

High Resolution Modelling of the Belgian Coastal Zone using NEMO

Auteur : Maertens de Noordhout, Guillaume

Promoteur(s) : Grégoire, Marilaure; Vandenbulcke, Luc

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UNIVERSITE DE LIEGE
FACULTE DES SCIENCES APPLIQUEES



High Resolution Modelling of the Belgian Coastal Zone using NEMO

MAERTENS DE NOORDHOUT Guillaume

Thesis presented to obtain the degree of :
Master of Science in Physics Engineering

Thesis supervisor :
GRÉGOIRE Marilaure

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Abstract

This master's thesis aims at modelling the hydrodynamics of the Belgian Coastal Zone (BCZ) using the Nucleus for European Modelling of the Ocean (NEMO), an European open-source model specialized in large scale oceanic flows from regional to global scales.

First, the configuration Atlantic Margin Model 7 km (AMM7) made by the English Met Office is assessed. It is a NEMO configuration covering the whole Northwestern European Shelf (NWES) with a resolution of 7 km. This model has been used widely and has been a product of the Copernicus Marine Environment Monitoring Service (CMEMS) for years. It is run on NEMO v5 and its results are analyzed against our current knowledge of the North Sea and against remote and *in-situ* observations.

Then, a configuration covering the Southern Bight of the North Sea (SBNS) with a resolution of 1 km, based on AMM7 is created and forced laterally by AMM7. The outputs of the model are analyzed in depth, compared with AMM7 and compared with available observations. The benefit of using a high resolution model is assessed on multiple features such as the residual currents, the Rhine/Scheldt salinity plume and a salinity front.

Statement of authorship

I, Guillaume Maertens de Noordhout, declare that I am the author of this thesis named "High Resolution Modelling of the Belgian Coastal Zone using NEMO".

I declare that all the provided scientific information that is not my work has been properly cited. All the figures have been either cited or made by my own codes.

All the thesis has been written only by me. I acknowledge that Chapter 2 is in large part inspired by the book of Geoffrey K. Vallis named "Atmospheric and Oceanic Fluid Dynamics". In particular its second chapter. The numerical part of Chapter 2 is widely issued from the NEMO reference manual.

Regarding the code, most of the preprocessing and postprocessing codes have been made by myself. The model code is the source code from the NEMO repository (<https://forge.nemo-ocean.eu/nemo/nemo>), coded by Gurvan Madec and the NEMO team. Some specific parts of my codes have been done thanks to ChatGPT. I also used ChatGPT to help me with Linux commands. Some plotting codes come from the Internet: The colormap used for the North Sea bathymetry has been found in <https://pyhogs.github.io/colormap-bathymetry.html>. The code to plot the vertical grids comes from the AMM7 repository https://github.com/JMMP-Group/CO_AMM7/tree/v9.0-alpha/scripts/analysis. Tidal ellipses have been made using the code from Zhigang Xu found in the repository https://github.com/KarenGuihou/tidal_analysis/blob/master/tidal_ellipse.py. Finally, my analysis code have used the Python libraries "caféine" and "mast-toolbox" developed by Jean-François Graillet at the MAST (unpublished yet). All my codes can be found on my Gitlab page <https://gitlab.uliege.be/Guillaume.MaertensDeNoordhout/North-sea-thesis>.

A handwritten signature in blue ink, reading "G. Maertens". The signature is stylized, with a large, looped "G" and a cursive "Maertens".

Acknowledgements

This master's thesis concludes one of the most important chapter of my life: my studies. Thus, I think that these acknowledgements should be dedicated not only to this work but to the five past years of my studies.

The first people that I want to thank is the Uliège Faculty of Applied Sciences who provided me, I am sure of it, an excellent background to start a career as a researcher. In five years, I learnt many topics of science and engineering that have nothing to do with oceanography. Still, I have never felt that it was useless, and I am sure that each single class contributed to shape my mind.

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Papy, ce travail t'es dédié.

Abbreviations

BCZ	Belgian Coastal Zone
NWES	North Western European Shelf
SST	Sea Surface Height
FIC	Fair Isle Current
WSC	Western Skagerrak Current
NCC	Norwegian Coastal Current
ROFI	Region Of Freshwater Influence
NAO	North Atlantic Oscillation
SST	Sea Surface Temperature
SSS	Sea Surface Salinity
NEMO	Nucleus for European Modelling of the Ocean
ROMS	Regional Oceanic Modeling System
POLCOMS	Proudman Oceanographic Laboratory Coastal Ocean Modelling System
GETM	General Estuarine Transport Model
HAMSOM	Hamburg Shelf Ocean Model
AMM	Atlantic Margin Model
CMEMS	Copernicus Marine Environment Monitoring Service
SGS	Subgrid Scale
TKE	Turbulent Kinetic Energy
GLS	General Length Scale model
MES	Multi-Envelope System
IO	Input Output
FRS	Flow Relaxation Scheme
RMSE	Root Mean Square Error
PB	Percentage Bias

SBT	Sea Bottom Temperature
SSS	Sea Bottom Salinity

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Part I

Introduction and state of the art

CHAPTER 1

Introduction

1.1 Context

Belgium is a small country with only a little maritime space lying between the Netherlands, the United Kingdom and France. Though this marine space is very little, Fig. 1.1 shows that it is an important place for economic activities such as energy, maritime shipping, fishing, industry and tourism.

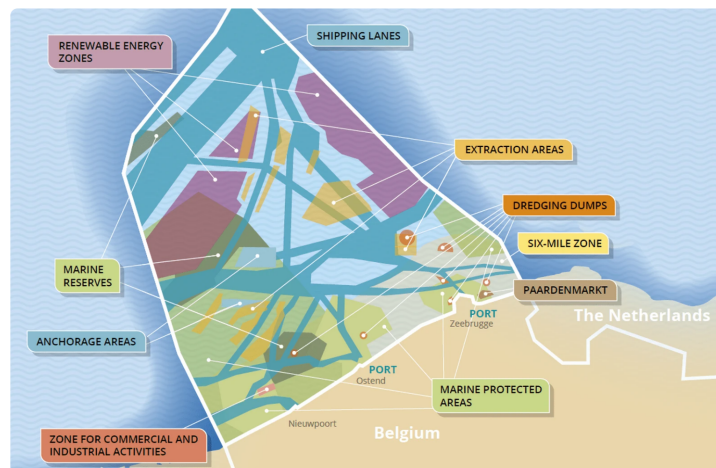


Figure 1.1. Current Belgian Marine Spatial Plan for the period 2026-2030. The map is taken from [1].

At the same time, the Belgian Coastal Zone (BCZ) is hosting some important ecosystems, for example in the Flemish Bank, that should be protected as much as possible from human activities. These places are also facing the climate change and are vulnerable to pollutants and eutrophication.

In order to build a Blue Economy, i.e. a sustainable maritime economy, policymakers should be aware of the potential impact of any future human activities and plan them in the least invasive way. Thus, they need a precise and quantitative view of these potential impacts.

One way to assess the impact of future human activities such as building new offshore windfarms is to use modelling tools. These tools, can provide an overview of the sea state over the whole 3D domain of interest and over time. However, to provide a realistic and accurate view on such a region, a high resolution model must be designed and must be able to represent the dominant processes of the region. Here, we propose to build a high resolution hydrodynamic model covering the BCZ and the Southern Bight of the North Sea (SBNS), using the general ocean circulation model NEMO [2].

1.2 General features of the NWES

The North Western European Shelf, here abbreviated NWES, is the shelf located northwest to Europe. It is shared by the United Kingdom, France, Belgium, the Netherlands, Germany and Denmark. This work will focus on the inner basin of the shelf, the part surrounded by UK, Belgium, the Netherlands, Germany and Denmark.

Topography

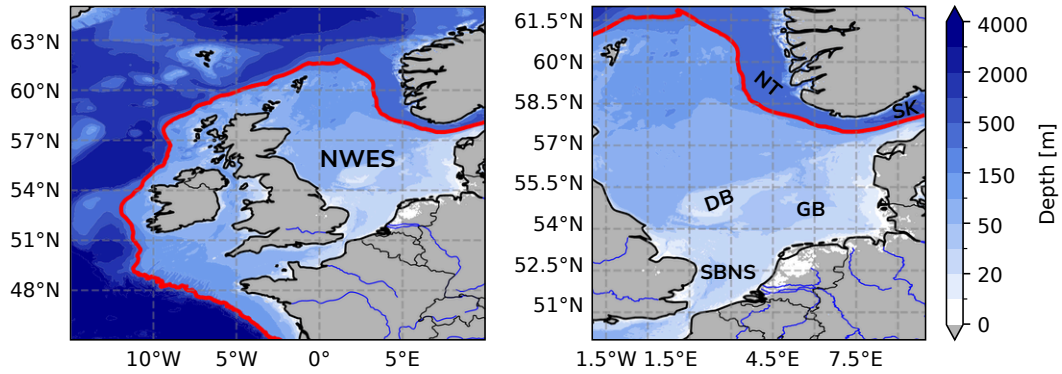


Figure 1.2. Main features of the NWES with its bathymetry coming from GEBCO 15 arc-second grid [3]. The red contour line shows the shelf break defined by the 200 m isobath. NWES: North Western European Shelf, SBNS: Southern Bight of the North Sea, GB: German Bight, DB: Dogger Bank, NT: Norwegian Trench, SK: Skagerrak.

Figure 1.2 shows the bathymetry of the continental shelf. It is shallow, about 80 m deep on average. The German and Southern Bight are particularly shallow with depths mostly around 20 m. The Dogger Bank is a sandbank located at the center south of the shelf that rises to a depth of less than 20 m. The northern part of the North Sea goes from 50 m north to the Dogger Bank to 200 m at the shelf break. At the very northeast of the shelf lies the Norwegian Trench falling to about 800 m deep. East of the Norwegian Trench lies the Skagerrak, a strait connecting the Kattegat and Baltic Sea to the North Atlantic.

The Southern Bight of the North Sea (SBNS) and particularly the BCZ are very shallow, about 40 m (Fig. 1.3). The BCZ is characterized by long sandbanks giving ripples parallel to the shore with shallow summits rising to about 10 m depth.

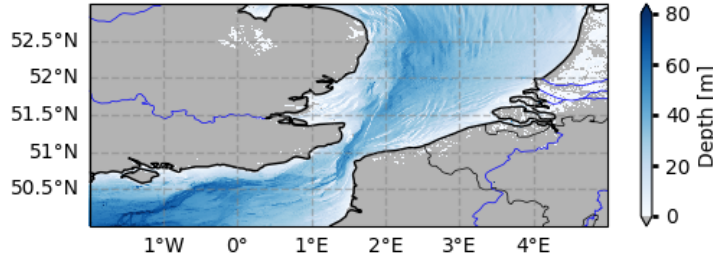


Figure 1.3. Bathymetry of the SBNS coming from GEBCO 15 arc-second grid [3].

Tides

In the North Sea, the circulation and the sea surface height (SSH) dynamics at hourly to daily timescales are driven by tides with the principal lunar M2 and solar S2 being the dominant constituents. Both lead to a semi-diurnal cycle and their combination leads to the usual neap and spring tidal pattern. The ratio of the neap to the spring tidal amplitude usually ranges from 40 to 50% but can exceed 80% in the German Bight for example [4]. The NWES presents three amphidromic points for both the M2 and S2 tide: Two in the central basin and one out of the shelf, nearby the southwestern Norwegian coast (Fig. 1.4).

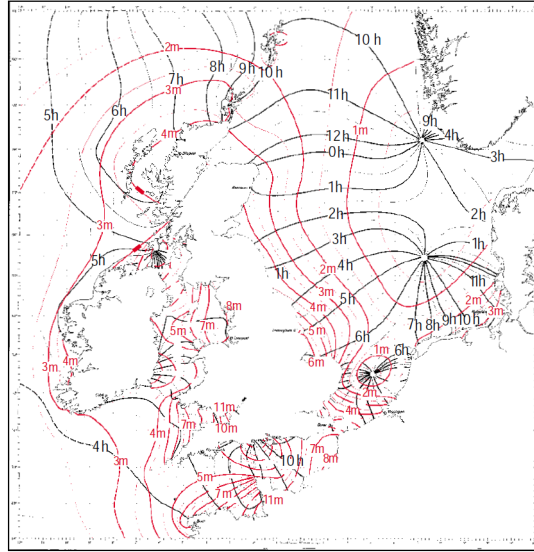


Figure 1.4. Spring tide co-tidal chart taken from [5]. The phase time is referenced to the time at which Moon crosses the Greenwich meridian.

Due to the small size of the shelf, the tidal potential has only a little influence on the dynamics [6]. Instead, it is the tidal waves coming from the North Atlantic that are important for the dynamics of the shelf. They enter from the northern side of the shelf (and in smaller extent through the Dover Strait [7]) and progress cyclonically around the shelf as Kelvin waves. On their way, the waves are dissipated by bottom friction. Then, the wave that is reflected from the English Channel has a lower amplitude in the northeast part of the shelf [6].

Because of the non-linear advection term in the Navier-Stokes equations, higher harmonics appear such as the M4 constituent [8]. More importantly, in shallow region such as the SBNS, non-linear effects coming from bottom friction, lead to a non-harmonic tidal flow [4], [6], [7]. When this happens, the tidal current averaged over a period bigger than the tidal period is different from zero. It means that there is a net residual current due to tides which is independent of the wind driven circulation for example. This residual current is one of the phenomenon responsible for sediments transport [4].

The tides have a very important dynamical effect on the North Sea. They drive tidal currents usually of the order of 10 cm/s [6] but that can reach 1 m/s or even 2 m/s at some places [4]. These currents are much stronger than the residual circulation, which is of the order of 1 cm/s on most of the shelf.

Tides play also an important role in mixing processes when the bottom friction at the seabed is considered. Indeed, the strong tidal current creates a bottom boundary layer with enhanced turbulence and thus enhanced mixing [9]. As an example, a tidal current of 1 m/s is as efficient as a current driven by winds of 25 m/s [4].

Residual circulation

The tidal processes are dominant when looking at the instantaneous current but the latter is null when averaging over several days (omitting anharmonic and long period constituents). On larger timescale than tidal scale, there is then a residual circulation, weaker than tidal currents but responsible for the net transport of water masses and tracers inside the shelf.

Many studies have described the various currents involved in the residual circulation of the NWES (see for example [10], [11], [12]). They differ mainly by the models or measurements they use. Here, a simple description of the main currents is presented.

On the shelf, the dominant winds called westerlies, come from the southwest. This pattern drives an overall cyclonic current traveling in the shelf. It can happen on episodic events that the dominant winds are reversed (easterlies), leading to an anticyclonic current. When winds are in between those two situations (wind coming either from the northwest or southeast), the circulation is no longer a strong current going around the shelf but is instead composed of smaller gyres appearing at different places on the shelf (see Fig. 4 in [6]).

Figure 1.5 shows the main currents involved in the NWES residual circulation. On the northern boundary, there is a strong slope current following the shelf break. Part of this current splits into three branches that will circulate inside the shelf and lead to Atlantic water inflows.

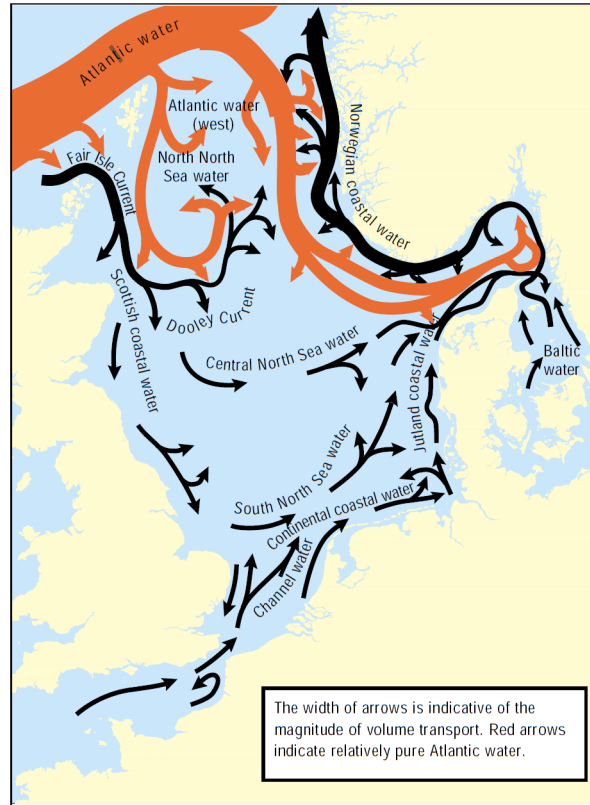


Figure 1.5. Main currents in the NWES. Red arrows represent Atlantic water while black arrows represent other water masses. Figure coming from Fig. 2.3 in [5] which is derived from [13].

The first Atlantic inflow is in between the Fair Isle islands and the Shetland islands. This current is known as the Fair Isle Current (FIC). A part of this current follows the UK shore and follows the known cyclonic current presented earlier. Another part follows the 100 m isobath and join the North Atlantic. It is known as the Dooley current, first described in [14]. A last part of the FIC follows the 40 m isobath north of the Dogger Bank [10], [12].

The second one enters the shelf just east of the Shetland islands. It goes south till joining the Dooley current. The third one, which has the biggest volume transport, called here the Western Skagerrak Current (WSC), enters in the western part of the Norwegian Trench and circulates in it with the shelf break to its right. At the end, this current joins the strong Norwegian coastal current (NCC).

All the water that enter entirely in the shelf joins at some point the Skagerrak and the NCC. The Baltic water from the Kattegat joins also the Atlantic by this current.

The residual circulation can be separated in three different types depending on their physical driver. The wind driven circulation is the circulation due to the wind blowing at the surface. The oceanic circulation is the circulation of water masses through the boundaries of the shelf, in particular the northern part of the shelf and the Dover Strait. The third one, which has received only little attention in early studies, is the density driven circulation. The latter is driven by density difference appearing for example around a tidal and fresh water plume fronts. Density driven flow can explain a significant part

of the overall circulation [15], [12]. At the end, the observed currents are a combination of those three types of circulation.

While the picture here about the circulation of the NWES does not show any vertical structure, there exists one. For example, the Skagerrak current is clearly separated in two layers, one of saline Atlantic water staying in depth and one upper layer of fresher water coming from the shelf [10]. A vertical structure can also appear in places where there is a density driven flow, at tidal fronts for example [15]. The wind driven circulation has also a natural vertical structure described by the Ekman layer.

Temperature and salinity

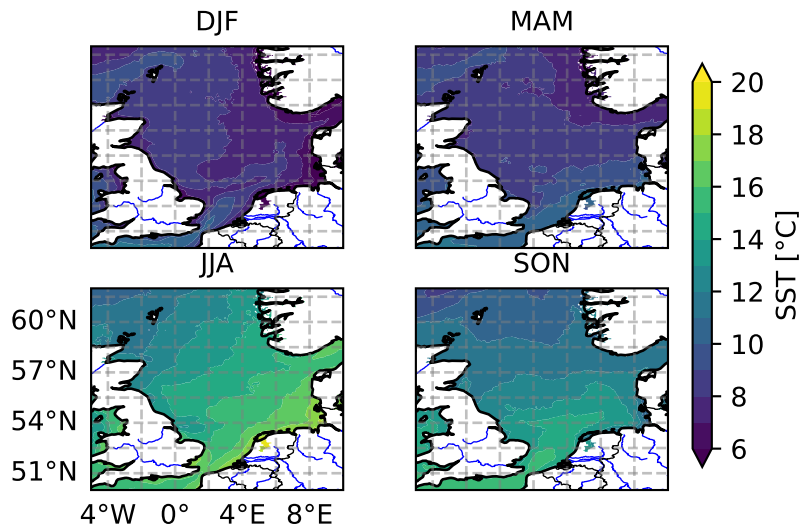


Figure 1.6. Seasonal average sea surface temperature (SST) on year 2007, coming from the Copernicus Marine Environment Monitoring service satellite product [16].

Figure 1.6 shows the seasonal sea surface temperature (SST) on the shelf. The temperature in the North Sea follows a clear annual cycle with the warmest temperature in August and coldest in February. Spatially, the temperature is higher in the southeast where there is the inflow of warm Atlantic water at the Dover Strait [5]. Interannual variation is related to the variability of the runoffs and the Atlantic inflows [17]. There is also a strong correlation between the temperature and the North Atlantic Oscillation (NAO) [5]. For example, during positive phase of the NAO index, the air temperature is usually warmer, affecting the air-sea heat flux and leading to a warmer SST [18].

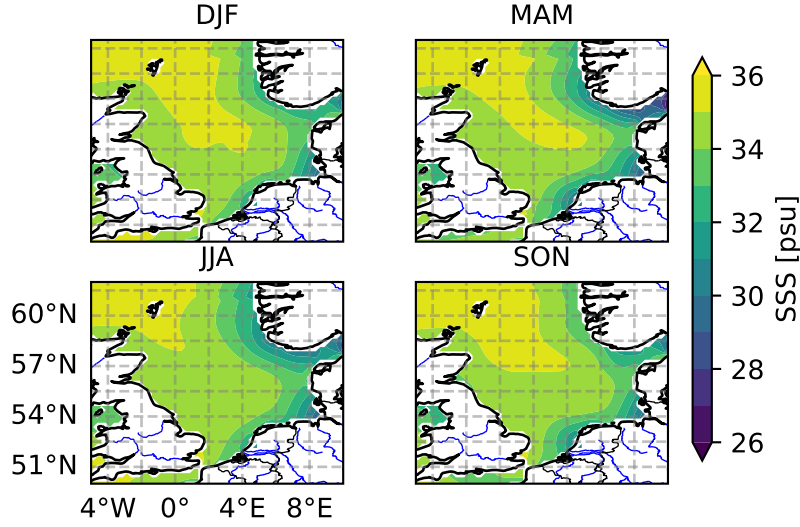


Figure 1.7. Seasonal average sea surface salinity (SSS) on years 2005 to 2014, coming from the World Ocean Atlas [19].

The salinity in the North Sea does not show significant annual variation but has a large spatial variability [20]. In regions of freshwater influence (ROFIs), the salinity is much lower due to the inflow of rivers like in the German Bight which receives not much Atlantic water but has a strong river inflow [10].

Stratification dynamic

The NWES presents some places with stratified water and some that are well mixed. The shallow southeastern part of the shelf is well mixed by tides and is then never stratified. On the contrary, in deeper region, tides may be not sufficient to mix the larger water column. In Summer, when the upper part of the column heats up, stratification starts then to occur. Both regions of stratified and mixed water are separated by a tidal front, a line where a strong gradient of temperature is present. This front first appears in the deep part of the shelf and progressively moves to the south in shallower region till reaching its permanent place at the end of Spring [10]. It then goes back progressively north at the end of Summer.

Other regimes of stratification do also exist. In the deep Norwegian Trench, the water is permanently stratified [5]. In ROFIs, the water can also be stratified [21].

Long term changes

North Sea presents a long term variability that can be separated in two parts the first one is linked to the natural oscillation of the climate. In this case the NAO is dominant. The second is linked to the global climate change due to anthropogenic pressure.

The NAO influences mainly the residual circulation through the variation of dominant wind patterns. Atlantic inflows are also influenced by the NAO since the whole Atlantic circulation is related to the NAO [6]. The SST is also closely correlated to the NAO index.

The global climate change will lead undoubtedly to change in the dynamic of the NWES. Temperature are expected to increase by 1 to 4°C. Due to the change in wind forcing, a more intense cyclonic circulation, with an associated geostrophic sea level increase of 20 to 40 cm is expected. Storm surge are also expected to be of the same mean intensity but more frequent. Those results have been shown in [22] by considering a scenario where the amount of atmospheric CO₂ is doubled from its reference state at that time (period 1979 - 1993). The tidal system is also expected to change with a shift of the amphidromic points [6].

1.3 An overview of NWES circulation models

The modelling of the North Sea has started in the seventies when computers and with numerical tools became able to solve mathematical models relevant for large scale geophysical flows. First models where designed either to model the short term periodic tidal circulation and storm surge processes [23], [24], [25] or the long term residual circulation [26], [27]. These models were all barotropic¹ due to the low computational resources available at that time.

Later, in the eighties and nineties appeared baroclinic models. In Liège, model for representing the high frequency [28] circulation and low frequency residual circulation [10] were developed. In Hamburg, the model now known as HAMSOM was successfully applied on the North Sea in [29]. First high resolution studies were also made for example with POLCOMS in [30]. These are only a few examples among a long list.

