Sediment dynamics in tidally dominated environments controlled by transport and turbulence: A case study for the East Frisian Wadden Sea

E. V. Stanev, G. Brink-Spalink, and J.-O. Wolff

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Various tide-related sediment transport mechanisms near a barrier island coast are addressed on the basis of analytical theory and numerical simulations carried out with a three-dimensional (3-D) numerical model. The theory proposed gives an explanation of the observed “twin peaks” in the concentration of suspended sediment; however, it does not adequately describe the whole variety of sediment responses. The major idea in this paper assumes that it is not only the change of the level of turbulence and advection individually but also the correlation between the two which controls the concentration of sediment. Because the horizontal gradients of sediment concentration are created not only by sources at the coast (e.g., fluxes from rivers) but also by the transport and turbulence, the system is highly nonlinear. Several simulations have been carried out aimed at revealing the nonlinear mechanisms in the sediment transport and individual contribution of turbulence and velocity field. An experiment in which the sediment model is driven only by turbulent kinetic energy fails to simulate most of the characteristic temporal and spatial sediment patterns. With another experiment it is demonstrated that the velocity gradient in the vertical gives an important mechanism shaping the patterns of temporal and spatial variability (and tidal asymmetries). Without shear diffusion, the sediment dynamics are largely governed by the “displacement mechanism”; that is, the system is more linear. One practical conclusion from the numerical simulations is that using simplified (2-D) dynamics in sediment transport modeling along with externally prescribed turbulence characteristics could lead to large artifacts.


1. Introduction

The transport of sediment controls the morphodynamics over large coastal areas of the Wadden Sea (and other near-coastal zones) and presents an important scientific challenge for sedimentologists, coastal engineers and physical oceanographers. A wide spectrum of approaches from physical oceanography could be applied to study the transport of sediment including simple process models and comprehensive three-dimensional (3-D) numerical models with realistic forcing. We demonstrate in this paper the usefulness of simple analytical and 3-D models in developing further understanding of the role of individual properties of transport patterns in sediment dynamics.

It is widely accepted that the phase difference between velocity and concentration of sediment can produce residual transports even in the case when there is no residual motion of water. There are two well known processes named scour and settling lags, which are used to explain the trends in transport of suspended matter. The scour lag is due to the difference between current speed which is required to erode the sediment and the one, which is sufficient to maintain it in suspension, the settling lag is measured by the time when the dynamics can no longer maintain the sediment in suspension and the time when the sediment reaches the bottom. Earlier models describing the lag effects were qualitative and explained the coastward migration of sediment particles as a result of the settling lag [van Straaten and Kuenen, 1958; Postma, 1961]. In this context it is important to note that the interplay between grain threshold velocity and tidal current works as a sorting mechanism leading to a progressive decrease of grain size landward. Later, Groen [1967] proved mathematically the possibility to develop a net (landward) sediment transport in a purely alternating flow (equal maximum ebb and flood currents) provided the tidal response was asymmetrical (shorter time is needed for the flood currents to reach maximum speeds than for ebb currents). Recently, Pritchard and Hogg [2003] revisited the theory of cross-shore trans-
port of sediment using a simple vertically integrated 1-D model and Lagrangian approach extending thus the concepts of Friedrichs and Aubrey [1996] about the equilibrium morphology of mud flats.

[4] Up to 7200 tons of mud and 4300 tons of sand are moved in and out of the tidal channel of Otzumer Balje (Figure 1) over one tidal cycle [Santamarina Cuneo and Flemming, 2000]. However, the net accumulation is small reflecting the precise balance during one tidal cycle between local erosion and deposition from one side, and transport from another. While there is a large number of studies addressing the quasi-equilibrium conditions, the governing dynamics, including the temporal variability of sediment transport, deserve further investigation. One important drawback in the existing theories is that they still do not adequately address the individual contribution of different properties of the velocity field (gradients, temporal asymmetries, etc.). It is also of great importance to compare the contributions of the horizontal transport and turbulence, both of which shape the patterns of sediment in a different way thus exerting the basic physical control on the morphodynamics.

