstenbil et al., 1993; Kromkamp and Peepe, 1995; Muylaert and Sabbe, 1999; Muylaert et al., 2000, 2009). Distinct riverine and coastal communities are imported into the estuary through river discharge and tidal mixing, respectively (Muylaert et al., 2009). To account for this process, a salinity threshold that allows distinguishing between marine and estuarine diatom communities was applied. It has been shown that coastal diatoms are generally characterized by a growth optimum at salinities between 25 and 33 (Brandt, 1984). However, some marine species tolerate lower salinities, but their growth rate generally decreases with increasing salinity (Brandt, 1984; Roubeix et al., 2008). Muylaert et al. (2009) showed that marine species mainly contribute to the species diversity in the polyhaline areas of the Scheldt estuary, while a complex mixture of estuarine, freshwater and coastal diatom species dominates at lower salinities. In addition, observations reveal that the spring phytoplankton development never intrudes further than 50 km (salinity ca. 2–15) into the estuary and spring maximum chlorophyll a concentrations are generally measured around km 40 (salinity ca. 15–25) (Van Damme et al., 2005). Based on these observations, the threshold value was chosen to allow for the intrusion of coastal diatom species down to salinities of 15. Parameter values for estuarine diatom growth (Table 1) where chosen according to Billen and Garnier (1997) and adjusted in order to better describe existing datasets on estuarine spring diatom biomasses (Van Damme et al., 2005).

2.3.3. Coupling SPM dynamics, light availability and primary production

The description of phytoplankton growth in the estuary–coastal zone continuum was modified to incorporate an explicit coupling with the suspended matter (SPM) dynamics. The depth-integrated photosynthetic activity \( \phi_i \) (mg Cm⁻³ d⁻¹) of phytoplankton species \( i \) is calculated according to:

\[
\phi_i = \frac{\kappa_i}{K_{\phi}} \int_0^{\text{max}} 1 - \exp \left( \frac{\alpha_i}{K_{\phi}} \cdot exp(K_d \cdot z) \right) dz
\]

where \( \kappa_i (d^{-1}) \) is the maximum rate of photosynthesis; \( \alpha_i (\text{mg C m}^{-3}) \) is the fraction of functional and structural metabolites in the biomass of species \( i \); \( \kappa_0 (\text{mg C m}^{-3}) \) is the incident photosynthetically active radiation; \( K_d (\text{m}^{-1}) \) is

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Estuary</th>
<th>Value coastal zone</th>
<th>Unit</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \phi )</td>
<td>Porosity</td>
<td>0.7</td>
<td>0.7</td>
<td>–</td>
<td>Estimated</td>
</tr>
<tr>
<td>( \rho )</td>
<td>Density</td>
<td>2650</td>
<td>2650</td>
<td>kg m⁻³</td>
<td>–</td>
</tr>
<tr>
<td>( M )</td>
<td>Erosion constant</td>
<td>5.5 × 10⁻⁴</td>
<td>5.5 × 10⁻⁴</td>
<td>m²</td>
<td>–</td>
</tr>
<tr>
<td>( \tau_{\text{coul}} )</td>
<td>Critical shear stress</td>
<td>0.2</td>
<td>0.3–0.5</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>( \rho_{\text{SPM}} )</td>
<td>Sediment density</td>
<td>1760</td>
<td>1760</td>
<td>kg m⁻³</td>
<td>–</td>
</tr>
<tr>
<td>( \eta )</td>
<td>Water viscosity</td>
<td>0.0013</td>
<td>0.0013</td>
<td>N s m⁻²</td>
<td>–</td>
</tr>
<tr>
<td>( \gamma )</td>
<td>Gravitational acceleration</td>
<td>9.82</td>
<td>9.82</td>
<td>m s⁻²</td>
<td>–</td>
</tr>
<tr>
<td>( \rho_{\text{w}} )</td>
<td>Water density</td>
<td>1000</td>
<td>1000</td>
<td>kg m⁻³</td>
<td>–</td>
</tr>
<tr>
<td>( k_i^{\text{DA}} )</td>
<td>Half-saturation constant for DIN uptake</td>
<td>5</td>
<td>0.8</td>
<td>mmol m⁻³</td>
<td>2.3</td>
</tr>
<tr>
<td>( k_i^{\text{Si}} )</td>
<td>Half-saturation constant for Si(OH)₄ uptake</td>
<td>7</td>
<td>0.4</td>
<td>mmol m⁻³</td>
<td>2.3</td>
</tr>
<tr>
<td>( k_i^{\text{PO}} )</td>
<td>Half-saturation constant for PO₄ uptake</td>
<td>0.5</td>
<td>0.3</td>
<td>mmol m⁻³</td>
<td>2.3</td>
</tr>
<tr>
<td>( k_0 )</td>
<td>Apparent background attenuation coefficient</td>
<td>1.4</td>
<td>0.188</td>
<td>m⁻¹</td>
<td>4.5</td>
</tr>
<tr>
<td>( k_{\text{SPM}} )</td>
<td>Specific attenuation of SPM</td>
<td>0.0992</td>
<td>0.024</td>
<td>mgl⁻¹ m⁻¹</td>
<td>4.5</td>
</tr>
<tr>
<td>( T_{\text{DA}} )</td>
<td>Optimal growth temperature</td>
<td>Days 1 to 160</td>
<td>25</td>
<td>5.5</td>
<td>°C</td>
</tr>
<tr>
<td>Days 160 to 365</td>
<td>25</td>
<td>15</td>
<td>°C</td>
<td>Estimated from 4.3</td>
<td></td>
</tr>
<tr>
<td>( d_{\text{DA}} )</td>
<td>DA temperature interval</td>
<td>Days 1 to 160</td>
<td>5</td>
<td>5</td>
<td>°C</td>
</tr>
<tr>
<td>Days 160 to 365</td>
<td>5</td>
<td>12</td>
<td>°C</td>
<td>Estimated from 4.3</td>
<td></td>
</tr>
</tbody>
</table>