In the past twenty years, the computational resources have become huge. It is then possible to solve the primitive equations of the ocean with a high spatial and temporal resolution for long time periods. While the ocean modellers used to build their own code for modelling their region of interest, they now use predefined engines like NEMO or ROMS. Some codes are suited for global simulations while others are made for regional modelling. These engines often present a lot of flexibility regarding the choice of the coordinate system, vertical and horizontal mixing parametrizations and many other numerical features. Table 1.1 shows part of the modern models applied to the North Sea for various purposes. The main differences in between each model lie in the horizontal resolution and the type of vertical coordinate system.

¹A barotropic model is a model solving only the 2D depth-averaged Navier-Stokes equations.

Engine	Horizontal resolution	Vertical discretization	Reference
NEMO	7 km	51 hybrid s - z levels	[31], [32]
NEMO	1.5 km	51 hybrid s - z levels	[33]
ROMS	1 km grid nested in a 5 km grid	15 s levels	[34]
ROMS	1 km grid nested in a 5 km grid	15 s levels	[7]
POLCOMS	12 km	42 s levels	[35], [36]
POLCOMS	1.8 km	34 s levels	[12]
GETM	20 km	s levels with density adaptation [37]	[38]
GETM	1.8 km	42 s levels with density adaptation [37]	[39]
HAMSOM	3 km	21 z levels	[40]
HAMSOM	2.7 km	30 z levels	[41]

Table 1.1. Non-exhaustive list of modern models of the NWES. z coordinates refer to vertical level aligned with the geopotential surface while s coordinates follow the bathymetry with a stretching to have a finer resolution at the surface (Sec. 2.2.2).

NEMO

The Nucleus for European Modelling of the Ocean (NEMO) [2] is a general circulation model suited for both global and regional modelling. The model is based on the hydrostatic Boussinesq primitive equations expressed on a curvilinear generic grid. The equations are solved by finite differences using a staggered Arakawa C-grid and centered difference. Any system of vertical discretization can be used which is a great advantage when it comes to model a shelf. NEMO includes a wide variety of parametrizations for unresolved processes in the vertical and horizontal physics. NEMO has also coupling capabilities with atmospheric, wave and biogeochemical models. Two-way nesting is also possible to couple two or more nested ocean configurations. More details will be presented in Sec. 2.

NEMO have been applied many times on the North Sea. Here a focus is given on the configuration AMM7 (Atlantic Margin model 7 km) from the English Met Office. This model was first described in [31] and have seen several upgrades [32]. It uses NEMO with a grid of approximately 7 by 7 km resolution. On the vertical, the latest version uses 51 terrain-following levels [42]. Its operational version that includes SST assimilation has been used for years by the Copernicus Marine Environment Monitoring Service (CMEMS) [43].

1.4 Goals of this thesis

While many general circulation models have been applied on the North Sea, only a few have focused on the BCZ. Here, a high resolution NEMO model will be configured on the SBNS. A downscaling approach is adopted for this thesis, meaning that a first low resolution model covering a large area is used to force a similar model but covering a smaller inner region with a higher resolution.

In this case, the low resolution model is AMM7. This configuration, accessible on a public Github repository will be run and its output will be analyzed. The quality of this model will be assessed by comparing the model results with our knowledge of the North Sea and with observations.

A zoom on the SBNS will then be made by creating a configuration similar to AMM7 but with a smaller domain covering only the SBNS with a resolution of 1 km. Its output will be compared to AMM7 to assess the benefit (or disadvantage) of using a high resolution model.

Chapter 2 will describe the mathematical and numerical model used to simulate the North Sea and SBNS physics. Chapter 3 will describe the method used to analyze the model results and the set of available remote and *in-situ* observations. Finally, chapter 4 and 5 will explore the various physical quantities modelled by the high and low resolution configurations.

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Part II

Material and Method

CHAPTER 2

Ocean general circulation model NEMO

In this chapter, the mathematical model used to simulate the circulation of the North Sea will be developed, starting from the primitive equations of geophysical dynamics. Then, the numerical model implemented in NEMO will be described. Since NEMO includes a wide variety of setting and equations, only the ones relevant for this thesis will be described.

2.1 Primitive equations of geophysical dynamics

In the case of the North Sea, and more generally for oceanic flow at the scale of a sea or an ocean, the governing physics is what is called the geophysical fluid dynamics. The latter is as specific branch of fluid dynamics focusing on large scale flow happening in our oceans and our atmosphere.

A geophysical flow has particular features making its physics unique:

First, it happens on a large spatial scale, at least several kilometers. On the vertical, the spatial scale is much smaller because the ocean depth goes from a few meters to a few kilometers. This difference of spatial scales, a high aspect ratio, leads to a flow that happens mainly on the horizontal. The high aspect ratio leads also to a flow that is mainly in hydrostatic balance.

The second characteristic is that because of the large spatial scale, the rotation of Earth cannot be neglected and has a strong dynamical effect and cannot be neglected. One good example for this is the geostrophic equilibrium that cause the atmosphere to move along isopressure line instead of moving in the direction of the pressure gradient.

Concerning the time scales, geophysical fluid dynamics deals with relatively slow motions. Speeds are of the order of cm/s to a few m/s. Transient effects go from the hourly to decadal time scales depending on the physical phenomenon that is explored. Here, the fastest phenomena considered will be tides having a period of few hours.

From a mathematical point of view, the motion of a geophysical fluid, here seawater, is governed by the Navier-Stokes equations expressed in the Earth rotating frame. This set of equations contains the mass, momentum and energy conservation equations. A

generic tracer conservation equation can also be drawn for any quantity that is advected and diffused in the fluid, for example, a dissolved ion.

Here, the equations will be briefly explained, skipping most of the mathematical details. Those details can be found in [1] which is the book over which this section is built. In particular, the sphericity of the Earth is here neglected to avoid lengthy equations involving metric terms.

2.1.1 Continuity equation

The first conservation equation states that the mass of a fluid is conserved. No mass of fluid can be created or destroyed, it can only be transported by the flow. In an Eulerian approach (i.e. when looking at a fixed location in space), this equation is

$$\frac{\partial \rho}{\partial t} = -\nabla \cdot (\rho \mathbf{v}), \quad (2.1)$$

with ρ the density of the fluid, $\mathbf{v}$ the 3D velocity field and $\nabla \cdot$ is the divergence operator. This equation says that the mass of a fixed volume in space and time can only increase or decrease through a flux of mass $\rho \mathbf{v}$. This equation can also be written in a Lagrangian approach with a material derivative as

$$\frac{D\rho}{Dt} = \frac{\partial \rho}{\partial t} + \mathbf{v} \cdot \nabla \rho = -\rho \nabla \cdot \mathbf{v}, \quad (2.2)$$

where $\frac{D}{Dt}$ is the material or Lagrangian derivative, giving the time evolution of a variable when following the path of a water parcel. ∇ is the gradient operator. This approach shows that the mass of a fluid parcel can only change by an expansion or contraction of its volume, indicated by the divergence of the speed field.

The density of seawater can be written as the sum of a constant reference density $\rho_0 = 1035 \text{ kg/m}^3$ and a density anomaly or variation $\Delta\rho$. In the ocean, this density anomaly is much weaker than the reference density itself. Knowing that, Eq. 2.2 is simplified as

$$\begin{aligned} 0 &= \frac{\partial \rho}{\partial t} + \mathbf{v} \cdot \nabla \rho + \rho \nabla \cdot \mathbf{v} \\ &= \frac{\partial \Delta\rho}{\partial t} + \mathbf{v} \cdot \nabla (\Delta\rho) + \rho_0 \nabla \cdot \mathbf{v} + \Delta\rho \nabla \cdot \mathbf{v} \\ &\approx \rho_0 \nabla \cdot \mathbf{v} \end{aligned} \quad (2.3)$$

leading to the incompressibility equation

$$\nabla \cdot \mathbf{v} = 0. \quad (2.4)$$

All the terms involving the density anomaly have been neglected due to the smallness of the density variation against the reference density. In geophysical dynamics, this approximation is called the Boussinesq approximation which leads to a volume conserving fluid instead of a mass conserving fluid [2].

2.1.2 Free surface

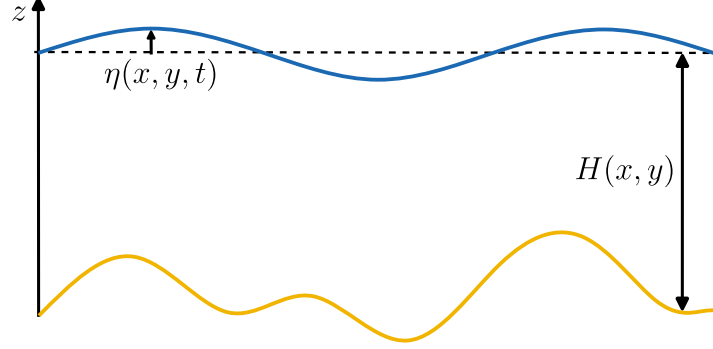


Figure 2.1. Sketch showing the free surface and bathymetry on a vertical plane. The free surface displacement is exaggerated for clarity. x and y are two horizontal independent coordinates. The dashed line is a reference geopotential surface at which the vertical coordinate $z = 0$.

An important part of the sea dynamics is the movement of the air-sea interface. Therefore, this boundary should not be approximated as a flat still surface but should instead be given by an evolution equation for the sea surface height (SSH). This equation can be obtained by a mass budget (or volume budget for the Boussinesq approximation) on the whole water column which gives

$$\frac{\partial \eta}{\partial t} = -\nabla_h \cdot ((H + \eta)\mathbf{U}) + P + R - E, \quad (2.5)$$

where η and H are the SSH and depth measured from the reference surface $z = 0$. $\mathbf{U}$ is the barotropic or depth averaged horizontal velocity

$$\mathbf{U} = \frac{1}{H + \eta} \int_{-H}^{\eta} \mathbf{u} \, dz. \quad (2.6)$$

Finally, P , R and E are respectively the precipitation, river inflow and evaporation flux accounting for an exchange of water with the exterior domain (expressed in a flow rate per surface unit).

2.1.3 Momentum conservation equation

The momentum conservation equation is based on the second Newton's law which states that the time derivative of the momentum in an inertial frame is the sum of the applied forces. The applied forces include Earth gravity, tidal forces and the fluid pressure. Here, seawater is assumed to be a perfect fluid, i.e. it has no viscosity. The equation is expressed in the Earth rotating frame. Thus, a fictitious force accounting for the constant rotation must also be added. Finally, the equation writes

$$\rho \frac{D\mathbf{v}}{Dt} + 2\rho\boldsymbol{\Omega} \wedge \mathbf{v} = \rho\mathbf{g} - \nabla p + \rho\nabla\Phi_t, \quad (2.7)$$

with $\boldsymbol{\Omega}$ the Earth rotation vector, aligned with the Earth south-north axis and with a magnitude $2\pi \text{ rad day}^{-1}$. It is assumed that the speed of rotation is constant and that the centrifugal term is included in the effective gravity acceleration field $\mathbf{g} = \mathbf{g}_0 - \boldsymbol{\Omega} \wedge \boldsymbol{\Omega} \wedge \mathbf{r}$ with $\mathbf{g}_0$ the Earth gravity acceleration and $\mathbf{r}$ the position vector of a water particle with respect to the Earth center. p is the pressure field. Φ_t is a tidal potential taking into account the gravitational force of the Moon and Sun.

This equation can be simplified first by using Boussinesq approximation. Assuming that the density anomaly is small, the latter can be neglected in each term except the gravity term, leading to

$$\frac{D\mathbf{v}}{Dt} + 2\boldsymbol{\Omega} \wedge \mathbf{v} = \frac{\rho}{\rho_0} \mathbf{g} - \frac{1}{\rho_0} \nabla p + \nabla \Phi_t. \quad (2.8)$$

A second approximation can be made when considering the large ratio of the horizontal and vertical scale of the flow. It can be shown with a scaling analysis that (i) the vertical speed is much smaller than the horizontal velocity and (ii) the flow is within a good approximation in hydrostatic balance. Thus, Eq. 2.8 projected along a vertical unit vector (normal to the geopotential surfaces) gives

$$\frac{\partial p}{\partial z} = -\rho g, \quad (2.9)$$

with z a vertical coordinate positive upward.

In a realistic geophysical flow, the non-linear term of the momentum conservation equation cannot be neglected. That is, the flow is by nature turbulent. It is composed of many features of different time and spatial scale, that interact together due to non-linearities. To resolve all these interacting scales, a model must have a spatial resolution going down to the Kolmogorov microscale where the smallest eddies are dissipated by molecular diffusion. In practice, we don't have the computational resources for such a model. Instead, a Reynolds decomposition is applied, splitting the flow in a mean resolved part and an unresolved perturbation. By averaging the continuity and momentum conservation equation, we obtain similar equations but for the mean, large scale flow. The only difference with the initial equations is that a term $\mathbf{D}^U$ must be added at the RHS of the momentum equation to take into account the coupling of small scale unresolved motion with the mean flow.

The major difference between the classical Kolmogorov theory of turbulence and the geophysical turbulence lies in the considered scales. In geophysical turbulence, the mean flow in the Reynolds decomposition is the flow at scales of more than a few kilometers while the unresolved flow is at smaller scale¹. This scale is much bigger than what is seen in classical turbulence, for example in a turbulent boundary layer. At this scale, the assumption of isotropic turbulence does not hold anymore and the energy cascade with a $-5/3$ log slope is not valid anymore. The rotation and stratification of the flow also make the classical theory invalid. The turbulent fluxes must then be parametrized accordingly.

¹Usually, the separation is made between the macroscale (100 - 1000 km) motion and the mesoscale motion (10 - 100 km). But nowadays, with the development of high resolution models, this separation is made between mesoscale and sub-mesoscale features (1 - 10 km).

Finally, the momentum equation can also be written in the useful vector invariant form

$$\frac{\partial \mathbf{u}}{\partial t} + (\mathbf{f} + \zeta \mathbf{k}) \wedge \mathbf{u} + w \frac{\partial \mathbf{u}}{\partial z} + \frac{1}{2} \nabla_h (|\mathbf{u}|^2) = -\frac{1}{\rho_0} \nabla_h p + \nabla_h \Phi_t + \mathbf{D}^U, \quad (2.10)$$

where $\mathbf{u}$ is the horizontal velocity vector (velocity along geopotential). $\mathbf{k}$ is the unit vertical vector (normal to the geopotential surfaces). $\mathbf{f} = 2\boldsymbol{\Omega} \cdot \mathbf{k}$ is the Coriolis parameter. $\zeta = (\nabla \wedge \mathbf{v}) \cdot \mathbf{k}$ is the vorticity of the flow projected on the vertical axis. w is the vertical component of the speed vector. ∇_h is the 2D gradient operator acting on the geopotential surface. This equation is very useful because it is valid in any coordinate system, in particular the spherical coordinate system.

2.1.4 Energy equation

To know the temperature field in the domain of interest, an energy conservation equation is needed, in this case, for the internal energy $I = c_{p,0}T$. I is the internal energy, $c_{p,0}$ the specific heat capacity of seawater, assumed constant and T the temperature. If viscous dissipation, pressure work and heat conduction are neglected, the internal energy evolves as

$$\frac{D}{Dt}(\rho_0 c_{p,0}T) = \rho_0 c_{p,0} \frac{DT}{Dt} = D^T + Q^T \quad (2.11)$$

with D^T a subgrid scale turbulence term that must be parametrized and Q^T a heat source term. The reference density is used because of the Boussinesq approximation. The source term is the radiative heat flux from the Sun that penetrates in the water column (See Sec. 2.2.6).

For convenience with the equation of state (Sec. 2.1.7), the *in-situ* temperature is replaced by a potential temperature. The potential temperature is defined as the temperature a water parcel would have if it was moved adiabatically to some reference depth, usually the surface.

2.1.5 Salinity equation

The seawater contains many dissolved ions (mainly chloride, sodium, sulfate and magnesium) that will influence its density and so the dynamics of the flow. These ions are transported and diffused with the flow thus having a concentration varying in space and time. An equation for the quantity of ions is thus needed.

Since the major ions dissolved in the water are in constant proportions, a single variable can be defined. The salinity denoted S is the amount of dissolved major ions in a fixed quantity of seawater

$$S = [Cl^-] + [Na^+] + [Mg^{2+}] + [SO_4^{2-}] + \dots, \quad (2.12)$$

where [...] denotes the mass concentration of a dissolved ion, expressed in mass of dissolved chemical per volume of seawater. Fortunately, the major ions written above are found in constant proportion in the oceans (Forchhammer principle). Then, from the sole knowledge of S , the concentration of each major ion can be found.

These major ions do not interact meaning there is no internal source or sink of salinity. The evolution equation thus writes

$$\frac{DS}{Dt} = D^S, \quad (2.13)$$

with D^S a subgrid turbulence term. Instead of the practical salinity, an absolute salinity is sometime used to be consistent with the employed equation of state (Sec. 2.1.7).

2.1.6 SGS parametrization and turbulence modelling

The terms $\mathbf{D}^u$, D^T and D^S contain unresolved small scale fluctuations that must be formulated in terms of the mean flow computed by our equation. This is called the turbulent closure problem.

SGS diffusion operators

The first step to close this problem is to assume that those terms are diffusion terms. The second assumption is that this diffusion-like process is not isotropic but is instead separated into a vertical and horizontal components. The SGS terms then separate as

$$\begin{aligned} \mathbf{D}^U &= \mathbf{D}^{lU} + \mathbf{D}^{vU} \\ D^T &= D^{lT} + D^{vT} \\ D^S &= D^{lS} + D^{vS} \end{aligned} \quad (2.14)$$

where l and v indicate the lateral and vertical terms.

On the vertical, the SGS terms write simply

$$\mathbf{D}^{vU} = \frac{\partial}{\partial z} \left(A^{vm} \frac{\partial \mathbf{u}}{\partial z} \right) \quad D^{vT} = \frac{\partial}{\partial z} \left(A^{vt} \frac{\partial T}{\partial z} \right) \quad D^{vS} = \frac{\partial}{\partial z} \left(A^{vt} \frac{\partial S}{\partial z} \right), \quad (2.15)$$

where A^{vm} and A^{vt} are the vertical turbulent viscosity and diffusivities that must be determined by a turbulent closure formulation. It is assumed that the diffusivity for temperature and salinity are the same.

In current models where the horizontal resolution is of several hundred meters to several kilometers, the lateral SGS operator is more a numerical term used to maintain the stability of the numerical model despite its low horizontal resolution. This term will thus be developed in the numerical model section 2.2.5

Vertical turbulent diffusivities

On the vertical, the turbulence is closed by either a one (TKE) or two equations model (GLS).

In the first case, the TKE model uses the turbulent kinetic energy $\bar{e}$ as a prognostic variable, governed by

$$\frac{\partial \bar{e}}{\partial t} = A^{vm} \left(\left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2 \right) - A^{vt} N^2 + \frac{\partial}{\partial z} \left(A^{vm} \frac{\partial \bar{e}}{\partial z} \right) - \epsilon. \quad (2.16)$$

The first term is a source related to the vertical shear of the horizontal speed field (u and v being the eastward and northward speed component) that enhance the turbulence. The second term contains the squared buoyancy frequency

$$N^2 = \frac{1}{\rho_0} \frac{\partial \rho}{\partial z} \quad (2.17)$$

characterizing the stratification of the water column. When the column stably stratified ($N^2 > 0$), buoyancy and gravity will tend to stabilize the column and prevent turbulence. A vertical mixing term diffuse vertically the turbulence with the momentum vertical turbulent viscosity. Finally, a dissipation term given by the turbulent dissipation ϵ represents the dissipation of turbulent eddies at the molecular viscous microscale. The two coefficients and the turbulent dissipation are parametrized by

$$\begin{aligned} A^{vm} &= C_k l_m \sqrt{\bar{e}} \\ A^{vt} &= \frac{A^{vm}}{Pr} \\ \epsilon &= c_\epsilon \frac{\bar{e}^{3/2}}{l_\epsilon}. \end{aligned} \quad (2.18)$$

C_k and c_ϵ are constant. Pr is the Prandtl number giving the ratio of the momentum turbulent viscosity A^{vm} and the tracers' turbulent diffusivity A^{vt} . This adimensional number can be chosen constant or dependent on the Richardson number

$$Ri = N^2 / \left| \frac{\partial \mathbf{u}}{\partial z} \right|^2. \quad (2.19)$$

l_m and l_ϵ are a dissipation and mixing length scale to be parametrized. They are basically the scale at which macroscale eddies are dissipated to feed the turbulence. The simplest parametrization is

$$l_\epsilon = l_k = \sqrt{2\bar{e}}/N \quad (2.20)$$

giving a length scale enhanced by turbulence but diminished by stratification.

A more advanced turbulent closure scheme is the General Length Scale (GLS) model. This model involves a similar equation for the turbulent kinetic energy but adds a second prognostic variable $\psi = C_{0\mu}^p \bar{e}^m l^n$. $C_{0\mu}$ is a constant. l is the mixing length and the triplet (m, n, p) is a set of real numbers chosen by the user. In the case of the triplet $(3, 3/2, -1)$, the classical $k - \epsilon$ closure is recovered.

The two quantities are governed by

$$\frac{\partial \bar{e}}{\partial t} = \frac{A^{vm}}{\sigma_e} \left(\left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2 \right) - A^{vt} N^2 + \frac{\partial}{\partial z} \left(A^{vm} \frac{\partial \bar{e}}{\partial z} \right) - \epsilon \quad (2.21)$$

$$\frac{\partial \psi}{\partial t} = \frac{\psi}{\bar{e}} \left\{ \frac{C_1 A^{vm}}{\sigma_\psi} \left(\left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2 \right) \right\} - C_2 A^{vt} N^2 + \frac{\partial}{\partial z} \left(A^{vm} \frac{\partial \psi}{\partial z} \right), \quad (2.22)$$

with σ_e , σ_ψ , C_1 and C_2 constant. The turbulent dissipation which couple both equations is expressed by

$$\epsilon = C_{0\mu} \frac{\bar{e}^{3/2}}{l}. \quad (2.23)$$

The turbulent viscosity and diffusivity are given by

$$\begin{aligned} A^{vm} &= C_\mu \sqrt{\bar{e}} l \\ A^{vt} &= C_{\mu'} \sqrt{\bar{e}} l, \end{aligned} \quad (2.24)$$

where C_μ and $C_{\mu'}$ are two stability functions depending on $\bar{e}$ and l . One possibility is the Canuto [3] stability functions.

2.1.7 Equation of state

Forgetting about the subgrid scale terms $\mathbf{D}^U$, D^T and D^S and assuming the forcing term Q^T is known, there is a total of 7 scalar unknowns: 3 velocities, the pressure field, the density, the temperature and the salinity. For the equations, there is a total of 6: 3 for the momentum equation, one for continuity, one for the temperature and one for the salinity. There is then one lacking equation that should be a thermodynamic relation linking the density of seawater to its temperature, salinity and pressure

$$\rho = \rho(T, S, p). \quad (2.25)$$

Unlike a perfect gas, seawater has a non-linear relation. Several empirical equations exist. NEMO provides three possibilities: a linear equation, EOS80 and TEOS-10. The last two are polynomials with various empirical coefficients.

2.1.8 Tidal forcing

On a continental shelf such as the NWES, tides are the dominant feature of the flow [4]. In the mathematical model, tides can appear in two ways: (i) as a tidal potential; (ii) at open lateral boundaries.

The main principle of tidal modelling relies on a linearity assumption. It is assumed that the tidal potential, the oscillating potential that adds to Earth gravity, is a sum of different tidal constituents such that

$$\Phi_t = \sum_{k=1}^{N_{tide}} A_{pot,k} \cos(\omega_k t + \phi_{pot,k}), \quad (2.26)$$

where N_{tide} is the number of considered constituents. ω_k is the angular frequency of the tidal constituent k . $A_{pot,k}$ and $\phi_{pot,k}$ are the amplitude and phase of the potential for the constituent k . Both depend on the horizontal position, leading to a potential varying horizontally and then an horizontal force applied on the fluid.