[5] The vertical transport of suspended particulate matter (SPM) has been extensively studied and a detailed description of the theory and observations is provided by Hill and McCave [2001]. The contribution of horizontal (water) transport to the sediment dynamics has also been previously addressed in a number of publications, where the phase relationships between sediment suspensions and tidal currents have been discussed using observations and modeling [Bass et al., 2002]. The correlation between SPM and current speed in estuaries is known for a long time [Allen et al., 1977; Officer, 1980].

[6] A simple conceptual model combining the displacement (advection) and resuspension processes is presented by Weeks et al. [1993] where it is suggested that tidal excursions explain the observed semidiurnal concentrations of SPM, while resuspension explains the quarter-diurnal variations. It is, however, not clear to what extent the superposition between the two giving the so-called “twin peaks” is valid in nonlinear tidal systems. As discussed by Jago and Jones [1998], conceptual models based on a depth-integrated description are not very successful in the whole water column (they match observations quite well at 10 and 20 m above the bed, but are less successful at 1 or 5 m above the bed), not to speak about possible problems in the horizontal direction where currents could largely vary.

[7] The major problem here is that in the “displacement theory” it is assumed that there is a horizontal gradient of SPM and by shifting the front back and forth the concentration increases (decreases) once per tidal period (we will call this in the following a one-modal oscillation). However, in a real sedimentary system the gradients from the coast to the open sea are not always the only important characteristic of SPM. The temporal evolution of sediment concentration away from the coast, over shallow topography plays an equally important role. There the horizontal gradients are created not only by sources at the coast (e.g., fluxes from rivers) but also by the interplay between erosion and sedimentation, this is controlled by the transport. This nonlinear coupling precludes simple speculations about possible responses.

[8] Our interest here is to further develop the above concepts using simulations with a 3-D circulation model coupled with a sediment transport model. This approach allows us to produce data from simulations over large areas. The latter is very important if we want to address the role of advection (we recall here that most of observations and theories on the relationship between sediment and dynamics are local). By comparing the simulations with local measurements in the area of the East Frisian Wadden Sea we aim at finding support of the results from 3-D modeling. This is very important because the temporal and spatial characteristics of turbulence which control sediment dynamics are largely unknown. However, measured SPM signals contain information about both the advection and turbulence, giving thus an important source of data for model validation.

[9] The paper is structured as follows. We first describe in section 2 the numerical model, followed in section 3 by analysis of circulation and turbulence. The major results are presented in section 4 where we discuss the individual contribution of different properties of the dynamics to the
establishment of the sedimentation pattern in the Wadden Sea.

2. Numerical Model

2.1. Circulation Model

[10] The present work uses results of numerical simulations with the General Estuarine Transport Model (GETM) coupled with a sediment transport module. GETM is a 3-D primitive equation numerical model [Burchard and Bolding, 2002] (hereinafter referred to as BB) assuming a hydrostatic pressure distribution, which means that it applies only to nearly horizontal flows. This is a justifiable assumption in the situation addressed in this paper. Furthermore, the model assumes a well-mixed system without density stratification, neither due to temperature and salinity, nor to gradients in sediment concentration. The momentum and continuity equations in Cartesian coordinates read:

\[
\frac{\partial u}{\partial t} + \gamma \left( \frac{\partial (u^2)}{\partial x} + \frac{\partial (uv)}{\partial y} + \frac{\partial (uw)}{\partial z} \right) = -g \frac{\partial \zeta}{\partial x} + A_H \nabla^2 u
\]

(1)

\[
\frac{\partial v}{\partial t} + \gamma \left( \frac{\partial (uv)}{\partial x} + \frac{\partial (v^2)}{\partial y} + \frac{\partial (vw)}{\partial z} \right) = -g \frac{\partial \zeta}{\partial y} + A_H \nabla^2 v
\]

