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The tidal asymmetries and residual flows in Ems Estuary

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Abstract A 3D unstructured-grid numerical model of the Ems Estuary is presented. The simulated hydrodynamics are compared against tidal gauge data and observations from research cruises. A comparison with an idealized test reveals the capability of the model to reproduce the secondary circulation patterns known from theoretical results. The simulations prove to be accurate and realistic, confirming and extending findings from earlier observations and modeling studies. The basic characteristics of dominant physical processes in the estuary such as tidal amplification, tidal damping, overtide generation, baroclinicity and internal mixing asymmetry are quantified. The model demonstrates an overall dominance of the flood currents in most of the studied area. However, the hypsometric control in the vicinity of Dollart Bay reverses this asymmetry, with the ebb currents stronger than the flood ones. Small-scale bathymetric characteristics and baroclinicity result in a very complex interplay between dominant physical mechanisms in different parts of the tidal channels and over the tidal flats. Residual flow reveals a clear overturning circulation in some parts of the estuary which is related to a mixing asymmetry between flood and ebb currents. We demonstrate that while areas close to the tidal river exhibit overall similarity with density controlled estuarine conditions, in large areas of the outer estuary barotropic forcing and complex

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bathymetry together with the density distribution affect substantially the horizontal circulation.

1 Introduction

The aim of this paper is to improve the understanding of tidal flows in the Ems Estuary, with a focus on the barotropic and baroclinic processes and associated flood and ebb asymmetry. This estuary is a well-mixed meso-tidal estuary (Talke et al. (2009), Dyer et al. (2000)) where the interplay between tidal forcing, complex geometry and longitudinal density gradient generates complex dynamics (de Jonge (2000)). This estuary, which is part of the Wadden Sea of the Northwest European shelf (Fig.1), is an intertidal area shielded from the open sea by a couple of barrier islands. The Ems River, with an average run off about $110 \text{ [m}^3/\text{s]}$ provides 90% of the total freshwater in the Ems Estuary (Chernetsky et al. (2010)). The distance between the barrier islands and the tidal weir in Ems River (see Fig. 1 for its position) is approximately 100 km. The major exchange between estuary and open sea occurs through the channel between Islands of Borkum and Rottumeroog. In the following we will refer to this channel as to main channel keeping in mind that it reaches up to the mouth of Dollart Bay. The remaining part, which extends up the river reaching the weir will be referred to as the tidal river.

The dynamics are primarily forced by the tidal Kelvin wave with its major amphidromic point centered at $55.5\text{N}/5.3\text{E}$ (Klein and Mittelstaedt (2001)). While the wave is of propagating character in the lower reaches, it acquires a standing character in the vicinity of the weir where it is reflected; the tidal range in the river part amounts to almost four meters (Krebs (2011)). The tidal excursion reaches up to 15 km (Heinzelmann and Heyer (2006), De Jonge (1992)).

The Ems estuarine system consists of 50 percent tidal flats (De Jonge (1992)). The Dollart Bay, which is situated half way between the barrier islands and the tidal weir is the major representative of these flat areas (Fig. 1). At the mouth of the Dollart Bay the flood flow bifurcates, with one part filling the bay, the other

part travelling up the Ems River. The volume of water residing in the estuary, which is a function of the water level changes over time, the difference between volumes at high and low water is known as tidal prism. The change tidal prism over time depends upon the bathymetry of inter tidal basins (in case of constant depth the volume is proportional to the sea level height). However complex bathymetry makes the volumetric characteristics of individual basins very diverse, which results in different 'hypsometric control' (Stanev et al. (2003)).

While in the Ems River the volume of residing water increases almost linearly with the surface elevation, in the Dollart Bay the relationship is exponential due to the larger proportion of tidal flats (Fig. 2). The ratio between the volume of intertidal storage versus the volume of deep channel at mean sea level is 0.6 for the Ems River and 2.9 for the Dollart Bay.

Large tidal flats in the Dollart Bay delay the propagation of the flood while enhancing the ebb-current (Dronkers (1986), Speer and Aubrey (1985)). This bay resembles the East Frisian tidal basins where the tidal prism also largely exceeds the permanent water storage (Stanev et al. (2003)) supporting the hypsometric control (Boon and Byrne (1981)) and triggering strong and short-lasting ebb current and a weak and long-lasting flood current (ebb dominance).

In the outer estuary and in the tidal river the tidal wave propagates faster during flood than during ebb because during ebbs the local depth is shallower. This asymmetry is known as flood dominance (Friedrichs and Aubrey (1988)). The interplay between flows controlled by different mechanisms makes the dynamics very complex; one example is at the confluence of flows from the tidal river and the mouth of Dollart Bay. Additionally, the current's amplitude is modified considerably by the along-channel density gradients (Jay and Musiak (1996)).

The spatial characteristics of Ems estuarine dynamics have been addressed by dedicated observations and modeling studies. Using more than a dozen cross-sections throughout the estuary De Jonge (1992) produced the first map of residual flow identifying circulation cells in front of the inlets, as well as circulatory flow be-

1 tween different channels or within one channel. This study demonstrated the over-
2
3 all flood-directed residuals over the tidal channels, which the author believes are
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5 hydraulically stable. The analysis of tidal asymmetry and residual circulation of
6
7 Van de Kreeke and Robaczewska (1993) using 2D numerical model confirmed the
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9 patterns derived from observations. In a more recent attempt to evaluate the
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11 possible locations of alternative dumping sites for dredged sediments Boon et al.
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13 (2002), using 2D curvilinear model with a resolution of 200 m, demonstrated the
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15 capability of the numerical modeling, but at the same time stressed the difficulties
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17 in simulating the longitudinal salinity distribution.

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19 A number of studies have been carried out (Ridderinkhof (1997)) addressing the in-
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21 teraction between hydrodynamics and suspended particulate matter (SPM) trans-
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23 port using analytical (Talke et al. (2009)) or two-dimensional (vertical and along
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25 channel direction) models (Chernetsky et al. (2010)). These studies contributed to
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27 understanding the evolution of the dynamics in the tidal river, including resonance
28 effects and changes in the M4 overtide.

29
30 Tidal asymmetries need further consideration in the area of Ems Estuary because
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32 they control the sediment dynamics. Earlier studies (e. g. Van de Kreeke et al.
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34 (1997)) stated that the main channel connecting the tidal flats and open sea was
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36 flood dominated. This is also the case in the area between the Dollart Bay and
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38 the tidal weir (Talke et al. (2009), Chernetsky et al. (2010)). However, the flood
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40 dominance in this area is to a certain extent a result of extensive dredging ac-
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42 tivities in the tidal river as well as in the lower reaches of the estuary (De Jonge
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44 (1992), Heinzelmann and Heyer (2006)). Dredging favours stronger currents dur-
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46 ing the flood period because of weaker frictional effects caused by the deepening of
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48 tidal river (Krebs (2011)). This has augmented the role of tidal pumping and the
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50 import of sediments (de Jonge (2000)), which is considered by many as a deterio-
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52 ration of the sediment system. Very relevant to the recent changes in the sediment
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54 system is the study of Winterwerp (2011), who, using a one-dimensional model
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56 and observations, addressed the effects of high sediment concentrations near bed
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1 and trapping of sediments in the tidal river. The above cited works gave good
2 examples of applicability of process studies based on simplified models.

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4 So far a complete 3D picture of the hydrodynamics in Ems Estuary unifying the
5 processes in physically different regions is missing. This is very important because,
6 although the barotropic dynamics are predominant in most of the tidal estuaries,
7 the baroclinic pressure gradients are of utmost importance (Nunes and Simpson
8 (1985), Lerczak and Geyer (2004)). However, the three-dimensional frame needed
9 to fully describe it has not been sufficiently addressed in the previous studies. We
10 hope that the analysis here extends the knowledge from previous two-dimensional
11 simulations and would be of interest to a wider international audience addressing
12 similar issues from the field of hydrodynamics related to natural and anthropogeni-
13 cally induced changes of coastal environment.

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15 With respect to the anthropogenically induced changes and other practical imple-
16 mentations the effects of the storm surge barrier are of particular interest. Tidal
17 gates can cause profound changes in the flow regime beyond the vicinity of the
18 construction (Bednarczyk et al. (2008)). However even if the tide gates are only
19 infrequently closed, as in the case of a storm surge barrier, the small constriction
20 of the flow by the supporting structure is of permanent character. The baroclinic
21 regime is expected to be very sensitive to such interference.