There exist a number of constituents, each accounting for some astronomical movement or non-linear effects acting on tides. The most important and basic constituents are the

semi-diurnal principal lunar M2 and solar S2 which account for a circular orbit of the Moon around Earth and the Earth around the Sun; The diurnal constituents P1 and O1 account for the declination of both orbit with respect to the Equator; The constituents L2 and N2 account for the ellipticity of Moon's orbit ... Each of these constituents have a distinct pulsation with diurnal constituents having a period close to 24 hours and semi-diurnal close to 12 hours.

By using again this linear decomposition, the importance of each constituent over the SSH or other quantities can be assessed. The SSH can for example be decomposed such that

$$\begin{aligned}\eta(t) &= A_0 + \sum_{k=1}^{N_{tide}} A_k \sin(\omega_k t) + B_k \cos(\omega_k t) \\ &= A_0 + \sum_{k=1}^{N_{tide}} A_{tot,k} \sin(\omega_k t + \phi_k),\end{aligned}\tag{2.27}$$

with $\eta(t)$ the SSH at some place and time t . A_0 is the residual amplitude of the signal, the steady one. A_k , B_k are the sine and cosine amplitudes while $A_{tot,k}$ and ϕ_k are the total amplitude and phase of each harmonic constituent. Given a signal at a given position, the coefficients can be found using a multi-linear regression analysis (App. A).

2.2 The numerical model NEMO

The Nucleus for European Modelling of the ocean NEMO [5] is an European consortium developing modelling tools for ocean related studies including forecasting and climatology. It is composed of the ocean physical model, a biogeochemical model and a sea-ice model. It can be coupled to other atmospheric, waves and biological models.

The ocean physical model solves the Navier-Stokes primitive equations described in Sec. 2.1 using the finite difference method. The large amount of available numerical settings and parametrizations makes NEMO suited for various application, from large scale global low resolution models to regional high resolution models. NEMO offers also other important features such as a data assimilation system, a stochastic module and an adaptative refinement tool (AGRIF). From a computational point a view, the code is written in Fortran90 and parallelized using the Message Passing Interface (MPI) library.

2.2.1 Mesh coordinate system

To overcome the drawback that finite difference must employ rectilinear grid, NEMO defines its primitive equations on a curvilinear grid. This coordinate system is *in-fine* imposed by the mesh provided by the user. The mesh can be a simple uniform mesh with constant spacing in latitude and longitude or can have a varying spacing in the zonal direction, for example to account for the change in zonal arc length at high latitude. Mathematically, each cell's center is defined by a set of three integer indices (i, j, k) and the mesh is defined by the relation

$$(\lambda, \phi, z) = (\lambda(i, j), \phi(i, j), z(i, j, k, t))\tag{2.28}$$

with i, j, k integer starting at one and ending at the mesh size. λ and ϕ are the longitude and latitude and z is the vertical coordinate. The horizontal grid is not allowed to change in the vertical while the vertical spacing can change in the three dimensions of space and in time. If the three relations $\lambda(i, j)$, $\phi(i, j)$ and $z(i, j, k, t)$ are continuous functions, then the set (i, j, k, t) can be seen as the new coordinate system in which the primitive equations will be expressed. By solving the equations in this coordinate system, the mesh will be purely rectilinear and uniform with a mesh size of one.

2.2.2 Vertical mesh

While a uniform grid spacing on the horizontal mesh is usually fine for regional applications, the vertical mesh can certainly not have a uniform spacing. Indeed, most of the physics happens in the surface layer (few tens of meters to a hundred meters) and the vertical resolution should then be increased there.

In coastal application and shallow places like the NWES, the bottom topography plays a crucial role in the sea circulation. Thus, it is of interest to have a good, smooth representation of the topography instead of a stair-like representation that would appear if the vertical levels were entirely flat (without horizontal dependence).

At the bottom, the bathymetry imposes a strong constraint on the flow because of the non-permeability condition. Further up in the water column, the flow will be less influenced by the topography but will be instead mainly constraint by the shape of the isopycnals (surface of constant density). In the surface layer, the water is well mixed with a density varying very weakly, the flow happens there on the geopotential [2]. Finally, the movement of the free surface must also be taken into account. This discussion shows then that a suited choice of a vertical coordinate system is not trivial and crucial to have a proper model. Stephen Griffies even says that it is "the single most important choice to be made when designing an ocean model" [2].

Free surface consideration

To represent correctly the external gravity waves, the non-linear equation 2.5 must be solved without any approximation and the variation of volume of the water column should be taken into account. To do that, NEMO basically adjust the vertical scale factor²

$$e_3(i, j, k, t) = e_{3,0} \left(1 + \frac{\eta(i, j, t)}{h_0(i, j)} \right) \quad (2.29)$$

with e_3 the scale factor defined by

$$e_3(i, j, k, t) = \frac{\partial z(i, j, k, t)}{\partial k}, \quad (2.30)$$

where $e_{3,0}$ the scale factor at rest ($\eta = 0$) and h_0 the water column height at rest. The vertical position of each level can be recovered by integration of the scale factor with

²The scale factors are rigorously defined when writing the spatial derivatives in terms of the mesh coordinate system. For short, they are basically the size of the cells in each direction

respect to the vertical coordinate k .

z coordinate

This is the simplest case of vertical coordinate system where the levels are independent of the horizontal position. They are then given by

$$z_0(i, j, k) = S(k), \quad (2.31)$$

where $z_0(i, j, k)$ is the vertical coordinate transformation at rest. $S(k)$ is a stretching function designed to have a varying resolution. An usual choice is

$$S(k) = h_{sur} - h_0 k - h_1 h_{cr} \log \left(\cosh \left(\frac{k - h_{th}}{h_{cr}} \right) \right), \quad (2.32)$$

where the h_x are user defined parameter tuning the levels position. Instead of tuning these parameters, the user can choose the depth and thickness of the bottom and surface cells. It gives 2 algebraic equations for the function S and 2 for its derivative. A fifth equation is added by choosing a value for h_{cr} . This parameter usually in between 3 and 10 defines how stretched are the levels with a higher h_{cr} giving a sharper stretching. At the end, there are 5 equations for the 5 parameters.

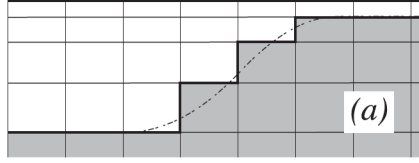


Figure 2.2. Example of 5 z levels on a sloping bathymetry coming from [5].

This coordinate system is the simplest but the main drawback is its poor representation of the bottom topography. Indeed, the seabed in this case is represented by a stairs profile (Fig. 2.2).

Terrain following s coordinates

s coordinates are designed to follow the bathymetry of the domain, written

$$z_0(i, j, k) = H(i, j)C(k), \quad (2.33)$$

where $C(k)$ is a stretching function going from 0 at the surface to -1 at the bottom and $H(i, j)$ is the depth at the horizontal position (i, j) .

We define the s coordinate as

$$s(k) = -\frac{k - 0.5}{n}, \quad (2.34)$$

with n the number of vertical levels. The term 0.5 is there because the depth of the levels are taken at the cell's centers (T point in Fig. 2.5). The simplest stretching function is

$$C(k) = s(k). \quad (2.35)$$

In this case, there is actually no stretching, and the vertical spacing is uniform in depth.

A more advanced stretching function is given by Song and Haidvogel [6]:

$$C_{SH94}(k) = (1 - b) \frac{\sinh(\Theta s)}{\sinh(\Theta)} + b \frac{\tanh\left(\Theta\left(s + \frac{1}{2}\right)\right) - \tanh\left(\frac{\Theta}{2}\right)}{2 \tanh\left(\frac{\Theta}{2}\right)}, \quad (2.36)$$

with $b \in [0, 1]$, $\Theta \in [0, 20]$ that control the stretching toward the bottom and the still sea surface. b close to 1 means a high stretching toward the bottom and a high Θ means a high stretching to the surface. This stretching function is known to have a poor surface resolution in deep region. Siddorn and Furner [7] have instead proposed a new stretching function to improve air-sea surface representation in deep water using

$$C_{SF12}(s) = -A\left(-s - \frac{1}{2}(s^2 + f(s))\right) - B(s^3 - f(s)) + f(s), \quad (2.37)$$

with

$$f(s) = (\alpha + 2)(-s)^{\alpha+1} - (\alpha + 1)(-s)^{\alpha+2}. \quad (2.38)$$

The parameters A and B can be derived if the user prescribes the surface and bottom cell thickness. The parameter α controls the shape of the stretching.

To avoid unnecessary fine cells at the surface of shallow regions, unstretched s coordinates or z coordinate are applied in region shallower than a critical depth h_c , both for SH94 and SF12 stretching.

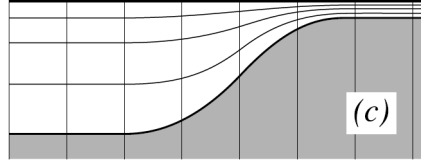


Figure 2.3. Example of 5 s levels on a sloping bathymetry coming from [5].

The main drawback of terrain following coordinates is that the isolevel surfaces can be very steep at places where the bathymetry is steep. This can cause truncation errors in the discretized horizontal pressure gradient and can lead to spurious currents [8], [9]. One way to tackle this problem is to replace the bathymetry $H(i, j)$ by a smoother and deeper bathymetry. The stiffness of the bathymetry is defined by

$$r_{bath} = \left| \frac{\delta H}{2\bar{H}} \right|, \quad (2.39)$$

where δH is the maximum depth difference between a bathymetry point and its neighbors. $\bar{H}$ is the average depth of the cell and its neighbors. The smoothing procedure finds each points where the stiffness r_{bath} exceeds a threshold r_{max} and iteratively deepens those points and their neighbors till reaching a smooth enough bathymetry. Since the new bathymetry is deeper than the original one, some levels are deeper than the true bathymetry and are then taken off the domain, leading to a partially stair-like representation of the topography. One possible smoothing algorithm is presented in [10]. Figure ?? and ?? shows $SF12$ s coordinates both with and without smoothing.

Similarly to the bathymetry stiffness, a grid stiffness can be computed at each grid point and in each horizontal direction with³

$$r_{grid} = \left| \left(\frac{\partial z_0}{\partial i} \right) \left(\frac{\partial z}{\partial k} \right)^{-1} \right| \quad (2.40)$$

which can be also computed in the j direction. For an unstretched s coordinate system, this stiffness is the same as r_{bath} but it is not the case for more general stretching functions.

Multi-envelope coordinates

The terrain following s coordinates represent well the bathymetry of the domain, but all the vertical levels are constrained to follow the bathymetry, even the upper levels where the flow is less influenced by the topography. This is a flaw of s coordinates because the upper part of the column of water usually moves along the geopotential and isopycnal surfaces meaning that z coordinates should be better in the upper layer. This can be achieved by having a mixed $z - s$ coordinates, and is possible since the addition of the multi-envelope coordinate system in NEMO v5.0.

The multi-envelope coordinate system (MES) was developed by Diego Bruciaferri in [11]. The basic idea is to have multiple layers of s coordinates system, each following a user defined surface called an envelope bathymetry. This surface can be the true bathymetry, a smoothed bathymetry or a flat surface for example. Then, one can define a first lower layer following the true bathymetry, giving s level on the lower layer and an upper layer following a flat surface, giving z levels. An example of a MES discretization is shown in Fig. 2.4 where 5 envelopes are used.

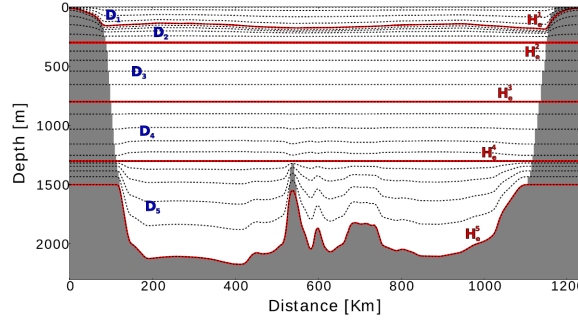


Figure 2.4. Example of a multi-envelope coordinate system on a varying bathymetry, with 5 envelopes. The red lines are the envelope bathymetries. The dashed lines are the levels. The dark gray shading is the model bathymetry. Figure coming from [11].

³This formulas was found by reading the NEMO code `domwri.F90` and could not be found in this precise form in the litterature.

2.2.3 Spatial discretization

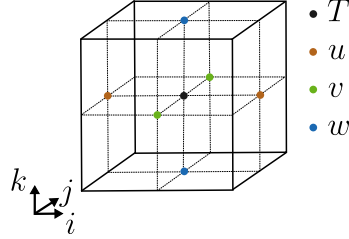


Figure 2.5. Grid cell in the mesh space (i, j, k) with the position of the speed and tracer variables according to the Arakawa C grid.

NEMO discretizes its primitive equations using a finite difference centered scheme on a staggered Arakawa C grid [12] shown in Fig. 2.5. The tracer points (salinity and temperature) are located at the center of each cell and referenced with integer (i, j, k) indices. Each speed components is located at the faces' center of the cell and thus have half integer indices.

Each differential operator appearing in the primitive equations can then be expressed in terms of the two basic discrete operators

$$\delta_i[q] = q(i + 1/2, j, k) - q(i - 1/2, j, k) \quad (2.41)$$

$$\bar{q}^i = \frac{1}{2}(q(i + 1/2, j, k) + q(i - 1/2, j, k)). \quad (2.42)$$

Both operators can also be taken in the j and k direction. The first one is the centered finite difference for order 1 derivatives. Because the grid is staggered, interpolation is sometimes needed to match the position of different quantities. This interpolation can be done by the averaging operator $\bar{q}^i$ or by more advanced advection schemes.

2.2.4 Time discretization

Leapfrog scheme

The general time stepping environment to discretize first order time derivatives in NEMO is a Leapfrog scheme. First, each prognostic equation (for the velocity, SSH and tracers) can be written in a general form

$$\frac{\partial q}{\partial t} = RHS(t) + RHS_{diff}(t), \quad (2.43)$$

where q is replaced by u, v, η, T or S . RHS and RHS_{diff} are two right-hand side terms depending both on space and the other modelled quantities. Here RHS_{diff} is a vertical diffusion term

$$RHS_{diff}(t) = \frac{\partial}{\partial k} \left(A^{vq} \frac{\partial q}{\partial k} \right), \quad (2.44)$$

with A some diffusivity constant. This term imposes the strongest stability constraint on the time step and should then be treated implicitly. Finally, the Leapfrog difference equation writes

$$q^{t+\Delta t} = q^{t-\Delta t} + 2\Delta t(RHS^t + RHS_{diff}^{t+\Delta t}), \quad (2.45)$$

where Δt is a time step and the superscript $()^t$ denotes the approximated quantity at time t . Since the RHS depends on the quantity $q(i, j, k)$ and its neighbors, at the time step $t + \Delta t$, a linear system must be solved to obtain the iterated a time $t + \Delta t$.

Split-explicit procedure

In the ocean, several phenomena occur at several time scales. Different types of waves for example travel at different speeds. Barotropic waves (free-surface gravity waves with uniform vertical structure) travel at a phase speed $\sqrt{gH}$ with g the gravitational acceleration and H the depth. Baroclinic waves, internal waves due to the stratification of the sea, have on the contrary a much slower phase speed.

In-fine, the time step should be chosen based on the phenomena modelled by the primitive equations that have the smallest timescale. Barotropic waves, governed by the free surface equation 2.5 and the depth averaged momentum, imposes then the strongest constraint on the time step.

One way to overcome this constraint is to split the flow into a depth independent barotropic mode which vary quickly and a baroclinic mode varying slowly. At each baroclinic time step, sub-iterations (usually 30) are made to solve the 2D barotropic mode. The latter is described by

$$\frac{\partial \eta}{\partial t} = -\nabla_h \cdot ((H + \eta)\mathbf{U}) + q_w \quad (2.46)$$

$$\frac{\partial \mathbf{U}}{\partial t} = -\mathbf{f} \wedge \mathbf{U} - g\nabla_h \eta - \nabla_h \Phi_t - \frac{c_b}{H + \eta} \mathbf{U} + \mathbf{G}. \quad (2.47)$$

The first equation is the free-surface equation with $q_w = P + R - E$ the precipitation/river/evaporation balance. The second equation is based on the depth integrated momentum equation, and describes the evolution of the barotropic velocity. The first terms on the right-hand side are the Coriolis acceleration, the surface pressure gradient, the tidal force and a bottom drag force. Those are the dominant forcing terms of the barotropic motion. Then, there is the term $\mathbf{G}$ containing the vertical average advection term, SGS and extra forcing terms like the wind stress. The term $\mathbf{G}$ is held constant during the barotropic iteration and is basically a coupling term between the baroclinic and barotropic motion.

Once the barotropic equations are solved till the next baroclinic time step $t + \Delta t$, the barotropic trend can be put in the barotropic equations (through the horizontal pressure gradient) and the baroclinic equations can be advanced to the new time step to obtain the speed field. The tracer equations are finally time stepped using the speed field at time step $t + \Delta t$.

2.2.5 Horizontal SGS parametrization

For the lateral SGS terms, it is necessary to recall the definition of the laplacian for a scalar and a vector quantity:

$$\begin{aligned}\Delta a &= \nabla \cdot (\nabla a) \\ \Delta \mathbf{A} &= \nabla(\nabla \cdot \mathbf{A}) - \nabla \wedge (\nabla \wedge \mathbf{A}),\end{aligned}\tag{2.48}$$

where Δ is the laplace operator. a and $\mathbf{A}$ are a scalar and a vector valued functions. The SGS terms can be built base on those operators:

$$\begin{aligned}D^{lU} &= \nabla_h(A^{lm}\nabla_h \cdot \mathbf{U}) - \nabla_h \wedge (A^{lm}\nabla \wedge \mathbf{U}) \\ D^{lT} &= \nabla(A^{lt}\mathcal{R}\nabla T) \\ D^{lS} &= \nabla(A^{lt}\mathcal{R}\nabla S).\end{aligned}\tag{2.49}$$

A^{lm} and A^{lt} are the lateral turbulent viscosity and diffusivity. The lateral momentum diffusion acts only on the horizontal plane, as shown by the use of the horizontal gradient ∇_h . For the tracer diffusion, a rotation matrix

$$\mathcal{R} = \begin{pmatrix} 1 & 0 & r_1 \\ 0 & 1 & r_2 \\ r_1 & r_2 & r_1^2 + r_2^2 \end{pmatrix}\tag{2.50}$$

is used with r_1 (resp. r_2) the slope between the vertical levels and the chosen diffusion surface in the $\mathbf{i}$ (resp. $\mathbf{j}$) direction. The user can then choose to diffuse the tracers on the levels ($r_1 = r_2 = 0$) but also on other planes such as geopotential or isopycnal surfaces.

Instead of using a laplacian diffusion operator, it is possible to apply it twice. This is possible using the bilaplacian operator. In this case, the diffusion constant is the square root of the eddy viscosity/diffusivity (in m^4/s). This operator is known to have a stronger scale selectivity because it diffuses more efficiently short scales disturbances than the laplacian operator. Indeed, for a simple infinite 1D diffusion problem with an initial condition being a sinus of wavelength λ , the characteristic time (or damping time) will scale as λ^2 for the laplacian diffusion and λ^4 . This can be shown by a simple scaling argument:

$$\underbrace{\frac{\partial q}{\partial t}}_{\sim \bar{q} T^{-1}} = \underbrace{\nu_{lp} \frac{\partial q}{\partial x^2}}_{\sim \bar{q} \nu_{lp} \lambda^{-2}} \iff T \sim \nu_{lp} \lambda^2\tag{2.51}$$

$$\underbrace{\frac{\partial q}{\partial t}}_{\sim \bar{q} T^{-1}} = \underbrace{\nu_{blp} \frac{\partial q}{\partial x^4}}_{\sim \bar{q} \nu_{blp} \lambda^{-4}} \iff T \sim \nu_{blp} \lambda^4,\tag{2.52}$$

where T is the characteristic time. q and $\bar{q}$ are a given quantity and its typical scale. ν_{lp} and ν_{blp} are the laplacian and bilaplacian diffusion constant (here constant). λ is the

length scale.

Horizontal turbulent diffusivities

On the horizontal, the diffusion is principally there to filter out spurious modes and prevent instability. The viscosity and diffusivities should be chosen as small as possible to limit artificial diffusion while being enough to have a stable model. Here, the viscosity and diffusivity are chosen constant and expressed in terms of a characteristic length scale and speed with

$$A^{lX} = \begin{cases} \frac{1}{2}U_{scl}L_{scl} & \text{for laplacian mixing} \\ \frac{1}{12}U_{scl}L_{scl}^3 & \text{for bilaplacian mixing} \end{cases}, \quad (2.53)$$

where L_{scl} and U_{scl} are a characteristic length scale and speed. X is m for the viscosity and t for the diffusivity.

2.2.6 Boundary conditions

The formulation of boundary conditions is a key choice when designing an ocean circulation model. In this case there are four interfaces where a boundary condition should be provided.

First, there is the air-sea interface where momentum, heat and water are exchanged. The wind drives the sea surface by the interface friction, and it enhances heat exchange in between the water and the atmosphere by adding a forced convective heat transfer. Water is also exchanged with a net flux given by the balance of evaporation and precipitation.

At the solid bottom and lateral boundaries, the land acts as a barrier that the flow cannot penetrate. Friction, particularly at the bottom, acts as a drag force that slows down the flow. Water can also be brought at the land-sea lateral interface in river mouths.

Finally, in regional configuration such as in this thesis, there are open boundaries where the inflow or outflow of water and the tracer concentrations must be specified.

Air-sea interface

At the sea surface, momentum, heat and water are exchanged. Those fluxes can be expressed by empirical formulations called bulk formulae.

The bulk formula for the momentum flux is a wind stress, expressed as

$$\boldsymbol{\tau} = \rho_{air}C_D\mathbf{U}_z|\mathbf{U}_z|, \quad (2.54)$$

with $\boldsymbol{\tau}$ the wind stress, ρ_{air} the air density, C_D a drag coefficient and $\mathbf{U}_z$ the horizontal wind velocity vector at a reference height z above the sea surface. This wind stress is applied at the boundary as

$$\boldsymbol{\tau} = A^{vm}\frac{\partial \mathbf{u}}{\partial z}\bigg|_{z=\eta}. \quad (2.55)$$

For the temperature equation, a heat flux must be provided. The latter is composed of a solar heat flux Q_{sr} that is transmitted in depth and a non-solar heat flux Q_{ns} directly transmitted at the surface.

The solar heat flux can be expressed in terms of the incoming shortwave radiation (Sun irradiation)

$$Q_{sr} = (1 - a)Q_{sw}, \quad (2.56)$$

with a an albedo coefficient quantifying the amount of reflected shortwave heat. Q_{sw} is the prescribed shortwave radiation. The solar heat flux is added to the temperature equation by mean of an irradiance $I(z)$ in the source term

$$Q^T = \frac{\partial I(z)}{\partial z}, \quad (2.57)$$

where the irradiance is such that $I(\eta) = Q_{sr}$. Several formulations exist to describe the absorption of light, taking into account the absorption of different light bands and possibly considering the amount of particles and chlorophyll in the water.

The non-solar heat flux Q_{ns} is a sum of three processes: the sensible heat flux Q_H , the latent heat Q_L and the net longwave (infrared radiation) heat flux Q_{ir} .

The sensible heat flux is a convective and conductive heat exchange between the air and water, enhanced by the wind. It writes

$$Q_H = \rho_{air} c_{p,air} C_H (\theta_z - \theta_s) |\mathbf{U}_z|, \quad (2.58)$$

where C_H is an heat transfer coefficient and $c_{p,air}$ is the specific heat of moist air. θ_z and θ_s are the air potential temperature at a reference height z and the potential sea surface temperature.

The latent heat flux is the heat taken to the water when it evaporates. It writes

$$Q_L = -L_v E, \quad (2.59)$$

with L_v the latent heat of vaporization and E the evaporation. The latter is expressed by the bulk formula

$$E = \rho_{air} C_E (q_z - q_s) |\mathbf{U}_z|, \quad (2.60)$$

with C_E an evaporation coefficient and where q_z is the specific humidity at height z and q_s is the saturation specific humidity of air.

Finally, the net radiative heat flux is a balance between the incoming longwave heat flux (radiation emitted by the atmosphere) and the heat radiated by the sea such that

$$Q_{ir} = \delta(Q_{lw} - \sigma T_s^4), \quad (2.61)$$

where δ is the emissivity of the sea surface, Q_{lw} is the incoming longwave heat flux and $\sigma = 5.67 \cdot 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$ is the Stefan-Boltzmann constant. T_s is the *in-situ* sea surface temperature.