1: (Arndt et al., 2007); 2: (Billen and Garnier, 1997); 3: (Lancelot, 1995); 4: (Desmit et al., 2005); 5: (Rousseau, 2000).
the light extinction coefficient and $\alpha_0$ (µE m$^2$ s$^{-1}$) is the photosynthetic efficiency. Eq. (7) is integrated numerically using a cost-efficient analytical approximation (Vanderborght et al., 2007). The extinction coefficient $K_d$ in Eq. (7) is an explicit function of the SPM concentration:

$$K_d = k_0 + k_{SPM} \cdot SPM$$

(8)

where $k_0$ is the apparent background attenuation and $k_{SPM}$ denotes the SPM specific attenuation constants, respectively. In the coupled model, SPM concentrations are calculated by a fully-formulated SPM transport model (Arndt et al., 2007). The erosion and deposition fluxes are expressed according to the formulations of Ariathurai (1974), Partheniades (1962) and Einstein and Krone (1962):

$$\frac{\partial h_{SPM}}{\partial t} + \frac{\partial}{\partial x} \left( v_x h_{SPM} \right) + \frac{\partial}{\partial y} \left( v_y h_{SPM} \right) = \frac{\partial}{\partial x} \left( h \cdot D_{spx} \frac{\partial}{\partial x} h_{SPM} \right) + \left( R_{ero} - R_{dep} \right)$$

(9)

where

$$R_{ero} = p_{ero} \cdot M \cdot SPM_b$$

(10)

$$R_{dep} = p_{dep} \cdot ws \cdot SPM$$

(11)

In Eqs. (10) and (11), $p_{ero}$ and $p_{dep}$ stand for the probabilities of erosion and deposition, respectively. $M$ denotes the erosion constant (m/s), $ws$ represents the settling velocity of particles (m/s) and $SPM_b$ refers to the concentration of total solids in the sediment. The settling velocity $ws$ depends directly on the diameter of suspended particles. The dynamical adaptation of the SPM diameter to processes such as deposition and erosion is described using the effective variable approach proposed by Wirtz and Eckhardt (1996) and Wirtz (1997). The wide range of different particle diameters is represented by a continuous gaussian grain size distribution, which is completely characterized by the dynamically adapting average grain diameter and its standard deviation (Arndt et al., 2007). Parameter values of the SPM transport model are given in Table 1.

2.4. Boundary conditions and forcings

Elevations and current directions are extracted from a coarser three-dimensional model (3D-MIRO&CO) for the year 1995 (Lacroix et al., 2007b, a) and imposed at the two open North Sea boundaries. In addition, the three-dimensional model provides the depth-averaged concentrations of biogeochemical state variables, temperature and salinity along the open sea boundaries. At the upstream boundary, the water flux through the estuarine cross-section, calculated by the one-dimensional estuarine model CONTRASTE (Regnier et al., 1998), is specified. Weekly measurements from the Hemiksen station (Regnier and Steefel, 1999) provide the boundary concentrations and determine the typical resolution of freshwater scalar inputs to the estuarine–coastal zone model. At the boundaries SPM concentrations are set to a constant background value of 50 mg l$^{-1}$ and 100 mg l$^{-1}$ at the open sea and estuarine limits, respectively. A two dimensional wind-field is specified by interpolated maps, which are extracted from the coarser three-dimensional model (Lacroix et al., 2007b, a). The incident light intensity $I_0$ (Royal Meteorological Institute of Belgium) is given as time-dependent external forcing, which resolves the daily variations in light regime.

2.5. Validation data

Time series of surface elevation (data source: Hydro Meteo Centrum Zeeland (HMCZ), www.hmcz.nl) at six stations along the estuary–coastal zone continuum and observed phase lags of the tidal wave in the Scheldt estuary (Horrevoets et al., 2004) are used to validate the hydrodynamics in the Scheldt estuary and the adjacent coastal zone. Transport is validated on the basis of salinity time series from three monitoring stations along the estuarine–coastal zone gradient (data source: Hydro Meteo Centrum Zeeland (HMCZ), www.hmcz.nl). Simulated SPM concentrations are compared to field observations at six different coastal and estuarine stations (data source: Ministerie van Verkeer en Waterstaat, www.waterbase.nl). In addition, time series from five estuarine and coastal zone monitoring stations (data source: Ministerie van Verkeer en Waterstaat, www.waterbase.nl) and reference station 330 (N51°26.05' E2°48.50', Breton et al. (2006)) are used for the validation of the seasonal dynamics of nutrients (NH$_4$, NO$_3$, Si(OH)$_4$, PO$_4$), phytoplankton mphChl a and O$_2$ concentrations.

3. Results and discussion

3.1. Hydrodynamics and transport

The hydrodynamics in the Scheldt estuary and the adjacent coastal zone is determined by strong tidal forcing. Over a spring/neap cycle, the tidal wave reaches an amplitude of 4.3/2.8 m in the coastal zone close to

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**Fig. 2.** Comparison of observed (gray, data from Hydro Meteo Centrum Zeeland (HMCZ)) and simulated (black) surface elevations at different locations (see Fig. 1) along the estuary–coastal zone continuum in March 1995. Surface elevations are given in the Belgian ordnance datum (TAW), which refers to mean low water level about 2.3 m below mean sea level.
Zeebrugge and 4.6/3.0 m near Oostende. In the estuary, the tidal range gradually increases due to convergence to a maximum amplitude of 5.8/4.6 m, until friction-induced dampening exceeds its amplification upstream of Antwerp. Fig. 2 compares measured and simulated surface elevations at different stations along the estuarine–coastal zone continuum. The time series reflect the dominant tidal forcing, on which local wind speed variations are superimposed. The average deviation between observations and simulations at all stations is ca. 6 cm and the simulated elevations are thus in good agreement with observations. The tidal wave propagates relatively fast (ca. 15 ms⁻¹) through the coastal zone and the lower estuary, where waters are still relatively deep (average depth 10 m). Close to the Dutch/Belgian border, the wave is partially reflected due to the sudden reduction in estuarine width. The superposition of the reflected and the incoming wave results in almost infinite propagation speeds. Further upstream, the propagation speed decreases (10 ms⁻¹) as a consequence of the decreasing average water depths (h = 4.5 m). The good agreement between simulated and observed propagation of the tidal wave in the estuary indicates that the model captures the main hydrodynamic features in this system Fig. 3.

