Numerical model study of the influence of the Rhine-Scheldt estuary on the water exchange with the Flemish harbours

Gilles Neerinckx

Promotor: prof. dr. ir. Peter Troch
Begeleider: Gijsbert Van Holland

Masterproef ingediend tot het behalen van de academische graad van Master in de ingenieurswetenschappen: bouwkunde

Vakgroep Civiele Techniek
Voorzitter: prof. dr. ir. Julien De Rouck
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Acknowledgements

This master thesis reflects the completion of my studies as civil engineer at Ghent University. The goal of this work is to improve the knowledge of the hydrodynamics in the Belgian coastal area.

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Gilles Neerinckx, juni 2012
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Numerical model study of the influence of the Rhine-Scheldt estuary on the water exchange with the Flemish harbours

The influence of the Rhine-Scheldt estuary on the salinity along the Belgian coast

doors

Gilles Neerinckx

Scriptie ingediend tot het behalen van de academische graad van
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Summery

The various processes that shape the ROFI (Region Of Fresh water Influence) in the Rhine-Scheldt estuary are described. More specifically the southern spreading of the fresh water is studied. First a literature study and measurement data analysis is carried out to understand the relevant processes. This knowledge is used to make a numerical model about our study area. Finally a sensitivity analysis is executed to examine the influence of the different parameters.

Keywords

Numerical model, ROFI (Region Of Freshwater Influence), Belgian coast, Salinity
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Gilles Neerinckx

Supervisor(s): Gijsbert Van Holland, Bart Verheyen, Peter Troch

Abstract—This article describes the various processes that shape ROFI (Region Of Fresh water Influence). More specifically the southern spreading of the fresh water is studied. First a literature study and measurement data analysis is carried out to understand the relevant processes. This knowledge is used to make a numerical model about our study area. Finally a sensitivity analysis is executed to examine the influence of the different parameters.

Keywords—Numerical model, ROFI (Region Of Freshwater Influence), Belgian coast, Salinity

I. INTRODUCTION

RIVER outflows are a major source of land-drained materials and of fresh water. The fresh water spreads as a surface buoyant layer (a river plume) away from the discharge point. The distribution of salinity within a continental shelf sea is determined by processes within the sea (transport and mixing), by boundary exchanges of fresh water (river discharge, precipitation and evaporation) and of salty water (at the oceanic boundary). Salinity is a handy tool to reflect the scale of freshwater influence and, in other words, to reflect the scale of dissolved conservative contaminants which behave like salinity.

This abstract will start with a short study of the literature and measurement data, where the relevant processes in the development of the ROFI (Region of Freshwater Influence) are described. Then a numerical model is developed to simulate this behavior. The focus of this thesis concerns the analysis of the model results in which the influence of different parameters on the fresh water exchange are examined. Finally conclusions are made.

II. LITERATURE

Previous numerical model studies [2] to the relative contribution of both the Schelde and the Rhine to the Belgian shore showed that the Rhine is largely responsible for the band of lower salinity water along the Belgian and Dutch coasts. The Rhine water spreads a considerable distance southward from the estuary mouth. On the other hand, Schelde water is limited mainly to the estuary mouth and to a lesser extent to near-shore and central Belgian waters.

The fresh water distribution is maintained by the residual circulation pattern of the North Sea, induced by the prevailing winds, the tide and by the density field itself [1]. These tidal and wind driven residual currents are responsible for a net transport northwards through the ROFI.

III. DATA ANALYSIS

Measurement data from the Hydrologisch Meteorologisch Centrum Zeeland [3] is used from the year 2006. It can be seen that the peaks in salinity generally correspond with high tide. This means that at high tide saltier water flows to the shore and during ebb, more fresh water flows out of the mouth.

When the salinity time series are plotted, an annual pattern is revealed. In the beginning of the year salinity values decrease until mid-June from where it slowly starts to increase and finally the salinity turns back into a decrease from November. This development corresponds to the drainage outflow of the rivers: in the winter months a larger flow rate exists and from the beginning of July this drain flow declines and starts back to rise in November.

Influences of the wind can be noticed in the salinity values. When a wind event in northern direction occurs, a decrease can be noticed in the salinity values. When on the other hand a wind event in southwest direction occurs, the salinity values increases.

IV. MODEL DESCRIPTION

For this master thesis, the numerical modeling tools from IMDC are used. These modeling tools consist of large-scale model of the Southern North Sea (KaZNo-model) and a detailed model of the Belgian coast and Schelde. The software package used is Delft3D, developed by Delft Hydraulics. This large-scale model is usually used for studies along the Belgian coast. The model is already able to calculate the tides with good accuracy for the French, Belgium and Holland coast.

Starting with initial water level and salinity values respectively 0 m and 35 ppt, the model has been spun up over the whole year 2006 to get the salinity distribution approximately right. Later simulations started with an equilibrated initial condition starting in 2006 for a shorter period.

V. MODEL RESULTS

By means of a sensitivity analysis, the influence of the different parameters on the fresh water transport can be examined.

Initial sensitivity experiments begin with a reference model run, over the period 7 March 2006–1 July 2006 using a time step of 15 s and a diffusivity coefficient of 10 m²/s. The anticipation of the tide results in a sort of circulating motion of the fresh water. By ebb the fresh water flows to the southwest, at turning point the river plume is pushed against the shore after which it flows back to the northeast with high tide. During the transition
from flood to ebb, the river plume expands in offshore direction. To get an idea of the residual flow pattern in the study area, a simulation is started with simulation time 12 u 30 min, the duration of one tide corresponding to the spring cycle. As an indicative measure of the residual current, the residual transport vector $v_{res}$ is taken. From the residual flow pattern, it becomes clear that, in general, a nett transport exists in northward direction. Closer to the shore irregularities such as eddies occur. There also arises a strip where a small residual flow to the southwest exists. The pattern shows that southward transport of salinity can be a result of advection processes as well and not only of diffusion. A new conceptual model for the southward transport of fresh water arises (figure 1): During ebb, sea water flows along the coast to the south west together with a part of the fresh water from Maasmond and Haringvliet (1, 3). During high tide, the fresh water is pushed back to the northeast. A fraction of the fresh water is hereby caught by the overlying estuary (2, 4). This fraction of fresh water flows further southwards during the following ebb (3, 5). This way fresh water gradually moves to the south. The distribution of the fresh water as a result of the tide is a slow process. Hence a long simulation time is needed to obtain a situation of equilibrium.

Fig. 1. New conceptual model of water transport to the south due to tidal process.

A series of tests is now carried out adjusting the discharge rate, diffusivity coefficient and adding wind influences. Experiments similar to the reference run are performed with different values for the diffusivity coefficient. The diffusivity coefficient is changed to respectively $1 m^2/s$ and $100 m^2/s$. A diffusivity coefficient that is too low, results in fresher water along the coast. An increase of the diffusivity coefficient results in higher salinity values near the coast and a less steep salinity profile. The discharge rate of the rivers is changed compared to the reference model run. Although the southward extend of the plume decreases with decreasing discharge rate, changes in the freshwater rate do not significantly alter the shape of the plume compared to the reference run. But for a lower discharge rate the plume is slower in progress. The behavior of the fresh water under wind load is tested on the basis of 4 simulations. A uniform wind field is imposed of $6 m/s$ with wind direction successively to the north, east, south and west. The results are consistent with what is learned from the literature. Wind direction to the north (downwelling) forces freshwater to the shore. The plume becomes narrower. With wind direction to the south (upwelling) the freshwater spreads offshore. Because the fresher surface water moves offshore, saltier water flows onshore along the bed. Hence less fresh water than in downwelling situation. Eastern wind direction causes an increase in salinity at the Belgian coast in contrast to the western wind direction, which induces a decrease in salinity. Additional simulations are carried out with consecutively setting the discharge of the Scheldt, Haringvliet and the Nieuwe Waterweg on and off. Logically, the most of the fresh water in the Belgian coastal waters is originating from the Scheldt. But the amount of fresh water in proportion to the distance of the fresh water source to the measuring point is low compared to Haringvliet and the Nieuwe Waterweg. The results also show that, despite the discharge rate of the Nieuwe Waterweg is 20 times greater than the discharge rate of Haringvliet, the contribution of Haringvliet appears to be greater than the contribution of the Nieuwe Waterweg at the Belgian coast.

VI. CONCLUSIONS

The three most important driving factors for the residual salinity profile are tide, wind and a density gradient. With regard to the southward distribution of the fresh water, the tide ensures the largest transport of fresh water. Wind processes and changes in discharge rates can only extend the ROFI a little further to the south. A new conceptual model of the southward distribution of the fresh water is suggested in this study.

REFERENCES

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Rhine-Scheldt estuary on the water exchange with
the Flemish harbours

Gilles Neerinckx

Supervisor(s): Gijsbert Van Holland, Bart Verheyen, Peter Troch

Abstract—This article describes the different processes that occur in
the ROFI (Region of Freshwater Influence). In addition, a numerical model
is presented to study the influence of the different parameters on
the salt balance. First, a reference run is made to study the
salt distribution as initial condition. Finally, the conclusions
are drawn from the different parameters under study.

Keywords—Numerical model, ROFI (Region Of Freshwater Influence),
Belgisch kustgebied, Saliniteit

I. INLEIDING

RIVIER discharges are a significant source of freshwater and ocean water.
The ocean water is discharged as a drier (river plume) on the coast of the
North Sea and the Schelde. Averaging over a year, the river plume
extends over a considerable distance along the Belgian and Dutch
coasts. The river plume is used to study the influence of the different
parameters on the salt distribution.

Deze samenvatting begint met een korte literatuurstudie en data
analyse. Hierin worden de relevante processen in de versprei-
ding van de ROFI (Region Of Freshwater Influence) beschreven. Daarna
wordt een numeriek model opgesteld dat deze processen
simuleert. De klemtoon van deze thesis ligt in de interpretatie
van de ROFI en dus de omvang van de opgeloste
contaminanten, die zich gedragen als het zoutgehalte, te weer-
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contaminanten, die zich gedragen als het zoutgehalte, te weer-
spiegelen.

II. LITERATUURSTUDIE

Eerder gemaakte numerieke studies [2] naar de relatieve bij-
drage van zowel de Schelde en de Rijn aan het Belgische zeewa-
toone aan dat de Rijn een belangrijke bron aan grondmate-
rielen en zoet water is. Het zoet water verspreidt zich als
een drijvende laag (rivierpluim) op het zoutere en dus zwaar-
der zeewater. De verdeling van het zoutgehalte in de Noord-
zee wordt bepaald door processen binnen de zee (adveictie en
menging), door uitwisseling van zoet water aan de grenzen (ri-
vierafvoer, neerslag en verdamping) en van zout water (aan de
offshore randvoorwaarden). Saliniteit is een handig hulpmiddel
om de omvang van de ROFI en dus de omvang van de opgeloste
contaminanten, die zich gedragen als het zoutgehalte, te weer-
spiegelen.