In the equation above, the quantities Q_{sw} , Q_{lw} , θ_z , q_z and $\mathbf{U}_z$ must be provided by the user and come most of the time from an atmospheric model. The quantities θ_s , T_s and A^{vm} are derived for the ocean model. There are the bulk coefficients C_D , C_H , C_E and a left that should be expressed. Several formulations exist like MFS, NCAR. In the present work, the ECMWF formulation will be used [13].

In principle, the boundary conditions should be updated at each time step. The limited storage capacity and high computational cost of IO (Input Output) operations makes it impossible to store the surface boundary conditions at each time step. Instead, the surface boundary conditions are stored with an hourly interval. NEMO then interpolates the value at each time step for its needs. Hourly forcings are sufficient to reconstruct most of the variability of the forcings relevant for the large scale and slow motion of the sea. The diurnal cycle for the heat flux is also well recovered. High frequency phenomenon like waves forced by the wind gust are effectively filtered out and are not represented by the model.

Seabed interface

At the seabed, a non-permeability of the fluid must be imposed. Here it implies that the vertical velocity (velocity along $\mathbf{k}$, no matter the coordinate system employed) is set to zero on the deepest w unmasked point.

The condition for the horizontal velocity is a Neumann type

$$A^{vm} \frac{\partial \mathbf{u}}{\partial z} \Big|_{z=-H} = \mathcal{F}, \quad (2.62)$$

with $\mathcal{F}$ a friction law. The friction law can be for example quadratic:

$$\mathcal{F} = C_D |\mathbf{u}_b| \mathbf{u}_b \quad (2.63)$$

with $\mathbf{u}_b$ the bottom horizontal velocity and C_D a bottom drag coefficient. The latter can be chosen constant or can change with the height using of the grid point, based on a log-layer law. In this case, the drag coefficient is not constant but given by

$$C_D = \frac{\kappa}{\log(z/z_0)}, \quad (2.64)$$

with $\kappa = 0.4$ the von Kármán constant, z_0 a roughness length and z the height of the u and v point just above the bottom.

For the salinity and temperature, bottom fluxes are assumed null, leading to

$$A^{vT} \frac{\partial T}{\partial z} \Big|_{z=-H} = 0 \quad A^{vS} \frac{\partial S}{\partial z} \Big|_{z=-H} = 0 \quad (2.65)$$

Solid lateral boundaries

At solid lateral boundaries, there is again a non-permeability condition and no tracer flux. For a boundary in the south-north direction, it writes

$$u = 0 \quad A^{iT} \frac{\partial T}{\partial i} = 0 \quad A^{iS} \frac{\partial S}{\partial i} = 0, \quad (2.66)$$

and similarly on an east-west boundary

$$v = 0 \quad A^{iT} \frac{\partial T}{\partial j} = 0 \quad A^{iS} \frac{\partial S}{\partial j} = 0. \quad (2.67)$$

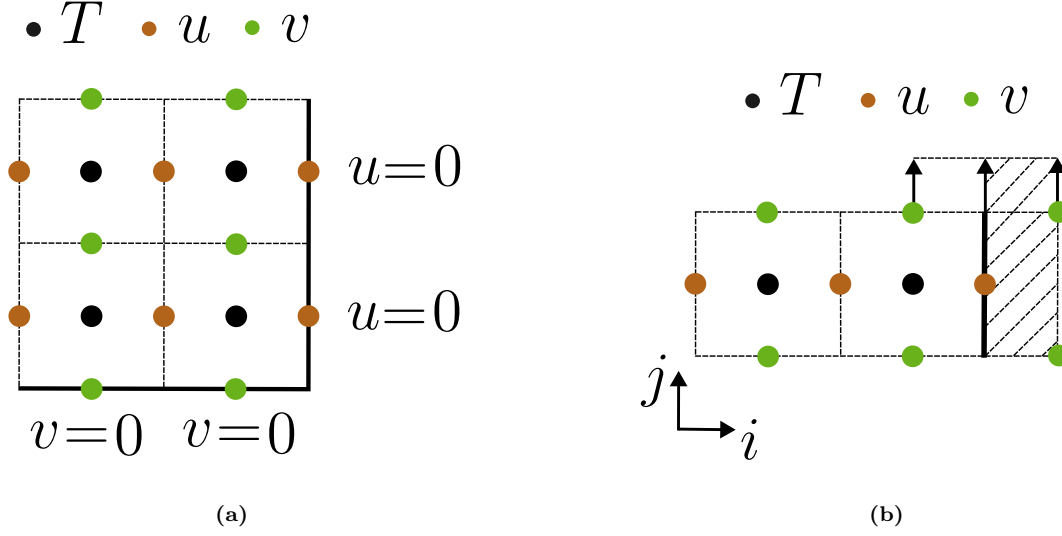


Figure 2.6. Non-permeability and slip boundary condition applied on a staggered Arakawa C-grid. The solid line indicates the solid boundary where both u and v should be computed. For a meridional boundary, the normal, u speed is set to 0, and the tangent, v speed is set to the offshore velocity. The slip boundary is effectively a zero gradient condition across the solid boundary.

For the tangential velocity, a slip condition is applied, meaning that the derivative of the tangential velocity is null:

$$\begin{aligned} \frac{\partial v}{\partial i} &= 0 \text{ for a south-north solid boundary} \\ \frac{\partial u}{\partial j} &= 0 \text{ for an east-west solid boundary} \end{aligned} \quad (2.68)$$

It means that at the boundary, the tangential speed is equal to the offshore tangential velocity (Fig. 2.6(b)).

Open lateral boundaries

At open boundaries, the tracers concentration, barotropic and baroclinic velocities must be provided. In this work, those fields are provided by a lower resolution models. In this case, simple Dirichlet condition are not suited for this kind of application since some processes are unresolved by the low resolution, giving a mismatch between the low and high resolution configuration. The boundary condition should ideally be able to let escape waves due to the initial condition and finer resolved processes coming from the nested domain [14].

One method is called the Flow Relaxation Scheme (FRS) [14], [15]. The principle is to impose the boundary condition not only at the boundary but also inside an inner layer in the domain. A discrete distance d is defined on each point of the Rim layer (Fig. 2.7).

Then, for each point of the layer, the boundary condition is applied as

$$\Phi(d) = \alpha(d)\Phi_e(d) + (1 - \alpha(d))\Phi_m(d) \quad (2.69)$$

$$\alpha(d) = 1 - \tanh\left(\frac{d-1}{2}\right), \quad d = 1, \dots, N, \quad (2.70)$$

where $\Phi_e(d)$ and $\Phi_m(d)$ are respectively the external field provided by the low resolution model and the field computed by the high resolution model. N is the number of layers in the relaxation zone. $\alpha(d)$ is a weight decreasing from 1 at the boundary to 0 far away from it. Because of this weight, the boundary condition is fully specified by the external model at the boundary and is progressively damped away from the boundary. This boundary condition can be employed for any modelled quantity and here will be used for the salinity and temperature. Usually, the relaxation area has about $N = 10$ layers.

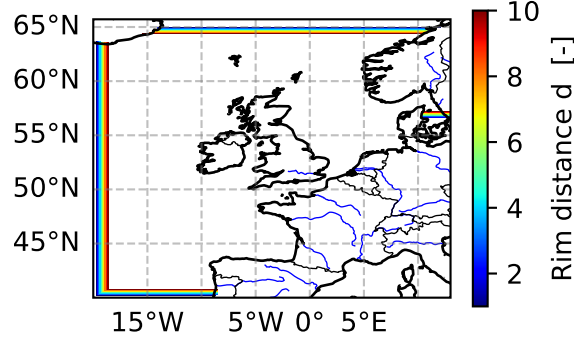


Figure 2.7. Open Atlantic and Baltic boundary for the AMM7 configuration of the North Sea. Each point is a T point with a color expressing the rim distance d from the boundary. $N = 10$.

For the barotropic velocities, the Flather condition [16] can be applied. This scheme prevents reflection of barotropic gravity wave by applying a correction on the barotropic velocity normal to the boundary. This is done using the gradient of the SSH as

$$U = U_e + \sqrt{\frac{g}{h}}(\eta - \eta_e), \quad (2.71)$$

where U is the applied barotropic normal velocity and U_e the one provided by the external model. h is the water column height at the boundary. η and η_e are the modelled SSH next to the boundary and the SSH given by the lower resolution model.

A last, much simpler boundary condition is a zero normal gradient at the boundary. This is unsuited for barotropic velocities since the barotropic motions contains most of the waves that should escape the domain but can be applied to the baroclinic velocities.

Because the tidal waves incoming from the lateral boundaries are among the most important forcing for coastal models, at least an hourly forcing is necessary to reconstruct correctly the tidal signal at the boundary. Instead of providing an hourly signal, it is possible to provide to the model the tidal amplitudes. The model is then able to reconstruct the tidal signal at its boundary for each time step.

River forcing

Strictly speaking, NEMO does not treat rivers has a special boundary condition. Instead, river discharge is added a source of mass in the continuity equation Eq. 2.4, leading to the new continuity equation

$$\nabla \cdot \mathbf{v} = q_r, \quad (2.72)$$

where q_r is a river field varying in the 3D space and time expressed by a flow rate per unit of volume $[(\text{m}^3/\text{s})/\text{m}^3]$. For a given river mouth, one can take its flow rate Q and choose a grid cell where to impose this source. Then, this flow rate can be divided by the area of the grid cell, giving the quantity

$$R = Q/A, \quad (2.73)$$

where A is the grid cell area. Finally, a depth profile for the q_r can be chosen. The default choice in NEMO is to impose the runoff at the surface cell, leading to

$$q_r = \frac{R}{e_{3,top}}, \quad (2.74)$$

inside the top grid cell and zero elsewhere. $e_{3,top}$ is the surface cell thickness. Another common choice is to spread the runoff evenly over depth as

$$q_r = \frac{R}{h} \quad (2.75)$$

over all the water column, where h is the water column height.

The river runoff is also added in the free surface equation Eq. 2.5.

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CHAPTER 3

Validation framework

This chapter describes the tools and the method used to analyze and validate both the low resolution NEMO configuration covering the whole NWES and the high resolution configuration of the SBNS and BCZ.

3.1 Validation method

The models presented in this study are analyzed, compared and validated using the method of the Global Ocean Data Assimilation Experiment, presented in [1]. This method is also followed by the Copernicus Marine Environment Monitoring Service (CMEMS) within its Quality Information Documents (QUIDs) and its Product Quality Dashboard [2]. This framework aims to assess the quality of a model using both a consistency assessment and a quantitative assessment. The first one compares the model outputs with our current qualitative knowledge of the sea while the second compares the results with systematic measurements. To do that, four classes of metrics are defined.

The class 1 metrics consist of gridded data interpolated on a reference grid containing the basics outputs of the model. In the present work this metrics contains the velocities, the salinity, the temperature and the SSH. For 3D data, the data is interpolated on slices at fixed depths, here 0, 30, 50, 100, 200, 400, 700, 1000, 1500, 2000, 2500, and 3000 m. The data, interpolated on the common grid and depths can then be compared to other such models and climatologies. The data can also be compared to satellite observations that have been interpolated on the common grid (L4 CMEMS products for example). In this work, class 1 analysis will be used mainly qualitatively to check if the model is consistent with our current knowledge of the North Sea. Comparisons with satellite measurements, interpolated on the model grid fall also in this class.

The class 2 metrics are again based on the basic variables of the model but focus on more precise locations than the whole domain. They consist of extracted profiles in specific regions and transects in particular sections. Extracted profiles and time series can also be compared to moorings data.

Class 3 metrics consist of quantities derived from the model basic variable. Here volume transports across some sections fall in this class.

Class 4 refers to data interpolated on observations, except moorings that are included in class 2. Observation include satellite measurements, ARGOS, gliders etc. This class is

used to assess the performance of the model to the "sea truth". In this thesis, class 4 metric will involve comparison of the model interpolated on CORA profiles [3].

In order to quantify the quality of the model against measurements, a set of statistics is defined. Those statistics give each a scalar quantity assessing the overall performance of the model.

The first employed statistic is the bias, measuring the average error of the model. It is defined by

$$Bias = \frac{1}{N} \sum_{k=1}^N (x_k^{mod} - x_k^{obs}), \quad (3.1)$$

where N is the number of measurements, x_k^{mod} the quantity predicted by the model and x_k^{obs} . k is an index representing either the time of the measurement, its vertical or its horizontal position. It depends if the compared data is a time series, a vertical profile or an horizontal grid.

The Root Mean Square Error (RMSE) is a measure of the quadratic error or the dispersion of the error. It writes

$$RMSE = \frac{1}{N} \sum_{k=1}^N ((x_k^{mod})^2 - (x_k^{obs})^2). \quad (3.2)$$

The correlation quantify how two variables evolve together. In case of a modelled and observed variable, both should evolve in the same way and thus have a correlation of 1. The correlation coefficient is defined by

$$Corr = \frac{\sum_{k=1}^N (x_k^{mod} - \bar{x}_k^{mod})(x_k^{obs} - \bar{x}_k^{obs})}{\sqrt{\sum_{k=1}^N (x_k^{mod} - \bar{x}_k^{mod})^2 \sum_{k=1}^N (x_k^{obs} - \bar{x}_k^{obs})^2}}, \quad (3.3)$$

with $\bar{\cdot}$ the average quantity

$$\bar{x} = \frac{1}{N} \sum_{k=1}^N x_k. \quad (3.4)$$

Using both the bias and the average, the percentage bias is defined by

$$PB = \frac{Bias}{\bar{x}^{obs}}. \quad (3.5)$$

According to [4] and [5], a correlation coefficient above 0.85 and a percentage bias smaller than 10% characterizes an excellent model. These numbers must be taken with care because they are still subjective. Depending on the application of a model, a bias larger can be acceptable or not. For example, for a climatic simulation characterizing the SST and the SSH trend, biases much smaller than 10% are wanted. On the contrary, if the model is just supposed to force a biogeochemical model for which temperature has a smaller effect, larger bias are acceptable.

To compare both a measured and modelled quantity, their position and their time should match. On the horizontal, the model grid cell nearest to the measurement is chosen. On the vertical, when comparing profiles, the observed profile is interpolated on the model grid using a cubic interpolation method.

3.2 Remote sensing and *in-situ* measurements

To assess the quality of the presented models of the North Sea and the SBNS, their outputs are compared to both remote and *in-situ* observations, each coming from a CMEMS product.

The SST will be compared to satellite measurements using the CMEMS L4 satellite product [6] covering the NWES and the Biscay Bay during the period 1982-2025. It provides a daily gap-free SST on a grid with a resolution of 0.05° or about 5.56 km.

To analyze the quality of modelled vertical profiles, *in-situ* profiles coming from the CMEMS CORA [3] dataset is used. This dataset gather temperature and salinity profiles from various instruments such as profiling float, Conductivity Temperature Depth (CTDs) sensors, gliders ... A quality control and adjustments have already been made in the dataset to provide the best measurements. Here, profiles with a quality flag of 1 meaning "Good data" and at least 10 valid vertical levels are used. Figure 3.1 shows the seasonal and spatial distribution of temperature profiles in 2007, the modelled year. It is shown that the amount of profiles in the SBNS is small, and that there is a small discrepancy in time with fewer profiles in summer and fewer profiles on the shelf in winter.

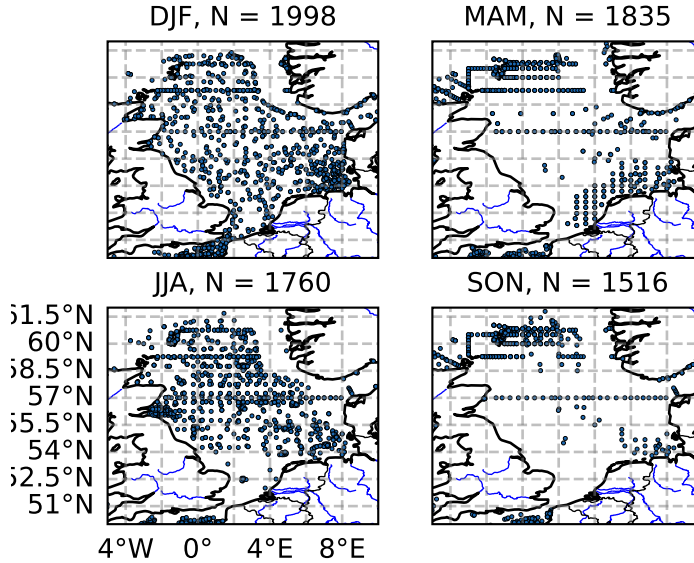


Figure 3.1. Spatial and seasonal distribution of CORA [3] temperature profiles on the North Sea domain in 2007. N is the number of profile for each season. Each profile has a quality flag 1 meaning "Good data" and has at least 10 vertical levels.

Finally, the tidal dynamics is assessed using tidal gauges from [7]. The tidal amplitude and phase of each constituent can be obtained from the time series by performing a multi-linear regression analysis (App. A) and then compared to the model. The tidal currents are assessed with the mooring platforms "BolVanHeistMP" and "ScheurWielingenMP" given by the CMEMS product [8].

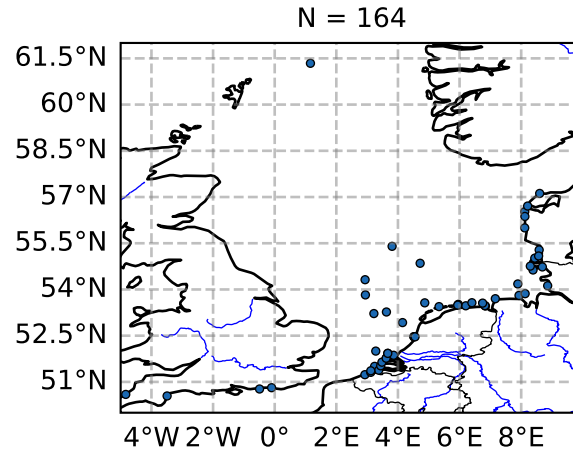


Figure 3.2. Spatial distribution of tidal gauges coming from the CMEMS product [7].

3.3 Model output

For each simulation, the model outputs daily averaged fields for temperature, salinity, barotropic and baroclinic currents and SSH. For qualitative analysis, the data is averaged monthly or seasonally and examined on its original grid.

When comparing with satellite images, both observation and model outputs are averaged monthly or seasonally. The satellite data is interpolated on the model grid using a bicubic method.

Profile are compared with model output at the same date. On the horizontal, the model grid cell closest to the observation is chosen. On the vertical, the observation is interpolated on the model vertical grid with bicubic interpolation.

In the tidal analysis, the model saves the SSH and barotropic velocities each hour. The tidal gauges measurements are compared to the closest model grid cells.

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Part III

Results

CHAPTER 4

The low resolution model

In this chapter, the AMM7 NEMO configuration developed by the English Met Office is first described. The various physical quantities derived by the model are studied and compared to our current knowledge of the North Sea (Sec. 1.2). Finally, the accuracy of the model is assessed by comparing its output to observations.

4.1 Configuration description

The AMM7 configuration is a NEMO configuration covering the entire NWES and a small part of the North Atlantic around the shelf. It was first described in [1] and got several upgrades in the following years. The last described version of the configuration can be found in [2] and [3]. The most recent version, used here, is present in the AMM7 repository [4] and forcing files were taken from the link <https://gws-access.jasmin.ac.uk/public/jmmp/AMM7> provided in the associated Github repository. The unique change done in this work is the update of NEMO from v4.2.1 to v5.0. All the numerical schemes, parameters, parametrizations and boundary conditions were kept unchanged when switching to v5.0¹.

4.1.1 Domain and mesh

The domain extends from 40° N to 65° and from 20° W to 15° E. There are 375 cells along the latitude and 297 along the longitude, leading to a resolution of approximately 7 km. This resolution is finer than the off shelf internal Rossby radius (about 14 km) but larger than the shelf radius (2 - 5 km) [5]²

The vertical discretization shown in Fig. 4.1 uses the multi-envelope coordinate system (Sec. 2.2.2) with two envelope bathymetries. The upper layer is composed of 26 *s* levels following the bathymetry till a maximum depth of 250 m where the levels become flat. The lower layer contains 25 *SF12 s* levels following the bathymetry deeper than 250 m. Then, on the shallow shelf there are always 26 levels following the topography while off

¹The migration toward NEMO v5.0 is not strictly necessary for this chapter. It is because the MES coordinate system is not available in the official NEMO v4.2.1 repository while it is needed to create the domain of the high resolution model in Chapter 5.

²The internal Rossby deformation radius is here based on a two-layers system. $R = \frac{1}{f} \sqrt{\frac{g' H_1 H_2}{H_1 + H_2}}$ where $g' = g \frac{\rho_2 - \rho_1}{\rho_0}$ is the reduced gravity. ρ_0 , ρ_1 and ρ_2 are the reference density, the density of the upper layer and the density of the second layer. H_1 and H_2 are the height of both layers. f is the Coriolis parameter.

the shelf there are 25 terrain following levels and 26 z levels. The envelope bathymetry of both layers is smoothed to avoid steep cells with $r_{max} = 0.1$ for the deep layer and $r_{max} = 0.07$ for the upper layer. The true bathymetry is extracted from GEBCO 15 arc-second product [6].

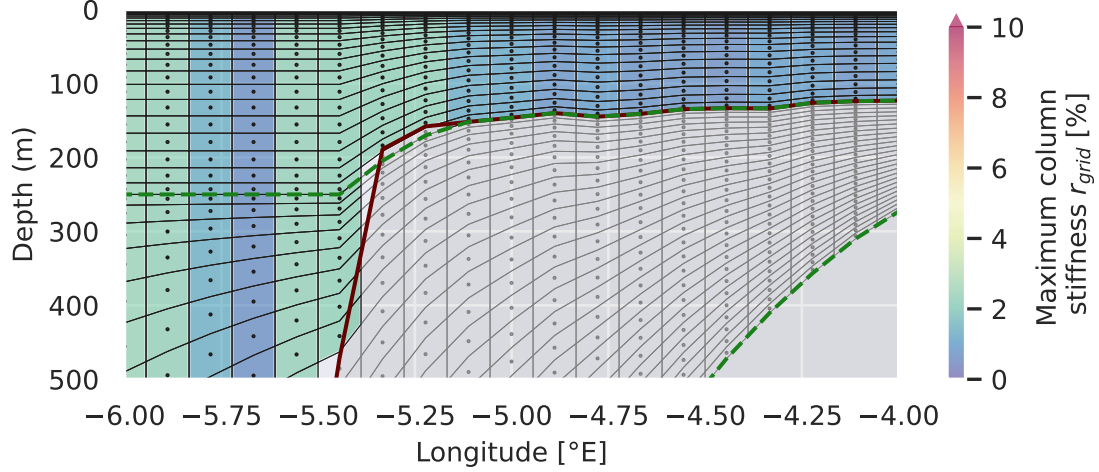


Figure 4.1. Vertical discretization of the AMM7 configuration at the shelf break (here at 47°N). The black points indicate the position of T points and the boxes are the cells boundaries. The color indicates the maximum value of the stiffness r_{grid} for each column and gray cells represent masked cells. The red line is the true bathymetry, and green lines indicate the envelope bathymetries. This chart has been made thanks to the Python code present in the AMM7 repository https://github.com/JMMP-Group/CO_AMM7/tree/v9.0-alpha/scripts/analysis.

On the time domain, the year 2005 to 2007 (date for which the forcings are available) are modelled. According to [7], on the same domain with a similar resolution, 15 months of simulation are needed to get rid of the unperfect initial condition. Here the years 2005 and 2006 are considered as spin up years and the last year is analyzed to get one seasonal cycle. The baroclinic time step is $\Delta t = 150$ s and there are 30 barotropic sub-iterations. The system is launched from rest with flat initial salinity and temperature profiles provided in the forcing files ³.

4.1.2 Forcings

On the air-sea interface, the ERA5 reanalysis [9], [10] provides the wind speed at 10 m, the saturation humidity at the surface, the air temperature at 2 m, the long- and shortwave surface heat fluxes, the precipitations and the surface atmospheric pressure. The ECMWF [11] bulk formulas are employed with cool-skin and warm-layer parametrizations.