The large tidal currents lead to a relatively well-mixed water column in the estuary and the adjacent coastal zone throughout the year. However, the freshwater discharge of the Scheldt produces a pronounced horizontal salinity gradient along the estuarine curvilinear axis and in a significant portion of the coastal zone. The salinity distribution depends on the freshwater discharge of the Scheldt, the tidal excursion and the prevailing wind speed and direction. Salinity increases gradually from the upstream boundary to the mouth. The less saline estuarine water then flows to the south–west along the coastline, forming the Scheldt plume (Fig. 4-a). In the estuary, the daily mean salinity variations reflect the dominant influence of freshwater discharge (Fig. 4-d). At the beginning of February, extremely high river discharges (>600 m³ s⁻¹) push less saline water far downstream and lead to a strong decrease in salinity at the estuarine stations (Fig. 4-c and -d). At the Baalhoek station, located roughly 50 km upstream from the estuarine mouth, salinity normally varies between 15 and 20. Yet, the winter freshwater pulse results in unusually low salinities, which drop down to 3 (Fig. 4-d). The offshore station shows much less variability (Fig. 4-b). Here, the salinity is mainly controlled by local wind conditions, which determine the extension of the Scheldt plume and the influx of less saline water through the northern boundary. Fig. 4-a illustrates this dominant influence of the local wind field (direction and speed) on the spreading of the estuarine plume. At the beginning of February, very strong south-westerly winds limit the extension of the plume, despite the extremely high freshwater discharge (>600 m³ s⁻¹) and are responsible for the very sharp longitudinal salinity gradient in the estuary (Fig. 4-a). On the other hand, well established northerly winds early in April push the Scheldt plume further south and thus lead to lower salinities in the coastal zone (Fig. 4-a, also shown by Lacroix et al. (2004)). In general, long-term residual transport processes determine the spatial structure of the estuarine plume. The Eulerian residual transport velocities (Fig. 5-a) reveal a general northwards circulation, resulting mainly from the large-scale sea surface slope across the shelf, and to a lesser extent, meteorological forcing (Delhez and Carabin, 2001). However, the Eulerian residual transport field does not explain the clockwise rotation of the Scheldt plume. Interactions between tidal currents and residual eddies, arising from tide-topography interactions, result in a strong stokes drift and, thus, in a pronounced difference between the Eulerian and the Lagrangian residual transports. Therefore, it has long been known that the Lagrangian residual transport provides a better representation of the coastal mean circulation in environments characterized by a complex, irregular bathymetry. However, the computation of the Lagrangian residual transport field is difficult due to its unsteady character. It is usually determined by releasing particles in the time-varying velocity field. Yet, the net-displacement of released particles strongly depends on the hydrodynamic history and thus on their release time. Fig. 5 shows the general structure of the Lagrangian residual transport field, according to a first-order approximation of the surface Lagrangian residual transport velocity calculated by Delhez and Carabin (2001) and the results of tracer experiments (Van den Eynde, 2004). In addition, it illustrates the variability of the transport field due to the effect of different wind directions by analyzing the tracks of seven particles, released at different locations in the estuary and the coastal zone at the end of January and in April 1995. The simulated tracks agree well with the first-order approximation of the surface Lagrangian residual transport velocity calculated by Delhez and Carabin (2001) and the results of tracer experiments (Van den Eynde, 2004). In the estuary (tracks 1 and 2; Fig. 5-c), most of the residual transport is limited to the main tidal channels. In addition, the oscillatory tidal forcing results in small residual velocities and, therefore, long residence times. Close to the mouth (track 3; Fig. 5-c), the residual circulation follows the deep navigational channel along the coast until the bathymetry induces an offshore drift close to Zeebrugge (Fig. 5-c). Further offshore (tracks 4 and 5; Fig. 5-c), the difference between Lagrangian and Eulerian transport is reduced by greater water depths and a more regular bathymetry. The Lagrangian circulation thus resembles the northeastern Eulerian circulation. A residual gyre forms offshore of the mouth, and increases the residence time in this area. In the southern part of the model area (track 6; Fig. 5-c), the residual transport is directed along the coastline in a narrow band close to the coast. Further offshore (track 7; Fig. 5-c), the complex bathymetry directs the particle northwards. Local wind conditions modify this general pattern and, therefore, result in differences between the simulated Langragian particle tracks recorded at the end of January and April 1995 (Fig. 5-c). At the end of January, predominantly southwesterly winds lead to a stronger alongshore residual transport in the southern part of the model area (track 6; Fig. 5-c), but limit the offshore transport at the mouth (track 3; Fig. 5-c). In April (Fig. 5-c), when northwesterly winds prevail, the particle released at the estuarine mouth is transported further offshore into the residual gyre (track 3; Fig. 5-c), while the residual transport in the southern area of the coastal zone is weaker (track 6; Fig. 5-c). Offshore, particles are less influenced by the wind conditions and are generally transported in northward direction (Fig. 5-c). However, well established southerly winds reinforce the residual transport, while weaker or northerly winds trigger a more complex transport pattern (tracks 4, 5, and 7; Fig. 5-c).

**Fig. 3.** Comparison of observed (dots, Horrevoets et al., 2004) and simulated (line) phase lags of the tidal wave along the estuarine gradient.
3.2. Suspended matter

Fig. 6 illustrates the spatio-temporal dynamics of SPM in the estuary and the adjacent coastal zone. The SPM distribution in the estuarine coastal-zone continuum is controlled by the interplay between hydrodynamic and meteorological conditions. In the estuary, strong tidal currents and shallow water depths result in relatively high turbidities (30–200 mg l\(^{-1}\)). A turbidity maximum forms in the vicinity of Antwerp due to a maximum in total energy dissipation (Chen, 2003; Chen et al., 2005; Arndt et al., 2007). In the adjacent coastal zone, turbidity is generally lower and falls within the range of observed SPM concentrations (Fig. 6-a to -d; Fettweis and Van den Eynde (2003), Fettweis et al. (2006), Ruddick et al. (2003). SPM concentrations show a characteristic distribution (Fig. 6-g) with higher concentrations (50–200 mg l\(^{-1}\)) in the shallow nearshore areas and decreasing concentrations (10–30 mg l\(^{-1}\)) in offshore direction (Fettweis and Van den Eynde, 2003; Ruddick et al., 2003). The decrease in residual transport between Oostende and Zeebrugge, as well as shallow water depths in this area and the difference between spring and neap tidal currents, lead to the formation of a coastal turbidity maximum between Oostende and the mouth of the Scheldt estuary (Nihoul, 1975; Gullentops et al., 1976; Fettweis and Van den Eynde, 2003; Fettweis et al., 2006; Van den Eynde, 2004).