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van de ROFI en dus de omvang van de opgeloste
contaminanten, die zich gedragen als het zoutgehalte, te weer-
spiegelen.

III. DATA-ANALYSE

In de saliniteitscurve is een jaarlijks patroon waar te nemen. In
het begin van het jaar wordt het zoutgehalte een daling ge-
waar tot aan midden juni van waar het langzaam terug begint
te stijgen. Tenslotte begint het zoutgehalte terug te dalen vanaf
november. Dit verloop komt overeen met de afvoer van de zoet-
waterbronnen: in de wintermaanden bestaat er een groter debiet,
vanaf begin juli neemt deze afvoer af en begint weer te stijgen
in november.

Er kunnen ook invloeden van de wind worden opgemerkt in de
meetgegevens. Als er een wind evenement in noordelijke richt-
ing optreedt, kan een daling worden opgemerkt in het zoutge-
halte. Wanneer daarentegen een wind in zuidwestelijke richting
plaatsvindt, neemt het zoutgehalte toe.

IV. NUMERIEK MODEL

Voor deze master scriptie kon gebruikt gemaakt worden van
het numierk instrumentarium van IMDC. Deze numerieke mo-
dellen bestaan uit een grootschalig model van de Zuidelijke
Noordzee (KaZNo-model) en een gedetailleerder model van de
Belgische kust en de Schelde. De gebruikte software is Delft3D,
ontwikkeld door Delft Hydraulics. De gebruikte modellen wor-
den veelal gebruikt voor studies in het Belgische kustgebied en
zijn reeds in staat om met goede benadering de getijden langs-
heen de Franse, Belgische en Nederlandse kust te modelleren.

Als initiële omstandigheden worden de waterstand en het zoutgehalte
respectievelijk de waarden 0 m en 35 ppt meegegeven. Met deze
beginccondities wordt een simulatie over het hele jaar 2006 uitge-
voerd om de saliniteitsverdeling min of meer realistisch te krij-
gen. Latere simulaties worden dan gestart met deze saliniteits-
verdeling als beginconditie.

V. MODELRESULTATEN

Door middel van een gevoeligheidsanalyse kan de invloed van
de verschillende parameters op het zoet watertransport worden
onderzocht. Eerst wordt een referentie model run gestart, over
De periode 07 maart 2006–1 juli 2006 met een tijdstap van 15 s. De inwerking van het getij resulteert in een soort circulerende beweging van het zoet water. Tijdens eb stroomt het zoete water naar het zuidwesten, op keerpunt wordt de rivierpluim tegen de kust geduwd waarna het terug vloeit naar het noordoosten tijdens hoog water. Tijdens de overgang van vloed naar eb zet de rivierpluim in offshore-richting uit. Om een idee te krijgen over het residuul stromingspatroon van het onderzoeksgebied wordt een simulatie gestart met simulatieperiod 12 u 30 min. Dit is de duur van 1 getij tijdens een springtij. Als maat voor de residuul stroming wordt de residuul transport vector \( v_{res} \) genomen.

Uit het residuul stromingspatroon blijkt dat er in het algemeen een nettotransport bestaat in noordelijke richting. Dichter bij de kust ontstaan onregelmatigheden zoals wervelstromen. Er ontstaat ook een strook waar een kleine reststroom naar het zuidwesten bestaat. Het stromingspatroon toont dat zot watertransport naar het zuiden ook een gevolg kan zijn van advectieve processen en niet enkel van diffusieve processen.

Een nieuw conceptueel model aangaande het zuidwaarts transport van zoet water ontstaat (figuur 1):

Tijdens eb stroomt zeewater, samen met een deel van het zoet water uit Maasmond en Haringvliet, langs de kust naar het zuidwesten (1, 3). Bij vloed wordt het zoetere water teruggedrongen naar het noordoosten. Een facietie van het zoet water wordt opgevangen door de bovenliggende monding (2, 4). Deze facietie van vers water stroomt verder zuidwaarts tijdens de volgende eb (3, 5). Op deze manier beweegt een deel van de zoetwaterafvoer zich geleidelijk naar het zuiden. De verdeling van het zoetwater als gevolg van het getij gebeurt langzaam. Vandaar dat een lange simulatietijd nodig is om een evenwichtstoestand te verkrijgen.

![Fig. 1. New conceptual model of water transport to the south due to tidal process.](image)

Een reeks simulaties worden nu uitgevoerd waarbij de afvoer, diffusiviteitscoefficient worden aangepast en wind-invloeden worden toegevoegd. Experimenten vergelijkbaar met de referentie run worden uitgevoerd met verschillende waarden voor de diffusie coëfficient. De diffusiviteitscoefficient wordt 10 keer kleiner en 10 keer groter genomen dan de diffusiviteitscoefficient van de referentiesimulatie. Een diffusiviteitscoefficient die te laag is, resulteert in vers water langs de kust. Een verhoging van de diffusiviteitscoefficient resulteert in hogere zoutwaarden dicht bij de kust en een minder steile saliniteitsprofiel.

In de volgende simulaties wordt het debiet van de rivieren ge- wijzigd ten opzichte van het referentimodel run. Hoewel de zuidelijke verspreiding van de pluim afneemt bij afnemende afvoer zorgen veranderingen in de zoetwaterafvoer niet voor significante veranderingen in de zuidelijke verspreiding van de pluim ten opzichte van de referentie run. Een lager debiet zorgt wel voor een tragere ontwikkeling van de rivierpluim.

Het gedrag van het zoetwater onder windbelasting wordt getest aan de hand van 4 simulaties. Een uniform windveld wordt opgelegd met amplitude \( \text{t/m/s} \) en windrichting achtereenvolgens naar het noorden, oosten, zuiden en westen. De resultaten zijn in overeenstemming met wat in de literatuur beschreven wordt. Windrichting naar het noorden (downwelling) dwingt zot water naar de kust toe. De pluim wordt smaller. Met de windrichting naar het zuiden (upwelling) wordt het zoetwater offshore verspreid. Omdat zot oppervlaktezoeewaterwaarts beweegt wordt dit in evenwicht gehouden door zot water dat via de bodem naar de kust toe stroomt. Zo ontstaat er, langs de kust, zouter water dan in een downwelling situatie. Oostelijke windrichting veroorzaakt een toename van het zoutgehalte aan de Belgische kust, in tegenstelling tot een windrichting naar het westen, wat een daling van het zoutgehalte induceert.

De invloed van de verschillende afvoeren wordt onderzocht door bijkomende simulaties uit te voeren waarin achtereenvolgens de afvoer van de Schelde, Haringvliet en de Nieuwe Waterweg aan en uit worden gezet. Logischerwijs is het grootste deel van het zoet water in de Belgische kustwateren afkomstig van de Schelde. Maar de hoeveelheid zot water in verhouding tot de afstand van het meetpunt tot de zoetwater bron is laag in verge lijking met Haringvliet en de Nieuwe Waterweg. De resultaten tonen ook aan dat, ondanks de het debiet van de Nieuwe Waterweg 20 keer groter is dan de afvoer van Haringvliet, de bijdrage van Haringvliet aan de Belgische kust toch groter blijkt dan de bijdrage van de Nieuwe Waterweg.

VI. BESLUIT

De drie belangrijkste processen die het saliniteitsprofiel bepalen zijn het getij, wind en de dichteitsgradient. Met betrekking tot de zuidelijke verspreiding van het zoetwater zorgt het getij voor het grootste transport van zoetwater. Wind processen en veranderingen in de afvoer kunnen de verspreiding verder beno venen. Tot slot wordt een nieuw conceptueel model van het zuidelijk transport van zoetwater voorgesteld in deze studie.

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Chapter 1

Introduction

1.1 Background

River outflows are a major source of land-drained materials and of fresh water. The coastal ocean, the recipient of these river waters, is a region of more saline and hence denser sea water. The fresh water spreads as a surface buoyant layer (a river plume) away from the discharge point. The distribution of salinity within a continental shelf sea is determined by processes within the sea (transport and mixing), by boundary exchanges of fresh water (river discharge, precipitation and evaporation) and of salty water (at the oceanic boundary). Satellite images show that these plumes tend to turn to the right in the Northern Hemisphere after entering the sea. This is due to the rotation of the Earth. As a result of this, a coastal current develops, that is capable of transporting the relatively fresh and buoyant plume over considerable distances along the coast. The salinity transport at the water surface and bottom is negligible as well as the variation in salinity through biological or chemical interactions. This makes salinity a conservative quantity transported mainly by advection and diffusion processes. Salinity is thus a handy tool to reflect the scale of freshwater influence and, in other words, to reflect the scale of dissolved conservative contaminants which behave like salinity. The region of fresh water influence (ROFI) also plays an important role in the transport and deposition of suspended particulate matter (SPM). SPM implies the transport of sediments, which influences the coastal morphology, siltation of shipping lanes and harbour entrances. An actual example is the turbidity maximum near Antwerp. This results in high suspended sediment concentration. Turbulence is related to the salinity profile which proves the importance of this thesis. SPM also consists of nutrients, plankton and larvae. In addition contaminants such as heavy metals can be adsorbed to the SPM.
Chapter 1. Introduction

At the same time the dispersive character of river plumes is increasingly used to flush away industrial and agricultural waste products. Consequently, the ability to model salinity is an essential step in the development of water quality models.

In this study, the influence of the Rhine-Scheldt estuary is modeled to understand the principles of the ROFI. Special attention is paid to the southern transport of the freshwater, examing the water exchange with the Flemish harbours.

1.2 Outline of the thesis

This study will start with a short literature study where the relevant processes in the development of the ROFI are described. Next an analysis is made of salinity data received from HMCZ. Before an analysis of the model results is made, first a description and validation of the model has been done. The focus of this thesis concerns the analysis of the model results in which the influence of different parameters on the fresh water exchange are examined. Finally conclusions and recommendations are made.
Chapter 2
Literature Study

2.1 Salinity

The salinity of an ocean or sea refers to the amount of salt in the ocean or sea. Until recently, a common way to define salinity values has been parts per thousand (ppt), or kilograms of salt in 1,000 kilograms of water. Today, salinity is redefined in the Practical Salinity Scale (PSS) with unit: practical salinity units (psu), a more accurate but more complex definition. Nonetheless, values of salinity in ppt and psu are nearly equivalent. PSS is defined, by UNESCO/ICES/SCOR/IAPSO Joint Panel on Oceanographic tables and Standards, as the ration of conductivity of a seawater sample to a standard KCl-solution. The salinity of seawater has on average a value of 35 PSS. Seawater can also be called as euhalien water (PSS-value of 30 to 35). Brackish water has a salinity of between 0.5 and 30 and metahalien water from 36 to 40. Water with a very high salinity is called pickle.