There are two lateral open boundaries: the large North Atlantic boundary around the shelf and the small connection to the Baltic through the Kattegat. Both are forced daily by the model GloSea5 [12]. The FRS scheme is applied on tracers, the Flather condition on barotropic velocities and a zero gradient condition on baroclinic velocities as explained in Sec. 2.2.6.

³There is no information on the repository telling the source of this initial dataset. It could come from a run of the FOAM 1/12° North Atlantic model [8].

The river forcing is provided daily by a climatology coming from [13], [14] and [15].

The tidal constituents M2, S2, T2, K2, L2, N2, Mu2, Nu2, 2N2, Q1, O1, S1, P1, K1 and M4 are considered. The tidal potential is updated at each model iteration. The tidal elevation and currents are added to the lateral barotropic forcings which contain only the residual signal. The tidal amplitudes at the boundaries are provided by the tidal model FES [16].

4.1.3 Miscellaneous parameters

The horizontal momentum diffusion uses a bilaplacian operator along the vertical s levels. The horizontal eddy viscosity is given by Equ. 2.53 using a characteristic velocity of 0.12 m/s and length of 10 km giving an eddy viscosity of $10^{10} \text{ m}^4/\text{s}$.

For the tracers diffusion, a Laplacian operator is applied on the geopotential surfaces. The diffusion should indeed be on the geopotential surfaces and not on the vertical levels since spurious vertical mixing could appear in steep regions [?]. The horizontal eddy diffusivity is obtained in the same way with a characteristic speed of 0.01 m/s and length of 10 km leading to a diffusivity of $50 \text{ m}^2/\text{s}$.

For the vertical physics, the GLS $k - \epsilon$ scheme is used with Canuto A stability functions.

For the tracer advection, a Flux Centered Scheme (fourth order on the horizontal and second order on the vertical) is employed. The Coriolis and vorticity term use an energy plus enstrophy conserving scheme. The horizontal pressure gradient uses a Density Jacobian formulation with a density profile recovered from a cubic spline.

4.2 Results

4.2.1 General residual circulation

The circulation of the NWES is shown in Fig. 4.2. The strongest current is the slope current along the continental slope which is a known feature of eastern slopes on Earth, induced by the Joint Effect of Baroclinicity And Bottom Relief (JEBAR) [17]. On the shelf, the cyclonic, weaker circulation is clear, with a stronger coastal current on the eastern part of the basin. Atlantic water enters the shelf trough three main sections: the Fair Isle current (FIC), the East Shetland Current (ESC) and the Western Skagerrak Current (WSC). The Dooley current, leaving the FIC and following the 100 m isobath is also present. Finally, the water leaves the shelf by joining the Skagerrak and following the NCC.

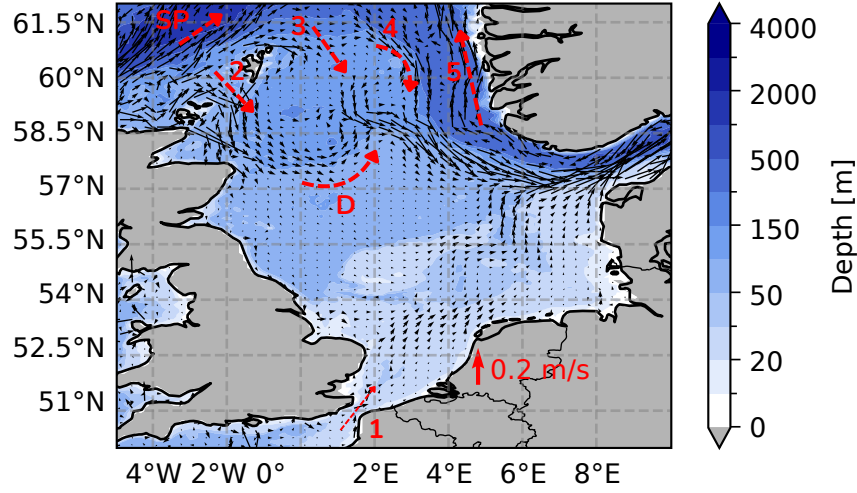


Figure 4.2. Bathymetry of the domain and vector field of the barotropic circulation. The velocity is averaged over the whole year 2007. For clarity, the vectors are plotted for one every third grid point. The red dashed arrows show the main currents. 1: Dover, 2: Fair Isle current (FIC), 3: East Shetland current (ESC), 4: Western Skagerrak current (WSC), 5: Norwegian coastal current (NCC), SP: Slope current, D: Dooley current.

To quantify those currents, the flow rate of water is computed across several sections shown in Fig. 4.3 and Table 4.1. The values found in the literature can vary strongly. Long term measurements are lacking and the few observations show that there can be a strong variability of these currents on a monthly and yearly scale. For most of the sections, the yearly averaged transport agrees well with the literature.

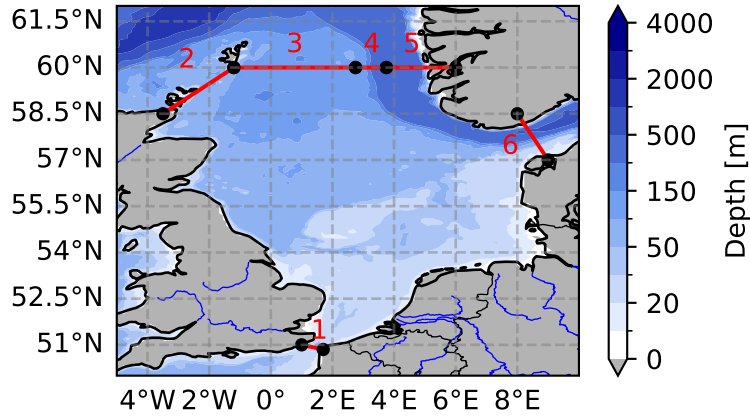


Figure 4.3. Sections where the transport of water is computed (see Table 4.1). The five first sections measure the currents shown in Fig. 4.2. 1: Dover, 2: Fair Isle current (FIC), 3: East Shetland current (ESC), 4: Western Skagerrak current (WSC), 5: Norwegian coastal current (NCC), 6: Skagerrak section.

Nb.	Name	Location		Modelled flow rate [Sv]	Literature [Sv]
1	Dover	(50.85°N, 1.7°E)	(51°N, 1°E)	-0.061	[-0.17, -0.06]
2	FIC	(58.5°N, 3.5°W)	(60°N, 1.2°W)	-0.721	[-0.5, -0.3]
3	ESC	(60°N, 1.2°W)	(60°N, 2.75°E)	-0.386	[-1.73, -0.9]
4	WSC	(60°N, 2.75°E)	(60°N, 3.75°E)	-0.761	
5	NCC	(60°N, 3.75°E)	(60°N, 6°E)	1.607	[1.34, 2.33]
6	Skagerrak	(58.5°N, 8°E)	(57°N, 9°E)	0.229	[-0.024, -0.015]

Table 4.1. Sections definition and their associated modelled transport averaged over the year 2007. A positive value means an outflow from the shelf. The literature values are ranges coming from [18], [19], [20] and [21]. In those references, the ESC and WSC are grouped together. For the Skagerrak section, literature value come either from a similar section or a section in the Kattegat. The sections are shown in Fig. 4.3.

There is one section having an unrealistic transport: the Skagerrak section. Normally, the Skagerrak should have a balance in between the strong flow of water coming from the shelf entering the south part of the section, the strong flow joining the NCC, and in a very small extent, the Baltic water that also joins the NCC. Here, the model predicts a flux coming into the Baltic of about 0.2 Sv instead of the 0.02 Sv inflow prescribed at the boundary. This mismatch is clear when looking at the prescribed and modelled barotropic speed in Fig. 4.4. This major difference in between the speeds is explained by the imposed SSH that seems to small (-20 cm on average), seen in Fig. 4.5. There is a general decreasing gradient of the SSH toward the Baltic boundary and the Flather condition has thus no other choice than imposing velocities going out of the domain. The reason for this unrealistic boundaries is at the end, a prescribed SSH that is too low. There is probably a mismatch in between the reference height of water in NEMO and the reference height of the boundary dataset provided by the Met Office.

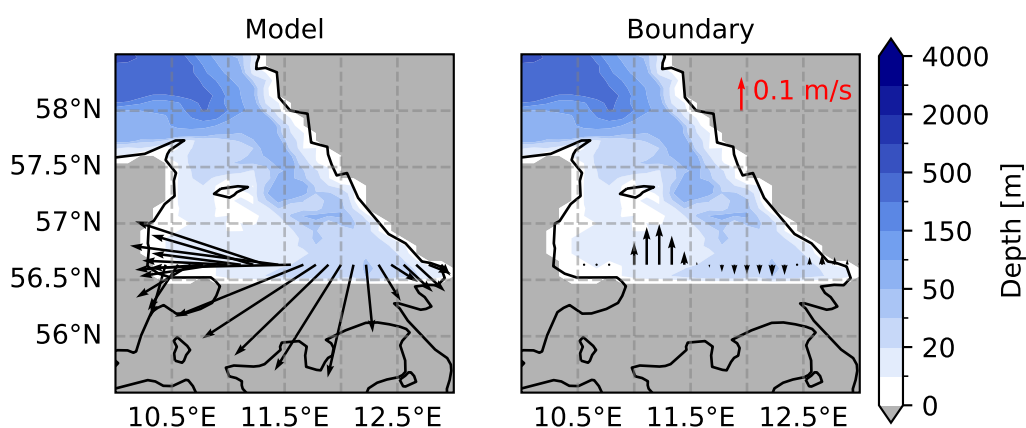


Figure 4.4. Modelled and prescribed barotropic speed field at the Baltic boundary of AMM7. The speeds are averaged over the whole year 2007. The contours show the bathymetry of the model, which is set to 0 out of the domain.

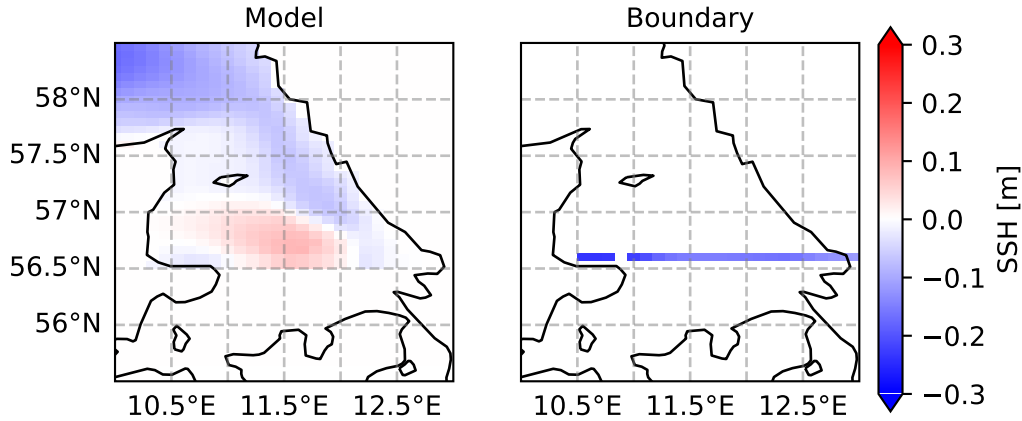


Figure 4.5. Modelled and prescribed sea surface height (SSH) at the Baltic boundary of AMM7. The SSH is averaged over the whole year 2007.

Concerning the SBNS and the BCZ, figure 4.6 shows that the current far from the shore is well organized, following the deepest isobath and joins the large coastal current in the German Bight. The bathymetry shows clearly that the shoreline are approximated very roughly. There is then a poor representation of the current near UK and Belgium. The current in the Dover Strait is quite messy, and it is hard to tell if it is well represented or not.

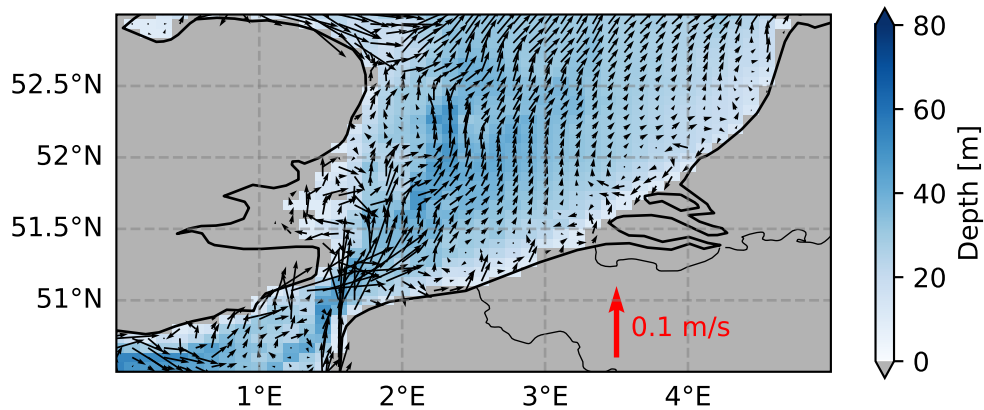


Figure 4.6. Model bathymetry and modelled barotropic speed field of AMM7 inside the SBNS. The speed is averaged over the whole year 2007.

4.2.2 Tides

Figure 4.7 shows the cotidal chart for the M2 tide modelled by AMM7. The three known amphidromic points are indeed present. The phase lines show that the tidal wave are circling around the two upper amphidromic points with a cyclonic path along the shelf. The biggest amplitudes are found near in the English Channel and near UK while the amplitude is smaller on the eastern basin because the Kelvin waves that are reflected in the shallow Southern Bight have experienced friction [22].

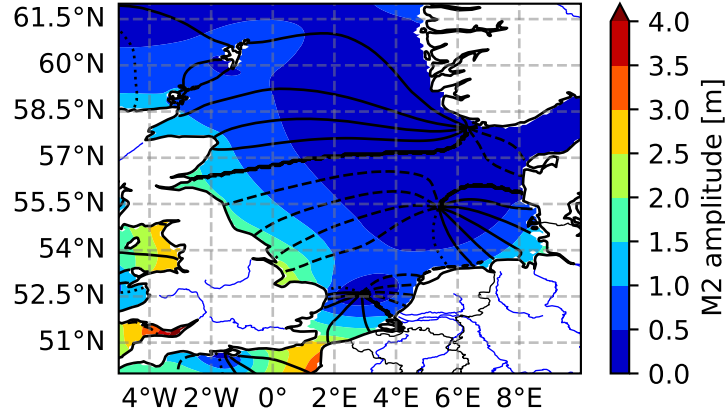


Figure 4.7. Co-tidal chart of the M2 tidal component modelled by AMM7. Co-phase lines have an interval of 30° . Positive phases are in solid lines, negative phases in dashed lines and the 0° phases are in dotted lines.

The averaged amplitude of each constituent are compared on the shelf and the SBNS in Fig. 4.8. The M2 and the S2 are, as expected, the dominant constituents. The amplitudes are generally higher in the SBNS because it is shallower. For the same reason, the non-linear constituent M4 is more important in the SBNS.

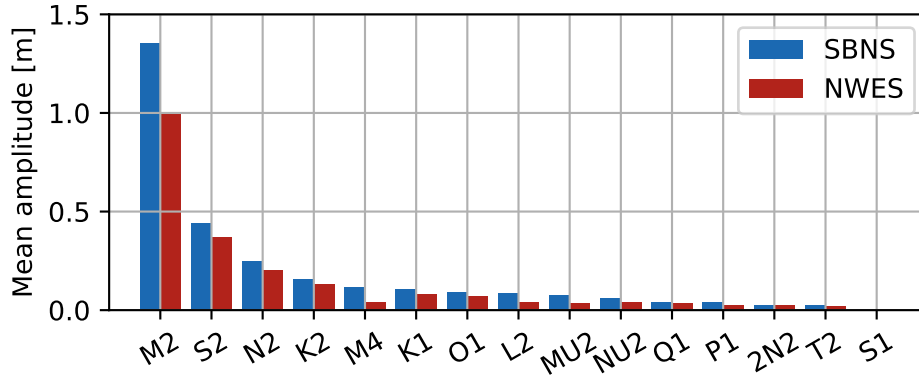


Figure 4.8. AMM7 tidal amplitudes for each modelled constituents averaged on the shelf (defined by the region shallower than 200 m) and on the SBNS (defined by the box $(50^\circ\text{N}, 2^\circ\text{W}) - (53^\circ\text{N}, 5^\circ\text{E})$).

The amplitude and phase of each constituent can be compared to tidal gauges measurements. Figure 4.9 shows the amplitude and phase bias for the M2 tide. For the amplitude, there is in general a very good agreement between the model and observation with a slight overestimation from the model, reaching at maximum 20 cm. The phase is also generally overestimated with bias of the order of 10° to 20° . The amplitude bias is generally smaller offshore where the observed amplitudes are smaller. This can be explained by the fact that the configuration is missing a wet and drying cell. In our model, each cell shallower than 10 m is considered as land cell to avoid cells with negative volume during low tides. This will be improved in further studies by configuring the wet and drying scheme from NEMO and changing the model bathymetry.

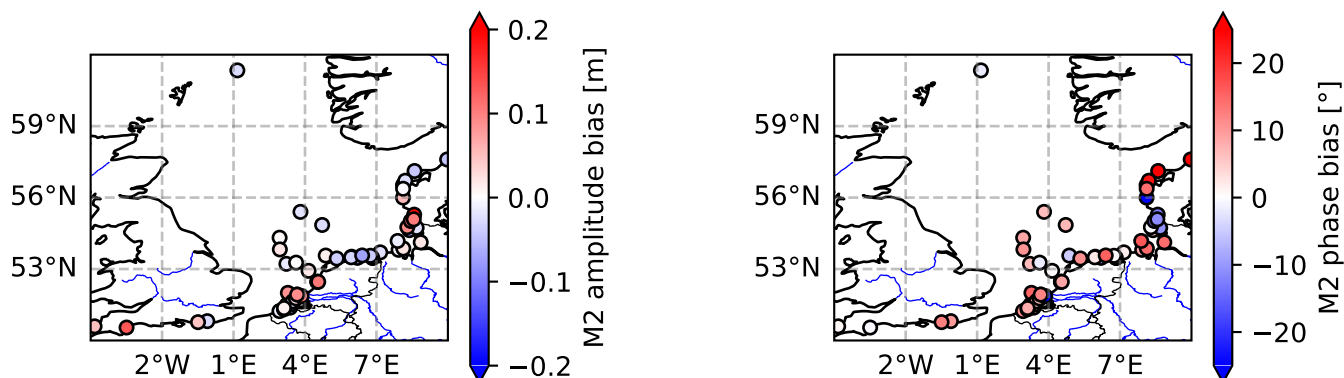


Figure 4.9. Amplitude and phase bias for the M2 tide modelled by AMM7 at each tidal gauge.

More quantitative results are given for each tidal constituents in Table 4.2. The predicted amplitude is generally excellent both on the shelf and the SBNS with the only exception of the M4 and K1 tides having a PB between 15 and 30 %. Since their amplitude is only of the order of 10 cm, this error is tiny compared to the total SSH. The modelled phase is generally in good agreement with measurements and with a bias usually smaller than 10° . The exception is again the M4 and K1 having biases between 10° and 20° on the SBNS. The model is a little bit less performant on the SBNS than on the whole shelf. This can be explained by the importance of the tidal wave entering the SBNS through the Dover Strait that is not well resolved by the low resolution model.

Const.	NWES			SBNS		
	Average am. [m]	Am. PB [%]	Ph. bias [$^\circ$]	Average am. [m]	Am. PB [%]	Ph. bias [$^\circ$]
M2	0.95	0.61	-5.70	1.22	1.81	-8.71
S2	0.28	-1.13	1.08	0.36	-1.18	13.34
N2	0.16	4.99	6.68	0.20	6.81	8.91
K2	0.11	-3.64	-17.07	0.14	-4.40	-1.41
M4	0.079	27.04	13.33	0.13	19.17	22.62
K1	0.074	14.15	-9.72	0.081	22.38	-7.10
O1	0.093	-6.52	3.77	0.11	-2.77	10.10
L2	0.071	2.86	8.39	0.091	5.21	8.57
MU2	0.052	0.95	6.05	0.086	7.60	14.78

Table 4.2. Average amplitude, amplitude bias and phase bias of the AMM7 model for the 9 dominant constituents. Those metrics are computed on the NWES and on the SBNS, based on tidal gauges measurements. The constituents are sorted according to Figure 4.8 on the SBNS. Const.: Constituent, Am.: Amplitude, Ph.: Phase.

The M2 tidal ellipse on the shelf and the BCZ are shown in Fig. 4.10 and 4.11. The size of major axis, giving the amplitude of the oscillating current, follows the same trend as the SSH amplitude with smaller currents offshore, close to the amphidromic

points. Their size also shows that the tidal current is roughly ten times bigger than the residual current, with amplitude reaching 1 m/s. Their eccentricity decrease toward the shore and their major axis get parallel to the shore. This result is expected since the non-permeability condition at land points should be respected. It also agrees with the classical Taylor solution for Kelvin wave in a rectangular basin [23]. Figure 4.11 shows more precisely the M2 tidal ellipses in the BCZ. It can be easily seen that the shoreline representation is much too coarse to predict correct tidal currents near the coast. The same argument holds for the thin Dover Strait which is resolved by less than five grid cells.

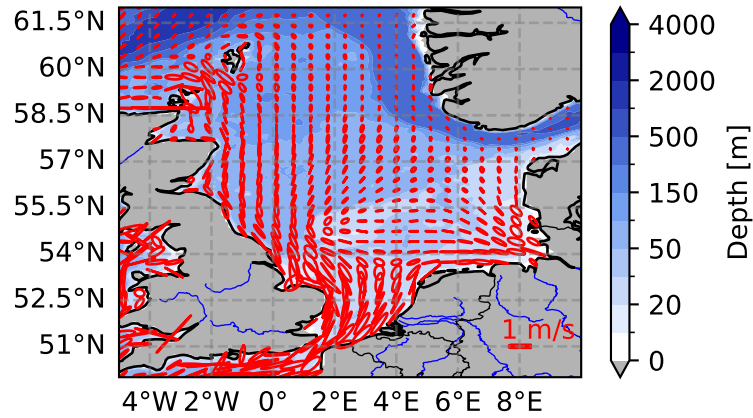


Figure 4.10. M2 tidal ellipses predicted by AMM7 on the NWES. The ellipses are plotted at every fifth cells for clarity.

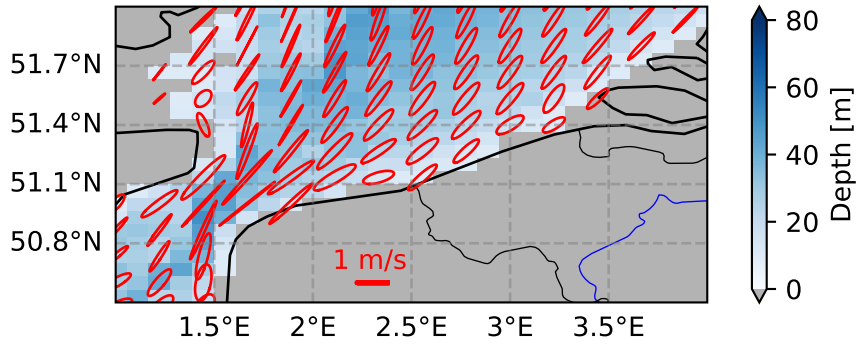


Figure 4.11. M2 tidal ellipses predicted by AMM7 on the BCZ. The ellipses are plotted at every two cells for clarity.

4.2.3 Temperature dynamics

Modelled temperature

The modelled SST is shown in Fig. 4.12. The seasonal trend with increasing temperature in summer is well represented. The general decrease of temperature northward is also present. Coastal waters, especially in the southeast have the largest SST variation

because those areas are very shallow and thus earn/lose heat much more easily than deep places [20].

The influence of Atlantic water can be seen by a tongue extending southward. In summer, the Atlantic water at the northern boundary is usually smaller than on the shelf. Thus, a cold tongue extending south appear. On the contrary, in winter the Atlantic water is warmer and gives a warm tongue on the shelf. This opposite behavior in summer and winter can be explained by the fact that the deep, off shelf water has a large thermal inertia and cools down at a slower rate than on the shelf [24]. In any season, there is an influence of the warm Atlantic water entering through the Dover Strait.