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23 In the present paper we use a well-validated, three-dimensional model based on
24 an unstructured-grid, which enables resolution of the multitude of processes with
25 different scales and the complex and fine-scale elements of topography. It will be
26 demonstrated that simulations with this model reveal further detail both in the
27 vertical and horizontal patterns. The major focus of the paper is on the area be-
28 tween tidal river and open-ocean, which constitutes the outer estuary. The paper
29 is structured as follows: After the model description in Section 2, model validation
30 is presented in Section 3 followed by the geophysical interpretation of the simula-
31 tions in Section 4. The discussion in Section 5 presents the major advancements
32 and remaining questions, followed by short conclusions.

2 The model

2.1 The equations

The Semi-Implicit Eulerian-Lagrangian Finite-Element model (SELFE - College of William and Mary branch) originates from a series of semi-implicit unstructured grid models that were designed to meet the challenge of cross-scale ocean modelling (Fringer et al. (2006)). These models overcame the limitations of the CFL-constraint by treating the barotropic pressure gradient, vertical viscosity in the momentum equations, and divergence term in the continuity equation implicitly (Casulli and Cattani (1994), Zhang and Baptista (2008)). Further the grid orthogonality required by finite-difference models has been overcome in SELFE by implementation of a finite-element scheme.

SELFE is based on the 3D shallow-water equations, assuming hydrostatic conditions and making use of Boussinesq approximation. The model solves for elevation, horizontal and vertical velocities, salinity and temperature.

Shear stress at the sea surface τ_w is computed according to a parameterization by Pond and Pickard (1998). This yields the boundary condition for the momentum at the surface:

$$\nu \frac{d\mathbf{u}}{dz} = \tau_w, \quad (1)$$

where $\mathbf{u}$ is the horizontal velocity with Cartesian coordinates x , and y , ν is the vertical eddy viscosity and τ_w is the wind stress. The bottom boundary condition reads

$$\nu \frac{d\mathbf{u}}{dz} = \tau_b. \quad (2)$$

Bottom stress is represented by

$$\tau_b = C_D |\mathbf{u}_b| \mathbf{u}_b, \quad (3)$$

where u_b is the velocity at the top of the bottom computational cell and C_D is the drag coefficient. Horizontal velocities in the bottom boundary layer are assumed to obey the logarithmic law. The bottom roughness length is set to $z_0 = 7.5 \times 10^{-4} [m]$ for the whole model domain.

2.2 Turbulence closure

The implementation of turbulence in SELFE uses the Generic Length Scale (GLS) closure by Umlauf and Burchard (2003). Turbulent kinetic energy (K) and a generic length-scale variable (ψ) are governed by the following equations:

$$\frac{DK}{Dt} = \frac{\partial}{\partial z} \left(\nu_k^\psi \frac{\partial K}{\partial z} \right) + \nu M^2 + \mu N^2 - \epsilon, \quad (4)$$

$$\frac{D\psi}{Dt} = \frac{\partial}{\partial z} \left(\nu_\psi \frac{\partial \psi}{\partial z} \right) + \frac{\psi}{K} (c_{\psi 1} \nu M^2 + c_{\psi 3} \mu N^2 - c_{\psi 2} \epsilon F_{wall}), \quad (5)$$

where $\nu_k^\psi = \frac{\nu}{\sigma_k^\psi}$ is a vertical turbulent diffusivity with constant Schmidt number σ_k^ψ , $M^2 = \left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2$ is the shear frequency and $N^2 = \frac{g}{\rho_0} \frac{\partial \rho}{\partial z}$ is the buoyancy frequency. The dissipation rate is given by $\epsilon = (c_\mu^0)^3 K^{1.5} l^{-1}$, where $c_\mu^0 = 0.3^{1/2}$ and l is the mixing length (Zhang and Baptista (2008)). In the above equations $\nu_\psi = \frac{\nu}{\sigma_\psi}$ is a vertical diffusivity with Schmidt number σ_ψ , $c_{\psi 1}$, $c_{\psi 2}$, $c_{\psi 3}$ are model specific constants (Umlauf and Burchard (2003)), and F_{wall} constitutes a wall proximity function. In eq. 4 and 5 the first term on the right hand side represents vertical diffusion of turbulence, and the second and third terms describe the production of turbulence.

The GLS variable is specified by the choice of the parameters p, m, n that correspond to the chosen turbulence scheme ($k - kl$ in our case):

$$\Psi = (c_\mu^0)^p K^m l^n. \quad (6)$$

Eddy viscosity is related to turbulent kinetic energy and the mixing length through

$$\nu = \sqrt{2} s_m K^{1/2} l, \quad (7)$$

and

$$\mu = \sqrt{2} s_h K^{1/2} l, \quad (8)$$

where s_m, s_h are stability functions.

Some important physical parameters and their values are given in Table 1.

Parameter	Value
Maximum drag coefficient	1.e-3
Bottom roughness	7.5e-4 [m]
Minimum diffusivity	$1.e - 5 [m^2 s^{-1}]$
Turbulence closure	$k - kl$

Table 1: Physical parameters of model setup. A maximum value for Cd is imposed in order to avoid singularity and very large values at shallow depths.

2.3 Model forcing

Surface elevation and horizontal velocities, water temperature and salinity are interpolated on the model mesh at the open boundaries from the output of the MyOcean operational ocean forecast system (O’Dea et al. (2012)). The MyOcean

product is based on the NEMO (Nucleus for European Modeling of the Ocean) model. The model uses a structured grid with horizontal resolution of 4 km and 32 sigma layers in the vertical. The output data are available with a temporal resolution of one hour, and were interpolated bi-linearly to every grid point at the open boundary (see the model area in Fig. 1). The model is forced by the meteorological forcing derived from bulk formulae using simulated SST, as well as wind, mean sea level pressure, air temperature and specific humidity taken from the hourly forecasts of the German Weather Service (DWD) COSMO-EU model with 7 km horizontal resolution. River discharge [m^3/s] is based on climatological daily averages and is introduced into the estuary at the tidal weir.

2.4 Horizontal, vertical grid and time stepping

The horizontal grid consists of approximately 80000 nodes. Its resolution varies between 3 km at the open boundary and 70 m along the coastline and in the main channel. The resolution approaches 50 m at the storm surge barrier (Fig. 1) and is coarser on the tidal flats. Cells in the main channel consist of split quads for ease of adjustment of grid resolution (Zhang and Baptista (2008)). The bathymetric data were provided by the German Hydrographic Office (BSH). Mapping on regular grid was done in the Helmholtz-Zentrum Geesthacht. In the vertical SELFE allows z-, S- and hybrid coordinates. In the current setup 25 S- layers are used with the resolution increasing towards the surface and the bottom. Flooding and drying of tidal flats is implemented in SELFE; a node is set dry if total water depth drops below 0.04 m.

The transport equations are discretized by an upwind scheme instead of a more expensive 2nd-order TVD scheme, because the vertical stratification in the estuary is generally weak. SELFE being a semi-implicit model, the implicitness factor has been set to 0.5. The time step is 100 s and results have been saved every 10 min. The simulation period includes the complete month of June 2012. The model is

fully parallelised and runs were carried out on 60 cores. Run time typically scaled to be 30-140 times faster than real time on different cluster we have access to.

In order to check the role of density gradients a barotropic run has been carried out where salinity and temperatures are set to constant values and all other forcing conditions are kept the same as in the baroclinic simulation.

3 Model validation

3.1 Water levels

The validation of numerical model performance against tidal gauge data is presented below for the period 1st – 30th of June 2012 at three stations: Borkum, Knock and Pogum. Gauge data have been provided by the German federal water and navigation administration (WSV). The gauges are located in the main channel, near the mouth of the tidal river in front of Dollart Bay and in the tidal river respectively (cf. Fig. 1). The sampling rate of the observations is one minute and the validation is performed at a rate of 15 min. The rms difference between observed and simulated total water elevations in these stations is 0.17 m, 0.11 m and 0.19 m, respectively and the corresponding tidal range is 2.6 m, 3.5 m and 3.8 m.