<table>
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<td>Fresh water</td>
<td>0.0–0.5</td>
</tr>
<tr>
<td>Brackish water</td>
<td></td>
</tr>
<tr>
<td>Oligohalien</td>
<td>Slightly brackish water</td>
</tr>
<tr>
<td>Mesoahalien</td>
<td>Moderately brackish water</td>
</tr>
<tr>
<td>Polyhalien</td>
<td>Very brackish water</td>
</tr>
<tr>
<td>Salt water</td>
<td></td>
</tr>
<tr>
<td>Euhalien</td>
<td>30.0–35.0</td>
</tr>
<tr>
<td>Metahalien</td>
<td>36.0–40.0</td>
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Generally, one can assume that the salinity transport between the boundaries air-sea and
sea-bottom is low. In addition, the variation in the salinity as a result of the biological or chemical interaction is negligible. Salinity can be considered as a conservative parameter, largely transported by advection and diffusion processes and is therefore a good tracer for water masses [Lacroix et al. 2004]. The salinity of ocean water is thus a useful tool to determine the fraction of water from river discharge and thus to display the extent of the freshwater influence.

2.2 Classification of the North Sea

On the basis of salinity and temperature data, measured and recorded for almost a century, following classification was made [Lee 1980]:

![Figure 2.1: Classification of the North Sea](image)

The different water masses are:

- A1: Scottisch coastal water
Chapter 2. Literature Study

- A2: English coastal water
- A3: Channel water
- A4: Continental coastal water
- A5 and A6 are transitional areas between A4 and B
- B: Northern North Sea water
- C: Skagerrak and the Norwegian Rinne
- D: transitional water to C

The figure shows that only the Channel water (A3), passing through the street of Dover in the central part of the North Sea, and the continental coastal water are relevant for the ROFI of the Rhine-Scheldt estuary. The continental coastal water is a strip of less denser water that begins east of Calais and drifts along the Belgian-Dutch-German coast. This strip can be seen as a coastal river that receives fresh river water, along with its nutrients, from the Seine, Scheldt, Rhine/Meuse and Elbe. From the measurements, one can observe a strong salinity gradient perpendicular to the Belgian coast. When observed more in detail, a clockwise residual gyre can be seen in the shallower region seaward of the river mouth. This current carries water from the Scheldt estuary south-westward and then offshore before merging with the north-eastward flow [Nihoul and Hecq, 1984]. Previous numerical model studies [Lacroix et al., 2004] to the relative contribution of both the Scheldt and the Rhine to the Belgian sea showed that the Rhine is largely responsible for the band of lower salinity water along the Belgian and Dutch coasts. The Rhine water spreads a considerable distance southward from the estuary mouth. On the other hand, Scheldt water is limited mainly to the estuary mouth and to a lesser extent to near-shore and central Belgian waters. Additional simulations were carried out with consecutively setting the discharge of the Scheldt on and off. Comparing results obtained with and without Scheldt discharge does not show significant differences in the salinity field of the Belgian waters (Belgian Economic Exclusion Zone, EEZ). The greatest difference is seen close to the Scheldt discharge. Even with no Scheldt discharge, the salinity within the Scheldt Estuary decreased. Although it is a conclusion based on a numerical model with lack of spatial resolution, this simulation does provide a warning that the conventional assumption that freshwater within the Scheldt Estuary originates solely from the Scheldt river may not be true.
Chapter 2. Literature Study

2.3 Residual current

The fresh water distribution is maintained by the residual circulation pattern of the North Sea, induced by the prevailing winds, the tide and by the density field itself (de Boer, 2008). The tidally induced residual flows are largest along the coastal regions where the Kelvin wave amplitudes are largest and are directed in the same direction as the propagation of the Kelvin wave (Nihoul and Ronday, 1975; Prandle, 1978; Teixeira, 1979; Prandle, 1984; Ye and Garvine, 1998). A second source for the residual current is the wind. The average WSW-wind, the predominant cyclonic rotation of the wind fields in combination with the bottom slope of the North Sea are responsible for a cyclonic residual current (Otto et al., 1990). These tidal and wind driven residual currents are responsible for a net transport northwards through the ROFI. The third source of residual flows, was mostly considered as negligible. However, closer to the coast where lower densities prevail, the cyclonic circulation pattern is significantly enhanced by density fields (de Boer, 2008).

To be able to describe the density driven residuals in the Southern North Sea, the knowledge of the density fields is required. These fields are governed by both salinity and temperature differences. The salinity difference is important near the coast, where the temperature effects are absent. In contrast with the central North Sea, where temperature is important and salinity differences are small. Temperature patterns show a pronounced seasonal cycle. In the winter the temperature of the Channel is 7 to 9°C, while the coastal waters are colder due to colder river runoff. In the summer time, the Channel water is relative cool (16 to 17°C) and the coastal waters are warmer due to warmer land runoff and more shallow depth. However, according to De Kok et al. (2001), the main density differences are due to the salinity structure and not to temperature. Pronounced temperature patterns do occur, but mainly as a response to haline stratification. The density structure in the Rhine ROFI is the result of the prevailing alongshore currents, which are mostly governed by the local density field. The dominant cross shore salinity gradient, in combination with Earth rotation (Coriolis effect), is responsible for this alongshore residual flow in northeast direction. This is called a coastal current (de Boer, 2008).

The water masses also have an important vertical structure with respect to density and velocity. In the Rhine ROFI the density driven flow dominates over tidal and wind driven

---

1 A Kelvin wave is an oceanic or atmospheric wave with long period and small amplitude. It moves parallel to a fixed boundary and is subjected to the Coriolis force, so that the amplitude of one of the two sides is larger. Kelvin waves always propagates with the coastline on their right side on the Northern Hemisphere (VLIZ, 2012).
residuals (de Boer, 2008). The description of the vertical velocity structure of a stationary
density driven flow is given by Heaps (1972). A northward velocity vector establishes
parallel to the coast in geostrophic equilibrium with the baroclinic and barotropic pressure
gradient perpendicular to the coastline. This is analogous to the principle of a gradient flow
(figure 2.2). The gradient force allows the water to flow from high to low pressure. Once
movement occurs, the Coriolis effect takes place forcing the gradient to flow eventually
across the gradient force, along the isobars (to the right in the Northern Hemisphere, to
the left on the Southern Hemisphere).

\[ \text{Figure 2.2: Geostrophic current (Wikipedia, 2012)} \]

de Boer (2008) represents the principle schematically in figure 2.3. The sum of the
barotropic and baroclinic pressure leads to an offshore force on the surface layers and an
onshore force on the bed layer. A barotropic pressure gradient (dotted line) would result
in a depth averaged offset of the alongshore velocity. In a frictionless case the cross shore
pressure gradient would be balanced only by the Coriolis force generated by an alongshore
velocity profile with constant shear. This limit case is called “thermal wind”. In contrast,
when friction is dominant, the cross shore pressure gradient would be balanced only by the
non-uniform velocity shear generated by the cross shore exchange current. This limit case
is called the “estuarine/gravitational circulation”. Heaps (1972) described the joint action
of the alongshore thermal wind and cross-shore estuarine circulation velocity profiles.
As mentioned before, the southern part of the North Sea is a shallow sea so that fric-

\[^2\]A baroclinic atmosphere is one for which the density depends on both the temperature and the pressure.
This is in contrast with barotropic atmosphere, for which the density depends only on the pressure.
tion must be taken into account. In the cross shore direction there is a gravitational flow comparable to the salt exchange current in an estuary.

Figure 2.3: Horizontal density gradients (de Boer 2008)
The horizontal density gradients are not uniform. On average the cross shore density gradients are one order of magnitude larger than the alongshore density gradients \cite{deBoer2008}. These cross shore density gradients diminish rapidly offshore. Because the density gradients determine the magnitude of the alongshore and cross shore currents, these currents tend to diminish also offshore.

Additionally, stratification can occur. \cite{Simpson1990} already discussed the influence of the tides and the cross shore density gradient. In addition, other processes such as surface heating and mixing produced through mechanical stirring by winds are also important with regard to stratification. On the figure below isolines of salinity are shown. These isolines are initially vertical at the start of the ebb. This arrangement is disturbed by differential displacement, with the lighter surface water moving faster seaward and overtaking heavier more saline water in the lower layers and thus a stable structure. Vertical mixing by wind stress near the surface and tidal stress at the bottom will tend to transform the structure into a two-layer profile with a sharp halocline \cite{Simpson1990}.

![Figure 2.4: Tidal Straining, Density Currents, and Stirring in the Control of Estuarine Stratification (Simpson et al., 1990)](image)

Observations in a region of Liverpool Bay \cite{Simpson1990} showed that, with the same frequency as the semidiurnal tide, stratification occurred alternating with periods of complete vertical mixing. These periods apply for several hours. The maximal stratification

---

\footnote{A halocline is the transition zone between water layers of different salinity. Fresh and salt water layers mix barely there, the fresh water layer “float” on the salt water. This creates a halocline or salinity jump.}
occurs close to low water slack. A continuous period of stratification of 3 days develops at the time of minimum mixing around neaps. Comparison of the phase of the relative displacement between the top and bottom current meters with that of the difference in salinity suggests the operation of the tidal straining mechanism. These observations have led to the development of a theory: Strain Induced Periodic Stratification (SIPS). In general, this theory shows that when the tidal flow is directed offshore, it acts on the offshore salinity gradient inducing to stratification whilst when the flow is onshore the shear in the tidal current structure acts to destabilize the water column causing overturning and destruction of the stratification. A 3D-modeling is thus necessary to take these effects into account.

2.4 River plume

We now have assumed for the general case that in the alongshore direction the density was homogeneous. In reality this is usually not correct (figure 2.5). Due to river discharge, meanders and baroclinic instabilities, significant along-shore gradients can occur in certain positions (De Kok 1996). The most important source for anisotropy is the river mouth itself. At the location where the river discharge takes place, an anti-cyclonically rotating bulge of less saline water is formed (figure 2.6). After a cyclonic turn this bulge feeds into a coastal current in the direction of the Kelvin wave i.e. with the coastline on its right side. The system of bulge and the coastal current is known as river plume.