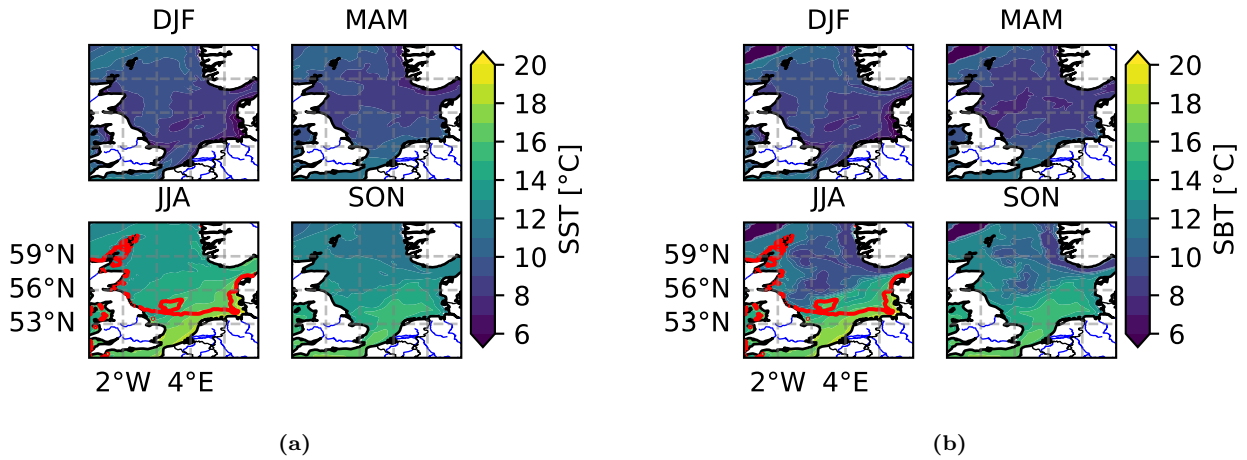


Figure 4.12. Seasonal (a) sea surface temperature (SST) and (b) sea bottom temperature (SBT) modelled by AMM7 on year 2007. The red contour line shows the summer tidal front defined by a difference of 0.5°C between the SST and SBT.

The seasonal sea bottom temperature (SBT) shown in Fig. 4.12 reflects the stratification of the water column. In winter, the shelf is well mixed and the SST is very close to the SBT except in the deep region past the continental slope. In summer, two distinct regions appear: the stratified region and the mixed region. In stratified regions, a thermocline exists and the difference of temperature can reach almost 7°C . The lower layer is isolated from the upper one and can be hardly heated, thus keeping a winter temperature. In mixed places, the temperature is almost homogeneous, with a difference smaller than 0.5°C . Those two regions are separated by the known tidal front, located just south to and around the Dogger Bank (red contour line on Fig. 4.12). Because the lower circulation is usually very weak, there is only very little horizontal mixing at the bottom. Large gradients of temperature, called a bottom front remain (Fig. 4.13). This front starts to settle in May and stay till the end of August. In September, the wind stress increases and become strong enough to mix progressively deeper places. Thus, the front moves north and disappears in November.

The vertical structure and the stratification of the water column are highlighted by Fig. 4.13. In winter, the temperature is uniform along depth while in summer, there is a clear vertical structure. The water is well mixed on the south part of the transect and

over the Dogger bank. On the edge of the Dogger bank, where the bottom gets deeper, a temperature gradient appears and gives a thermocline of 3 to 6°C. The mixed and stratified regions are separated by bottom fronts. There are characterized by sharpening and narrow isopleths showing a strong horizontal gradient at the bottom.

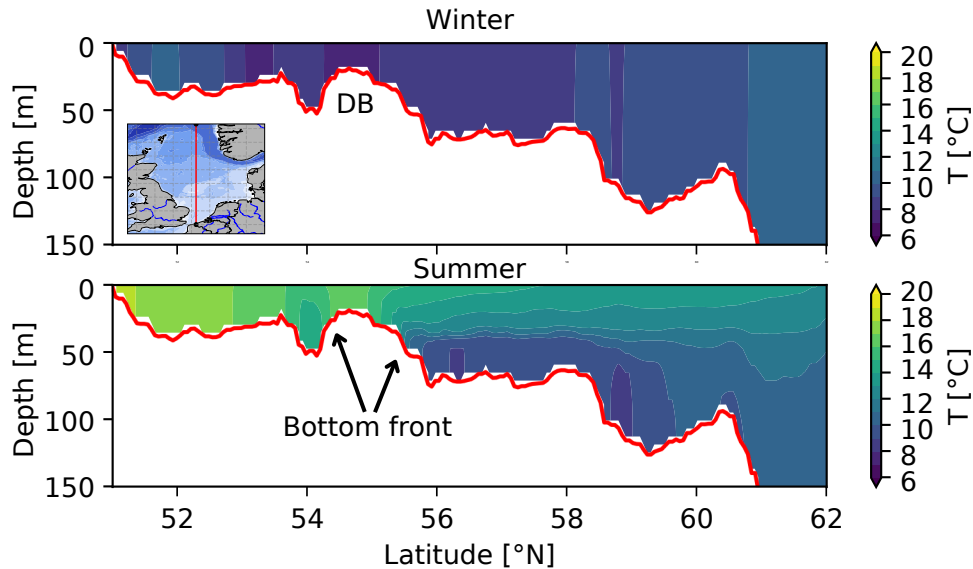


Figure 4.13. AMM7 modelled temperature averaged in winter and summer 2007 across a south-north central section, extending from the SBNS (51°N, 2.5°E) to the continental slope (62°N, 2.5°E) and crossing the Dogger Bank (DB). The red contour is the model bathymetry. The white spaces between the contour lines and the bathymetry are a plotting artifact coming from the interpolation of the model output on a common flat vertical grid, necessary for this plot.

Comparison with observations

The quality of the modelled temperature is assessed thanks to satellite measurements and CORA profiles. The seasonal SST bias with respect to satellite observations is plotted in Fig. 4.14. Although the general temperature pattern is consistent with what we know about the North Sea, there is systematically a warm bias of 0.88°C on average, that can reach 2°C on the shelf. The sole exception is a cold bias in summer over the Norwegian Trench. Comparisons with CORA profiles show a similar bias both on the SST, SBT and on the whole profiles (see Table 4.3). On the SBNS, the statistics are similar for satellite measurements but much better for CORA profiles. This is because the majority of profiles are measured in winter and autumn (see Fig. 3.1) where the bias is smaller than in summer. In any case, the correlation is excellent, showing that the spatial and temporal patterns are well described by the model but offset by a constant bias.

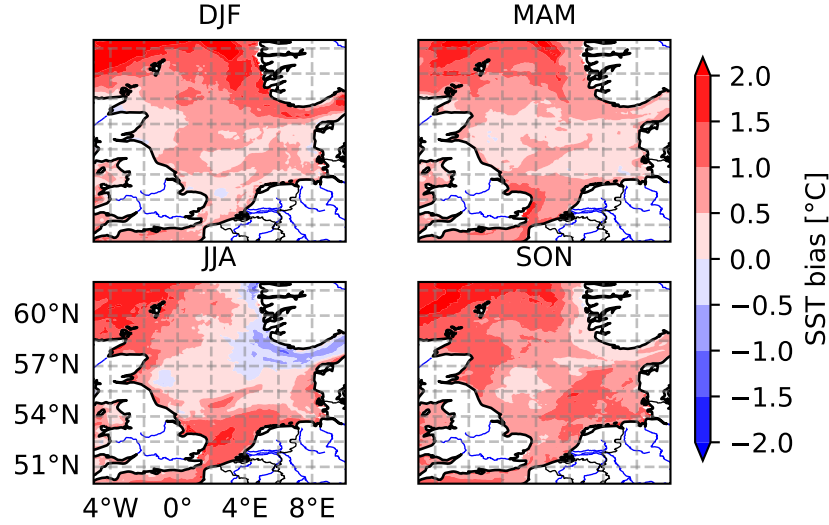


Figure 4.14. Seasonal seas surface temperature (SST) bias between AMM7 and CMEMS satellite L4 product on year 2007.

	NWES				SBNS			
	Bias [°C]	PB [%]	RMSE [°C]	Corr [-]	Bias [°C]	PB [%]	RMSE [°C]	Corr [-]
Sat. SST	0.88	7.52	1.05	0.98	0.90	7.13	1.01	0.99
CORA SST	0.66	6.29	0.93	0.98	0.37	3.35	0.53	0.99
CORA SBT	1.03	10.94	1.43	0.95	0.32	3.30	0.37	0.99
CORA profiles	0.89	8.29	1.20	0.95	0.33	2.87	0.45	0.99

Table 4.3. Main metrics for the modelled shelf and SBNS temperature compared to observations. The sea surface temperature (SST) is compared both to satellite observations and CORA profiles. The sea bottom temperature (SBT) is assessed with CORA. The statistics of the overall profiles are also presented.

It should be noted that the presented results are surprisingly different from the ones presented in [2] and [25] while they should come from the same model. In both papers, the model has a cold bias while, here, a warm and stronger bias is present. The only explanation found is that the forcings and namelist provided in the official repository [4] are partially incorrect. We trust the surface forcing and the employed scheme. A simulation without river showed that the general bias was not modified significantly except in ROFIs. The suspect boundary is thus again the lateral boundary. If the Atlantic boundary has an erroneous SSH like at the Baltic, it is expected that the water transport through the boundary is wrong and thus the lateral heat fluxes.

4.2.4 Salinity

Modelled salinity

The modelled salinity is shown in Fig. 4.15. As expected, there is an increase of salinity northward to 36 psu because of the North Atlantic influence. In ROFIs, and

more generally along the east coast, the salinity is much smaller, going down to less than 26 psu.

There is only a very weak annual variation of salinity over time on most of the shelf as shown by the amplitude in Fig. 4.15(b). The only places where the salinity can vary significantly are in ROFIs and in the Norwegian Trench. The salinity there can vary a lot, sometimes by more than 5 psu. This is mostly explained by the variability of rivers discharge.

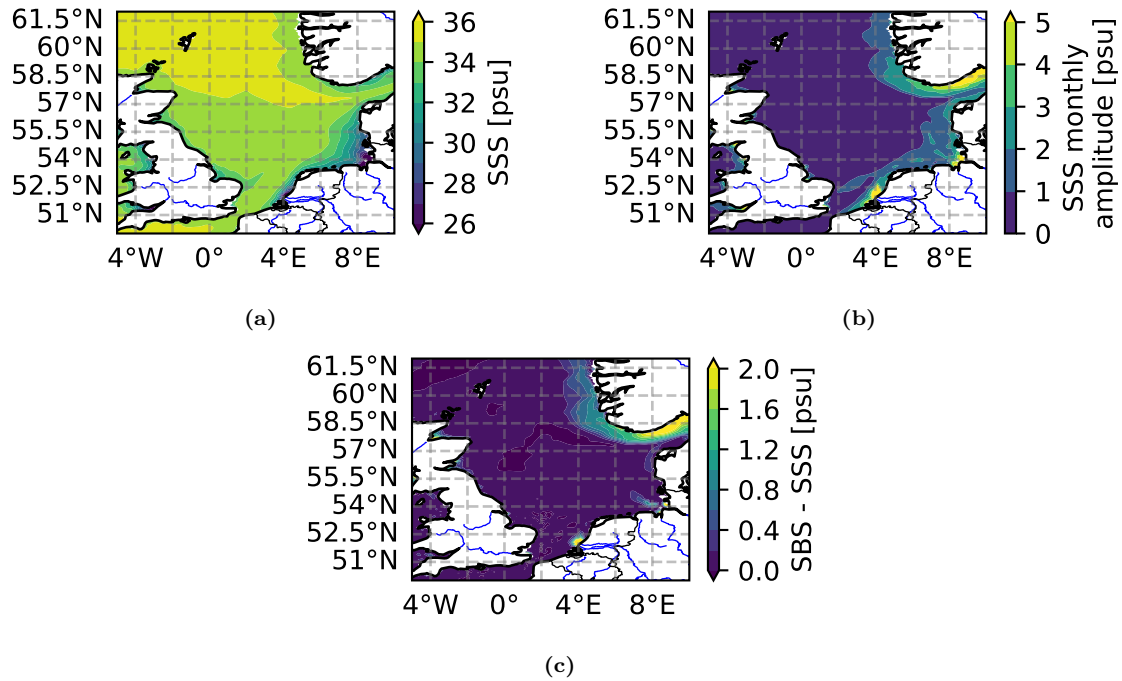


Figure 4.15. Practical salinities modelled by AMM7. (a) Sea surface salinity (SSS) averaged on year 2007. (b) Amplitude of the surface salinity time variation, computed by a difference between the maximum and minimum monthly salinity of the year. (c) Bottom to surface salinity difference averaged over year 2007.

The ROFIs can again be distinguished by looking at the difference of bottom (SBS) and surface salinities (SSS). On the shelf, this difference is most of the time very weak while in ROFIs, a large salinity stratification exists. In the Rhine plume, this difference can exceed 2 psu beside the shallowness of the region. There is also a large salinity stratification in the Norwegian Trench where saline North Atlantic water lays bellow the fresher North Sea water coming out of the shelf.

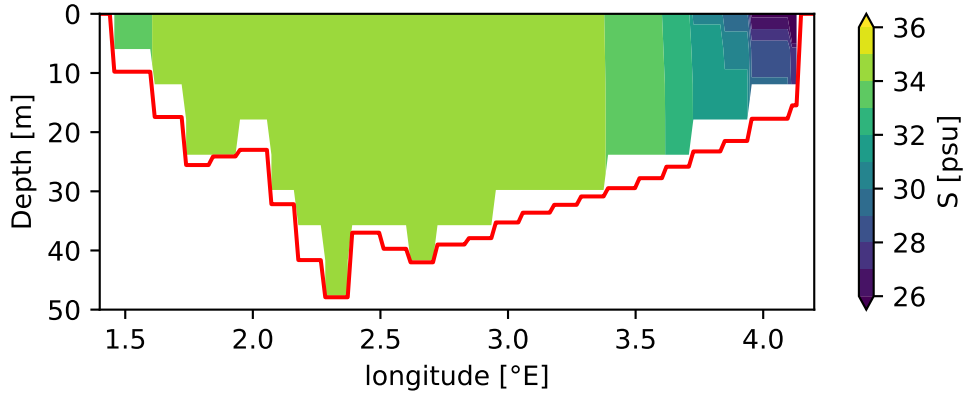


Figure 4.16. Modelled salinity averaged on year 2007 across a west-east section, extending from UK (52°N, 1°E) to the Rhine ROFI (52°N, 4.5°E). The red contour is the model bathymetry. The white spaces between the contour lines and the bathymetry are a plotting artifact coming from the interpolation of the model output on a common flat vertical grid, necessary for this plot.

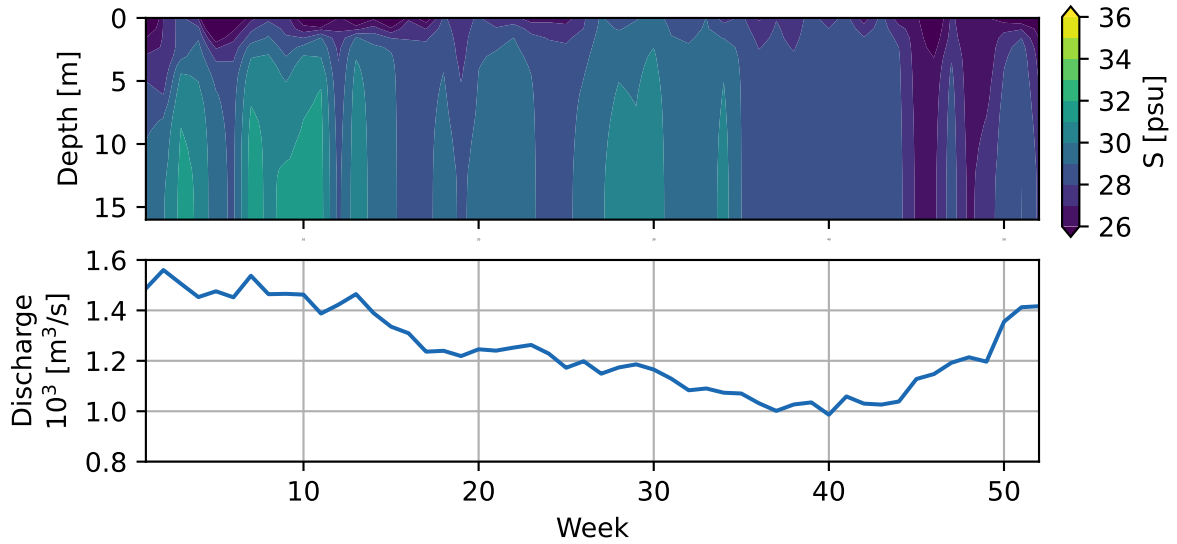


Figure 4.17. Top: Time series of the modelled salinity in the Rhine ROFI (52°N, 4°E), weekly averaged during the year 2007. Bottom: Weekly averaged discharge of the Rhine prescribed to the model.

Because the Rhine freshwater partly extends south till the BCZ and thus influences strongly its physics [26], here is a deeper a look at its salinity. A transect of the modelled salinity from England to the Rhine freshwater plume is shown in Fig. 4.16. The vertical structure has two distinct behaviors. A uniform profile, far offshore in the center of the SBNS and a stratified profile near the coast and the Rhine mouth. Figure 4.15(b) shows a strong variability of the surface salinity. The variability of the Rhine discharge can explain this variability (see Fig. 4.17). Indeed, there is a negative correlation (-0.46) in between the surface salinity and the Rhine inflow. Part of the SSS variability can be explained by the Rhine discharge. The other factors are the others neighboring rivers and the variability of the eastward current bringing more saline water close to the coast.

The latter can also explain why the bottom salinity experiences a strong time variability that seems independent to the river discharge.

Comparison with observations

Salinity is compared with CORA profile in Table 4.4 for the overall NWES and SBNS. In general the bias is slightly positive in surface and negative at the bottom. The bias is reasonably good but its deviation (given by the RMSE) reaches 1 psu, strongly affecting the density structure of the flow. On the SBNS, the overall statistics are of similar quality but with a negative bias. The correlation is good, but less than the temperature and is a little bit better at the bottom where the salinity varies less.

	NWES				SBNS			
	Bias [psu]	PB [%]	RMSE [psu]	Corr [-]	Bias [psu]	PB [%]	RMSE [psu]	Corr [-]
CORA SSS	0.12	0.35	1.13	0.70	-0.31	-0.89	0.66	0.77
CORA SBS	-0.25	-0.71	0.79	0.93	-0.51	-1.47	0.59	0.94
CORA profiles	0.10	0.85	0.53	0.79	-0.34	-1.00	0.47	0.85

Table 4.4. Main metrics for the modelled shelf and SBNS salinity compared to CORA profiles.

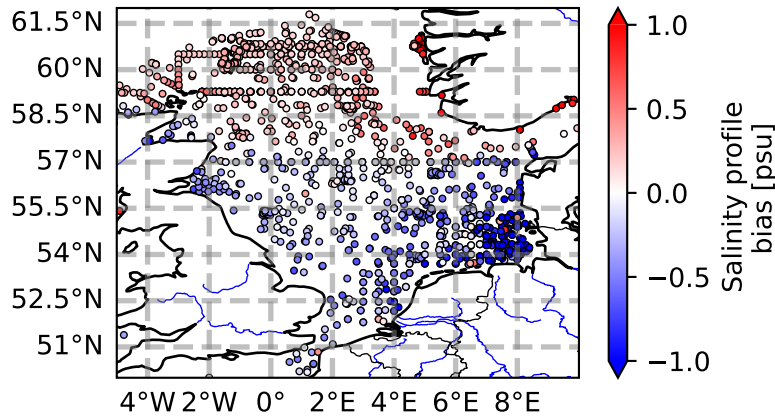


Figure 4.18. Salinity bias computed on between AMM7 and each CORA profile.

Figure 4.18 shows the bias of each compared profiles. The German Bight and SBNS present a fresh bias that was already described in [2]. This bias is caused by a too large input of freshwater at the Rhine and other rivers. Actually, the river runoff imposed to the model is probably accurate but the assumption that the freshwater input has no salinity is wrong. This is because in this model the estuaries and coastal regions such as the Wadden Sea are not resolved by the model and are replaced by point source of freshwater. In reality, the water at the outlet of the estuaries is partially mixed with seawater and has thus a non-zero salinity. One possible way to improve the salinity of

the model, while keeping estuaries unresolved, would be to consider river source with some salinity [27].

4.3 Concluding remark and possible improvements

It has been shown toward this chapter that AMM7 is able to represent most of the features presented in Sec. 1.2 from a qualitative point of view. From a quantitative point of view, comparison with remote and *in-situ* measurements have shown that the model performs usually well but can be improved in some places.

The main currents and transports are well resolved by the model. The major place where the model does not perform well is at the Baltic boundary where the predicted flow is in the wrong direction, causing an unrealistic flow toward the Baltic. This inaccuracy is due the SSH which is wrongly prescribed at the Baltic boundary. The solution to this problem would be to change the boundary data, either by applying a simple offset to the SSH or by using data from a new model such as a North Sea - Baltic model.

The tides are especially well predicted by AMM7 with very small biases for the dominant M2 and S2 tides. A possible improvement would be to add a "wet and drying" scheme allowing the model to handle cells with very small depth that would be land cells during low tides and water cells during high tides.

Although the configuration is less performant for the temperature and salinity fields, it is still able to represent crucial patterns of the NWES both on the horizontal and the vertical planes. It is again suspected that the SSH data provided at the boundary is incorrect and affects the heat budget of the domain. To have a better representation of ROFIs and coastal areas, the river schemes could be improved in several ways. First, the river discharge can be applied on several cells along depth instead of providing the whole flux to the surface cell. A heat exchange could also be implemented by specifying an inflow with a different temperature than the SST. Adding a non-zero salinity to the river runoff could decrease the fresh bias in the German Bight and would be more representative of the condition at the boundary between the open sea and an estuary.

Concerning the SBNS and the BCZ, the model has a similar performance to the overall shelf. However, it is clear that the resolution is insufficient to capture smaller scale features that are caused by the coastline and the complex topography. The transition in between the coastal current and the offshore current is thus poorly represented. The same conclusion applies for the thin Dover Strait which is resolved by less than five grid points. It is clear that a finer model is lacking to study practical topics such as the impact of human activities on the Belgian Coastal Zone hydrodynamics and biogeochemistry.

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CHAPTER 5

The high resolution model

In this chapter, a high resolution NEMO configuration covering the SBNS and forced by AMM7 is established. The configuration, its domain, and its forcings are first presented. Then, the model is run for one year. The various physical quantities derived by the model are shown, and the results are compared to the previous low resolution configuration presented in Chapter 4.

5.1 Configuration description

The configuration, here named SBNS10 is essentially the same configuration as AMM7 but with a high resolution zoom on the SBNS. Most of the parameters, numerical schemes and parametrizations are kept unchanged and will not be recalled here. For more details about this configuration, the reader should refer to Sec. 4.1.

5.1.1 Domain

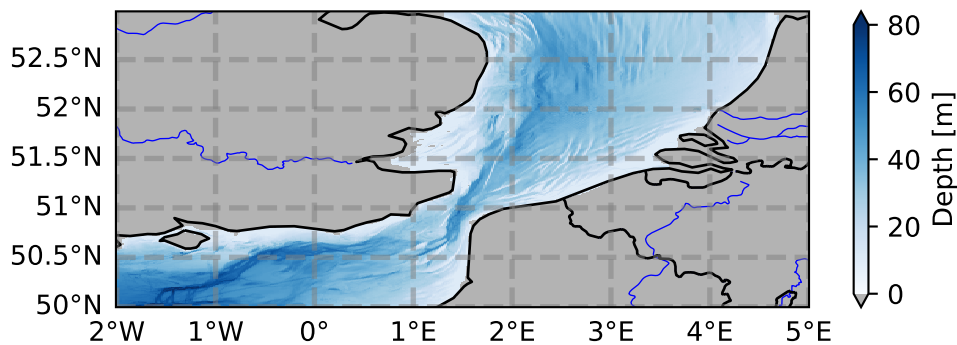


Figure 5.1. SBNS10 model bathymetry, coming from GEBCO 15 arc-second product [1]

The domain covers the whole SBNS from 50°N to 53°N and 2°W to 5°E. The horizontal grid has a resolution of 1 km, giving 441 cells on the zonal direction and 315 on the meridional direction. The bathymetry is interpolated from the GEBCO 15 arc-second product [1] on the domain grid. On the vertical, the exact same coordinate system as AMM7 is employed (Sec. 4.1). Since the domain covers only shallow places, the vertical

grid consists at the end of the upper part of the multi-envelope system, giving a layer of 26 s levels following the bottom. The model bathymetry and a section of the mesh near the Belgian coast are shown in Fig. 5.1 and Fig. 5.2.