SPM concentrations show a significant temporal variability (Fig. 6-a to -e). Previously, a spectral analysis of estuarine SPM time series revealed that fluctuations are mainly aggregated at tidal frequencies (Arndt et al., 2007). In the coastal zone, the temporal variability of SPM concentrations is not only controlled by the amplitude of the tide, but also by wind conditions. In general, maximum daily mean concentrations are significantly higher during spring tides (Fig. 6-a to -e, Fettweis et al. (2006)). However, the seasonal trend in the SPM record is mainly induced by long-term variations in wind speed. In autumn, winter and spring, higher wind speeds lead to increased maximum SPM concentrations, while summer concentrations are generally lower (Fig. 6; Fettweis et al., 2006).

3.3. Spatio-temporal trends in nutrient, \(O_2\) and Chl \(a\) concentration

Fig. 7 compares the observed and simulated spatio-temporal variability in nutrient (NH\(_4\), NO\(_3\), Si(OH)\(_4\), PO\(_4\)), phytoplankton Chl \(a\) and \(O_2\) concentrations at selected locations along the estuarine–coastal zone continuum over the year 1995. In general, nutrient and \(O_2\) concentrations show a clear seasonal trend characterized by high values in winter, a progressive decrease during the productive spring and summer period, and a replenishment of nutrient levels in late summer and autumn (de Galan et al., 2004; Van der Zee and Chou, 2005; Lancelot et al., 2005). In spring, nutrient depletion is significant for NH\(_4\), Si(OH)\(_4\) and PO\(_4\), while NO\(_3\) shows a similar but damped evolution (Fig. 7-a to -d). Chl \(a\) time series reflect the development of characteristic phytoplankton spring and summer blooms along the estuary–coastal zone continuum that coincide with the periods of low nutrient levels (Fig. 7-a to -d). Yet, the magnitude and the timing of Chl \(a\) maxima differ along the estuary–coastal zone continuum. The upper estuarine reaches (Fig. 7-a) are characterized by a relatively broad early summer Chl \(a\) maximum whereas a sharp spring maximum, followed by a smaller summer phytoplankton development are observed and simulated in the lower estuary (Fig. 7-b to -d) and the adjacent coastal zone (Fig. 7-e and -f). These general seasonal trends in nutrient and Chl \(a\) concentrations are superimposed by the characteristic oscillatory tidal variability in the estuary, while stochastic fluctuations induced by the wind forcing become more important close the estuarine mouth and in the coastal zone. A pronounced decrease in nutrient concentrations can also be observed along the estuary–coastal zone continuum due to dilution and consumption along the salinity gradient. At the mouth, nutrient-enriched estuarine waters generally turn southwards and form the Scheldt plume, following the Lagrangian residual transport along the coastline (Fig. 5). It subsequently mixes with nutrient-poor Atlantic waters offshore, resulting in a pronounced onshore/offshore and north–east/south–west gradient (Van der Zee and Chou, 2005; de Galan et al., 2004). However, during nutrient depletion in the vicinity of the estuarine mouth, a reversed gradient in Si(OH)\(_4\) and NH\(_4\)
concentrations can be observed, with slightly higher concentrations offshore. \( \text{O}_2 \) concentrations are generally low at the upstream limit of the model domain (Fig. 7-a), where nitrification and respiration rates exceed the uptake capacity of oxygen from the atmosphere. In downstream direction, an intensification of mixing processes progressively improves water column oxygenation and ultimately results in oxygen concentrations close to saturation (Fig. 7-b and -f) or even slightly oversaturated in the coastal zone and estuarine mouth in early spring (Fig. 7-d to -f). Later in the year, concentrations drop to lower values due to enhanced respiration rates induced by the increased availability of fresh organic matter from the declining phytoplankton blooms (Fig. 7-c and -f).

The model performance is quantitatively evaluated on the basis of different statistical measures (e.g. Allen et al., 2007). Correlation coefficients, model efficiencies, percentage model biases and cost functions are calculated as defined in Allen et al. (2007) on the basis of tidally-averaged simulation results and the corresponding field data for the entire estuary–coastal zone transition (Fig. 8). Correlation coefficients and model efficiencies are high for macro-nutrients and somewhat lower for phytoplankton and oxygen (Fig. 8-a and -b). The percentage model bias shows that the model has the tendency to slightly overestimate ammonium and underestimate phytoplankton (Fig. 8-c). The cost function quantifies the difference between model results and measurement data and reveals a very good fit for all analyzed variables. The summary of basic model data fit metrics (e.g. Allen et al., 2007) for the spatio-temporal dynamics of macronutrients \((\text{NH}_4, \text{NO}_3, \text{Si(OH)}_4, \text{and PO}_4)\), phytoplankton and oxygen (Fig. 8) reveals that the model has skill for all variables analyzed. It indicates that model performance slightly deteriorates from nitrate, silicate to ammonium, phosphate and finally to phytoplankton and oxygen. However, all evaluated variables reach good to excellent results and the model thus performs well in capturing the spatio-temporal dynamics along the estuary–coastal zone.

### 3.4. Seasonal evolution of estuarine and coastal primary producers

Fig. 9 illustrates the evolution of the simulated, volume-integrated phytoplankton biomass in the coastal zone (model grid 1) and the estuary (model grids 2 and 3). In both the estuary and the coastal zone, the evolution of the phytoplankton biomass is characterized by a spring and a summer bloom (Fig. 9-a and -b). Simulation results reveal that in 1995, the annual \textit{Phaeocystis} and diatom biomass
represent 58% and 38% of the total annual, coastal phytoplankton biomass, while nanophytoflagellates only contribute 4%. The estuarine phytoplankton community is almost entirely dominated by diatoms.