![Figure 2.5: Density current alongshore (de Boer 2008)](image-url)
Chapter 2. Literature Study

Figure 2.6: Bottom and surface current in a river plume

The flow at the surface follows the isopycnals \(^4\) whereas near the surface there is a cross-isopycnal current (de Boer 2008). Near the surface there is an anti-cyclonic outward spiraling and near the bottom there is a cyclonic inward spiraling flow.

Chao and Boicourt (1986) used a simple 3d-model to simulate the behavior of a plume. It took the plume 10 days to evolve. By modeling Kourafalou et al. (1996) found that in the absence of a surrounding current, the bulge continued to grow indefinitely. Fong and Geyer (2001) saw that in this case the coastal current wasn’t transporting the whole river discharge, the remaining part recirculated in the continually growing bulge. They found that a steady state could be achieved only in the case of an ambient, alongshore current.

Kourafalou et al. (1996), used an idealized boxmodel. The parameters that were most influential in determining the shape of the plume were: Horizontal eddy viscosity/diffusivity, vertical eddy viscosity/diffusivity, the freshwater discharge rate and the depth of the receiving basin. They described the shape of river plume based on \( \lambda = L/L_c \). \( L \) is defined as the seaward extent of the bulge (taken as cross-shore distance from the channel mouth to the plume boundary) and \( L_c \) as coastal current width (taken as distance from the coast to the plume boundary across the head of the coastal current, where the northward velocity is maximum). If \( \lambda > 1 \) the plume is called supercritical and if \( \lambda < 1 \), the plume is called subcritical. When \( \lambda \) is close to 1, the plume exhibits a diffusive behavior, the bulge and

\(^4\) An isopycnal is a surface of constant potential density of water.
the coastal river have merged \( (L = L_c) \) and the plume dissipates energy in the offshore direction.

Without any imposed mechanisms such as wind and tides, a meandering pattern appeared. This result agreed with the numerical study on riverine buoyant plumes by Oey and Mellor (1993). They concluded that the meandering of the coastal current had two stages of development: a barotropic instability stage characterized by short wavelengths (20 km) and a baroclinic instability stage characterized by long wavelengths (60 km).

If the behavior of the buoyant river plume under wind load was tested, it was found that upwelling increases the vertical stratification. See figure 2.7 for a schematic representation of upwelling. Upwelling forces fresher surface waters offshore, in order to compensate, saltier water flows onshore along the bed. For a sufficient wind velocity and a certain time scale, Hickey and Hamilton (1980) found that a plume could be detached from the coast. The plume stretches out in the upper layer of the sea, thereby increasing the surface of the plume affected by the wind. Downwelling has an opposite effect, the plume becomes narrower and deeper, with a decreased vertical stratification. In addition, these winds accelerate the already downstream propagating coastal current. Crossshore winds only increases or decreases the discharge from the estuary or river mouth.

![Figure 2.7: Upwelling principle](image)

Previous numerical modeling (De Ruijter et al. 1987) have shown that the wind parameter (constant northwest wind of 3.5 m/s and a constant southwest wind of 4.5 m/s) can bring a significant southwest spreading of the Rhine plume. Other model results show that strong wind activities can move the entire band of fresher coastal water towards the northeast or southwest, or push it against the coastline or advect it offshore with associated frontal meandering and eddy formation (De Kok 1997). Many papers have studied the behavior
of the river plume. These studies all affirm the conclusion of [Chao (1988)] that the plume response to the wind can be described by Ekman drift \(^5\) even when the area is shallow and frictional.

### 2.5 Conclusions and Hypotheses

Based on the literature and previous studies found, some conclusions and hypotheses can be drawn. The Rhine seems to be the biggest influence on the freshwater quantity along the Belgian and Dutch coastline. There is much literature available on the development of the Rhine ROFI along the Dutch coast. But only little is known about the transport towards the Scheldt and the salinity distribution in the Belgian coastal area. All the relevant processes in the freshwater distribution lead to a northward flow. Yet there is an appreciable influence south of the river mouth. Using numerical modeling tools will help us understand how this southward transport works. The three most important driving factors for the residual salinity profile are tide, wind and a density gradient. We have learned that an offshore density gradient causes a northward current. A hypotheses can than be drawn that this southward transport is caused by wind effects and the tide. Which process has the biggest influence can than be tested by means of numerical simulations.

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\(^5\)The movement of ocean water caused by wind blowing steadily over the surface; occurs at right angles to the wind direction (Northern Hemisphere)
Chapter 3

Data Analysis

3.1 Introduction

To simplify the procedure of modeling and the analysis of the model results, first an analysis of the measurement data has to be fulfilled. In that way, an overview of the points of attention can be made. Using the data analysis helps to discover trends in the measurement data and lists interesting events. It supports the model setup and helps directing the sensitivity analysis. The Hydrologisch Meteorologisch Centrum Zeeland (HMCZ) provides all sorts of measurements in the Belgian and Dutch coast. The most important parameters for this study are salinity measurements, water level, and wind data. Noteworthy is that raw measurement data is used in which some unrealistic jumps are noticeable what may be due to algae growth on the sensors.

3.2 Measuring stations

The different measuring stations of the data provided by HMCZ are displayed in figure 3.1. For this thesis, some stations are very interesting. By comparing the salinity values of Haringvliet (HA10), Brouwershavensegat (BG2) and Vlakte van de Raan (VR), we can see the evolution of freshwater transport to the south. The variation in salinity across the Oosterschelde and Westerschelde is given respectively by the stations Oosterschelde (OS4) Marollegat (MRG) and Hoofdplaat (HFPL), Baalhoek (BAAL). As measurements were most abundant for the year 2006, this period was chosen for the salinity simulations.
Chapter 3. Data Analysis

3.3 Minimum, maximum and average values

The monthly minimum, maximum and average salinity values from the three offshore measurement locations Haringvliet, Brouwershavensegat and Vlakte van de Raan are given in Table 3.1, Table 3.2 and Table 3.3. The values for the salinity indicates the polyhalien and euhalien salt regimes. Also, it is seen that on average, the northern measurement station has a salinity that is somewhat lower than the middle measurement station which on his turn is lower than the southern station. The minimum salinity values show a wide range of values. This is not surprising, because a minimum, as well as a maximum, is very sensitive to outliers in the dataset. Nevertheless, the general occurrence of minimum salinities is significantly lower than the value normally found at sea (34 to 35 ppt) indicating the occurrence of fresh water into the North Sea. The minimum salinities are still mostly in the polyhalien regime. The maximum salinities on the other hand, do not show a large range of values, but generally have values around 33 to 34 ppt, taking into account the maximum salinity at sea.
### Table 3.1: Minimum

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### Table 3.2: Maximum

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From the measurement data of the three locations Haringvliet, Brouwershavense Gat and Vlakte van de Raan it can be noticed that the difference between the salinity at the bottom and at the surface is largest at Haringvliet and becomes smaller at Vlakte van de Raan. It seems that the more to the south, the more the fresh water is mixed with the salt sea water. This makes a 3D-simulation less needed at Vlakte van de Raan and, more extended in the Belgian coastal area.

Figure 3.2 illustrates the salinity in the Westerscheldt. What can be noticed is that, the closer to the mouth, the more the salinity values are subjected to the tide. The salinity curve of Baalhoek is much wider than the curve of Hoofdplaat which on its turn is wider than the curve of Vlakte van de Raan. In the figure below (figure 3.3) a detail of the salinity curve of Baalhoek is displayed with the water level. It can be seen that the peaks in salinity generally correspond with high tide. This means that at high tide saltier water flows to the shore and during ebb, more fresh water flows out of the mouth. The salinity values in the Westerscheldt range from 10 PSU at Baalhoek up to 34 PSU at Vlakte van de Raan.

![Figure 3.2: Salinity plots of 3 stations in the Westerscheldt: Baalhoek, Hoofdplaat and Vlakte van de Raan](image)
Figure 3.3: The salinity curve of Baalhoek displayed with the water level in Baalhoek.

The salinity curves of the three measurement stations are showed in figure 3.4. The first point of notice is that the salinity curve in Haringvliet is the most variable and lies under the salinity curves of the other two measuring points. This seems obvious because Haringvliet lies the closest to the mouth of the Rhine. The salinity measured in Brouwershavenssegat and Vlakte van de Raan is more static and the difference between the two points is small. When salinity is plotted together with the water level (a detail of the situation in Haringvliet is shown in figure 3.5), it can be noticed that the salinity has a slight growth around neap tide and declines around springtide. This seems the opposite with was mentioned before, saltier water flows to the shore at high tide. Nevertheless, this pattern can be explained when the flow velocities are taken into account. The flow velocities during neap tide are much smaller compared to the flow velocities during spring tide. So it is possible that net more fresh water flows in the sea during spring than during neap tide. A second point of notice is that the salinity does not change with the semidiurnal high and low tide. The change in salinity values has thus a lower frequency than the tide.
When the salinity is plotted, an annual pattern is revealed. In the beginning of the year salinity values decrease until mid-June from where it slowly starts to increase and finally the salinity turns back into a decrease from November. For this pattern no correlation to the
other physical events like wind or tide were found, but it was useful to look at the discharge rates of the rivers (figures A.10, A.12 and A.14). The discharge rate of the Scheldt, measured in Schelle, and Rhine/Meuse, measured in the Maasmond and at the Haringvliet locks, were obtained respectively from the Hydrologisch Informatiecentrum Vlaanderen (HIC, 2012) and the Rijkswaterstaat (Rijkswaterstaat, 2012). The development of the salinity previously described corresponds to the drainage outflow of the rivers: in the winter months a larger flow rate exists and from the beginning of July this drain flow declines and starts back to rise in November.

3.4 Lower frequency phenomena

In order to improve studying the effects of lower frequency phenomena on tidal records, it is advantageous to first filter out the energy at tidal frequencies. A Doodson XO Filter was used to eliminate tidal energy from observed data. The Doodson X0 filter is a simple filter designed to damp out the main tidal frequencies. It takes hourly values, 19 values either side of the central one. A weighted average is taken with the following weights:

\[
101001011011020112012112011020110100101
\]

The Doodson Filter is one of the earliest and most commonly applied tidal filter, used to eliminate tidal energy from observed water level data. The Doodson Filter eliminates 99,94% of the tidal energy at the semidiurnal frequencies, 99,79% of the tidal energy at the diurnal frequencies and 99,38% of the tidal energy at the overtide frequencies (Grove, 1995).

The remaining variation in the salinity is due to other influences like wind, fresh water discharge,... . To show the influence of the tide, a plot of raw and filtered data is shown in figure 3.6. A large part of the variation in the salinity can therefore be explained by the tide.
Figure 3.6: Raw and filtered data at Haringvliet to show the influence of the tide.