(2)

\[
\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0
\]

(3)

where \(u, v\) and \(w\) are the velocity components with respect to the \(x\) (east), \(y\) (north) and \(z\) (upward) direction, respectively, \(f\) is the Coriolis parameter, \(g\) is the acceleration due to gravity, \(\zeta\) is the sea surface height, \(A_H(k, \varepsilon, \gamma)\) is a generalized form of the vertical eddy viscosity coefficient, \(k\) the turbulent kinetic energy (TKE) per unit mass and \(\varepsilon\) the eddy dissipation rate (EDR) due to viscosity. The lateral eddy viscosity \(A_H(x, y)\) has been introduced to suppress nonphysical noise along the open boundaries in a 3 grid point wide sponge layer, where it changes exponentially from its boundary value of \(10^3 \text{ m}^2 \text{ s}^{-1}\) to \(1/\varepsilon\) of this value. In the interior of our model domain the dissipation is dominated by vertical friction. This is a valid assumption for the present model, in which the vertical eddy viscosity coefficient is \(\sim 10^{-1} \text{ m}^2 \text{ s}^{-1}\) [Stanev et al., 2003a] (hereinafter referred to as SWBBF), the horizontal resolution is \(\Delta x = 200 \text{ m}\), and the coarsest most vertical resolution \(\Delta z \sim 1 \text{ m}\). Even for extremely large magnitudes of the horizontal eddy viscosity coefficient (for areas of such dimensions) of \(\sim 10^3 \text{ m}^2 \text{ s}^{-1}\) the ratio between horizontal and vertical mixing terms \((A_H / A_{\text{vertical}}) \leq 1\).

[11] The success of near coastal models in tidal basins depends largely on the capabilities of models to account adequately for the process of drying and flooding. One example is given by Oey [2005] describing the drying and flooding algorithm applied recently to the Princeton Ocean Model (POM). In GETM, this process is incorporated in the hydrodynamic equations through a parameter \(\gamma\) which equals unity in regions where a critical water depth \(D_{crit}\) is exceeded and which approaches zero when the thickness of the water column \(D = H + \zeta\) tends to a minimum value \(D_{min}\):

\[
\gamma = \min \left(1, \frac{D - D_{min}}{D_{crit} - D_{min}}\right)
\]

(4)

where \(H\) is the local depth (constant in time), taken as the bottom depth below mean sea level in the model area. The minimum allowable thickness \(D_{min}\) of the water column is 2 cm and the critical thickness \(D_{crit}\) is 10 cm (BB and SWBBF). For a water depth greater than 10 cm \((D > D_{crit}\) and \(\gamma = 1\), the full physics are included. In the range between critical and minimal thickness (between 10 and 2 cm) the model physics are gradually switched toward friction domination, i.e., by reducing the effects of horizontal advection and Coriolis acceleration in equations (1) and (2) and varying the vertical eddy viscosity coefficient \(A_H\) according to

\[
A_H = \nu_t + (1 - \gamma)\nu_h
\]

(5)

where \(\nu_h = 10^{-2} \text{ m}^2 \text{ s}^{-1}\) is a constant background viscosity. The eddy viscosity \(\nu_t\) is obtained from the relation

\[
\nu_t = c_{\mu} k^2 \varepsilon
\]

(6)

where \(c_{\mu} = 0.56\) [see, e.g., Rodi, 1980]. The above description demonstrates that the drying and flooding parameterization does not simply switches off all dynamics below a certain water depth, but uses a dynamic approach by introducing the factor \(\gamma\). Furthermore, it is demonstrated by Stanev et al. [2007] that dynamics associated with flooding and drying tend to oppose the Stokes transport, the latter being an important factor establishing asymmetries of tidal response in shallow waters.