In the coastal zone, the onset of the spring phytoplankton bloom occurs in March with the development of a diatom bloom (Fig. 9-b). Shortly after, both a small nanophytoflagellate and a pronounced Phaeocystis colony bloom start to develop (Fig. 9-b). Coastal diatom biomass peaks in mid-April before it rapidly decreases to low values (Fig. 9-b). The simulated Phaeocystis colony biomass reaches several maxima in late March and at the end of April and slowly declines to lower values in May. A summer/fall diatom bloom follows the decline of the Phaeocystis colonies (Fig. 9-b). The simulated succession of coastal phytoplankton community is clearly dominated by a bloom of Phaeocystis colonies (Lancelot et al., 1987, 1998; Cadee and Hegeman, 1991; Lancelot, 1995; Rousseau et al., 2000). Diatoms are usually present throughout the productive period and dominate the summer phytoplankton bloom (Lancelot et al., 1998; Rousseau et al., 2000).

In the estuary, maximum spring diatom biomass coincides with the spring bloom in the coastal zone (Fig. 9-a). Observed and simulated Chl a time series reveal that diatoms mainly develop in the lower estuary (Vlissingen-Hanswert, see Fig. 1), while, further upstream, environmental conditions do not meet the requirements for a spring diatom bloom (Fig. 7). Multiple year monitoring results show a similar, repetitive pattern with spring Chl a maxima of variable magnitude (20–100 μg Chl a l⁻¹) in the lower Scheldt estuary (<km 60) (Van Damme et al., 2005). The simulated and observed spatio-temporal pattern thus indicates that the increase in estuarine spring biomass is initiated by the import of euryhaline coastal diatoms, which encounter suitable growth conditions in the relatively nutrient-rich saline estuary. Observations indicate that typical coastal phytoplankton species contribute up to 20% to the total diversity of phytoplankton.
Fig. 7. Annual evolution of measured (points) and simulated (gray line) nutrient, phytoplankton and oxygen concentrations at different stations in the model area (see Fig. 1). Source: Ministerie van Verkeer en Waterstaat, www.waterbase.nl except for station 330 Rousseau (2000), Lancelot et al. (2005), Breton et al. (2006).

Fig. 8. Model performance summary statistics for the entire estuary–coastal zone transition (Fig. 7) data set, a—correlation coefficient, b—model efficiency, c—model bias, and d—cost function.
species in the polyhaline zone (salinity 30–18) of the Scheldt estuary Muylaert et al. (2009). Similarly, Phaeocystis colonies are imported into the estuary by tidal mixing at the time of their spring maxima. The simulated import leads to a small accumulation of Phaeocystis biomass in the vicinity of the estuarine mouth (Kromkamp and Peepoe, 1995). In summer, diatom biomass reveals a second increase (Fig. 9-a). This increase coincides with the peak stage of the periodically observed freshwater diatom bloom in the upper reaches of the Scheldt estuary (Muylaert et al., 2000; Van Damme et al., 2005; Muylaert et al., 2005). High summer Chl a concentrations at the upstream boundary, as well as a rapid downstream concentration decrease (Fig. 7) indicate that the simulated estuarine summer maximum is mainly triggered by the import of freshwater diatoms (Arndt et al., 2007). In what follows, these general trends in primary production dynamics are discussed in the light of the constraints set by the seasonal evolution of nutrient availability along the estuary–coastal zone continuum.

3.5. Nutrient cycling along the estuary–coastal zone continuum

Fig. 10A–C (a–g) summarizes the DIN, Si(OH)₄, and PO₄ dynamics along the estuary–coastal zone continuum, respectively. The estuarine domain comprises the area between the estuarine mouth in the vicinity of Vlissingen and the upstream boundary (grids 2 and 3, Fig. 1), while the coastal domain covers the entire coastal area between the estuarine mouth and the open sea boundaries (grid 1, Fig. 1). Fig. 10A–C, illustrate the seasonal evolution of the low-pass filtered (cut-off frequency f_c = 16d) nutrient flux through the upstream boundary at Hemiksen (a), the low-pass filtered flux across the estuarine mouth in the vicinity of Vlissingen (d) as well as at the low-pass filtered flux through the open sea boundaries (g). In addition, the spatially integrated, low-pass filtered net nutrient transformation and the respective contribution of the various individual processes (c and f) are shown together with the instantaneous and low-pass filtered mass variations in each subsystem (b and e). The following paragraph discusses the evolution of nutrients fluxes and transformations from the estuary (Fig. 10A–C (a–d)) to the coastal zone (Fig. 10A–C (d–g)).