Figure 3.7: Salinity of Haringvliet, Brouwershavensegat and Vlakte van de Raan filtered by the Doodson Filter.
3.5 Single events

By looking more into detail on certain fragments of the graphs, one could notice some special events in the salinity curves. So are the salinity values dependent of the wind. A decrease can be noticed which coincides with an increase in wind amplitude in northern direction (april 5, april 27). On the other hand, when wind is blowing in southwest direction, the salinity is increased (may 20, august 2). At the end of march, the salinity in Brouwershavensegat decreases. This coincides with an increase of the discharge in Haringvliet and Maasmond.

![Figure 3.8: Brouwershavensegat, detail march. Data retrieved from HMCZ.](image-url)
Figure 3.9: Brouwershavensegat: salinity compared to wind data. Data retrieved from HMCZ.

Figure 3.10: Vlakte van de Raan: Wind event starting on may 19. Data retrieved from HMCZ.
Figure 3.11: Brouwershavensegat, a decrease due to a Northern wind event.
Chapter 4

The numerical model

4.1 Introduction

In order to fully understand the principles of the ROFI, the ability to model salinity is an essential step. For this master thesis, the numerical modeling tools from IMDC could be used. These modeling tools consists of large-scale model of the Southern North Sea (KaZNo-model) and a detailed model of the Belgian coast and Scheldt. The software package used is Delft3D, developed by Delft Hydraulics. It is capable of 2D and 3D simulations of ocean basins, coastal seas and rivers and consists of several modules, each executing a particular process. The influence of waves as well as climatic conditions (temperature, wind and atmospheric pressure) can be included in the model. Here only the FLOW-module will be used. The flow-modeling program Delft3D-FLOW solves the shallow water equations, consisting of the momentum and the continuity equations. The program includes among others the following effects:

- earth rotation
- horizontal momentum exchange due to eddy viscosity and turbulence
- density differences
- bed friction
- wind influences

A more extensive description of Delft3D-FLOW and its numerical aspects can be found in the user manual citepdelftmanual. This chapter starts with a description of the governing equations used in Delft3D. Next, the model setup used in this study is described.
Chapter 4. The numerical model

4.2 Governing equations

The numerical implementation of the boundary conditions, as well as the coupling of mixing and stratification related processes, play an important role in this study. This explains the need to understand the governing equations and some numerical aspects of the model. First the hydrodynamic equations need to be discussed, concerning the conservation of mass and continuity.

4.2.1 Hydrodynamic and transport equations

Delft3D-FLOW solves the Navier Stokes equations for an incompressible fluid, under the shallow water and the Boussinesq assumptions. In the vertical momentum equation the vertical accelerations are neglected, which leads to the hydrostatic pressure equation. In 3D models the vertical velocities are computed from the continuity equation. The set of partial differential equations in combination with an appropriate set of initial and boundary conditions is solved on a finite difference grid.

The Reynolds averaged equations of motion are presented in this section. The hydrostatic\(^1\) and Boussinesq\(^2\) approximations have been made. Delft3D employs a sigma coordinate transformation in the vertical. The vertical grid consists of layers bounded by two sigma planes, which are not strictly horizontal but follow the bottom topography and the free surface (see 4.1).

![Figure 4.1: Example of σ- and Z-grid Delft3D Flow Manuel](image)

However, for simplicity a Cartesian coordinate system \((x, y, z, t)\) is adopted here, in which

---

1. In the \(\sigma\) coordinate system the depth is assumed to be much smaller than the horizontal length scale. For such a small aspect ratio the shallow water assumption is valid, which means that the vertical momentum equation is reduced to the hydrostatic pressure relation. Thus, vertical accelerations are assumed to be small compared to the gravitational acceleration and are therefore not taken into account.

2. The effect of variable density is only taken into account in the pressure term.
the $x$-axis is oriented to the east, the $y$-axis to the north and the $z$-axis is positive up away from the bed. The $z$-axis ranges from $-h(x,y)$ at the bed, to $\eta(x,y,t)$ at the free surface, where $t$ is time. At the free surface $h = 0$. The continuity and momentum equations in $x$ and $y$ direction are respectively given by:

$$\frac{\partial u}{\partial t} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \quad (4.1)$$

$$\frac{\partial u}{\partial t} + \frac{\partial u^2}{\partial x} + \frac{\partial uw}{\partial z} - fv + \frac{1}{\rho_0} \frac{\partial P}{\partial x} - F_x - \frac{\partial}{\partial z} \left( v_v \frac{\partial u}{\partial z} \right) = 0 \quad (4.2)$$

$$\frac{\partial P}{\partial z} = -\rho g \quad (4.3)$$

where $u, v$ and $w$ are the velocity components in the horizontal $x, y$ and vertical $z$-direction and $P$ is the pressure. For simplicity in the following the $f$-plane approximation has been made in which it is assumed that the Coriolis term, $f$, is constant with latitude. The Coriolis term is given by $f = 2\omega \sin \Phi$, where $\Phi$ is the latitude and $\Omega$ the angular velocity of the Earth. In the following $\Phi$, is given the value occurring at the centre of the numerical basin. Furthermore $\rho_0$ is the density of air taken as $1,25 kg m^{-3}$, $\rho$ the in-situ density, $g$ the gravitational acceleration and $v_v$ the vertical turbulent eddy viscosity. The horizontal friction terms $F_x$ and $F_y$ are given by:

$$F_x = \frac{\partial}{\partial x} \left( 2v_H \frac{\partial u}{\partial x} \right) + \frac{\partial}{\partial y} \left( v_H \left( \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right) \right) \quad (4.4)$$

$$F_y = \frac{\partial}{\partial y} \left( 2v_H \frac{\partial v}{\partial y} \right) + \frac{\partial}{\partial x} \left( v_H \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) \right) \quad (4.5)$$

where $v_H$ is the horizontal turbulent eddy viscosity.

The pressure at a certain depth $z$ can be obtained by integrating equation 4.3 from $z$ to the free surface, $\eta$:

$$P(x,y,z,t) = P_{atm} + g \rho(\eta) \eta + \int_z^\eta b \, dz \quad (4.6)$$

where $b$ is the buoyancy given by $b = -g(\rho - \rho(\eta))/\rho(\eta)$. This pressure gradient contains respectively an atmospheric term ($P_{atm}$), a barotropic term and a baroclinic term. The atmospheric term is considered to be zero in this study. The water surface elevation $\eta$, measured from the undisturbed water surface, varies in time with $x$ and $y$. Integrating the continuity equation over depth and using kinematic boundary conditions at the free surface:

$$w = \frac{\partial \eta}{\partial t} + u \frac{\partial \eta}{\partial x} + v \frac{\partial \eta}{\partial y} \quad (4.7)$$
and bed:

$$w = u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y}$$

leads to the free surface equation:

$$\frac{\partial \eta}{\partial t} + \frac{\partial}{\partial x} \int_{-h}^{\eta} v \, dz = 0$$

Hereby, the total water depth is the sum of $h(x, y)$ and $\eta(x, y, t)$.

The conservation equations for salinity ($S_{sal}$, in PPT) and temperature ($T_{temp}$, in °C) can be written as:

$$\frac{\partial S_{sal}}{\partial t} + \frac{\partial u S_{sal}}{\partial x} + \frac{\partial v S_{sal}}{\partial y} + \frac{\partial w S_{sal}}{\partial z} - F_S - \frac{\partial}{\partial z} \left( D_V \frac{\partial S_{sal}}{\partial z} \right) = S_{ss}$$

$$\frac{\partial T_{temp}}{\partial t} + \frac{\partial u T_{temp}}{\partial x} + \frac{\partial v T_{temp}}{\partial y} + \frac{\partial w T_{temp}}{\partial z} - F_T - \frac{\partial}{\partial z} \left( D_V \frac{\partial T_{temp}}{\partial z} \right) = \frac{1}{\rho} Q_H + S_{SS}$$

where the term $D_V$ represents the vertical turbulent diffusivity coefficient, $S_{SS}$ represents the sources and sinks and $Q_H$ the heat exchange. $F_S$ and $F_T$ are given by:

$$F_S = \frac{\partial}{\partial x} \left( D_H \frac{\partial S}{\partial x} \right) + \frac{\partial}{\partial y} \left( D_H \frac{\partial S}{\partial y} \right)$$

$$F_T = \frac{\partial}{\partial x} \left( D_H \frac{\partial T}{\partial x} \right) + \frac{\partial}{\partial y} \left( D_H \frac{\partial T}{\partial y} \right)$$

in which $D_H$ represents the horizontal turbulent diffusivity coefficient.

The density $\rho$ is calculated according to an equation of stat, which is a function of the temperature and the salinity.

$$\rho = \rho(S_{sal}, T_{temp})$$

The standard version of Delft3D uses the sigma coordinate transformation given by

$$\sigma_z = \frac{z - \eta}{h + \eta}$$

in which the water column is divided into a constant number of layers. This means that the layer thickness differs (in shallower areas the vertical resolution is higher than in deeper parts).
4.2.2 Turbulence

In the above the Reynolds averaged equations of motion are described. Reynolds averaging generates more unknowns than equations. To be able to solve these equations nevertheless, turbulence closure models are used that make assumptions about some of these unknowns. These turbulence closure models are briefly discussed here. It should be noted here that molecular viscosity and diffusivity have been neglected in the equations presented above. In addition separate horizontal and vertical turbulence closure models have been introduced. Since we are dealing with hydrostatic flows it is considered appropriate to use separate horizontal and vertical turbulent viscosities and diffusivities. In the horizontal either constant values of eddy viscosity eddy diffusivity coefficients can be selected or a Smagorinsky formulation \(^3\) can be used. For calculating the vertical eddy viscosity and diffusivity an algebraic, \(k - L\) or \(k - \epsilon\) turbulence closure model is used. The turbulent kinetic energy \((k)\) and the turbulent dissipation \((\epsilon)\) are determined by transport equations, \(L\) is the mixing length. Delft3D has the option to specify a background vertical eddy viscosity and diffusivity coefficient. In this way some account can be taken of unresolved mixing, for example due to internal waves. However, it is still somewhat subjective as to what value to select. This background value acts as a minimum value to which the model resorts in the situation that the sum of the calculated 3D- and molecular coefficients is less than the minimum prescribed by the user. The vertical eddy diffusivity \((D_V)\) and eddy viscosity \((V_V)\) are defined as:

\[
D_V = \max(D_{\text{background}}, D_{3D} + D_{\text{mol}}) \tag{4.16}
\]

\[
V_V = \max(V_{\text{background}}, V_{3D} + V_{\text{mol}}) \tag{4.17}
\]

where \(D_{\text{background}}\) and \(V_{\text{background}}\) are the background eddy diffusivity and eddy viscosity; \(D_{3D}\) and \(V_{3D}\) the 3-dimensional diffusivity and viscosity components; \(D_{\text{mol}}\) and \(V_{\text{mol}}\) the molecular values. Horizontal values are much larger than the vertical ones. In the horizontal direction constant values are used to model the eddy viscosity and diffusivity.