[12] The momentum equations (1) and (2) and the continuity equation (3) are supplemented by a pair of equations describing the time evolution of the TKE and EDR

\[
\frac{\partial k}{\partial t} - \frac{\partial}{\partial z} \left( \nu_t \frac{\partial k}{\partial z} \right) = P - \varepsilon
\]

(7)

\[
\frac{\partial \varepsilon}{\partial t} - \frac{\partial}{\partial z} \left( \nu_t \frac{\partial \varepsilon}{\partial z} \right) = \frac{\varepsilon}{k} (c_1 P - c_2 \varepsilon)
\]

(8)

where \(\sigma_k\) and \(\sigma_\varepsilon\) are the turbulent Schmidt numbers, \(c_1 = 1.44\), and \(c_2 = 1.92\) [see Rodi, 1980; BB; SWBBF]. In the above equations the advection of TKE and EDR is ignored, which is a widely used simplification used to describe model systems of the type studied here.

[13] The vertical shear production \(P\) is a function of the shear frequency \(S\):

\[
P = \nu_s S^2
\]

(9)
with

\[ S^2 = \left( \frac{\partial u}{\partial x} \right)^2 + \left( \frac{\partial v}{\partial z} \right)^2 \]  

(10)

[14] There is clear observational evidence that the velocity profiles close to the bottom in the East Frisian Wadden Sea can be adequately described by a logarithmic boundary layer [Davis and Flemming, 1991; Antia, 1993]:

\[ u(z')/u'_w = \frac{1}{\kappa} \ln \left( \frac{z' + z_0}{z_0} \right) \]  

(11)

where \( u'_w = \sqrt{k\tau_b} \) is the friction velocity at the seafloor, \( \tau_b = \rho_w u'_w \sqrt{\frac{\rho_w}{\rho}} \) is the bed shear stress, \( z' \) is the distance from the bed, \( z_0 \) is the bottom roughness length, \( \kappa \) is the von Karman constant, and \( \rho_w \) is the water density. Furthermore, close to the bed, TKE and EDR are governed by the law of the wall with

\[ k = \frac{(u'_w)^2}{c_v^{1/2}}, \quad \varepsilon = \frac{(u'_w)^3}{\kappa(z' + z_0)} \]  

(12)

[15] Boundary conditions for equations (7) and (8) are described in detail by BB and SWBBF along with the used parameterizations. It is noteworthy that we cannot simply set the lowermost velocity to zero. This would be physically correct for a point directly at the bed, but the lack of resolution in some locations precludes this possibility. To solve that problem we assume that the lowermost grid box is fully submerged within the logarithmic boundary layer. This does not mean that we directly trigger a log layer in the whole water column, but only that

\[ u_1 = \frac{u'_w}{\kappa} \ln \left( \frac{0.5h_l + z_0}{z_0} \right) \]  

(13)

where \( h_l, u_1 \) are the thickness and velocity of the deepest box, respectively.

[16] The bed shear stress is formed by two important mechanisms: tides \( (\tau_t^n) \) and wind waves \( (\tau_w^n) \), where indices \( t \) and \( w \) stand for tide and wind. These two mechanisms control the hypsometries (distribution of horizontal surface area with respect to depth) in tidal basins, as demonstrated by the theory of Friedrichs and Aubrey [1996]. The convex hypsometry in the Wadden Sea [Dieckmann et al., 1987; Stanov and Wolff, 2003] gives an indication that the contribution of tides is dominant. Because (1) the impact of wind waves on the sedimentary system in the same area is addressed elsewhere [Stanov et al., 2006], (2) bed stress induced by wind waves could mask individual effects of tidal response in some areas, and (3) this study has the character of a process study focused on the tidal response, we will consider here only the tidal factor.