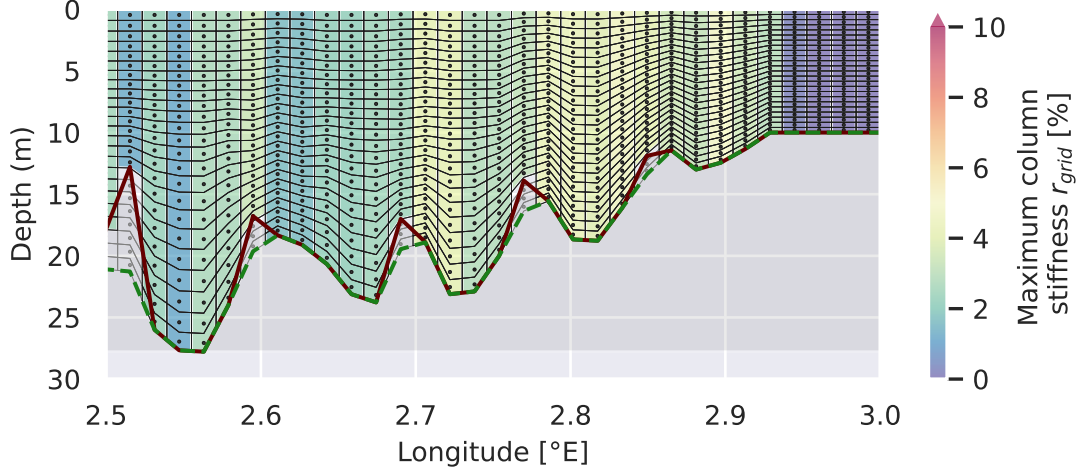


Figure 5.2. Vertical mesh of the SBNS10 model near the Belgian coast (51.3°N). The red line is the model bathymetry. The green line is the envelope bathymetry, with $r_{max} = 0.07$. The color map indicates the maximum stiffness r_{grid} of the column at each horizontal position. Grey cells are masked cells. This chart has been made thanks to the Python code present in the AMM7 repository https://github.com/JMMP-Group/C0_AMM7/tree/v9.0-alpha/scripts/analysis

On the time domain, the same scheme as in AMM7 is employed. The Leapfrog scheme is applied with 30 barotropic sub time steps. The baroclinic time step is reduced by a factor 7 and rounded down to the closest divider of an hour¹, giving $\Delta t = 20$ s. The initial condition is the same as AMM7 but interpolated onto the new grid.

5.1.2 Forcings

Air-sea interface

The surface forcings and parameters are kept unchanged from AMM7 to the high resolution SBNS configuration. They still come from the ERA5 reanalysis [2], [3]. The data is interpolated on the domain grid with a bicubic interpolation, and is provided at every simulation hour to NEMO. The bulk parametrization is again from the ECMWF bulk formulas [4], with cool-skin and warm-layer parametrizations.

Open boundaries

The same lateral conditions as AMM7 are applied: FRS for the tracers, Flather for the barotropic velocities and null gradient for the baroclinic velocities. For the FRS scheme, a Rim layer of 10 rows is applied and the temperature and salinity of AMM7 are interpolated on these grid points using a bicubic method. The barotropic velocities and the SSH are also provided by AMM7 with a bicubic interpolation over the boundaries. Each field is provided daily with the tidal signals that have been filtered out.

¹This is needed by NEMO to manage hourly and daily output.

The tidal amplitudes of the SSH and barotropic velocities are computed by AMM7 and interpolated on the boundaries of SBNS10 so that the model can reconstruct the full signals at any time, given the residual and the tidal amplitudes.

Rivers forcing

Because a new domain is created for the high resolution model, the river data provided by the AMM7 repository is not usable and a new river grid must be given to our model. River data from EFAS [5] is provided daily to the model. EFAS river data is a product from the Copernicus Emergency Management Service that use the open hydrological model LISFLOOD covering Europe with a resolution of 1.5 km. This model is generally in very good agreement with measured flow rate for major rivers [6]. The rivers discharge, averaged over 2007 are shown in Fig. 5.3. The logarithmic scale shows that there are large differences in between the rivers and that the Rhine is by far the most important river of the region, with an average discharge of about $2300 \text{ m}^3/\text{s}$.

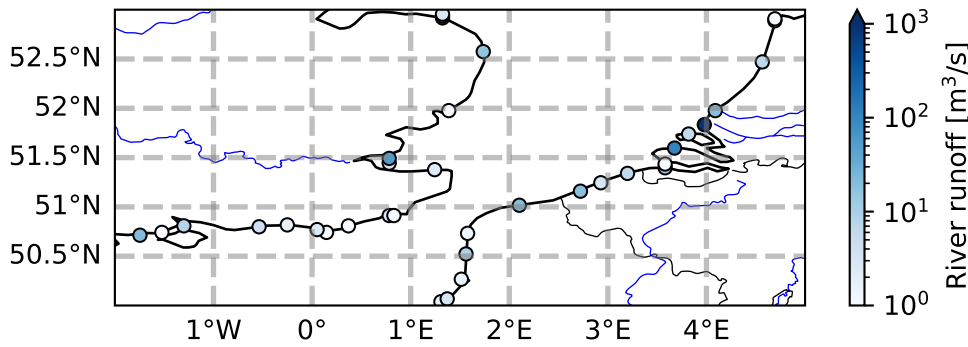


Figure 5.3. River discharges computed from the river model EFAS [5] averaged over 2007 and provided to the SBNS10 configuration.

The EFAS data is not directly compatible with NEMO because it provides rivers discharge on land, on a grid different than SBNS10. Preprocessing is thus needed to make them compatible. First, each EFAS grid point with an ocean neighbor cell and with an average flow rate bigger than $1 \text{ m}^3/\text{s}$ are isolated. Then, for each of those cells, we find the closest water cell in the SBNS10 grid and give it the daily flow rate.

In the end, there are some important differences between the AMM7 river dataset and the EFAS dataset. The annual average river discharges are shown for both model in the Rhine and Thames region in Fig. 5.4. First, the Thames discharge is reduced from $131 \text{ m}^3/\text{s}$ to $55 \text{ m}^3/\text{s}$. The neighboring rivers have also a reduced discharge. Secondly, the principal Rhine discharge has been increased from $1249 \text{ m}^3/\text{s}$ to $2078 \text{ m}^3/\text{s}$. The second main mouth discharge has been decreased from $636 \text{ m}^3/\text{s}$ to $112 \text{ m}^3/\text{s}$. AMM7 has an additional mouth for the Rhine with $119 \text{ m}^3/\text{s}$. Thus, both models have a similar total discharge for the Rhine but spread differently in the domain. A last difference between AMM7 and SBNS10 is the addition of the Scheldt and some weaker river along the Belgian coast and the Netherlands.

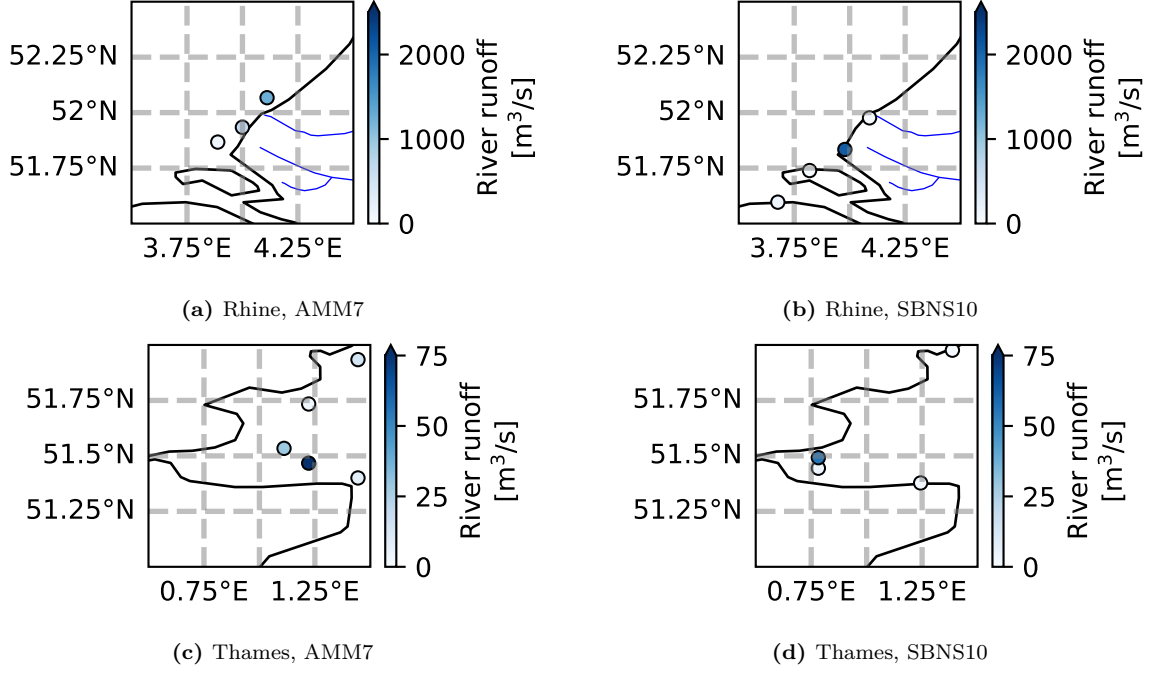


Figure 5.4. Comparison of the annual averaged river discharges between AMM7 and SBNS10 in the Thames and Rhine region. The location of each mouth is different because of the change in the domain and coastline from one configuration to the other.

5.1.3 Miscellaneous parameters

Most of the numerical parameters are kept unchanged in between AMM7 and SBNS10. The only difference is for the horizontal viscosity and eddy viscosity. Since the resolution is higher, smaller scale processes are now resolved by the model and should then not be included in the diffusion term of the momentum and tracer equations. Practically, it means that the eddy viscosity and diffusivity should be smaller. We therefore divide the characteristic length scales by a factor 7, giving a viscosity of $2.9 \cdot 10^7 \text{ m}^4/\text{s}$ and a diffusivity of $7.15 \text{ m}^2/\text{s}$.

5.2 Results

5.2.1 Residual currents

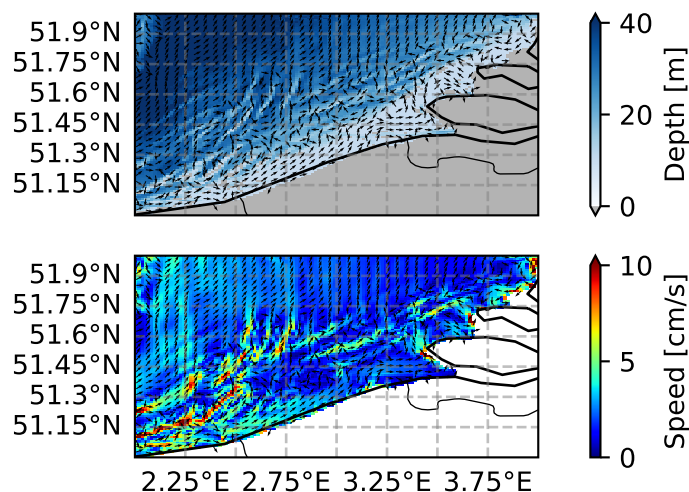


Figure 5.5. Barotropic velocities modelled by SBNS10 on the BCZ, averaged over year 2007. The arrows have all the same size and indicate the direction of the currents while the colormap shows the model depth (top) and the magnitude of the current (bot.).

The residual currents modelled by SBNS10 on the BCZ is shown in Fig. 5.5. The current far from the shore is not significantly different from what is predicted by AMM7. On the contrary, new features appear when moving toward the coast. There is a current of medium intensity flowing along the shoreline. Then further out, there is a transition between the current far offshore and close to the shore. In this transition layer, the current flows in several directions that often follow lines of constant bathymetry. It is especially stronger at the summit of each ridge.

Further investigations show that the intense residual current appearing above the sandbanks ripples is due to non-linear interactions between the tidal currents and the topography. Zimmerman has shown in [7] that the oscillating flow past sandbanks can lead to peaks of residual vorticity when the length scale of the ripples is of the same order of the tidal displacement that a water parcel experiences during a tidal cycle. A run of SBNS10 without any tidal forcing at the boundaries, neither in the bulk show that the intense currents disappear, proving that they come from tides. Fig. 5.6 shows also that the model, with tides, shows indeed a residual vorticity, maximum at the summit of the ridges (Fig. 5.7).

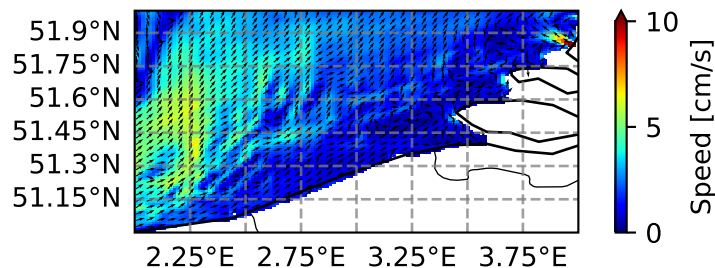


Figure 5.6. Barotropic velocities modelled by SBNS10 on the BCZ without any tidal forcing, averaged over year 2007. The arrows have all the same size and indicate the direction of the currents while the colormap shows the magnitude of the current.

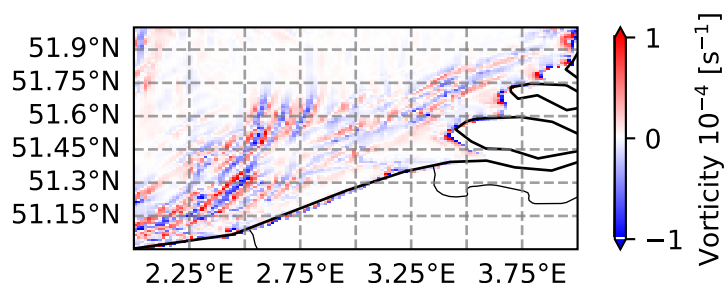


Figure 5.7. Depth averaged vertical vorticity modelled by SBNS10 on the BCZ, averaged on year 2007.

Unlike [8], there is no residual flow going southward along Belgium, coming from the Scheldt estuary. We obtain a circulation closer to [9] and [10] that predict a northward flow out of the estuary. The currents following the sandbanks are a new feature that is probably due to a more accurate and recent bathymetry.

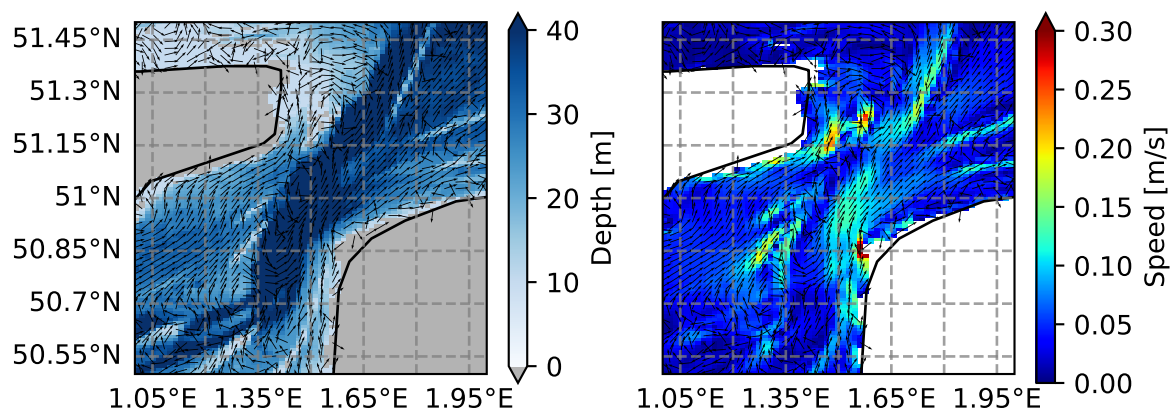


Figure 5.8. Barotropic velocities modelled by SBNS10 in the Dover Strait, averaged over year 2007. The arrows have all the same size and indicate the direction of the currents while the colormap shows the model depth (left) and the magnitude of the current (right).

The Dover Strait, upstream of most of the current flowing over the BCZ, is shown in Fig. 5.8. The flow is now resolved much more accurately than with AMM7. The

Atlantic inflow can be separated into four branches, each following a coastline or an isobath. There is a coastal current to the west, flowing around UK. An intense current flows in the center of the strait to the northeast along a shallow isobath (20 m) confined in between two deeper areas (40 m). Finally, two branches go to the northeast, joining the Belgian coast, following also two shallow isobath lines.

5.2.2 Tides

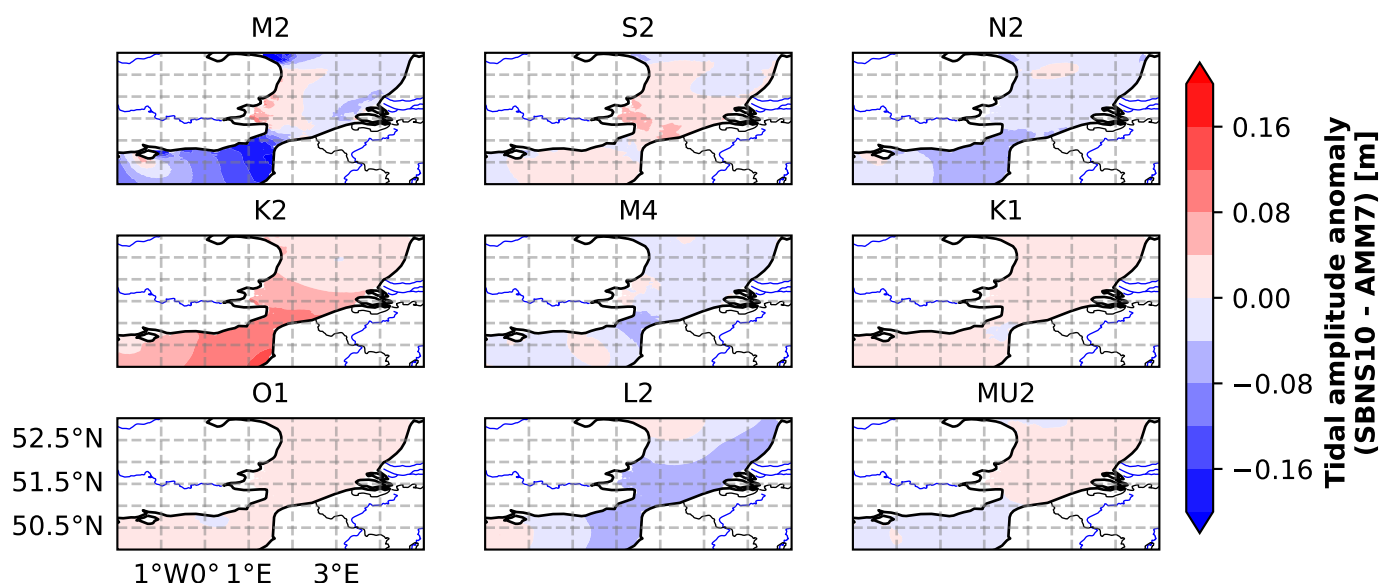


Figure 5.9. Difference of sea surface height (SSH) tidal amplitude between SBNS10 and AMM7. The AMM7 tidal output is interpolated on the SBNS10 grid using bilinear interpolation.

The tidal elevation modelled by SBNS10 is not substantially different but there are some variations in between the two configurations. Figure 5.9 shows that some differences appear, mainly in the English Channel. There, the M2 constituent is reduced with a difference that reaches 20 cm. The K2 tide shows similar but positive changes. We attribute these changes to the increased resolution in the English Channel and at the Dover Strait. While the low resolution model had enough grid cells to cover the wavelength of the semi-diurnal and diurnal tides², the resolution was insufficient to represent correctly the topography and the coastline that shape the eigenmodes of the tidal waves. The increased resolution should also lead to a better representation of non-linear interactions and processes such as bottom friction.

The tidal performance of the model is compared to tidal gauges, with statistics shown both for SBNS10 and AMM7 in Table 5.1. Except for the M2 and M4 tides that have seen an improvement, most of the constituents decline in the high resolution model. This decrease of performance can seem a high threat but actually, the four dominant

²The phase speed of Kelvin waves is $c = \sqrt{gH}$ and at least 9.9 m/s at the shallowest places of the domain. The smallest tidal period is about 12 hours for the semi-diurnal tides. It leads to a minimal wavelength $\lambda = cT = 427.9$ km.

constituents accounting for most of the SSH variability, are still well predicted by the model. This decline of performance is again explained by the lacking "wet and drying" scheme that is even more needed in high resolution coastal application. This missing feature reflects in the statistics because the majority of tidal gauges are located very close to the coastline in places shallower than 10 m. At those places, the model bathymetry is cropped to 10 m and does not represent well the true domain anymore. It has been shown in [11] that the "wet and drying" scheme implemented in NEMO can significantly improve the tidal dynamics. In this case, a high resolution model of 1.5 km on the whole North Sea was used, and the tidal performance was assessed in the Wadden Sea. We expect similar improvements in our domain if we add this feature.

Const.	Average am. [m]	AMM7		SBNS10	
		Am. PB [%]	Ph. bias [°]	Am. PB [%]	Ph. bias [°]
M2	1.29	1.81	-8.71	-1.38	-12.06
S2	0.39	-1.18	13.34	2.13	7.27
N2	0.22	6.81	8.91	-10.28	2.83
K2	0.14	-4.4	-1.41	29.43	117.97
M4	0.14	19.17	22.62	6.01	-46.34
K1	0.079	22.38	-7.10	50.25	-27.61
O1	0.11	-2.77	10.10	18.16	-42.02
L2	0.093	5.21	8.50	-56.10	-93.54
MU2	0.088	7.60	14.78	13.79	-110.56

Table 5.1. Average amplitude, amplitude bias and phase bias of the SBNS10 and AMM7 model for the 9 dominant constituents, on the SBNS. The constituents are sorted according to Figure 4.8 on the SBNS. Const.: Constituent, Am.: Amplitude, Ph.: Phase.

The tidal currents are not significantly different from AMM7. The main improvement is seen close to the coastline that is now well represented by the model. Thus, the coastal tidal current follows well the coastline (Fig. 5.10). The tidal ellipses of AMM7 and SBNS10 are compared to the two mooring platforms "BolVanHeistMP" and "ScheurWielingenMP". Figure 5.11 shows their respective M2 tidal ellipses. In both cases, the measured ellipses have no eccentricity because the platforms stand very close to the shore. The SBNS10 give slightly better ellipses due to the refined representation of the coastline.

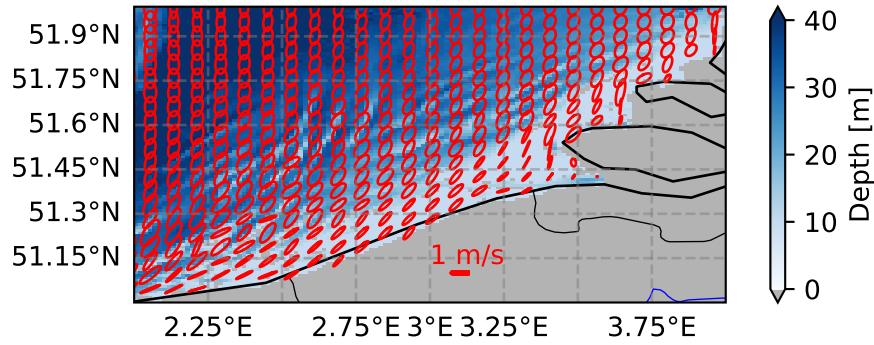


Figure 5.10. Tidal ellipses for the M2 tide modelled by SBNS10 along the Belgian coast. Ellipses are plotted one every two grid points for clarity.