Further detail about agreement between simulations and observations and evolution of tidal oscillations is given below in terms of characteristics of tidal spectrum. Tidal analysis has been applied to one month of observed and simulated tides. During this period atmospheric conditions were moderate. Both in the observations and numerical simulation the four most important components in descending order were M2, N2, S2 and M4 (Fig. 3). At the Borkum station the observed amplitude of M2 amounts to 1.1m approximately, whereas the N2 amplitude equals to little more than 20 % of the M2 value. Model performance improves at the Knock station, but the model overpredicts the amplitudes of M2, M4 and S2 at Pogum station. M6 is underestimated at Pogum station by about 20%. The overall con-

clusion is that the transformation of the tidal wave and the generation of overtides within the region are reproduced by the model.

In order to evaluate the variability of the barotropic pressure gradient the observed and simulated water level differences between Knock and Pogum stations (see Fig. 1 for locations) are compared below. The maximum surface slope appears shortly after the low slack water at Pogum, at the same time the flood has already established at Knock (Fig. 4). The surface slope between Knock and Pogum is positive during very short time-period both in observations and simulations. Accordingly, phases and magnitudes agree well. The greatest negative slope (barotropic gradient pointing towards the ocean) occurs at the beginning of the ebb which is identified by the zero-crossing of the surface slope. Both in the observations and simulations the slope between these two stations agrees well, in particular during the end of ebb phase, before the slope becomes positive again. The difference between simulations and observations is that in the former the direction of the gradient reverses from positive to negative earlier (end of flood in the simulations and high water in observations).

3.2 Surface velocities

During 07th – 15th of June 2012 a field campaign was carried out in the Ems Estuary. Horizontal velocities were measured with an on-board ADCP using the motorboat Otzum. The survey on 7th of June 2012 covered the main estuarine channel and tidal river up to the storm surge barrier (Fig. 5a, cf. Fig. 1). The average boat-speed amounted to 4 kn. ADCP data were collected by a RH instruments Sentinel profiler. Vertical sampling was applied over a maximum of 45 bins, but on average 25 bins because most part of the sampled region was quite shallow. In order to compare the observations and simulations, model results have been interpolated linearly in time and space onto the locations of observations. The green line in Fig. 5b,c displays the simulated sea level at the respective ship position, giving an idea about the timing during part of the tidal cycle. The tran-

sect started at HW slack at Borkum Island and then proceeded along the main estuarine channel (cf. Fig. 5a). In this area the model grid is particularly fine. The boat followed the channel thalweg and therefore the observed currents are projected on a reference frame parallel to the channel axis so that u_a refers to the along-channel component of flow and u_c refers to the lateral component of flow. There is a good overall agreement between observations (blue) and simulations (red) in Fig. 5b,c. This is particularly the case for the along-channel flow, for which velocity ranges and acceleration were well represented in the simulations. It is noteworthy that there is a slight underestimation of lateral velocities in the second part of the transect (between 15:00 h and 15:45 h). However, during this time the boat was downstream of the mouth of Dollart Bay, where the bathymetry and thus the flow regime notably changes, making simulations particularly demanding (cf. Fig. 5a). The temporal and spatial variability along the main channel gives an integrated view of the processes in a large part of the estuary. The above agreement between observations and simulations during a relatively long part of the tidal cycle supports the overall credibility of the simulation.

3.3 Vertical profiles of horizontal velocity

The second part of the measuring campaign started on the 12th of June 2012 and consisted of more than 100 crossings of the channel near Pogum within 12.5 hours (see Fig. 5a for the position of this transect), i.e. one full tidal cycle. The sampling frequency of the ADCP was 0.1 s. It took approximately a quarter of an hour for the boat to cross the channel back and forth, and therefore the temporal resolution of the record in the middle of channel was 7.5 min. Because the frequency with which the model output was produced was 15 min, observations were also averaged with 15 min intervals between them.

Observed and simulated velocity profiles (Fig. 6) demonstrate substantial flood dominance in this location. Not only was the maximum flood current higher than the maximum ebb current, but also it accelerated much faster. The vertical shear

of ebb currents revealed clearly a surface intensified jet and was more pronounced in the observations. It was generally weaker during flood in the majority of water column, which was well revealed in the simulation and observations.

3.4 Surface salinity

Estuarine salinity structure is very informative about the estuarine circulation (Geyer (2010)). It identifies the position of salinity front and its dynamics, giving thus the overall view on the estuarine mixing (Monismith et al. (2002), Monismith (2010)). During the survey on 7th June 2012 CTD casts have been performed along the transect line in Fig 5a. For the same period observed and simulated surface salinity (SSS) is plotted in Fig. 7a. The agreement between simulations and observations can be seen from the comparison between the background (time versus position diagram from simulations) and observations on the track. Perfect match would make the transect line in Fig. 7a invisible. The slope of salinity isolines (spatial versus temporal changes) represents the propagation speed of the tidal wave as seen by salinity field (max. 2 [m/s]).

The longitudinal salinity range along the chosen transect measured as the difference between maximum and minimum salinity in Fig. 7a reaches about 30[psu] in about 70 km. Most of this change (between 5 and 25[psu]) occurs in the area of salinity front which is about 25 km wide displaced back and forth between km 30 and 70. This indicates that the tidal excursion of isohalines (see the black and white isolines) amount to more than 15 km. Furthermore, the general pattern of salinity front is overlain by several anomalous features demonstrating a rather complex frontal behaviour and intrusions, such as the events at 10.5h seen at km 47 and km 57.

The comparison between observations and simulations (Fig. 7b) proves that the numerical model represents well the horizontal and temporal changes of salinity. The salinity range in this figure is given by the distance between the two red dashed lines. It is noteworthy that the survey started by high water when salinity around

1 Borkum Island was at maximum. By this time the observations were closer to the
2 upper dashed line. The ship was moving further into the direction of tidal river
3 that is against the ebbing water (see Fig. 7a), and therefore in the second part
4 of transect, observations were closer to the lower dashed line. The fact that the
5 difference between observations and simulations was much smaller than the salinity
6 range demonstrated that the model was capable to replicate the basic temporal and
7 spatiale variability patterns (Fig. 7b).
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21 The results illustrated above compare well with those of Monismith et al. (2002).
22 The estuary can be roughly considered as composed of three parts: outhter estuary
23 with salinities above 25[psu], which is identified beyond km 30, salinity-front area
24 between km 30 and km 55, and tidal river, which is upstream of km 55. In the first
25 area salinity range is relatively small and one could expect that barotropic pro-
26 cesses are dominant there. In the second area, salinity range increases up to km 55,
27 which is approximately where the storm surge barrier is. The expectation is that
28 the estuarine circulation in this area is dominated by stratification/baroclinicity.
29 At ebb fresher water enters the main channel at high rates near the sea surface;
30 consequently the sharp halocline prevents vertical mixing. It will be shown in the
31 following that during flood the tidal straining dominates the estuarine dynamics
32 in this area. These contrasting situations require a correct representation of strat-
33 ification; otherwise substantial phase differences could build up between model
34 results and observations (Jay and Musiak (1996)). In the third zone tidal range
35 rapidly decreases upstream of the storm surge barrier, which is also consistent with
36 other estuarine cases (e. g. see Monismith et al. (2002)). This zone is not covered
37 by the observations presented in this study and will be subject of future analyses
38 using new observations and model simulations with even finer resolution.
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4 Geophysical interpretation

4.1 Tidal characteristics along the main channel

The tidal range increases from 2.6 m in the main channel inlet between Borkum and Rottumerog (km 0 in Fig. 8a,b) to 3.8 m in the tidal river (km 65 in the same figure) (Fig. 8a,b). The rate of increase is almost constant in the lower estuary (between km 0 and 50 in Fig. 8a,b) and decreases in the tidal river (between km 55 and 65 in Fig. 8a,b). This can be explained by the change in the shape of estuary: a funnelling (convergence) in the outer estuary and almost parallel coast on both sides of tidal river (cf. Fig. 1). Tidal range even drops slightly on the lee side of the storm surge barrier (km 61), which demonstrates that friction there dominates over the convergence.

The ratio between tidal amplitude and depth a/h can be considered as a measure of distortion of the tidal wave. Associated with this ratio are the effects of generation of overtides and asymmetric response of tidal currents (Boon and Byrne (1981)), as well as reduction of tidal range by friction (Speer and Aubrey (1985)). This ratio is below 0.1 around Borkum Island (Fig. 8b). It attains maximum in the vicinity of the storm surge barrier characterized by small depths and sectional area, which indicates an enhancement of non-linearity and friction.