4.2.3 Boundary conditions

At a coastal wall the normal velocities are set to zero, while for the tangential velocities a free-slip condition is applied. No flux conditions are applied for salt and temperature.

\(^3\)Among many others, Joseph Smagorinsky proposed a useful formula for the eddy viscosity in numerical models, based on the local derivatives of the velocity field and the local grid size:

\[
\nu_t = \Delta x \Delta y \sqrt{\left( \frac{\partial u}{\partial x} \right)^2 + \left( \frac{\partial v}{\partial y} \right)^2 + \frac{1}{2} \left( \frac{\partial u}{\partial x} \frac{\partial u}{\partial y} \right)^2}
\]
Chapter 4. The numerical model

The boundary conditions at the free surface \( z = \eta(x, y) \) are given by

\[
\begin{align*}
V \frac{\partial u}{\partial z} &= \frac{\tau_s^x(x, y, t)}{\rho_0} \\
V \frac{\partial v}{\partial z} &= \frac{\tau_s^y(x, y, t)}{\rho_0}
\end{align*}
\] (4.18)

with \( \tau_s^x \) and \( \tau_s^y \), the surface wind stress terms which are calculated assuming a quadratic friction law. It is assumed that the heat flux, evaporation and precipitation are zero; therefore a zero flux condition on temperature and salinity is also imposed at the surface.

At the bed \( z = -h(x, y) \) the momentum flux is balanced by quadratic bottom stress computed using the velocity nearest to the bottom. Therefore the corresponding boundary conditions at the bed are

\[
\begin{align*}
V \frac{\partial u}{\partial z} &= \frac{\tau_b^x(x, y, t)}{\rho_0} \\
V \frac{\partial v}{\partial z} &= \frac{\tau_b^y(x, y, t)}{\rho_0}
\end{align*}
\] (4.20)

A logarithmic velocity profile has been assumed near the bed. A zero flux condition on temperature and salinity is also imposed at the bed. The transport of dissolved matter such as salt is described by the advection-diffusion equations, as shown in equations 4.10 and 4.11. When the concentration of the boundary during outflow differs from the inflow, discontinuities occur during the turning of the tide. Therefore it is possible to impose a return time, for which the model remembers the concentration of the outflow. In Delft3D this return time is called the Thatcher Harleman time lag, which describes the variation of the concentration as a halfcosine (Thatcher and Harleman, 1972).

Attention needs to be paid to the open boundaries. At an inflow boundary values need to be specified, for example surface levels or flux densities perpendicular to the boundary, as well as the temperature and salinity. At an outflow boundary disturbances must be allowed to leave the basin. Open boundaries very often generate reflections. There are several different types of open boundaries available in Delft3D to overcome this problem, which will be discussed briefly.

- Riemann boundaries prescribe the Riemann invariant or wave-characteristic at the boundary perpendicular to the basin. At the upstream boundary this type of boundary can force a certain relation between the water level and the velocity, at the downstream boundary it acts as a type of absorbing boundary.
Chapter 4. The numerical model

- Dirichlet boundaries impose either the water level or the velocity perpendicular to the boundary, thereby leaving the other dependent variable to develop freely in time. A velocity boundary is more or less coherent with a discharge boundary. These boundaries can be used in case of a supercritical flow or to force certain water levels or velocities to the upstream side of a basin. In case of a sub-critical flow, also the downstream value of one of the two variables must be determined. Due to wind set-up or Coriolis tilting, water levels and velocities vary along the open boundaries. When using Dirichlet boundaries, these variations have to be calculated beforehand. When this is not done properly, circular flow through the boundary may occur, disturbing the internal properties of the basin.

- Recently a new type of boundary condition has been implemented in Delft3D, where the alongshore water level gradient can be imposed at the lateral boundaries. It may be specified as a time varying or harmonic function and thus be used in tidal simulations in combination with other forcings. The advantage of this kind of boundary is the flexibility to allow the cross-shore water level and velocity distribution to develop freely in time and space. These boundaries differ with so-called Neumann boundaries, which impose water level or velocity gradients that are zero.

- Finally a basin can be “nested” into a larger model. After running this larger model, the output is directly interpolated to the boundaries of the smaller model. However, this is time-consuming because every time a property (bottom roughness, tidal amplitude, etc.) in the area of interest is varied, the larger model has to be run first to obtain the open boundary data for the smaller basin.

4.3 Model setup

4.3.1 KaZNo Model

Figure 4.2 presents the large-scale model (2D) used by IMDC. It includes the southern North Sea and the English Channel. The boundaries of the model are approximately 200 km north and 240 km west of Boulogne-sur-Mer. The model has a curvilinear grid and consists of 63,724 meshes.

The bathymetry of the large-scale model is shown in figure 4.2. It is the result of the navigation maps of the British Admiralty (UK Hydrographic Office). Depths are shown relative to the Mean Sea Level (MSL).
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The boundary conditions of the large scale model are astronomical tides. These are based on astronomical components from the tide model TOPEX for the region of the European continental shelf.

This large-scale model is usually used for studies along the Belgian coast. The model is already able to calculate the tides with good accuracy for the French, Belgium and Holland coast. Figure 4.3 shows the correspondence of the model with measurements for the main tidal harmonic component (M2) in different ports.
Figure 4.3: Comparison of the harmonic component of the M2 tide between the large-scale model and measurements along the French, Belgian and Dutch coast (IMDC 2011)
4.3.2 Coastal Model

In the large-scale KaZNo model a more detailed coastal model was nested. This model extends from the France/Belgium border to the border of South-Holland/North-Holland. Through a domain decomposition (DD) technique, a refinement of the grid towards the Belgian coast and Scheldt was carried out. The resolution varies from 27 m – 549 m at the small-scale Schelt domain; from 64 m – 945 m in the Belgian coast domain and from 331 m – 1346 m in the large-scale offshore domain. At the Maasvlakte, the port of Rotterdam is included in the grid as dry points as well as the little islands in the Eastern Scheldt. This total package of domains is further referred to as the coastal model (Kuststrook-model).

The coastal model is driven by Riemann edges on the offshore northern and western boundary and by a current edge on the eastern side (downstream boundary). Freshwater discharges of the Scheldt and the Rhine/Meuse into the Southern North Sea are implemented through a discharge point in Schelle, Haringvliet and Maasmond (upstream boundary). In horizontal direction the domain decomposition of the hydrodynamic grid results in three computational grids, see figures 4.4, 4.5 and 4.6.

Figure 4.4: Overview of the different hydrodynamical grids.
The water density of sea water is set on 1030 kg/m³, temperature on 15 °C. For spring and autumn this comes close to reality, for summer and winter the temperatures are warmer and colder. The model is run with a time step of 15 s. Starting with initial water level and salinity values respectively 0 m and 35 ppt, the model has been spun up over the whole year 2006 to get the salinity distribution approximately right. Later simulations started with an equilibrated initial condition starting in 2006 for a shorter period. The observation stations used to evaluate the model results are displayed in figure 4.7.
Figure 4.7: Overview of the observation stations implemented in the model.
Chapter 5

Sensitivity analysis of the model

5.1 Introduction

By means of a sensitivity analysis, the influence of the different parameters on the fresh water transport can be examined. From the knowledge of this analysis the most relevant parameters for the process can be determined and which parameters are most useful for model adjustment. Furthermore if there is great sensitivity to a particular input, this shows that greater knowledge of this parameter may be necessary for improved model accuracy. This is a necessary step in the model development process to provide accurate simulations of an actual period in the future.

5.1.1 Reference model run

Initial sensitivity experiments began with a reference model run, over the period 7 March 2006 – 1 July 2006 using a time step of 15 s. The different characteristics are displayed in table 5.1. Previous test simulations showed that a long simulation period is needed for the model to converge in a periodic solution that remains stable. In contrast with the water levels, the salinity profile requires a much longer simulation time. To compensate this, a separate run was executed in order to generate adequate initial conditions. A restart file from this run was chosen as initial conditions to shorten the simulation time. The test run consisted of the same characteristics and run over the whole year 2006 with initial conditions: a uniform salinity of 35 ppt and water level 0 m.

As discharge rates at the Scheldt, Haringvliet and the Nieuwe waterweg, the average values of the obtained data for that period were used. For the Scheldt the discharge data in Schelle was available [HIC 2012]. For Haringvliet and the Nieuwe waterweg, only data
Table 5.1: Characteristics of the standard model run

<table>
<thead>
<tr>
<th>Simulation:</th>
<th>Kuststrook01</th>
</tr>
</thead>
<tbody>
<tr>
<td>Period</td>
<td>7 March 2006 – 1 July 2006</td>
</tr>
<tr>
<td>Diffusivity coefficient</td>
<td>10</td>
</tr>
<tr>
<td>Scheldt discharge</td>
<td>100</td>
</tr>
<tr>
<td>Haringvliet discharge</td>
<td>1194</td>
</tr>
<tr>
<td>Nieuwe waterweg discharge</td>
<td>1898</td>
</tr>
<tr>
<td>Wind</td>
<td>/</td>
</tr>
</tbody>
</table>

derived from the SOBEK-model was available. These discharge rates seemed rather large, especially compared to the discharge values used in the model in Leyssen et al. (2011) (Scheldt: 65 m$^3$/s; Haringvliet: 65 m$^3$/s; Nieuwe waterweg: 1200 m$^3$/s). The viscosity and diffusivity coefficient are respectively 1 and 10 m$^2$/s as in the original KaZNo-model. The wind was turned off. At the boundaries, a Thatcher Harlemaan (see section 4.2.3) correction with time lag 420 min is used. In a previous test run we used a time lag of 180 min. Because there was a discontinuity noticed in the top right corner of the eastern boundary, the following simulations all use a Thatcher Harlemaan time lag of 420 min to compensate this.

In figure 5.1, the development of the salinity field from the standard model run is shown. It can be seen that the region of fresh water influence grows in time. No equilibrium situation is obtained, the salinity in the Eastern Scheldt is still ascending at the end of the simulation as displayed in figure 5.2.

1The SOBEK-model is a 1D model of the Northern Delta Bassin.
Chapter 5. Sensitivity analysis of the model

Figure 5.1: Evolution of the spatial salinity plot of the standard model run.