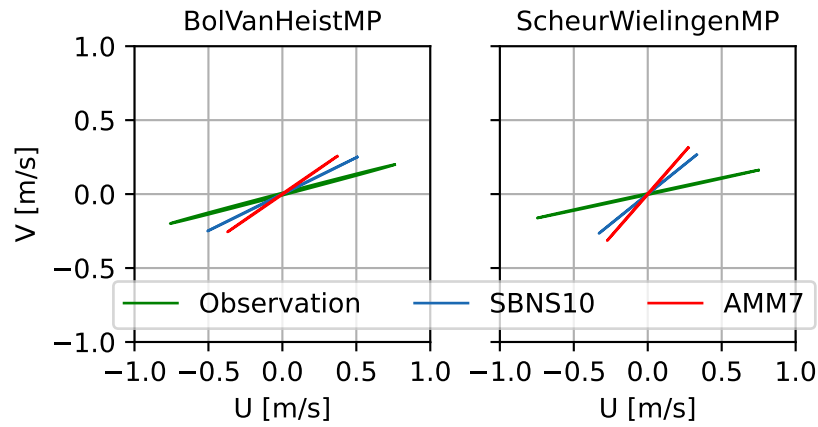


Figure 5.11. M2 tidal ellipses modelled by AMM7, SBNS10 and observed at the mooring stations "BolVanHeistMP" and "ScheurWielingenMP". The ellipses are degenerated since the moorings are located nearby the shore.

5.2.3 Temperature

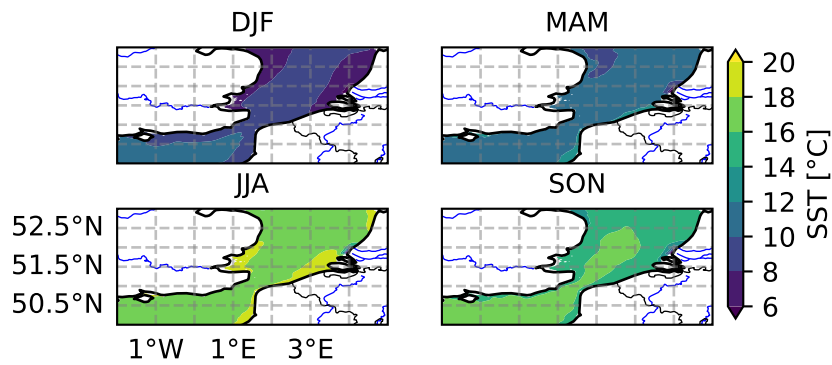


Figure 5.12. Seasonal sea surface temperature (SST) modelled by SBNS10 on year 2007.

The seasonal temperature predicted by SBNS10, shown in Fig. 5.12, is in appearance not much different than what is predicted by AMM7. The water is never thermally stratified and the warm Atlantic water tongue mainly visible in autumn and winter is still present.

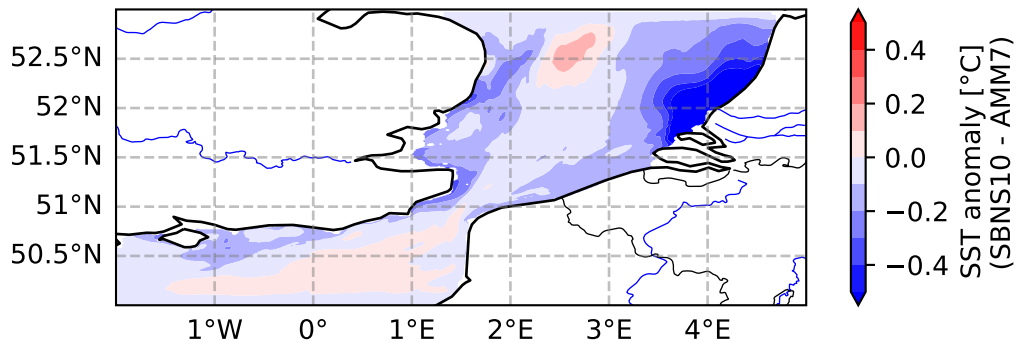


Figure 5.13. Difference of the modelled sea surface temperature (SST) in between AMM7 and SBNS10 averaged over year 2007. The AMM7 output was interpolated on the SBNS10 grid using bilinear interpolation

There is actually a general difference of the magnitude of the SST in between SBNS10 and AMM7 with temperatures generally smaller in SBNS10, especially around the Rhine mouth. (Fig. 5.13). This major change close to the Rhine mouth can only be explained by the change of resolution since the total Rhine discharge has not changed from AMM7 to SBNS10. This inaccuracy is actually due to the combination of the small size of the grid cell and of the large river runoff that is forced only to the surface cell. A simulation with the river forced evenly over the whole water depth shows that the area of cold water disappears and an annual SST cycle is recovered (not present when the river is only provided at the surface).

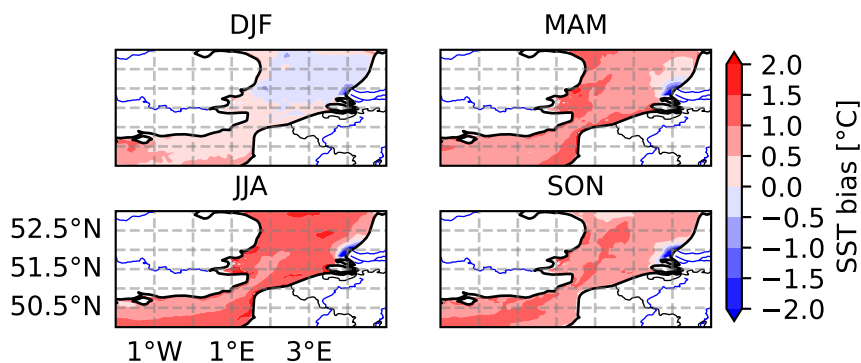


Figure 5.14. Seasonal sea surface temperature (SST) bias between SBNS10 and CMEMS satellite L4 product on year 2007.

The SST is compared to satellite measurements in Fig. 5.14. The bias is in general similar to AMM7. A cold bias appears at the Rhine mouth because of the bad river forcing. Table 5.2 confirms that the bias is smaller but this improvement is exaggerated by the large Rhine cold bias, cancelling partly the general warm bias. The RMSE do not

change in between the two configurations, showing that there is no true improvement. The CORA profiles reflect the same statistics.

It can be surprising that the temperature is not improved from the low to the high resolution configuration. In such a small domain, the heat content is mainly fed by lateral heat fluxes instead of surface fluxes. Because the forcing configuration AMM7 has already a large warm bias, the heat fluxes at the boundaries are inaccurate and the heat budget of the domain is thus badly modelled.

	AMM7				SBNS10			
	Bias [°C]	PB [%]	RMSE [°C]	Corr [-]	Bias [°C]	PB [%]	RMSE [°C]	Corr [-]
Sat. SST	0.90	7.13	1.01	0.99	0.71	5.63	1.00	0.99
CORA SST	0.37	3.35	0.53	0.99	0.27	2.42	0.53	0.99
CORA SBT	0.32	3.30	0.37	0.99	0.08	0.83	0.36	0.99
CORA profiles	0.33	2.87	0.45	0.99	0.31	2.67	0.47	0.99

Table 5.2. Main metrics for the modelled temperature compared to observations for AMM7 and SBNS10. The sea surface temperature (SST) is compared both to satellite observations and CORA profiles. The sea bottom temperature (SBT) is assessed with CORA. The statistics of the overall profiles are also presented.

5.2.4 Salinity

There are two important changes in the new configuration concerning the salinity. There is a first change due to the change in river forcings at the Thames and the Rhine that is highlighted by Fig. 5.15. The positive anomaly near the Thames is clearly caused by the smaller river inflow imposed to SBNS10. A similar positive anomaly appears near the Rhine because two main Rhine mouths have been swapped (Fig. 5.4). In SBNS10, the main Rhine mouth is now located more to the south compared to AMM7. Thus, there is a positive anomaly north to this mouth and on the contrary, a negative anomaly south to it.

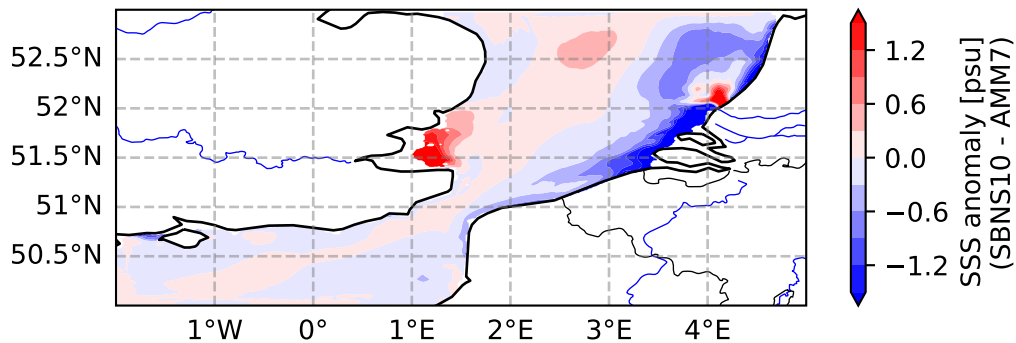


Figure 5.15. Comparison of the yearly averaged sea surface salinity (SSS) modelled by AMM7 and SBNS10 on year 2007. The AMM7 salinity was interpolated on the SBNS10 grid with a bilinear interpolation.

Since the total Rhine inflow (summing the discharge of each neighboring mouth) is not much different in between AMM7 and SBNS10, the change of river forcing cannot explain

the large fresh anomaly appearing at the Rhine and extending south along Belgium. The change of resolution in between the two configurations must instead explain this change. Because the cells are finer, the salinity near the coastline and near river mouths must be better resolved. The finer resolution allows a representation of larger gradient so the salinity plume can also have a smaller salinity. On the contrary, bigger grid cells in AMM7 mean that the fresh water is diluted in larger volumes of water and thus has a larger salinity.

The finer representation of the Scheldt-Rhine plume in the BCZ is highlighted in Fig. 5.16. It has been shown that the Rhine and Scheldt freshwater is advected toward the southwest along Belgian, beside the residual Eulerian transport being in the other way. This is due to the non-linear nature of the flow, making the Lagrangian displacement, the true transport, being opposed to the Eulerian averaged transport [12]. In the low resolution model, it seems that this transport is too weak to advect freshwater till Belgium. The Scheldt plume is not present at all because this river is missing in AMM7. Another important feature represented by the high resolution model is the detachment of the plume from the coast, turning in a clockwise direction due to a residual gyre [13]. This finer scale structure is typically not resolved by low resolution model such as in [14] while being present in finer models such as [9], [8] and [15].

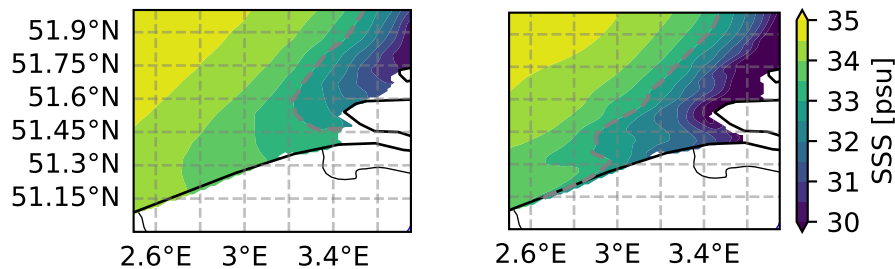


Figure 5.16. Annual SSS modelled by AMM7 (left) and SBNS10 (right) on the BCZ, averaged on year 2007. Grey contour lines show the 33 psu isoline, showing an example of the plume detachment in the SBNS10 model.

The finer horizontal resolution of the model allows also a better representation of the frontal structure in surface and at depth. This can be seen near the Rhine mouth in the transect of Fig. 5.17 that should be compared to Fig. 4.16. In AMM7, the stratification is already present but the front is poorly resolved. Now, with SBNS10, the front is much clearer. The front curving toward the coast at the bottom shows that there is a residual current bringing some salty water under the fresh Rhine water. This pattern is a typical feature of an estuary outlet [12].

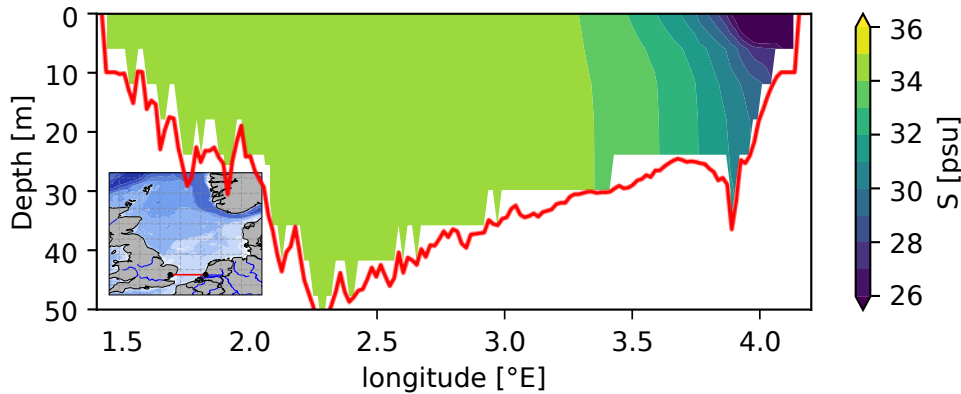


Figure 5.17. Salinity modelled by SBNS10 and averaged on year 2007 across a west-east section, extending from UK (52°N, 1°E) to the Rhine ROFI (52°N, 4.5°E). The red contour is the model bathymetry. The white spaces between the contour lines and the bathymetry are a plotting artifact coming from the interpolation of the model output on a common flat vertical grid, necessary for this plot.

The modelled salinity is assessed against CORA measurements in Table 5.3. Because of the clear lack of measurements in the ROFIs and near the Belgian coast, there is no clear improvement. The high resolution model is similarly performant compared to AMM7 with fresh biases around 0.3 psu, RMSE around 1 psu and correlations about 0.8. A way to correct this fresh bias would be again to consider river sources with non-zero salinities.

	AMM7				SBNS10			
	Bias [psu]	PB [%]	RMSE [psu]	Corr [-]	Bias [psu]	PB [%]	RMSE [psu]	Corr [-]
CORA SSS	-0.31	0.66	0.66	0.77	-0.35	-1.01	0.71	0.79
CORA SBS	-0.51	0.59	0.59	0.94	-0.41	-1.20	0.57	0.88
CORA profiles	-0.34	0.47	0.47	0.85	-0.30	-0.98	0.49	0.84

Table 5.3. Main metrics for the modelled salinities compared to CORA observations for AMM7 and SBNS10.

5.2.5 Concluding remark and possible improvements

In the chapter, the model AMM7 was downscaled to the Southern Bight of the North Sea, refining the model to a resolution of 1 km on the horizontal and keeping the same vertical grid. The surface forcings were kept unchanged while the lateral boundaries were provided by AMM7. The river forcings were switched to EFAS data. Most of the numerical parameters were kept unchanged except the diffusion length scales that were changed according to the new horizontal resolution.

The main improvements that the model SBNS10 shows relies on the better representation of the coastline and the bathymetry. The residual current is more precise along the Belgian coast, and we could show the appearance of a residual current constrained by the bathymetry which is due to a non-linear interaction of the tides and the topography. The tidal currents are also more consistent along the coast.

The tidal elevation is slightly improved for the dominant M2 tide and the M4 but the other constituents are not necessarily better resolved by the model. There are especially large biases for the phase of the constituents. The addition of a "wet and drying" scheme to the model is expected to improve those biases. This scheme should also improve most of the coastal processes as there will be an even more accurate representation of the bathymetry. Indeed, currently the bathymetry is cropped at 10 m to avoid cells with negative volumes during ebb.

The temperature is in general not much different to what is predicted by AMM7. There is a strong cold bias near the Rhine mouth that is due to the river forced only at the surface cell. A further version of the model will distribute the runoff over depth to prevent this unphysical feature³. The bias elsewhere is not substantially improved in the high resolution model because the lateral boundaries that contribute a lot to the heat content of the domain have already a large warm bias coming from AMM7.

For the salinity, there is a clear improvement of the combined Rhine-Scheldt plume extending along the Belgian coast which is consistent with other studies. The fronts are also now well resolved both at the surface and at depth. Though we could see an improvement of the plume structure, we could not assess it because of lacking measurements in ROFIs.

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³Due to time constraint, we could not get a new full year run that includes the new river scheme.

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Part IV

Conclusion

CHAPTER 6

Conclusion

6.1 Thesis summary

The Belgian Coastal Zone is a marine area located in the Southern Bight of the North Sea (SBNS), confined between larger countries like France, United Kingdom and the Netherlands. This territory hosts a lot of economic activities that are in expansion on of the sensible ecosystems present in the region. In particular, Belgium plans to build new offshore windfarms to double its offshore wind power by 2030. To assess the possible impact of such structures and to plan efficiently their layout, a high resolution model simulating accurately the hydrodynamics of the region is needed.

The goal of this thesis was to develop a high resolution model using the ocean circulation model NEMO. We adopted a downscaling approach by choosing first a low resolution model covering a larger area, the NWES. This model was then used to force a similar model but of higher resolution covering only the SBNS.

The low resolution model

For the low resolution model, the AMM7 model developed by the English Met Office was used because it was well documented and because it has been used widely by CMEMS. This configuration was run and its output quantities were analyzed. It was shown that most major features of the North Sea are well predicted by the model.

The main currents such as the slope currents, the Fair Isle current, the Dooley current, etc. are recovered by the model and their associated transports are consistent with other model and observational studies. The only bottleneck is the Baltic boundary which is not well simulated because of the provided SSH that seems too low.

Tides are especially well resolved by the model with biases smaller than 10 cm for the amplitude and 10° for the phase. The tidal currents are also consistent.

The temperature modelled by AMM7 follows a good seasonal pattern and its spatial distribution is physically consistent. The stratification dynamics, with the tidal front settling in summer is well simulated by the model. Although the temperature looks physically accurate, comparison with satellite measurements and *in-situ* profiles show a persistent warm bias of about 1°C. A possible cause would be again the open boundaries. If the SSH is incorrect at the Atlantic boundary, the currents flowing through the boundary can be wrongly simulated and affect the heat budget of the domain.

For the salinity, the main patterns with regions of fresh water influence (ROFIs) are recovered by the model but with usually a large fresh bias. This is caused by the river forcing which considers river sources without salinity, representing badly the outlet of estuaries and large rivers such as the Rhine.

The high resolution model

The high resolution model SBNS10 was created by using most of the same parameters as AMM7. The domain covers the whole SBNS with a resolution of 1 km on the horizontal and keeps the same vertical discretization. AMM7 data is provided at the lateral boundaries. The surface forcing is unchanged. The river forcing was changed using EFAS data.

Because of the higher resolution, new features appear in the residual circulation. On the BCZ, there is a clear transition between the coastal current and the offshore current. More interestingly, some currents following the Belgian sandbanks are present. These currents are due to the interaction between the tidal current and the topography and agree with a theoretical model.

The tides are the bottleneck of this model. The amplitudes are still reasonably predicted by the model, but the high resolution model is not better than AMM7, except for the M2 tide. The phase prediction is worse than AMM7. To improve the tidal prediction, a "wet and drying" scheme should be configured in NEMO in order to consider depths shallower than 10 m in the domain. It would then improve the representation of the coast.

The temperature prediction is not improved in the high resolution model. The problem comes from the lateral forcing that already includes a large warm bias. We also highlighted the importance of the river forcing that should be prescribed over depth and not only at the surface to avoid an unrealistic cold area at rivers outlets.

Though there are not enough data to assess the quality of the modelled salinity, we could show that the Scheldt and Rhine freshwater plume is better resolved by the high resolution model and agrees with other similar studies. The frontal structure looks also better because finer cells are able to represent stronger temperature and salinity gradients. It is expected that the complex frontal circulation is improved due to the finer representation of the front.

6.2 Perspectives

In this thesis, we managed to have a low resolution model covering the NWES that forces a high resolution model of the SBNS. We showed that there are possible improvements but we could not implement them due to time constraint. Thus, the first perspective is to improve both models. For AMM7, the boundary SSH data should be improved to avoid a wrong representation of the Baltic boundary and, maybe, a warm bias appearing due to wrong heat transport at the Atlantic boundary. For SBNS10, there are two direct improvements: (i) Implement the "wet and drying" scheme which is possible in NEMO

but currently not well documented. (ii) Apply the river forcing over depth. The latter is straightforward.

Once the models are improved, the door is open to study practical topics in the BCZ. The hydrodynamic model can be used to assess the physical effect of offshore windfarms on the BCZ if one add a drag force to the model. By coupling a biogeochemical model to NEMO, the effect of offshore windfarms on the sediments distribution and the general ecosystems can be studied. Other challenges such as the eutrophication and pollutants transport can be studied with the same model.

Appendices

APPENDIX A

Linear regression for tidal analysis

To perform a tidal analysis, we want to find the sine and cosine amplitude of each tidal constituent, given a discrete observed signal $\eta(t_n)$ approximated by

$$\tilde{\eta}(t_n) = \tilde{\eta}_n = A_0 + \sum_{k=1}^{N_{tide}} A_k \sin(\omega_k t_n) + B_k \cos(\omega_k t_n) \quad (\text{A.1})$$

$$= A_0 + \sum_{k=1}^{N_{tide}} A_{tot,k} \sin(\omega_k t + \phi_k), \quad (\text{A.2})$$

with N_{tide} the number of tidal constituents considered in the regression, $n = 0, 1 \dots N-1$ and N the number of observations. Since the signal is linear in terms of the tidal amplitudes A_k and B_k , a linear regression method can be employed, here using the least squares method.

The least squares method finds the coefficients A_k and B_k that minimize the quadratic error

$$e^2 = \sum_{n=0}^{N-1} |\eta(t_n) - \tilde{\eta}(t_n)|^2. \quad (\text{A.3})$$

Adopting a tensorial formalism, each equation can be written as

$$\boldsymbol{\eta} = (\eta_0, \eta_1 \dots \eta_{N-1})^T, \quad (\text{A.4})$$

$$\tilde{\boldsymbol{\eta}} = (\tilde{\eta}_0, \tilde{\eta}_1 \dots \tilde{\eta}_{N-1})^T, \quad (\text{A.5})$$

$$\boldsymbol{\beta} = (A_0, A_1, B_1 \dots A_{N_{tide}}, B_{N_{tide}})^T, \quad (\text{A.6})$$

$$\mathbf{X} = \begin{pmatrix} 1 & \sin(\omega_1 t_0) & \cos(\omega_1 t_0) & \sin(\omega_2 t_0) & \dots & \cos(\omega_{N_{tide}} t_0) \\ 1 & \sin(\omega_1 t_1) & \cos(\omega_1 t_1) & \sin(\omega_2 t_1) & \dots & \cos(\omega_{N_{tide}} t_1) \\ \vdots & \vdots & \vdots & \vdots & \ddots & \vdots \\ 1 & \sin(\omega_1 t_{N-1}) & \cos(\omega_1 t_{N-1}) & \sin(\omega_2 t_{N-1}) & \dots & \cos(\omega_{N_{tide}} t_{N-1}) \end{pmatrix}, \quad (\text{A.7})$$

$$\tilde{\boldsymbol{\eta}} = \mathbf{X}\boldsymbol{\beta}, \quad (\text{A.8})$$

$$e^2 = \|\mathbf{X}\boldsymbol{\beta} - \boldsymbol{\eta}\|^2. \quad (\text{A.9})$$

By cancelling the partial derivative of the quadratic error with respect to each component of $\boldsymbol{\beta}$, one can show that the tidal amplitudes that minimize the error are given by

$$\boldsymbol{\beta} = (\mathbf{X}^T \mathbf{X})^{-1} \mathbf{X}^T \boldsymbol{\eta}. \quad (\text{A.10})$$

Once the sine and cosine coefficients are known, the total amplitude and phase can be obtained by the relations

$$A_{tot,k} = \sqrt{A_k^2 + B_k^2} \quad (\text{A.11})$$

$$\tan(\phi_k) = \frac{B_k}{A_k}. \quad (\text{A.12})$$

To avoid an ambiguity with the tangent function here above, the phase will always be taken in the range -180° to $+180^\circ$ throughout all this thesis.

In order to have an accurate regression, the number of observations and their frequency must be large enough. We found that 3 months of hourly output are enough to have converged coefficients for the dominant M2 and S2 tides while a whole year is needed for smaller tides such as MU2.