One way to represent tidal asymmetries is by analysing the duration of ebb and flood. Along the main channel the rising tide becomes more than one hour shorter than the falling tide (red curve in Fig. 8b). The duration asymmetry attains maximum around 30 km, which is half way between the inlet and the tidal river. Of utmost importance is the change in tidal wave near Knock (at about km 40, see also Fig. 1), which coincides approximately with the confluence zone. In this region the flood-dominance weakens, the ebb-period surpassing the flood one by only about 0.5 hours. The processes there are strongly influenced by the interaction between the tidal river outflow and the outflow from the Dollart Bay. East of the confluence zone the duration of ebb increases, which demonstrates that Dollart

Bay region is one specific area changing the tidal wave locally due to hypsometric controls (see Fig. 2).

One fundamental issue of estuarine dynamics is associated with the role of density. In order to evaluate its effect we compare in the following the results of a baroclinic and barotropic run (see Section 2.4). In the former the de-tided sea level along the transect-line (solid green line in 8b) shows differences of about 10 cm. The sea level slope increases strongly in the river mouth and tidal river.

The sea level difference between baroclinic and barotropic runs (the pure effect of density anomalies as illustrated by the green dashed line in Fig. 8b) is very instructive. In the zone between the barrier islands and km 35 the tidally-averaged sea-level slope is very small (the sea-level differences between the two experiments are larger than the sea level difference between barrier islands and km 35 in the baroclinic experiment). This demonstrates that the pressure gradients due to baroclinicity there could exceed those caused by the barotropic forcing. In the zone between km 35 and km 55, which is the zone of salinity front, the time-averaged sea level shows an increase in the up-river direction, which is comparable with the the sea level differences between the two experiments. In a third zone, which is beyond km 55 that is in the tidal river the sea level drastically increases, however the difference between the baroclinic and barotropic experiments in this zone reduces because salinity there is very low and almost constant (see 7). The above representation of the possible role of baroclinicity is instructive to where to look for the largest effects of density (see for more detail the comparison with the results from the barotropic run presented in sections 4.2.2,4.4,4.8).

4.2 Patterns of tidal currents and tidal-current asymmetry

4.2.1 Horizontal patterns

Depth-averaged mean flood and ebb currents (Fig. 9a,b) reveal that tidal channels and their branches play a major role in transporting water back and forth between

the open sea and estuary. Velocities over the tidal flats remain small. Particularly at ebb the small-scale channel contours are clearly seen in the pattern of current magnitude. Ebb velocities are well aligned to the channel axis, while flood currents are much wider. Ebb currents mirror well the extension of the tidal deltas North of the main channel and Osterems inlets (cf. Fig. 1). In the open ocean currents of about 0.5 [m/s] are substantially weaker than those in the tidal channels and deltas (higher than 1 [m/s]), and the horizontal patterns of magnitude are rather smooth.

The ratio between maximum flood and maximum ebb current magnitude (Fig. 9c) displays clearly the different regional responses of tidal currents to the tidal forcing originating from the open sea. North of the tidal inlets and tidal deltas this ratio slightly exceeds unity indicating that flood and ebb are almost symmetric.

Figure 9c demonstrates that Ems Estuary is similar to other estuaries in the region which are forced by a flood-dominant tide (Van der Spek (1997), Dronkers (1986)). Upon approaching the back-barrier islands from the North the dominance of flood current gets more pronounced (Fig. 9c). The patterns there are almost zonally-oriented, following the topography (see Fig. 1 and Fig. 8a for topography map). Another area of pronounced flood-dominance is the vast inter-tidal zone, which falls dry during part of the tidal cycle, i.e. during part of the tidal period the velocity is zero there. These areas are blanked grey on the map.

The patchiness in the tidal delta demonstrates a sequence of ebb and flood dominated zones, in particular in front of the main channel and Osterems Channel. Ebb-dominated currents are located over the deep regions of the tidal deltas, and this in turn identifies the areas dominated by estuarine outflows.

The flood currents along the major estuarine channel are amplified over the channel banks, and this is most notable for the shoals situated to the right of the flood current flow direction. This holds true also for the Osterems Inlet. The major difference between the two channels in their areas connecting the open ocean with the tidal flats is that the major western channel is flood-dominated over

its complete section, whereas in the Osterems ebb currents prevail at the eastern bank.

A very pronounced pattern demonstrating the relationship between topography and tidal asymmetry is the partitioning of flood and ebb dominance in the Alte Ems and Randzelgat Channels (see Fig. 1 for their locations and Fig. 8a for the topography map). In these regions flood and ebb flows are primarily routed in specific channels or parts of each channel. The most pronounced asymmetry appears in the Alte Ems Channel where the ebb dominance is located in its eastern branch and flood dominance is in the western one. A similar partition of asymmetries is also clearly seen in the two-channel system in the area of Dukegat (cf. Fig. 1 for the position).

The asymmetric regime becomes very different around the confluence zone (Fig. 9c) where the branch originating from the tidal river is strongly ebb dominated while the one originating from Dollart Bay splits up into an ebb-dominant and a flood-dominant part. Ebb-dominance is basically due to the hypsometric control (see Fig. 2) and is also clearly represented in Fig. 9c by the proportion of tidal flat area which illustrates the vast storage capacity of this basin. The detailed analysis of the flow dynamics revealed that at the beginning of ebb the Dollart Bay region (including the Dollart part of the tidal river) constitutes a single water surface that drains off towards the Northern part of the confluence zone. In the second part of the ebb period the basins separate along the Geisedam (see Fig. 1) with branched outflows. Only the Southern bank of the confluence zone is flood-dominated.

Upstream of the Dollart Bay along the tidal river, the asymmetry changes slightly in favor of flood dominance, but unlike the areas discussed above the magnitude of maximum ebb currents almost equals that of maximum flood currents. For this area, however duration of flood is shorter (cf. Fig. 8b). It is noteworthy that the flood and ebb dominance is considered here in relationship to the maximum flow speed rather than the duration of the tidal phase. Detailed explanations on the

effects which tidal asymmetries have on the sediment transport and morphodynamics are provided by Brown and Davies (2010) and Robins and Davies (2010). In summary, the estuary can be longitudinally divided into three basic parts: a flood-dominated part where the properties of open-ocean forcing are conserved or further amplified in the funnel-shaped main channel, an ebb-dominated part represented by the Dollart Bay and another flood-current dominated part in the tidal river. These general regularities are interrupted by numerous small-scale patterns, mostly associated with the complex bathymetry in the region.

4.2.2 Vertical patterns

In order to investigate tidal asymmetry in more detail we examine the vertical distribution computed as the ratio between $\max(u_{flood})/\max(u_{ebb})$ of along-channel velocity across four sections (see Fig. 8a for their position), where the maximum is estimated over the period of flood and ebb, respectively. These sections are representative of the complex bathymetry of the main channel. Three of them (i.e. CS I,II,III – cf. Fig. 8a), which are representative of the different channel-structure in the outer estuary, illustrate well the intensification of flood current dominance over the channel flanks (Fig. 10). Overall the patterns are almost vertical which gives an indication of frictional control establishing a similar structure of the distribution of ebb and flood currents in the vertical. The variety of patterns is noteworthy: while there is only a tiny ebb dominance in the upper layer of Borkum inlet towards the left bank (looking landwards) (CS I) the ebb dominance in the central channel at CS II (Fig. 10c,d) is extremely strong in its left side. However ebb asymmetry at CS III (Fig. 10e,f) occurs in the right part of its left channel. Obviously the multi-channel structure of Ems Estuary creates multiple zones with different tidal characteristics, which is expected to support complex transport patterns of suspended sediment. The influence of baroclinicity in this region is indicated by the intensification of asymmetry in the ebb-dominated zones (10c,d and e,f).

There is a slight similarity between CS I and CS IV (Fig. 10a,g) in the sense that in both cases the estuary consists of only one channel and the right side (looking landwards) of it is strongly flood dominated. However unlike the case in CS I, the one in CS IV shows very clear ebb dominance in its left side. Furthermore asymmetries are clearly structured in the vertical revealing that the ebb currents dominate a large part of the surface layers in the left half of the channel in CS IV. This is due to density stratification, which is demonstrated by a comparison with the barotropic run where tidal asymmetry patterns do not clearly display any layers (Fig. 10b,d,f,h). This issue will be addressed further when we consider the vertical distribution of residual flows.

The basic conclusion from this part is that the estuarine dynamics exemplified by the four transects supports the expectations formulated in previous sections. The outer estuary is largely dominated by the barotropic forcing, while the area of salinity front exemplified by CS IV resembles the known cases of estuarine circulation. In the latter the coupling between tidal and fresh water forcing is dominant.