In the Scheldt estuary, the freshwater reaches are characterized by short residence times and dominated by advective transport. At the upstream boundary, nutrient fluxes thus correlate well with the riverine inflow and well-defined winter maxima coincide with peak river discharges (Fig. 10A–C (a)). However, this correlation decreases in summer, when enhanced process rates consume a larger fraction of the nutrient inventory in the freshwater reaches, upstream of the model domain (Arndt et al., 2009). Sharp summer minima in Si(OH)₄ and PO₄ inflow can be observed (Fig. 10B-a and C-a) and result from the nutrient consumption upstream of the model domain during the freshwater diatom bloom (Arndt et al., 2009). DIN consumption also occurs and is largely dominated by denitrification, which maintains the DIN inflow at low values between late spring and early autumn (Fig. 10A-a). In the estuary, DIN transformation is controlled by aerobic respiration and denitrification which are of similar absolute magnitude in terms of carbon but result in a net consumption of DIN. This net consumption is amplified by the DIN uptake during the spring, and to a lesser extent, during the summer diatom bloom (Figs. 10A-c and 9-a). Si(OH)₄ consumption correlates with the primary production signal and thus reveals a pronounced peak in spring followed by smaller peaks in summer (Fig. 10B-c). Silica dissolution partly compensates the uptake during the summer bloom and results in slightly reduced net-consumption rates during this period. PO₄ transformation is dominated by net production, due to organic matter degradation in the estuarine domain (Fig. 10C-c). Yet, a net consumption is simulated during the spring diatom bloom in the saline estuary when diatom PO₄ uptake largely exceeds PO₄ regeneration. The nutrient riverine inflow is thus significantly modified by biogeochemical processes during the estuarine transit and low-pass filtered nutrient fluxes across the estuarine mouth differ markedly from the fluxes at the upstream boundary (Fig. 10A–C–d). This variability is however not only the result of the evolution in the magnitude of biogeochemical processes, but also results from flux imbalances, which arise from the time-lagged response of the nutrient scalar fields to hydrological perturbations (Regnier and Steefel, 1999; Arndt et al., 2009). The magnitude of this imbalance strongly depends on the specific shape of the respective nutrient scalar fields. The increase in residual transport driven by the pronounced winter discharges results in a storage of Si(OH)₄ (Fig. 10B-b). During winter and spring, Si(OH)₄ fluxes (Fig. 10B-d) therefore show a damped and smeared response to input variations (Fig. 10B-a). The time-lagged release of the stored Si(OH)₄ initially supports the spring diatom bloom in the coastal zone. However, the enhanced nutrient consumption, due to the coastal diatom bloom intruding into the estuary in late spring leads to a pronounced Si(OH)₄ and PO₄ mass decrease in the estuarine domain (Fig. 10B-b and C-b). During the terminal stage of the estuarine diatom bloom, the lower estuary becomes completely depleted in Si(OH)₄ and PO₄ (Fig. 7) and a small influx of these elements occurs across the estuarine mouth (Fig. 10B-d and C-d). In early summer, estuarine Si(OH)₄ and PO₄ pools are partly replenished by the influx through the upper boundary in combination with net PO₄ in-situ production (Fig. 10B and C). Therefore, a net Si(OH)₄ and PO₄ export to the coastal zone is again simulated after the decline of the estuarine bloom. Yet, in the case of Si(OH)₄ the estuarine summer phytoplankton bloom develops at the time when the riverine influx is low, resulting in another period of Si(OH)₄ import from the adjacent coastal zone in late summer (Fig. 10B-d and C-d). DIN fluxes are less affected by the spring bloom, which results in a relatively large DIN outflux to the coastal zone (Fig. 10A-d). Later in the year, the DIN summer consumption significantly depletes the DIN inventory in the estuarine domain and results in a decreasing net flux, which reaches a minimum in late summer (Fig. 10A-b to d).

The adjacent coastal zone is characterized by comparatively long residence times that allow intense biogeochemical transformation processes (Fig. 10-e to g). The nutrient uptake during the early stage of the spring phytoplankton bloom leads to a complete depletion of the coastal Si(OH)₄ and PO₄ levels and reduces the DIN inventory (Fig. 10A–C–e). This reduction triggers a net nutrient influx through
the open sea boundaries (Fig. 10A–C–g). However, the increased availability of labile organic matter and biogenic silica after the decline of the spring bloom fuels an intense nutrient recycling (Fig. 10A–C–f). As a result, the coastal Si(OH)₄ inventory slowly increases, resulting in a net outflux of this nutrient (Fig. 10B–e to -g). The intense consumption of Si(OH)₄ during the coastal diatom bloom in summer leads again to low levels (Fig. 10B–e to -g) and, thus, to an influx through the open boundaries. In the case of DIN, the uptake by Phaeocystis, combined with estuarine export fluxes, maintain low coastal stocks (Fig. 10A–e). A net DIN influx is thus simulated during the summer month, until the flux variability increases in autumn due to the intensification of the meteorological forcing (Fig. 10A–g). PO₄ transformations show a similar pattern (Fig. 10C–f). Yet, unlike DIN, persistent PO₄ influxes through the open sea boundaries and the estuarine mouth result in an increase in PO₄ stocks after the early spring bloom (Fig. 10C–d, -e and -g).

The evolution of nutrient limitation factors provides important insights into the functioning of the phytoplankton succession in the coastal zone. Fig. 11 summarizes the seasonal evolution of nutrient limitations for the dominant estuarine and coastal phytoplankton groups. The timing of the respective phytoplankton concentration maxima is indicated by dashed lines. Nutrient limitation in the estuarine and coastal-zone sub-domains is assessed by calculating the species-specific limitation factors, formulated as \( N_{\text{limit}} = \frac{N_{\text{PO4}} \cdot \text{DIN} \cdot \text{Si(OH)}_{4} \cdot \text{f}}{N_{\text{ph}} + \frac{K_{n}}{K_{s}} + N_{k} + K_{s}} \) based on specific half-saturation constants (\( K_{n} \), Table 1) and volume-averaged concentrations. N. Low values indicate no or little nutrient limitation, while high values reflect a strong limitation of primary production. Thus, they represent an integrative measure of the control of nutrient availability on phytoplankton succession. At its early stages, the spring diatom bloom is mainly limited by PO₄ availability. Yet, the exponential phytoplankton growth almost completely depletes Si(OH)₄ (Fig. 10B–e), causing a shift towards a silica limitation around day 100 (Fig. 11–a). Similarly, Van der Zee and Chou (2005) report, based on nutrient measurements at different stations in the Belgian coastal zone, a PO₄ limitation of the spring diatom production in the southern and offshore parts of the coastal zone, co-limited by Si(OH)₄ close to the estuarine mouth. Previous modeling studies show similar trends even though the predicted PO₄ limitation was weaker (0.3 for 1995) for the spring phytoplankton bloom and simulated spring Si(OH)₄ concentrations were generally overestimated (Lancelot et al., 2007; Gypens et al., 2007). The intense coastal diatom bloom depletes the PO₄ inventory (Fig. 10C–e), which results in a strong PO₄ limitation of phytoplankton growth. The simulation predicts a PO₄ limitation (0.4) of the spring Phaeocystis colony bloom, which is in agreement with previous modeling results (Gypens et al., 2007). However, the transient simulation results reveal that this limitation is short-lived, due to the very fast recovery of PO₄ inventories by transport and in-situ regeneration (Fig. 11–b). Therefore, the production of Phaeocystis colonies, which are also able to use organic phosphorous for growth, is virtually unaffected by low PO₄ concentrations (Figs. 10C–e, 11–b). A moderate DIN limitation (around 0.3) is simulated at the late stage of the bloom (Fig. 11–b). Later in the year, the nutrient influxes through the boundaries and the in-situ recycling increase the nutrient availability in the coastal zone (Fig. 10A to C). The coastal summer diatom bloom is thus weakly limited by nutrient availability (Fig. 11–a), indicating that temperature, light conditions and grazing are more important controlling factors. In the estuary, the intruding coastal spring diatoms, which have comparably low nutrient requirements, encounter excellent growth conditions. Therefore, the estuarine spring bloom is characterized by a weak nutrient limitation factors close to 0.1 (Fig. 11–c). Salinity stress, combined with the decreasing availability of light in the turbid estuary put more severe environmental constraints on the intruding coastal diatoms and limit their upstream propagation (Fig. 7, Van Damme et al. (2005)). The freshwater diatoms, which are imported from the tidal river during the summer months, are adapted to nutrient-rich conditions and, therefore, are seriously limited by the decrease in estuarine nutrient levels along the salinity gradient (Fig. 11–d). The summer freshwater diatom bloom thus reveals a strong co-limitation of Si(OH)₄ and PO₄. DIN concentrations are less affected by dilution and consumption within the estuarine domain and, as a consequence, the limitation by this nutrient is weaker.