[17] The subgrid-scale parameterizations are of utmost importance for the sediment transport, and the analysis of simulations by SWBBF demonstrates that the general characteristics of turbulence in the bottom boundary layer are consistent with the requirements formulated by Dyer and Soulsby [1988] for correct prediction of sediment transport rates. Two of them are the logarithmic velocity profile and parabolic-like distribution of \( A_S \):

[18] The parameter \( z_0 \), which gives a general representation of the bottom roughness is taken constant over the whole area (SWBBF). This simplification does not account for some bed forms (e.g., ripples), which are subgrid-scale features for the model. We admit that they are important elements of bed load transport (bed load transport can be estimated by measuring bed form propagation). Because this study does not address morphodynamics and other details associated with bed load transport, but rather larger-scale balances, we will not introduce very complex physical parameterizations.

2.2. Sediment Transport Model

[19] We consider two particle classes, one being mud and the other sand of a grain size of \( d = 63 \mu m \) and \( d = 200 \mu m \), correspondingly. The sediment transport model uses a standard diffusion-advection equation for the concentration of both particle classes \( c \) where settling is added on the right hand side:

\[ \frac{\partial c}{\partial t} + u \frac{\partial c}{\partial x} + \frac{\partial c}{\partial z} + w \frac{\partial c}{\partial z} = \frac{\partial}{\partial z} \left( A_s \frac{\partial c}{\partial z} \right) + \frac{\partial}{\partial z}(w_s c) \]  

(14)

[20] In the above equation \( w_s \) is the settling velocity of the sediment in suspension. For the cohesive sediments, which include mud particles and parts of the silt fraction, the settling velocity is concentration-dependent. Neglecting the horizontal diffusion in equation (14) is justified by the same arguments as have been used for the horizontal friction in section 2.1.

[21] In the following we present first the parameterizations associated with the cohesive sediments, to which we refer to as fine SPM \( (d < 63 \mu m) \). Higher concentrations result in a formation of larger aggregates which in turn have a larger settling velocity. Experiments reveal a strong increase of the settling velocity with the sediment concentration [van Leussen, 1988] which is expressed by the formula

\[ w_s = k_s c_{nd}^{m_s} \]  

(15)

where \( c_{nd} = \frac{c}{c_{mn}} \) is a nondimensional concentration of the grain size in question, \( c_{mn} = 1 \text{ kg m}^{-3} \), \( k_s \) and \( m_s \) are empirical constants. These are chosen to be \( k_s = 0.017 \text{ ms}^{-1} \) and \( m_s = 1.33 \) in agreement with the measurements summarized by van Leussen [1988].

[22] The sediment flux at the seabed

\[ \left( A_s \frac{\partial c}{\partial z} + w_s c \right)_{z=-H} = E - D \]  

(16)

which is the bottom boundary condition of equation (14), is based on well-known parameterizations of deposition and erosion rates \( D \) and \( E \). The deposition rate given by Einstein and Krone [1962] is

\[ D = w_s c_s \left( 1 - \frac{\tau_s}{\tau_d} \right) \]  

(17)
where $c_h$ is the fine SPM concentration near the bottom, $\tau_b$ is the bed shear stress and $\tau_d$ is the critical shear stress for deposition. The bottom concentration is extrapolated from the concentrations of the two lowest layers ($c_1$ and $c_2$, both located in the middle of the layer) as $c_h = c_1(c_1/c_2)^s$. For equidistant resolution $q \approx 2$. This extrapolation is consistent with the Rouse profile and is superior to the linear extrapolation, the latter is very inaccurate in the areas of high bottom gradients.

[23] The erosion rate is computed using the formula of Partheniades [1965]:

$$ E = \alpha M_e \left( \frac{\tau_b}{\tau_c} - 1 \right) $$

(18)

where $M_e$ is an empirical constant giving the erosion rate at twice the critical shear stress for erosion. The parameter $\alpha$ specifies the fraction of the grain size in question in the bottom sediment and is 0.5 for both sediment types. The value used for $M_e$ is $3.7 \times 10^{-6}$ kg m$^{-2}$s$^{-1}$. This is somewhat smaller than the values suggested by other authors [e.g., Mehta, 1988; Puls and Sündermann, 1990; Clarke and Elliot, 1998] ranging between $6 \times 10^{-6}$ and $4 \times 10^{-3}$ kg m$^{-2}$s$^{-1}$, but showed good results in initial model runs.