4.3 Horizontal patterns of residual circulation

Water transport across channel sections in Ems Estuary has been analysed by De Jonge (1992) based on measurements. This work, as well as the study of Van de Kreeke and Robaczewska (1993) provided the first descriptions of vertically and tidally averaged flow patterns in the Ems Estuary. The present numerical simulations (Fig. 11a) support these earlier studies and extend the understanding of regional dynamics in vertical and horizontal direction. In contrast with the earlier studies, our analysis includes the inundation of flats and shoals. Of particular importance for the following considerations is the analysis of the difference between baroclinic and barotropic simulations.

Two distinct circulation regimes dominate the area of our numerical simulations: small residual currents north of barrier islands and complex circulation in the

shallow ocean. The vertically integrated residual flow shows however that largest transports occur in the open sea and tidal channels (Fig. 11b). For the sake of clarity the features of residual flow that we introduce in the following paragraphs will be labeled by capital letters (A,B,C...) and marked accordingly in Fig. 11a.

The vertically averaged current pattern is dominated by relatively large-scale residual circulation cells North of the tidal inlets (A) as well as by smaller residual vortices placed either (1) around the main channel, or (2) shaped by the complex multiple channel structure. One example of situation (2) is the recirculation starting in Randzelgat Channel (B) (cf. Fig. 1) and propagating to the south until this channel merges with the Alte Ems channel. At the merging point the circulation reverses to North, follows the Alte Ems channel until the inlet of Borkum, where the direction changes to southward. This circulation pattern is coherent with the ebb-flood dominated pattern in the same area seen in Fig. 9c indicating that the residual circulation is closely related to the pattern of tidal current asymmetry. Another coherence between residual flow and tidal asymmetries can be well detected in the circulation patterns in the tidal deltas (directed towards open ocean) where asymmetry is ebb-dominated.

The case of the anticyclonic circulation in Dukegaat- Hond Paap (C) is geometrically similar, but dynamically very different. In this area one sees a combination of a deep channel (dominated by flood directed current) and tidal flats east of it. The numerical simulations demonstrated that the compensating circulation occurs in the shallows west of Dukegat. This result does not support the view of De Jonge (1992) and some other estuarine studies claiming that under hydraulically stable conditions flood-directed residuals are found in shallow regions. Similar “deviation” from the earlier understanding is the case with the cyclonic vortex to the east of the point connecting Dukegat and Alte Ems, which extends to the east up to the Osterems Channel (D). The conclusion here is that residual flows along the main channel are coupled to the vortex-type circulations in the tidal flats generating on both sides of channel vortices of opposite direction.

One fundamental question associated with the above considerations is about the role of fresh water in establishing these circulation patterns. The comparison between baroclinic and barotropic runs demonstrated that all circulation features in Fig. 11a were simulated in the barotropic run, while the flow magnitudes differ noticeably (Fig. 11c, see also the difference between the two experiments in Fig. 11d). One clear difference is the stronger eastward transport in the open sea simulated in the barotropic run, in particular in the western part of model area. Another difference is the strong northward current on the tidal flats in the eastern part of model area, seen as a southward flow in the difference figure. The intensification of circulation, in particular in the shallow sea demonstrates that the variable density in the baroclinic run tends to reduce the magnitude of the vertically averaged flow. Of utmost importance is that this reduction is small in the area of salinity front and in the tidal river. This brings us to the first important result, which demonstrates that the impact of baroclinicity on the horizontal circulation is subject to specific balance of terms in momentum equation. It appears that density gradients can shut down horizontal circulation, something which has been addressed by Chant and Wilson (1997), Seim and Gregg (1997), Chant (2010), Lerczak and Geyer (2004). Chant (2010) provides an extensive review on how the secondary flows are shut down by the buoyancy effects of vertical density stratification. The second important result is that in the outer estuary the impact of baroclinicity is not negligible. The difference map demonstrates that density field in this area tends to change the circulation pattern. This result is consistent with the discussion presented in section 4.1 on the difference between baroclinic and barotropic experiments seen in the tidally-averaged sea level (see Fig. 8b). It was shown there that in the outer estuary the impact of density can exceed that of the along-channel barotropic pressure gradient.

4.4 Vertical patterns of residual circulation

There is a number of studies on the dynamical controls in estuaries (Wong (1994), Valle-Levinson et al. (2003), Burchard et al. (2011)). Simplified models consider the role of frictional and rotational control represented by the Ekman ($Ek = A_V^M / [fH^2]$) and Kelvin ($Ke = B/R_d$) numbers respectively (Valle-Levinson (2008)), where A_V^M is the vertical eddy viscosity calculated from the model results, H is channel depth, B is the channel width and R_d is the the internal Rossby radius of deformation.

High Ekman number flow ($Ek > 1$) is fully controlled by friction and the exchange flow pattern is purely horizontally sheared (Valle-Levinson (2008)). However, the effect of the Earth's rotation can lead to an inclined shearing plane (Kasai et al. (2000)). This is the case at intermediate Ek , that is $0.1 < Ek < 1$. Under low frictional conditions ($Ek- > 0$) exchange flow is two-layered for small Ke and horizontal shear emerges at $Ke > 2$ (Valle-Levinson (2008)).

Along-channel flow across the sections I - IV (see Fig. 8a for their positions) reveal diverse patterns (Fig. 12), which are difficult to explain in the frame of the above theoretical considerations. We distinguish below three different regimes.

1. One-channel estuary

In CS I (Fig. 12a) and CS IV (Fig. 12g) the bottom is geometrically similar (single channels). In these channels the landward current occupies the right bank (looking landwards) and reaches the deepestmost part of the channel. The outflow fills the left side of the channel, and in the case of CS IV extends almost to the opposite bank in the surface layer. Obviously there is a flow separation and a tilting of the interface between the residual inflow and outflow revealing a typical case of estuarine circulation. As seen from the the results of barotropic run (Fig. 12b,h) flows are clearly delineated to the left and right part of the channel; the separation line being vertical. However without the effect of density in this experiment no tilting of the interface is being observed. There is an overall agreement with the patterns of tidal asymmetry (compare Fig. 10). The sloping contours in CS IV, which do

not occur in CS I give a clear view on the role of density stratification, a known effect which is illustrated in the numerical simulations by Lerczak and Geyer (2004).

2. Circulation controlled by multiple-channel estuary

This case is illustrated in Fig. 12c for CS II and displays two almost equally big channels. In one of them the transport is directed landwards (over most of the section) and in the other one towards the open sea. At the western most part of the section there is a third channel, which is very small and shallow. The transport patterns agree with the asymmetry patterns in Fig. 10c. Comparisons with the barotropic simulation (see Fig. 12d) reveals that there are almost no differences between the two simulations at this section, i.e. the effects of density here do not result in qualitative changes. However, the magnitudes of land-ward flow in the left channel are higher in the barotropic run and those of ocean-ward flow are lower. This suggests that density tends to enhance the outflow and reshape the horizontal circulation.

There is only a weak flow-separation in this section in the western part of the central channel. This situation is in agreement with the theories of friction-dominated controls. However the main specificity of CS II is that the overall dynamics in this area is associated with the separation of opposite-directed flows and not with flow separation in the channel itself.

3. Flow separation within multiple channels

This situation, which is characteristic for CS III gives two different appearances of the inflow and outflow (Fig. 12e,f). We have now two different channels (small and big one). The bigger one displays typical estuarine circulation dominated by friction. Inflow is to be found on the shallows and in the very deep part of the channel, and the outflow fills most of the interior part of channel. However comparison with the barotropic experiment reveals that the flood-directed flow in

the deepest part of channel exists in the two experiments, and so it is not a result of density stratification.

The left (smaller) channel is divided almost into two equal parts showing inflow along the left coast and outflow along the right coast. The difference between barotropic and baroclinic experiments again reveals that in the outflow dominated area the outflow increases and covers larger sectional areas in the barotropic experiment.

Ekman numbers indicate high-frictional dominance throughout the main channel (CSI: 1,0; CSII:0.9; CSIII: 1,9; CSIV: 2,0) and this supports the overall prevalence of the barotropic residual flow pattern. Here we stress the possible importance of lateral shear in the residual flow and the change of horizontal circulation caused by density effects. Thus the horizontal circulation cells could constitute an efficient mechanism for the distribution of matter, e.g. pollutants, in the estuary (see also Geyer and MacCready (2014)).