Figure 5.2: Time series of the salinity values in Roompot, an observation station in the Eastern Scheldt.
When the salinity plots of the three measuring stations Haringvliet, Brouwershavensegat and Vlakte van de Raan (figure 3.4) are compared to the model results on the same location (figure 5.3), this shows that the model values are too low. The deviation becomes smaller away from the Rhine/Meuse discharge. A reason for this deviation can be the large values of the discharge at Haringvliet and the Nieuwe waterweg. Similar as noticed with the measured data analysis, is that the lowest salinity values are in Haringvliet and the more saline water in Vlakte van de Raan. The closer to the mouth, the more the salinity values are influenced by the tide.

![Figure 5.3: Model values of the three measuring stations Haringvliet, Brouwershavensegat and Vlakte van de Raan.](image)

The anticipation of the tide results in a sort of circulating motion of the fresh water. By ebb the fresh water flows to the southwest, at turning point the river plume is pushed against the shore after which it flows back to the northeast with high tide. During the transition from flood to ebb, the river plume expands in offshore direction.

### 5.1.2 Residual flow

To get an idea of the residual flow pattern in the study area a simulation is started with simulation time 12u 30 min, the duration of one tide corresponding to the spring cycle. As an indicative measure of the residual current, the residual transport vector \( \vec{v}_{res} \) is taken.
This is a weighted sum of the velocity vectors $\vec{v}$ for the simulation period with the water depth $H$ as weight.

$$v_{res} = \frac{\sum (\vec{v} \times H)}{\sum H} \quad (5.1)$$

In figure 5.4 it becomes clear that, in general, a nett transport exists in northward direction. Closer to the shore irregularities such as eddies occur. There also arises a strip where a small residual flow to the southwest exists. The figure shows that southward transport of salinity can be a result of advection processes as well and not only of diffusion.

![Figure 5.4: Residual transport vector over one tide.](image)

The magnitude of the residual flow in this study area, order of magnitude $10^{-1} m/s$, corresponds with the values discussed in [Leyssen et al. (2011)].
5.2 Sensitivity analysis

A series of tests was now carried out adjusting the discharge rate, diffusivity coefficient and adding wind influences.

5.2.1 Variable horizontal diffusivity coefficient

Experiments similar to the reference run were performed with different values for the diffusivity coefficient. For simulation “Kuststrook02” (figure 5.5(a)) and “Kuststrook03” (figure 5.5(c)) the diffusivity coefficient was changed to respectively $1 \, \text{m}^2/\text{s}$ and $100 \, \text{m}^2/\text{s}$.

The results, figure 5.5, seemed rather surprising. The higher the diffusivity coefficient, the smaller the region of fresh water influence seems. For higher values of the diffusivity coefficient, no distinct coastal current was observed. The plume expanded mainly in the offshore direction. An increase in the diffusivity coefficient results in increased salinity values close to the shore and a less steep salinity profile. A reduction of the diffusivity coefficient in turn leads to a reduced salinity on the coast and a steeper salinity profile. This can be explained by the following. In simulations with a low diffusivity coefficient, salinity is mainly transported by advection. Because of the low water velocities, the fresh water will remain concentrated around the river outflow. Because more and more fresh water is discharged into the North Sea, accumulation of the fresh water occurs and the region of fresh water expands. In simulations with a high diffusivity coefficient on the other hand, diffusivity is enhanced and transports the salinity much quicker. The density gradient will decrease because the salt content will diffuse rapidly in the fresh water until an equilibrium arises. Hence the cross shore density gradients will diminish rapidly offshore. Because the density gradients help determine the magnitude of the alongshore currents (see chapter 2), these currents tend to diminish also in offshore direction. This results in a salinity field

<table>
<thead>
<tr>
<th>Table 5.2: Characteristics of simulation Kuststrook02 en Kuststrook03</th>
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<tbody>
<tr>
<td>Simulation</td>
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<tr>
<td>Period</td>
</tr>
<tr>
<td>Diffusivity coefficient</td>
</tr>
<tr>
<td>Scheldt discharge</td>
</tr>
<tr>
<td>Haringvliet discharge</td>
</tr>
<tr>
<td>Nieuwe waterweg discharge</td>
</tr>
<tr>
<td>Wind</td>
</tr>
</tbody>
</table>
without a pronounced coastal current. Because the salinity at the open boundaries are set on the constant value of 35 ppt, the salinity will increase close to the shore.

A quantitative estimate that could be made was the effective flow of salt through the systems west and north boundaries. In the following figures (figure 5.6) the salt flux across the north boundary and west boundary of the model is shown. The figures show that advection across the boundaries is two or four orders of magnitude greater than the corresponding diffusion. As expected, much more salt was transported through diffusion in simulation “Kuststrook03”.

### 5.2.2 Variable river discharge

In simulation “Kuststrook04” the discharge rate of the rivers was changed compared to the reference model run. Because of the high discharge rates and therefore fresher water in the reference simulation, there is decided to impose lower discharges, the same discharge rates as in [Leyssen et al. (2011)](#).

<table>
<thead>
<tr>
<th>Table 5.3: Characteristics of simulation Kuststrook04</th>
</tr>
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<tbody>
<tr>
<td>Simulation</td>
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<td>Period</td>
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</tr>
<tr>
<td>Scheldt discharge</td>
</tr>
<tr>
<td>Haringvliet discharge</td>
</tr>
<tr>
<td>Nieuwe waterweg discharge</td>
</tr>
<tr>
<td>Wind</td>
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</tbody>
</table>

Although the southward extend of the plume decreased with decreasing discharge rate (figure 5.7(b)), changes in the freshwater rate did not significantly alter the shape of the plume compared to the reference run. But for a lower discharge rate the plume was slower in progress. In figure 5.8 the salinity plots in Brouwershavensegat are displayed for “Kuststrook01” and “Kuststrook04”. The salinity plot from “Kuststrook01” drops more quickly to its equilibrium than the one from “Kuststrook04” with a lower discharge rate.
Chapter 5. Sensitivity analysis of the model

(a) diffusivity coefficient 1 $m^2/s$

(b) diffusivity coefficient 10 $m^2/s$

(c) diffusivity coefficient 100 $m^2/s$

Figure 5.5: Effect of changing the diffusivity coefficient.
Chapter 5. Sensitivity analysis of the model

Figure 5.6: Effective flow of salt through the North- and West boundary.

Figure 5.8: Salinity plot at Brouwershavensegat for simulation “Kuststrook02” and “Kuststrook04”.
Chapter 5. Sensitivity analysis of the model

(a) Result of simulation “Kuststrook01” with a high discharge rate

(b) Result of simulation “Kuststrook04” with a lower discharge rate

**Figure 5.7:** Effect of changing the discharge rate.
Chapter 5. Sensitivity analysis of the model

5.2.3 Imposing wind load

The behavior of the fresh water under wind load was tested on the basis of 4 simulations. As a reference, simulation results from the month May from simulation “Kuststrook04” were used. This was a simulation without any wind influence. In the following models, a uniform wind field was imposed of $6\text{m/s}$ with wind direction successively to the north, east, south and west. More wind could provide a stronger mixing of fresh and salt water and influence the currents. A comparison on a global scale is made by means of (horizontal) salinity profiles (figure 5.11). The results were consistent with what was learned from the literature. Wind direction to the north (downwelling) forces freshwater to the shore. The plume becomes narrower. In figure 5.9(b) the salinity in a cross section is shown along a grid line perpendicular to the coast (figure 5.9(a)). In case of wind to the north, a fresher and narrower coastal current arises because less fresher water is advected offshore compared to a southern wind direction.

![Diagram](image)

(a) Cross section along grid line $m=19$. (b) Salinity values along a cross section in case of wind direction to the north and to the south.

**Figure 5.9:** Cross section perpendicular to the coast.

The coastal currents is accelerated by the wind. With wind direction to the south (upwelling) the freshwater spreads offshore. Because the fresher surface water moves offshore, saltier water flows onshore along the bed. Hence less fresh water than in downwelling situation. Eastern wind direction causes an increase in salinity at the Belgian coast in contrast to the western wind direction, which induces a decrease in salinity. In figure 5.10 the difference in salinity at Vlakte van de Raan is plotted in case of a uniform wind field with direction to the east, west and in case no wind field is imposed.
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Figure 5.10: Salinity plot at Vlakte van de Raan for respectively eastern, western and no wind conditions.

Table 5.4: Characteristics of the simulations imposed with wind fields.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Kuststrook05</th>
<th>Kuststrook06</th>
<th>Kuststrook07</th>
<th>Kuststrook08</th>
</tr>
</thead>
<tbody>
<tr>
<td>Diffusivity coefficient</td>
<td>10</td>
<td>10</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>Scheldt discharge</td>
<td>65</td>
<td>65</td>
<td>65</td>
<td>65</td>
</tr>
<tr>
<td>Haringvliet discharge</td>
<td>65</td>
<td>65</td>
<td>65</td>
<td>65</td>
</tr>
<tr>
<td>Nieuwe waterweg discharge</td>
<td>1200</td>
<td>1200</td>
<td>1200</td>
<td>1200</td>
</tr>
<tr>
<td>Wind</td>
<td>6 m/s North</td>
<td>6 m/s Easth</td>
<td>6 m/s South</td>
<td>6 m/s West</td>
</tr>
</tbody>
</table>

5.2.4 Influence of the different discharges

Additional simulations were carried out with consecutively setting the discharge of the Scheldt, Haringvliet and the Nieuwe Waterweg on and off. Figure 5.13 shows the situation at the end of the simulation. What is striking is that all three freshwater discharges have impact on the Belgian coast. Logically, the most of the fresh water in the Belgian coastal waters is originating from the Scheldt. But the amount of fresh water in proportion to the
Chapter 5. Sensitivity analysis of the model

(a) End result of simulation “Kuststrook04” with no wind field imposed.
(b) End result of simulation “Kuststrook05” with a northern uniform wind field of 6 m/s.

(c) End result of simulation “Kuststrook06” with an eastern uniform wind field of 6 m/s.
(d) End result of simulation “Kuststrook07” with a southern uniform wind field of 6 m/s.

(e) End result of simulation “Kuststrook08” with a western uniform wind field of 6 m/s.

Figure 5.11: Horizontal salinity gradient in case of imposed wind fields.
distance of the fresh water source to the measuring point is low compared to Haringvliet and the Nieuwe Waterweg (figure 5.12). Figure 5.12 also shows that despite the discharge rate of the Nieuwe Waterweg is 20 times greater than the discharge rate of Haringvliet, the contribution of Haringvliet appears to be greater than the contribution of the Nieuwe Waterweg at the Belgian coast. It should be noted that for a clearer result, the simulation time must be much longer. As the decreasing curves in figure 5.12 show, there is not yet a state of equilibrium established.