### 3.6. Patterns in ecosystem structure

The residual transport field and in-situ turnover rates control the local nutrient availabilities and the emergence of distinct spatial patterns in ecosystem structure throughout the productive period. Fig. 12 illustrates the spatio-temporal evolution of the dominant phytoplankton groups and bacteria along the estuary-coastal zone continuum. They show a general pattern in agreement with previous field observations (Nihoul and Hecq, 1984; Borges and Frankignoule, 1999, 2002; Muylaert et al., 2006). This pattern is characterized by high primary production rates in the estuarine plume, the development of zooplankton along the deflected plume and an intense recycling around the coastal residual gyre (Nihoul and Hecq, 1984).

Simulation results indicate that the coastal spring phytoplankton bloom is spatially constrained by the species-specific nutrient requirements. Even though the early spring diatom bloom is initiated in the southern part of the Belgian coastal zone, highest biomass rapidly clusters around the nutrient-rich Scheldt plume, where Si(OH)₄ is available. The bloom thus follows the movement of the estuarine plume along the coast line and maximum biomasses are simulated between Vlissingen and Zeebrugge (Fig. 12-a, stage 1). The exponential phytoplankton growth in late March progressively depletes coastal Si(OH)₄ and PO₄ levels which ultimately become limiting for diatoms (Fig. 11-a). At the same time, tidal pumping, caused by the asymmetry between flood and ebb currents, transports euryhaline diatoms upstream into the estuary, where they encounter good growth conditions. The enhanced estuarine diatom production significantly reduces the nutrient fluxes across the estuarine mouth (Fig. 10-d), exerting a negative feedback on the coastal diatoms and further limiting their production to the estuarine domain (Fig. 12-a, stage 2). The release of labile organic matter during the decline of the spring phytoplankton blooms stimulates bacterial growth and the recycling of DIN and PO₄ in the coastal zone (Fig. 10A and C). Maximum bacterial biomass drifts with the residual gyre (Fig. 12-c, stage 1). In addition, the residual transport field is characterized by a decrease in velocities between Oostende and Zeebrugge, which leads to an accumulation of organic matter close to Oostende (results not shown) and the formation of remineralization patches of DIN and, to a lesser extent, PO₄. Field observations are consistent with our simulations and show a spring minimum in O₂ saturation, which coincides with maxima in pheopigment concentrations, organic carbon and pCO₂ oversaturation in this area (Nihoul and Hecq, 1984; Borges and Frankignoule, 1999, 2002; Van der Zee and Chou, 2005). Low DIN concentrations moderately limit Phaeocystis colony growth in the coastal domain (Fig. 11-c). In early spring, the maximum in Phaeocystis colony biomass is situated close to the open boundaries (Fig. 12-b, stage 1), where they benefit from the inflow of nutrient-rich waters from the English channel and, occasionally, from the Rhine plume (Lacroix et al., 2004). Yet, the progressive nutrient depletion of...
coastal waters, especially with respect to Si(OH)₄ and PO₄ (Figs. 10 and 11) ultimately favors the onset of a Phaeocystis colony bloom along the coast (Fig. 12-b, stage 2). During this period, the estuarine nutrient fluxes to the coastal zone are strongly reduced by the diatom consumption in the lower estuarine reaches and DIN, as well as PO₄ are mainly supplied by import through the open boundaries, in combination with local DIN recycling (Fig. 10A and C). Therefore, maximum Phaeocystis biomass clusters along the outer edge of the estuarine bloom and oscillates between the area of enhanced nutrient recycling close to Zeebrugge and the north-eastern part of the coastal domain, which occasionally profits from the southward spreading of the Rhine plume (Fig. 12-b, stage 2). Observational evidence confirms the dominance of Phaeocystis colonies in the outer Scheldt plume, while diatoms and Phaeocystis equally contribute to the Chl a spring concentrations close to the estuarine mouth (Muyllaert et al., 2006). At the end of May, nutrient depletion leads to a decline of the spring diatom and Phaeocystis blooms (Fig. 12). Nevertheless, Phaeocystis colonies are still imported through the northern boundary (Fig. 12-b, stage 3). During summer, microbially-mediated heterotrophic processes drive a nutrient recycling loop around the coastal residual gyre (Figs. 10-f and 12-c, stage 2).

Concomitantly, a summer diatom bloom develops in the freshwater reaches of the Scheldt estuary (Muyllaert et al., 2000; Meire et al., 2005) and leads to a significant import of freshwater diatoms by the residual estuarine downstream transport (Fig. 12-d). Yet, these species are generally adapted to high nutrient and light conditions (Muylaert et al., 2005) and phytoplankton growth is thus limited to the area upstream of the Dutch/Belgian border (Fig. 12-d). The development of this summer diatom bloom further reduces the summer nutrient export fluxes that are already low due to the summer decrease in river discharge (Fig. 10-a and -d). Therefore, the coastal summer diatom bloom starts in the southern coastal zone, where low turbidities (Fig. 6) and intense nutrient recycling favour production (Fig. 12-d, stage 1). In late summer, the decline of the freshwater diatom bloom and the associated increase in nutrient export fluxes leads to a shift of maximum coastal diatom biomass towards the nutrient-rich estuarine plume (Fig. 12-d, stage 2).