4.5 Profiles of residual flow in the channel axis

The overturning residual circulation over the deepest part of the channels in CS I,III and IV is illustrated in Fig. 13. Apparently the near-bottom inflowing component has relatively smaller magnitude in the outer estuary (CS I). Further upstream (CS III,IV) it becomes more substantial. In this region however the vertical overturning is accompanied by a sideways tilting of the residual inflow and residual outflow (cf. Fig. 12g), which reflects the basic properties of estuarine flows (Lerczak and Geyer (2004), Chant (2010)).

The extreme variability of vertical structure of horizontal flow leads to the question of the driving mechanisms behind the vertical overturning circulation. In the classical theory of estuarine exchange flow vertical overturning circulation has been attributed to the direct effect of longitudinal and vertical density gradients (Hansen et al. (1965)). Even if the estuary was partially mixed by tides, straining induced stratification could lead to a density driven residual regime

(Nunes and Simpson (1985), Simpson et al. (1990), Burchard et al. (2011)). Several studies extended the traditional understanding by emphasizing the effect of a mixing asymmetry between flood and ebb (Jay and Musiak (1996)) that drives a vertical overturning circulation. These authors suggested that the flood tide momentum would be preferentially mixed downward more effectively than the ebb tide momentum. This is because the longitudinal density variations lead to a reduced eddy viscosity during the ebb whereas during the flood relatively more momentum is mixed from the surface to the near-bottom region (Stacey et al. (2010), Burchard et al. (2011)). This supports a two-layered exchange flow with inflow at the bottom and outflow near the surface in a weakly stratified estuary (Cheng et al. (2011)).

Several physical numbers have been used to classify estuarine exchange flow (see overview in Geyer and MacCready (2014)). The Simpson number (Si) reads

$$Si = \frac{b_x H^2}{C_D U_T^2}, \quad (9)$$

where b_x is the along-channel derivative of buoyancy $b = -g(\rho - \rho_0)/\rho$ and U_T is the amplitude of vertically-averaged along-channel velocity (U). The number reflects the competition between baroclinic and barotropic forcings (Stacey et al. (2010)). At values of Si (> 0.15) the estuary is partially mixed due to tidal straining (Burchard et al. (2011)). For $Si < 0.15$ the estuary is permanently mixed. Nevertheless the vertical distribution of eddy viscosity may vary over the tidal cycle, thereby controlling the effectiveness of mixing (Cheng et al. (2011)). The effectiveness of tidal mixing may be quantified by a mixing number M (cf. Geyer and MacCready (2014)):

$$M_{f,e}^2 = \frac{C_D U_{\langle f \rangle, \langle e \rangle}^2}{\omega \bar{N} H_{\langle f \rangle, \langle e \rangle}^2}; \quad (10)$$

where the subscripts $\langle f \rangle, \langle e \rangle$ refer to average flood and ebb conditions, respectively, and ω is the frequency of the M-2 tide. $\bar{N}^2 = (\frac{-g}{\rho_o} \frac{d\rho}{dz})$ is the buoyancy frequency with top-bottom density gradient $d\rho/dz$.

The values of Si, M_f, M_e in the positions of the vertical profiles in Fig. 13 and of the observed profile in Ems River (in Fig. 6, for location see Fig. 5a) identify the dominant dynamic balances (Tab.2). The Ems Estuary is found to be well-mixed in all cross-sections, except CS IV where the simulations reveal Strain Induced Periodic Stratification (SIPS) regime. Therefore the vertical overturning circulation in the outer estuary (CS I, CS III) cannot be attributed to baroclinicity directly. The quantification of mixing however yields differences between flood and ebb, whereas the values of M are bigger for flood at these locations. Thus the asymmetry in tidal mixing is the plausible explanation for the overturning circulation in these locations. The same applies to CS IV except that Si implies a direct contribution of SIPS. In Ems River mixing numbers indicate a slight flood-ebb mixing asymmetry (cf. Fig. 8b). Nevertheless residual flow is entirely ebb-directed. This can be explained by (1) the large contribution of tidally-induced Stokes drift (Stanev et al. (2007)) during the flood that requires a prolonged ebb (2) river outflow that notably increases ebb-directed residuals in the narrow channel.

4.6 Along-channel profile of residual flow and tidal current asymmetry

An along-channel profile of the longitudinal residual flow (Fig. 14a) reveals a typical pattern of estuarine circulation up to km 45 where the time-averaged current

	Si	M_f	M_e
CSI	0.04	1.5	1.4
CSIII	0.05	1.8	1.6
CSIV	0.19	1.5	1.3
River	0.10	2.0	1.9

Table 2: Si and mixing numbers M_f, M_e for profiles of residual flow in Fig. 13 with drag coefficient $C_D = 0.001$.

in the upper layer follows the direction of river outflow. In the bottom layer currents are up-river. The pattern is patchy between km 45 and the Borkum Island with several local maxima and minima of landward-directed currents. The tidal current asymmetry pattern (Fig. 14b) demonstrates the partitioning of the main channel into two flood-dominated regions between km 0 and 35 as well as between km 50 and 65, and the ebb-dominated zone between km 4 and 50, which occurs in the area of confluence. Between km 35 and 45 the sign of asymmetry reverses as the transect passes through a sequence of flood and ebb dominated zones (see the inset Fig. 9c).

The residual flow is partially congruent with the tidal asymmetry pattern in the outer estuary whereas in the Ems River it is mostly opposing the asymmetry. The overturning circulatory cells between km 30 and 45 span a few kilometers each. Residual flow in this region reveals thus a small-scale phenomenon both in the vertical and the horizontal direction (cf. Fig. 11 for the respective horizontal pattern).

4.7 Vertical salinity distribution

The difference between along channel salinity during flood and ebb (Fig. 15a, d) demonstrates an increasing range with increasing distance from the barrier islands. This range increases rapidly between km 50 and km 62 that is in the region of frontal intrusions (compare with Fig. 7). Even sharper increase is observed in front of the storm surge barrier (cf. Fig 1) giving an indication that this construction greatly affects estuarine dynamics. During flood longitudinal salinity gradient flattens and the water column is only partially or weakly stratified. Under ebb conditions the longitudinal gradient steepens particularly sharply upstream of km 60 and strong stratification establishes in the upper half of the transect, including Ems River and the region affected by the outflow from Dollart Bay (Fig. 15d). Locally stratification attains maxima over sharply varying bottom topography as seen in the cross-sections displaying vertical salinity anomaly (Fig. 15e,f). This

is the case between the Knock region and Dollart Bay where depth reduces by several meters over a distance of a few kilometers only (between km 40 and 45, and also around km 55 in Fig. 15c).

The halocline is particularly strongly stratified at the storm surge barrier during ebb. This reflects the classical tidal situation where relatively fresh water is advected over saltier water, thereby establishing a two-layer flow. The maximum salinity difference between sea surface and channel bed amounts to more than 6 [psu]. A few kilometers downstream of the storm surge barrier stratification weakens and the top-bottom salinity difference does not exceed 1 [psu]. As seen by Fig. 15a the storm surge barrier at km 60 also significantly modifies the longitudinal density gradient which is seen as a steep change at this location. This could be considered as an important control for the rate of intrusion of freshwater into the main estuarine channel. Another local maximum of stratification is in the confluence zone (km 40 - 45) which is the region where the Si number indicates a contribution of the SIPS to residual circulation (see Section 4.5).

In the outer estuary the water column is practically well mixed in both flood and ebb (see previous sub-sections). Zero isohaline in Fig. 15e gives actually the position of mean salinity (red curve in Fig. 15a) along the section line. During ebb this line outcrops at the surface or bottom and the salinity pattern is rather patchy in particular between km 0 and 35. During flood the 0-isoline is located at about the mid water column, suggesting a weakly layered structure.