[24] The critical shear stress for erosion of fine SPM is set constant to 0.2 Nm$^{-2}$. The critical shear stress for deposition is chosen to be equal to that for erosion. This means that either deposition or erosion occurs and no region of transition exists where none of the two processes are active. This assumption can be justified if only the tidal timescale is considered (the case in this paper), but has to be relaxed if longer timescales are important.

[25] For the noncohesive sediment (sand) the settling velocity is determined by Stokes formula:

$$ w_s = \frac{g}{18\mu} (\rho_s - \rho_w) d_{50} $$

(19)

where $\rho_s$ is the density of sediment, $d_{50}$ the median grain size and $\mu$ the coefficient of molecular viscosity (we recall here that for a larger grain size the settling velocity is slightly overestimated by Stokes formula).

[26] The exchange of sand between the bed and the water column is parameterized on the basis of the following considerations. We first assume that under stationary conditions an equilibrium distribution of suspended sediment tends to be established in the water column. Under non-equilibrium situations the exchange rate at the bottom is parameterized as proportional to the difference between the actual concentration and the equilibrium concentration:

$$ E_s - D_s = w_s (c_e^* - c_s) $$

(20)

where $c_e^*$ is the equilibrium concentration of the sand fraction (the index $s$ stands for sand). This parameterization ensures that when the near-bed concentration exceeds the equilibrium value, net deposition occurs and vice versa. Now we have to find an expression for the equilibrium concentration. One simple solution is given by the Engelund and Hansen [1967] transport formula, from which we find for the vertically integrated sediment content the following expression:

$$ h c_e^* = \frac{0.05 u_d^3}{\sqrt{g\Delta c d_{50}}} \rho_d $$

(21)

where $c_e^*$ is the depth-averaged equilibrium concentration, $\Delta = (\rho_s - \rho_w)/\rho_w$ is the relative density and $C$ is the Chezy friction coefficient ($50$ m$^{1/2}$s$^{-1}$). However, we need in equation (20) the concentration in the bed layer, but not $c_e^*$. Taking for simplicity for the equilibrium profile the Rouse formula

$$ c_e^*(z) = c_e^*(\frac{z}{z_r})^{\frac{w_s}{z_r}} $$

(22)

assuming that the reference level $z_r \ll h$ and that $w_s$ is large enough compared to $ku^*$ gives

$$ c_e^*(z) = c_e^*(\frac{z}{z_r})^{\frac{w_s}{z_r}} $$

(23)

where $\beta = \frac{w_s}{z_r (\frac{u^*}{\mu} - 1)}$; that is, the reference concentration (the concentration of the first model level) is proportional to the depth-averaged equilibrium concentration given by equation (21). Obviously, the parameter $\beta$ is subject to calibration in the model ensuring simulated values, which are comparable with the observations.

[27] The physical consistence of equation (20) becomes clear if we vertically integrate equation (14). Then the right-hand side of the equation for the vertically averaged concentration $c_s$

$$ \frac{\partial c_s}{\partial t} + ... = w_s (c_e^* - c_s) $$

(24)

describes a relaxation process with a time parameter $T = \frac{1}{w_s}$ [Gerritsen and Berentsen, 1998].

[28] The alternative to this simple model would be to derive the bed load transport and the local equilibrium concentration $c_e^*$ from algebraic sediment transport formulas. One candidate would be the transport formula of van Rijn [1984]. Because in this paper we focus on basic processes resulting from different sediment properties (and the transports associated with them), rather than on parameterization problems, we decided to use the simpler approach based on total load transport and to apply the Engelund-Hansen formula for the equilibrium concentration. Additional motivation for this choice is found in the observations analyzed by Dyer and Soulsby [1988] demonstrating that during most of the spring tide cycle bed load transport is much smaller than the suspended sediment transport (in the simulations described in this paper we focus on the sediment dynamics forced by spring tide). Furthermore, suspended sediment transport is much more amenable to conventional fluid mechanics than is bed load transport [Dyer and Soulsby, 1988].