Figure 5.12: Difference in salinity in Zeebrugge by consecutively setting the discharge rate of the Scheldt, Haringvliet and the Nieuwe Waterweg on.

Table 5.5: Characteristics of the simulations with every discharge successively switched on.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Kuststrook09</th>
<th>Kuststrook10</th>
<th>Kuststrook11</th>
</tr>
</thead>
<tbody>
<tr>
<td>Period</td>
<td>7 March 2006 – 1 July 2006</td>
<td>7 March 2006 – 1 July 2006</td>
<td>7 March 2006 – 1 July 2006</td>
</tr>
<tr>
<td>Diffusivity coefficient</td>
<td>10</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>Scheldt discharge</td>
<td>65</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Haringvliet discharge</td>
<td>0</td>
<td>1200</td>
<td>0</td>
</tr>
<tr>
<td>Nieuwe waterweg discharge</td>
<td>0</td>
<td>0</td>
<td>65</td>
</tr>
<tr>
<td>Wind</td>
<td>/</td>
<td>/</td>
<td>/</td>
</tr>
</tbody>
</table>
Chapter 5. Sensitivity analysis of the model

(a) End result of simulation “Kuststrook09” with only a fresh water discharge at Schelle.

(b) End result of simulation “Kuststrook10” with only a fresh water discharge at the Maasmond.

(c) End result of simulation “Kuststrook11” with only a fresh water discharge at Haringvliet.

Figure 5.13: Influence of the different fresh water sources.
Chapter 6

Discussion

In the previous chapter (chapter 5: Sensitivity analysis of the model), the residual flow pattern is charted by determining the weighted average of the velocity vectors over one tide. Analysis of the simulations where every discharge is separately set to zero together with the residual flow pattern indicate that not only a north-eastward residual current is present. A North-eastward current would, after all, advect fresh Rhine water away from the Belgian coast. It proves that processes such as horizontal diffusion of freshwater and tidal advection acts in both alongshore directions. This is supported by previous studies of Lacroix et al. (2004). In their simulation more than 1% of the water found at the French-Belgian coastal border originates from the Rhine estuary. Even the salinity within the Scheldt Estuary was significantly affected by freshwater from the Rhine intruding through the estuary mouth. They present a new conceptual model of the origin of the water masses (figure 6.1) which corresponds to our experiences.

Figure 6.1: New conceptual model of water masses and dispersion of river water in the Southern Bight of the North Sea. Arrows denote dispersion paths, not residual currents. (Lacroix et al. 2004)
A sensitivity analysis of the numerical model provides an indication of the effect of the diffusivity coefficient in the reproduction of the ROFI of the Rhine-Scheldt estuary. A diffusivity coefficient that is too low, results in fresher water along the coast. This is due to the accumulation of fresh water since limited mixing occurs with the salt water from the North Sea and almost no horizontal diffusion takes place. A higher diffusivity coefficient leads to a limited river plume and coastal current. As Kourafalou et al. (1996) observed already, a plume where the bulge and the coastal river have merged exhibits a diffusive behavior and the plume dissipates energy in the offshore direction.

An interesting quantitative estimate that could be made was the effective flow of salt through the systems boundaries. It became clear that advection across the boundaries is considerably more important, by two to four orders of magnitude, than diffusion. This is a phenomenon that Jones and Howarth (1995) also encountered in their study.

Although the southward extend of the plume decreased with decreasing discharge rate (figure 5.7(b)), changes in the freshwater rate does not significantly alter the shape of the plume compared to the reference run. Despite the discharge rate of the Nieuwe Waterweg is 20 times greater than the discharge rate of Haringvliet, the contribution of Haringvliet at the Belgian coast appears to be greater than the contribution of the Nieuwe Waterweg (figure 6.2). It is clear that a good distribution of the discharge between the Nieuwe Waterweg and Haringvliet appears to be important for proper modeling of salinities near the Belgian coast.

![Image of a map showing the distribution of salinity values along a cross section in case all three discharges are turned on separately.](image)

**Figure 6.2:** Salinity values along a cross section in case all three discharges are turned on separately.
From the literature study, a hypothesis could be drawn that the southward transport of salinity is caused by wind effects and the tide. From the numerical simulations it becomes clear that diffusivity is also an important parameter for this southward transport and a good estimation of the diffusivity coefficient is important to obtain accurate results. As already mentioned before, diffusivity is 2 to 4 orders of magnitude lower than advection, so, although an important process, diffusivity is not the main transport medium.

Wind events could directly be observed in the salinity values of the measured data. The influence of the different imposed wind fields in the simulations is been investigated for the four wind directions. It is clear that imposed wind events have a fast impact on the salinity values, while a change in discharge rate has a much slower effect. Observation of the results of the simulations with wind forces concludes that the shape of the river plume, more specifically the southern spread, can be altered as a result of different wind fields (figure 6.3).

Figure 6.3: Salinity values along the Belgian coast to display the difference in southern spreading of the river plume between a situation with imposed wind directed to the north and wind directed to the south.

The main cause of the southern transport is however the tide. A new conceptual model imposes itself. See figure 6.4 for a schematic representation of this southern transport. During ebb, sea water flows along the coast to the south west together with a part of the fresh water from Maasmond and Haringvliet (1, 3). During high tide, the fresh water is pushed back to the northeast. A fraction of the fresh water is hereby caught by the
overlying estuary (2, 4). This fraction of fresh water flows further southwards during the following ebb (3, 5). This way fresh water gradually moves to the south. The distribution of the fresh water as a result of the tide is a slow process. Hence a long simulation time is needed to obtain a situation of equilibrium.

Figure 6.4: New conceptual model of water transport to the south due to tidal process.
Chapter 7

Concluding remarks and recommendations

The processes related to the distribution of the fresh water along the Belgian and Dutch coast have been examined. First, the relevant processes are investigated by means of a thorough literature study. Thereafter the evolution of time series, obtained by salinity measurements, are explained and interesting events are listed. The measurement data of the year 2006 is used. With this knowledge gained, a series of sensitivity experiments are designed to test the physical role of the diffusivity coefficient, fresh water discharge and wind fields. The existing numerical tools of IMDC are used for this purpose.

The three most important driving factors for the residual salinity profile are tide, wind and a density gradient. With regard to the southward distribution of the fresh water, the tide ensures the largest transport of fresh water. Wind processes and changes in discharge rates can only extend the ROFI a little further to the south. A new conceptual model of the southward distribution of the fresh water is suggested in this study. Fresh water from the southern discharges flows in direction of the southwest during ebb. During high tide the fresh water is pushed back to the north-east. A fraction of the fresh water is caught by the overlying estuary. During the next low tide this fraction of fresh water flows further to the south. Hence fresh water gradually moves to the south.

The sensitivity analysis of the numerical model provides an indication of the effect of the diffusivity coefficient in the reproduction of the ROFI of the Rhine-Scheldt estuary. An increase of the diffusivity coefficient results in higher salinity values near the coast and a less steep salinity profile. This is because lower salinity gradients are obtained with a higher diffusivity coefficient. A reduction of the diffusivity coefficient in turn leads to a
reduced salinity on the coast and a steeper salinity profile.

In previous studies caused the inclusion of realistic wind fields in the calculation a higher salinity in the Belgian-Dutch coastal zone. More wind can provide a stronger mixing of fresh and salt water. Since the wind direction is mainly inland, it can provide a net transport of saltier sea water to the coast. This is tested by means of a simulation whereby a uniform wind field is imposed. A wind field with direction to the east causes indeed higher salinity values along the Belgian coast.

Although the southward extend of the plume decreases with decreasing discharge rate (figure 5.7(b)), changes in the freshwater rate does not significantly alter the shape of the plume compared to the reference run. Still, a good ratio of the discharges is necessary to obtain actual salinity values. This can be noticed in the influence of the fresh water discharge in Haringvliet on the salinity values in the Belgian coastal area. Despite the discharge rate of the Nieuwe Waterweg is 20 times greater than the discharge rate of Haringvliet, the contribution of Haringvliet appears to be greater than the contribution of the Nieuwe Waterweg at the Belgian coast.

In contrast with the water levels, the salinity profile requires a much longer simulation time. On top of that, small deviations in a salinity field accumulate throughout the simulation and thus cause larger deviations. To take the stratification into account, a 3D model would be necessary. Due to an increased calculation time, this was beyond the scope of this thesis.
Appendix A

Figures of the data obtained from 2006

A.1 Introduction

In this chapter, all the obtained data from measurements are plotted.

A.2 Salinity measurements

The salinity measurements data are all obtained from HMCZ [HMCZ 2012]. The monitoring network operates according to the Rijkswaterstaat Meetinformatiestructuur, a standard so that hydro-meteorological data and the different services are measured and processed the same way. In addition, the measurement networks are linked to each other for the exchange of data. There are also links with the English monitoring network and the monitoring network Vlaamse Banken in Belgium. In figure A.1 is a map displayed where all the salinity measuring stations of HMCZ are indicated.
Figure A.1: Overview of the different measuring stations in the Scheldt estuary [HMCZ 2012].

Figure A.2: Salinity measurements in Haringvliet (HA10). Data retrieved from HMCZ.
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Figure A.3: Salinity measurements in Brouwerhavensegat (BG2). Data retrieved from HMCZ.

Figure A.4: Salinity measurements in Vlakte van de Raan (VR). Data retrieved from HMCZ.
Figure A.5: Salinity measurements in Hoofdplaat (HFPL). Data retrieved from HMCZ.

Figure A.6: Salinity measurements in Marollegat (MRG). Data retrieved from HMCZ.
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Figure A.7: Salinity measurements in Baalhoek (BAAL). Data retrieved from HMCZ.

Figure A.8: Salinity measurements in the Oosterschelde (OS4). Data retrieved from HMCZ.
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Figure A.10: Discharge rate measured in Schelle.
Figure A.11: Location of the measuring station Maasmond

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Figure A.13: Location of the measuring station Haringvliet locks

Figure A.14: Discharge rate measured at the Haringvliet locks. Data retrieved from Waterbase (Rijkswaterstaat 2012)
A.4 Wind measurements

**Figure A.15:** Wind amplitude and direction in Brouwerhavensegat (BG2). Data retrieved from HMCZ.

**Figure A.16:** Wind amplitude and direction measurements in Vlakte van de Raan (VR). Data retrieved from HMCZ.
A.5 Water level measurements

Figure A.17: Water level measurements in Haringvliet (HA10). Data retrieved from HMCZ.

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