4.8 Transversal circulation

The importance of transversal circulation for the estuarine dynamics has been clearly demonstrated by Lerczak and Geyer (2004), Chant (2010), Geyer and MacCready (2014). This circulation is essentially a secondary flow in a plain which is normal to the main along-channel flow. It is about ten times weaker than the along channel flow and it results from the interplay between lateral mixing, along channel dynamics and dispersion. Modelling the secondary flows is challenging and good

1 model performance can serve as a prove of model realism. Deviations from simpli-
2 fied theoretical cases, as will be shown in the following, could identify specificities
3 of the regional balances dominating estuarine circulation. It is thus expected that
4 in the Ems Estuary the large variety of cross-channel variations in topography
5 drives secondary flows through differential advection (Nunes and Simpson (1985),
6 Lerczak and Geyer (2004)).
7

11 The three-dimensional velocity field at CS I (a single channel section) is presented
12 in Fig. 16a, b, c during flood, ebb and slack after ebb conditions. Flood situation
13 supports the concepts of Nunes and Simpson (1985), who demonstrated that con-
14 vergence and downward motion is observed in the middle of channels during the
15 flood phase. Ebb conditions (Fig. 16b) reveal a divergent flow and an upward mo-
16 tion in the channel interior. The pattern of vertical circulation cell differs from the
17 one seen in the results from Lerczak and Geyer (2004), the latter demonstrating
18 a two-cell circulation for the same tidal phase.
19

26 The comparison of the baroclinic and barotropic run (Fig. 16d) demonstrates that
27 the vertical circulation at CS I is largely due to the baroclinicity. However the
28 differences between the simulations presented here and the idealized experiments
29 are indicative that the specific circulation pattern is primarily governed by the
30 bottom topography and curvature effects (see also Chant (2010)). The large dif-
31 ferences between the simulations discussed above and theoretical cases such as the
32 one discussed by Lerczak and Geyer (2004) beg the question of whether the used
33 model is capable of simulating secondary circulation. With the following numer-
34 ical experiment we will give an answer to this question. The model set-up used
35 above was applied to a tidally-driven idealized estuary with parabolically shaped
36 cross-sectional bathymetry. The estuary's width decays exponentially from 12 km
37 at the open sea to 1 km at the landward end. The length of the estuary is 100 km,
38 which is comparable with the length of the Ems Estuary, the depth decreases from
39 25 m at the mouth to 5 m at the head. The model is initialized from rest with a
40 constant along-channel density gradient. The forcing consists of an M2 tide, con-
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stant salinity of $30[psu]$ at the open boundaries and $0[psu]$ at the end of channel. A constant inflow of $80[m^3/s]$ which also corresponds to the conditions in Ems Estuary is prescribed. The model output analysed below has been sampled after 30 days of simulation during which a quasi-steady state with zero trend has been reached.

The results discussed below are for a cross-section situated near the head of the estuary where the salinity front is well pronounced. Because the performance of the realistic model during slack after flood and slack after ebb did not reveal well classical patterns known from the theory we show below the results from the idealized experiments only during these periods (Fig. 17a,b), which are identified by the height of sea level in the plots.

The two-cell circulation at slack water after flood is similar to the pattern simulated by Lerczak and Geyer (2004) with a downwelling in the middle of the channel. This feature is very important because it supports the tidal straining. However the transversal circulation is not fully symmetric, which results from the finite dimensions of the estuary (shorter than in some other theoretical studies), specific aspect ratios and specific balances at this particular transect. The asymmetry during slack after ebb is even stronger, but the two cells still rotate in the sense described by Lerczak and Geyer (2004).

We did not perform experiments with different parameters of tidal channel or forcing in order to increase the consistency with some well known previous simulations. This was not aimed here because some previous simulations were essentially two-dimensional (e. g. Burchard et al. (2011)); in some others tidal channels were very long, which did not correspond to the shape of Ems Estuary. The key finding here is that (1) the model is capable of reproducing dynamic characteristic for the estuarine circulation, and (2) the large differences between the realistic simulations and the idealised cases justify the use of the former as tools to better understand

the circulation in individual estuaries. Thus we conclude that the present study has achieved one of its basic objectives formulated in the Introduction.

5 Discussion

The ability of the model to realistically simulate sea surface elevation in the estuary has been demonstrated by inter-comparison between tidal analysis of observations and model results. Inside the estuary the performance turns out to be particularly satisfactory. The reproduction of the sea surface slope between two stations in the inner estuary (Fig. 4) has been achieved with high accuracy even though the tidal wave encounters a complex basin geometry.

For the first time a model setup of a Wadden Sea estuary has been validated against a large dataset of ADCP observations that extend in time and space, continuously covering different morphological environments and the majority of a tidal cycle. This validation proves that a state-of-the art hydrodynamical model can effectively cross different physical scales in one setup without resorting to nesting techniques. The same setup has been successfully validated against 12-hour profiles of along-channel velocity in the tidal river. This profile illustrates the internal tidal asymmetry of flow characterized by strong vertical mixing of momentum during flood compared with a more vertically sheared flow at ebb. These and other properties of tidal flow, including maximum velocities, phasing, acceleration and deceleration of the current, have been satisfactorily reproduced by the model.

The ship survey along the main estuarine channel also yielded CTD-observations of surface salinity, revealing a strong gradient in the converging region of the estuary. In view of the tidal variation of the salinity field, model results compared fairly well against the observed longitudinal salinity gradient. This is an important result since the steepness of the longitudinal salinity gradient is strongly related to the degree of stratification of the water column. It has been demonstrated that during

ebb the tidal river and the upper region of the outer estuary are notably stratified. A particularly strong stratification is induced by the storm surge barrier.

In the second part of the results section the focus has been on tidal asymmetry and residual or exchange flows. The vertical tides in the main channel are characterized by a landward rising tidal range and an overall negative duration asymmetry, which is significantly reduced in the vicinity of Dollart Bay by the effect of hypsometric control. Other than the storage area of tidal flats and shoals, variations of the channel topography have only minor effects on the vertical tides. Horizontal tides behave very differently. It has been shown that vertically-averaged mean flood and mean ebb currents exhibit distinct patterns. The differential evolution of flood and ebb has been further illustrated by the ratio of maximum flood and maximum ebb currents, which yielded the first ever 3D mapping of tidal current asymmetry in the Ems Estuary. Tidal current asymmetry is shown to strongly dominate the horizontal and vertical patterns. Of utmost importance was the demonstration that the tidal mixing asymmetry acted as the basic physical mechanism controlling the gravitational circulation. The comparison between barotropic and baroclinic runs provided evidence of the impact of density gradients on vertical and horizontal patterns of residual flow, which vary greatly for the analysed sections. The vertical overturning residual circulation was attributed to the density-induced mixing asymmetry (see Figs. 12-15). It is noteworthy that in a recent modeling study Li et al. (2010) emphasized the role of baroclinic pressure gradients as a driver of the tidal mixing asymmetry. The comparison between the baroclinic and barotropic experiments, which were driven by the same barotropic forcing, revealed that in the barotropic case velocities were overall higher than the baroclinic case. This could indicate that part of the horizontal momentum imported from the boundaries was lost in the baroclinic case because of the buoyancy effect and reduction by stratification mixing opposing the barotropic forcing.

The dominance of flood currents in the outer estuary and in Ems River acts as a major driver of tidal pumping of sediments (Winterwerp (2011)). Local deviations

from this basic dominance have been identified in the Dollart Bay. This shallow area functions as a big natural retention basin that induces ebb-current dominance in a region that extends several kilometers seawards from the mouth of the bay. Another morphological configuration that apparently controls the balance between flood and ebb dominance is the bifurcation of the main channel into multiple channels. One basic result from the analysis of numerical simulations is that the Ems Estuary has to be considered as a multiple-channel system characterized by complex patterns of tidal asymmetry and recirculations. Although the dynamics of the outer Ems Estuary is strongly driven by the barotropic forcing, the density distribution substantially modifies the patterns of circulation (Fig. 11d). The constriction of the channel by tide gates modifies the outgoing tide more than the incoming tide. At ebb an above average stratification is found downstream of the storm surge barrier. This leads to a weakening of ebb-currents and associated transport of passive tracers in the near-bottom region whereas ebb-dominance is induced in the surface layer. We believe that this specific impact will be of interest beyond this local study.