[29] The sediment source at the bottom is taken (as a first-order approximation) to be inexhaustible for sand and fine SPM everywhere. Effects of erosion or deposition on the bathymetry are not considered in our simulations. If the water depth equals $D_{min}$ the location is considered dry and the sediment equations are not solved anymore until the water depth rises again. The lack of special treatment of suspended matter on the mud flats is one substantial simplification in the present model, which has to be avoided
in future studies, in particular the ones focusing on mud flats. The present study addresses larger scales and fundamental physical balances, which, we hope, are not much affected by the simple treatment of sedimentation in cases when the water column is thinner than \( D_{\text{min}} \).

2.3. Model Area, Discretization, and Forcing

[30] The first application of this model to the area of our study is described by SWBBF and we refer to this paper for more details. The model uses terrain-following equidistant vertical coordinates \((\sigma-\text{coordinates})\). The vertical column extending from the bottom \(-H(x, y)\) to the surface \(\zeta(x, y, t)\) is divided into ten nonintersecting layers. With a horizontal resolution of 200 m we resolve the area in Figure 1 with 324 and 88 grid points in the zonal and meridional direction, respectively.

[31] The forcing at the open boundaries is taken from the simulations with the operational model of the German Bight provided by the German Federal Maritime and Hydrographic Agency (Bundesamt für Seeschifffahrt und Hydrographie (BSH)) [Dick et al., 2001]. The sea level data (one value every 15 min) have been interpolated in time and space onto the grid points along the boundaries of our regional model. The output of the BSH model incorporates the main elements of the regional circulation, which is the coastal wave associated with the well known amphidromy at \(\sim(55.5^\circ\text{N}, 5.5^\circ\text{E})\). The tidal signal crosses the model area from west to east in \(\sim50\) min. The vertical motion of the sea level at the open boundary and its slope provide the major driving force for the model.

[32] Lateral boundary conditions for concentration are taken to be zero for inflowing water. This agrees with satellite and direct observations [Gemein et al., 2006] demonstrating that concentrations along the northern model boundary are small (at least 5 times smaller than in the tidal basins). While for short-time integrations, like in the present paper, this approach seems acceptable, more care has to be given to the open boundary condition in long-term simulations. Outflowing water results in a sediment flux of \(u_{\text{in}}c\) out of the model area, where \(u_{\text{in}}\) is the velocity normal to the boundary.

[33] At slack water the sand settles fast enough to be completely deposited whereas the fine SPM is accumulated in the water column and an equilibrium is not reached before 4 tidal periods. Thus the SPM field after 4 tidal cycles of a run started with \(c = 0\) was taken as the initial field for all other runs. This method reduces the time of adjustment.

[34] Here we will discuss only the results of simulations for the period 16–18 October 2000, which are representative for the general conditions during spring tide. Larger timescales (neap-spring, seasonal, interannual) are not addressed here not only because of short integration times, but also because of missing processes acting at these timescales (air-sea exchange, wave action, storminess, temperature dependence of the settling velocity, consolidation, etc.).

3. Circulation

[35] Most of the present study is based on the same simulations as reported in the study of SWBBF and Stanev et al. [2003b] (hereinafter referred to as SFW) where an extensive validation has been carried out using continuous measurements (10-min records) of sea level and currents (ADCP transects across the tidal channels). The general agreement between observations and simulations demonstrates that the model captures the basic dynamics of the model area. In a more recent model validation in the Spiekeroog Island [Stanev et al., 2007] the emphasis has been put on analysis of ADCP transects in the Otzumer tidal basin (see Figure 1) using EOFs. The results demonstrate that although the horizontal resolution in the straits could be further increased, the agreement between observations and simulations is already very good.