Simulation results on nutrient dynamics and ecosystem structure reveal that the spatio-temporal variability in the distribution and transformation of carbon and nutrients is strongly controlled by the constraints set by the physical environment. The estuary is characterized by a highly-transient one-dimensional spatial structure, which is controlled by the residual downstream transport. The land-derived nutrient export to the coastal zone shows a strong seasonal variability, which results from the evolution in the magnitude of riverine inputs and estuarine biogeochemical transformations. In addition, large transient fluctuations in estuarine nutrient inventories arising from the time-lagged response of the scalar fields to upstream hydrological perturbations influence the timing of the nutrient delivery to the coastal zone. The coastal zone, on the other hand, exhibits a complex two-dimensional pattern. This pattern is mainly controlled by the Lagrangian residual transport field, whose complex structure determines the distribution of nutrient-rich estuarine waters and defines the residence times. High primary production rates are observed in the nutrient-rich estuarine plume and drive the development of bacteria and higher trophic levels along the plume. The results show also that the Lagrangian residual transport field is essentially unsteady and subject to continuous hydrodynamic adjustments, which lead to large variations in the spreading of the estuarine plume and, therefore, in the overall structure of the coastal ecosystem over short periods of time. In addition, the formation of a well-developed residual gyre increases the residence times in the nearshore coastal zone and favors an enhanced nutrient recycling in the vicinity of the gyre. The specific nutrient requirements of the various phytoplankton groups further constrain the detailed spatial distribution of nutrient fluxes to the coastal zone.
structure of the ecosystem. Maximum diatom concentrations develop in the immediate proximity of the estuarine mouth where nutrient concentrations are generally higher than elsewhere in the coastal zone. This bloom progressively depletes $\text{PO}_4$ and $\text{Si(OH)}_4$ at the interface between the estuary and the coastal zone. This local depletion strongly limits the development of *Phaeocystis* in the vicinity of the estuarine mouth. Therefore, they preferentially develop in areas where enhanced recycling rates or import fluxes through the open sea boundaries provide suitable growth conditions.

4. Conclusions

A mechanistic model accounting for the coupled biogeochemical transformation processes and fluxes of carbon and macro-nutrients

![Fig. 12. Spatio-temporal dynamics of daily maximum phytoplankton biomasses and bacteria during the productive period in spring and summer 1995. a—spring diatoms, b—*Phaeocystis*, c—bacteria, and d—summer diatoms.](image-url)
along the entire mixing zone of the shallow, tidally-dominated Scheldt estuary and adjacent coastal zone has been developed. The spatial and temporal resolution of the model allow to include the short-term transient forcings triggered by tides, freshwater inflows and wind stress, thus providing a detailed representation of the hydrodynamic and transport fields along a topographically-complex system. The results give important insights into the processes and environmental factors that control the spatial structure and evolution of autotrophic and heterotrophic processes affecting the C, N, P and Si cycles at the intra-annual timescale.

The land-derived nutrient export to the coastal zone is characterized by a strong seasonal variability, which results from the evolution of riverine inputs as well as large transient fluctuations in estuarine nutrient inventories and biogeochemical transformation processes. The spatial pattern of the ecosystem in the coastal zone is strongly controlled by the constraints set by the local physical environment. The latter are mainly determined by the Lagrangian residual transport field, whose complex structure control the location and residence time of land-derived, nutrient-rich waters within the coastal zone and, thus, areas of high primary production. Furthermore, the unsteady nature of the Lagrangian residual transport field and its continuous hydrodynamic adjustments lead to large variations in the spreading of the estuarine plume and, therefore, in the overall structure of the coastal ecosystem over short periods of time. The formation of a residual gyre is another distinct feature of the local physics. It increases the residence times in the nearshore coastal zone and favors an enhanced nutrient recycling in the vicinity of the gyre. The specific nutrient requirements of the various phytoplankton species further control the spatial structure of the ecosystem.

The interplay between estuarine nutrient inputs and physical constraints suggests that optimum phytoplankton growth conditions are encountered near the dynamic, yet poorly-surveyed, estuarine mouth. Marked spatial concentration gradients develop in this narrow zone and episodically lead to a reversal of the material fluxes, from the coast into the estuary. Therefore, the estuary does not operate independently from processes in the coastal zone. Euryhaline coastal diatoms, for instance, intrude far upstream into the saline estuary where favorable growth conditions are encountered during distinct episodes of the productive period. This intrusion reduces the estuarine nutrient concentrations and export fluxes, thereby reinforcing the nutrient limitation in the coastal area. Furthermore, the mixture of marine and estuarine phytoplankton may sustain an intense carbon recycling loop. The interdependency of the biogeochemical dynamics in these two ecosystems and the high process rates which occur at their transition strongly supports the continuum approach adopted here. In the past, monitoring programs and process-based laboratory studies have nevertheless largely restricted their domain of investigation to either the estuary or the coastal zone. As a consequence, the highly dynamic estuary–coastal zone interface has been poorly surveyed and comprehensive data sets for the transition zone are lacking. An improved description of the spatial structure of coastal, estuarine and freshwater phytoplankton communities from the shallow, nutrient-rich estuary to the deeper, nutrient-starved coastal zone is thus hampered by the lack of monitoring strategies encompassing the entire estuary–coastal zone ecosystem. Process-based experiments investigating the spatial variability in rate constants and model parameters of the carbon and nutrient cycles along the continuum are also essentially missing. In the future, joint observational and monitoring programs could help advance our understanding of the pronounced and fast temporal variations in biogeochemical dynamics along the continuum.

The design and the parameterization of the coupled model mainly build on the mechanistic understanding of the biogeochemistry in the Scheldt estuary and the adjacent coastal zone. It could thus be applied with minimum parameter adjustment to neighboring systems in the North Sea or other systems with similar physical and climatological characteristics. However, the set-up of the fully-coupled estuary–coastal zone model requires a detailed bathymetry, transient wind field maps and a comprehensive set of boundary conditions, compromising therefore its application to poorly surveyed environments. Although the computational time (ca. 3 days per simulated year) does not allow for an extensive sensitivity analysis, it allows for simulations at inter-annual timescales if a data set is available to constrain boundary conditions and external forcings over these periods. Therefore, the coupled model is ideally suited to explore the effect of global change and/or water quality management on coastal eutrophication. It could also provide useful guidelines for the incorporation of simplified representations of the land–ocean transition in biogeochemical models of the global coastal ocean.

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