6 Conclusions

The validation of a numerical model and its particular setups against observations and theoretical concepts demonstrated its realism and applicability to the studied estuarine situation. This realism was partially attributed to using very fine resolution enabled by the unstructured grids, which is more optimal than using multiple nested models. The good model performance gives also hope that sediment dynamics and associated specific ecological problems in this estuary, including response to dredging, can be well tackled in the future with this model. The differences from the idealised simulations demonstrated the complex nature of dynamics in the Ems Estuary (Fig. 9c) and justified using the model to advance the understanding of regional processes. We demonstrated that over large areas in the outer estuary the dynamics are almost barotropic. However, even relatively weaker, the

1 baroclinicity supports complex circulation between barrier islands and tidal river.
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3 In the area between the tidal river and the outer estuary the model reveals clearly
4 that concepts of estuarine circulation known from earlier theoretical and observa-
5 tional studies apply. However the complex regional conditions are responsible for
6 the relatively large deviations of present simulations from idealized models.
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10 11 12 13 14 15 16 17 18 19 20 21 22 23 24 25 **7 Acknowledgements** 26

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Fig. 1: The model area as part of the German Bight (upper right panel). The panel in the middle displays details in the bottom topography along with the major geographical names referred to in the text. Three magnified panels are also given with model grid illustrating the resolution in different areas. The position of the storm surge barrier is identified by the ellipse. The positions of tide gauge stations providing data for model validation are indicated with arrows in the respective insets.

Fig. 2: Volume of the Dollart Bay (the area South of Geise Dam) in green and of the tidal river (including also the section North of Geise Dam) in blue plotted as a function of mean sea elevation. The red curve gives the sum of green and blue curves.

Fig. 3: Amplitudes of observed (blue) and simulated (red) tidal constituents. Analysis of sea-level uses t-tide package (Leffler and Jay (2009)), the analysis time covers two spring-neap-cycles, from the 1st of June 2012 until the 30th of June 2012.

Fig. 4: Water levels at Knock (black) and Pogum (red) in observations (a) and simulation (b). The difference of water levels between Knock and Pogum is plotted in (c) where the blue line corresponds to observations and the red line to simulations. The analysis period starts on the 10th of June 2012 at 09:50h and ends on the 12th of June 2012 at 01:50h.

Fig. 5: (a) Ship transect taken on the 7th of June 2012 with motor boat Otzum from 12:30 h to 16:45 h (red line) with bathymetry [m] as a background. Selected times illustrate the timing of continuous measurements with a time interval of 0.1 s. At the location marked by the black cross (at Pogum) extensive cross channel sections have been performed with the same motor boat on the 12th of June 2012 from 05:00 h to 17:30 h. (b,c) Comparison of ADCP observations of horizontal surface velocity [m/s] (blue dots) and model simulations (red dots) along the ship's transect (red line) in (a). The along-channel component (b) has positive values in the flood direction and the cross-channel component (c) has positive values directed to the left of along-channel velocity. Simulated surface elevation (green dotted line) is also given in order to better represent the timing of change in surface current.

Fig. 6: Time versus depth diagram of along-channel velocity [m/s] as (a) simulated by the model and (b) observed by an ADCP (see the area near Pogum in Fig 1 and also Fig. 5a). Positive velocities are directed toward the River Ems (flood). The profile is vertically limited by the ADCP's blank-after-transmit distance of 1.4 m at the sea surface and a blank zone of 1.3 m at the bottom. Observations have been taken on the 12th of June 2012 from 05:00 h to 17:30h. The isolines (white) of $|u| = 1 \text{ m/s}$ are also given.

Fig. 7: (a) Comparison of CTD observations of sea surface salinity [psu] to model results along the ship's transect (red line) in Fig. 5a. Model results are represented in the form of a Hovmoeller plot with the y-axis corresponding to the timing and the x-axis to the distance from the beginning of the transect. Observations have been carried out with motorboat Otzum on the 7th of June 2012 starting at 12:25 h and ending at 16:45 h. Data are projected onto the the closest point in the time-space plane that is spanned by the model results. The starting point of the transect is indicated by the green cross and its end is marked by the red cross. The white and black contour lines identify the 25 [psu] and 15 [psu] isohalines, respectively. (b) Time series of observations (blue line) and sampled from the simulations in the observational locations salinity (black dashed line). The maximum and minimum salinity (red dashed lines) illustrate the local tidal range estimated by the model.

(a)

(b)

Fig. 8: (a) Transect (blue) and cross-sections (CS) I-IV along which no measurements have been taken but along which analysis of simulations is presented further in the text. The background shows the bottom topography. (b) Tidal range (black line) [m], ratio between tidal range and depth (blue line), tidally averaged water level (green solid line) [m], difference between the tidally-averaged water level between the baroclinic run and the barotropic run (green dashed line) [m] and duration difference between flood and ebb (red line) [h] along the main channel (blue transect line in (a)).

Fig. 9: Time averaged flood (a) and ebb (b) magnitude of depth-averaged horizontal velocities [m/s]. (c) is the ratio between maximum flood and maximum ebb depth-averaged velocity magnitudes. Values greater than one indicate flood current dominance, values lower than one indicate ebb current dominance. Locations that fall dry are blanked out with grey color. The data has been sampled from one tidal cycle of model output (baroclinic run) from the 6th of June 2012. The insets in (a,b,c) give a zoomed-in view of the region of confluence (see the box in (c)).

Fig. 10: Ratio between maximum flood and maximum ebb velocity magnitudes along cross-section lines in Fig. 8a. The results are given for the baroclinic run (a) - CS I; (c) - CS II, (e) - CS III; (g) - CS IV, and for the barotropic run (b) - CS I; (d) - CS II, (f) - CS III; (h) - CS IV. Values greater than one indicate flood current dominance, values lower than one indicate ebb current dominance. The data has been computed from two tidal cycles of model output dating from the 6th and 7th of June 2012.

(a)

(b)

(c)

(d)

Fig. 11: (a) Vertically averaged residual circulation [m/s], (b) vertically integrated residual transport [m²/s], (c) vertically averaged residual circulation [m/s] of the barotropic run, and (d) difference between the baroclinic run (a) and the barotropic run (c) [m/s]. The bathymetry [m] is given a background. The analysis time covers two tidal cycles beginning on the 6th of June 2012 at 05:00 h. The inset in (a) gives a zoomed-in view of the residual circulation in Dollart Bay. The red labels (A,B,C,D) in (a) correspond to the physical patterns that are introduced in section 4.3.

Fig. 12: Residual flow [m/s] across cross-section lines in Fig. 8a: The results are given for the baroclinic run (a) - CS I; (c) - CS II, (e) - CS III; (g) - CS IV, and for the barotropic run (b) - CS I; (d) - CS II, (f) - CS III; (h) - CS IV. The data has been computed from two tidal cycles of model output dating from the 6th and 7th of June 2012. Positive values are flood-directed.

Fig. 13: Vertical profiles of along-channel residual velocity in the deepest parts of tidal channels. The averaging is shown for the channel axes of three cross-sections (see legend and Fig. 8a for the positions of cross-sections). Average velocities are given as a function of depths greater than low water. The analysis time covers two tidal cycles starting on the 6th of June 2012 at 05:00 h.

Fig. 14: (a) Along-channel profile of residual flow [m/s] in the main channel (cf. Fig. 8a). (b) is the ratio between maximum flood and maximum ebb along-channel velocity. Values bigger than 1 indicate on flood-dominated asymmetry, residual velocities are flood-positive. The data has been computed from two tidal cycles of model output dating from the 6th and 7th of June 2012.

Fig. 15: Along-channel transect (see the blue line in Fig. 8a for the channel location) of vertically averaged salinity (a) during flood and (b) ebb, as sampled from the model output from the 6th of June 2012 at 14:00 h and 20:50 h, respectively. The vertical salinity sections are given for (c) flood and ebb (d). The deviations from the vertical averages (profiles in a and b subtracted) along the same cross-section are plotted in (e) and (f) for flood and ebb conditions, correspondingly. Units are [psu].

Fig. 16: Flow field in CS I (see Fig. 8a for location of section) during (a) flood, (b) ebb and (c) slack after ebb in the baroclinic run. The difference between the baroclinic and the barotropic run at times of slack after ebb is also given (d). The big colorbar represents the along-channel velocity [m/s], positive values being flood directed. The small colorbar represents the difference in along-channel velocity between the baroclinic run and the barotropic run in (d). Here positive values are indicative of baroclinic velocities being larger than the barotropic velocities. Vectors illustrate transverse circulation. The vertical component has been scaled by a factor of 100. All the vectors in (c) and (d) have additionally been scaled by factor 5. Model output is from 6th of June 2012.

Fig. 17: Three-dimensional flow field in an parabolically shaped estuary. The colorbar represents the along-channel velocity [m/s], positive values being flood directed. The vectors illustrate transverse circulation. The vertical component has been scaled by factor 100. The sub-panels correspond to snapshots taken at (a) slack after flood and (b) slack after ebb. The cross-section is situated near the head of the estuary.