[36] The simulated wet area (Figure 2) varies with time demonstrating that the model resolves one of the most characteristic features of the Wadden Sea, which is the process of flooding and drying. The tidal prisms for the individual basins vary from \(30 \times 10^6\) m\(^3\) to \(170 \times 10^6\) m\(^3\). The water exchange during one tidal period between open ocean and back barrier basin of the Spiekeroog Island amounts to \(~140 \times 10^6\) m\(^3\) during spring tide. The volume of water residing in the tidal basin during low water (\(39 \times 10^6\) m\(^3\)) is much smaller than the tidal prism, revealing that the dynamics are strongly controlled by hypsometry.

[37] The circulation (Figure 2) is dominated by westward transport during ebb and eastward transport during flood. The simulated magnitudes compare well with the observations of Santamaria Cuneo and Flemming [2000]. It has been demonstrated by SWBBF that, while the transport through the straits is mainly controlled by the up-and-down tidal oscillations, the alongshore circulation, as well as the circulation in the intertidal areas, are controlled by the spatial properties (back-and-forth propagation) of the forcing signal. The vertically integrated transport on the tidal flats is very small because of the small depths.

[38] The TKE (Figure 3) is equally important for the sediment dynamics as is the circulation. The absolute maxima are observed in the tidal inlets and their funnel-like extensions in the region of the tidal deltas (north of the inlets). The minima are in the back-barrier area (Figures 3a and 3b) coinciding approximately with the watersheds separating the individual basins (cf. Figure 1). The patches with small TKE values, along the northern coasts of the islands and in the tidal basins are favorable for the accumulation of sediment, although wave action in the real basins would in some cases oppose this effect [Stanev et al., 2006]. Another large area of low level of TKE is located along the southeastern shores of the barrier islands, and is (in most cases) connected with the area of minimum TKE on the watersheds. These complex temporal and spatial patterns contribute to the formation of complex friction properties in the model area which have a number of practical consequences, in particular the physical control of sediment transport.

[39] The evolution of the tidal signals over time is illustrated below with the help of diagrams in Figure 4 (for the locations, see Figure 1). The simulated transport and turbulence characteristics in Figure 4 reveal two velocity maxima occurring at every tidal period. The first maximum corresponds to the flood and the second one to the ebb. During most of the time the entire water column shows relatively strong vertical gradients in velocity and therefore a high level of turbulence. Only during slack water (dua-
tion of \(\sim 1\) hour) the vertical gradients of velocity vanish and the level of turbulence diminishes. The erosion capabilities can be roughly estimated by taking the critical bed shear stress as \(0.4\, \text{N m}^{-2}\) which is sufficiently high to trigger the erosion of sediment from the bottom [Santamarina Cuneo and Flemming, 2000]. This value corresponds to a friction velocity of \(\sim 2\, \text{cm s}^{-1}\) [40]. The maximum ebb velocity is observed in the tidal channels shortly before the rate of sea level fall reaches its maximum. However, the maximum flood velocity is delayed in the tidal channels by \(\sim 2\) hours with respect to the maximum rate of sea level rise, which is explained by SFW as due to the hypsometric control of basins with time-variable horizontal area. This asymmetry is also known as “shorter-falling” asymmetry (in the terminology of...
Friedrichs and Madsen [1992]). Because asymmetries in the tidal response [Dronkers, 1986; Friedrichs and Aubrey, 1988; van de Kreeke and Robaczewska, 1993] are indicative for the transports during flood and ebb, shorter-falling and shorter-rising asymmetries are used sometimes as synonymous of ebb and flood dominated regime, correspondingly.

The usual hydrodynamic definition is that a flood-dominated tidal regime is one in which the peak current on the flood is greater than the peak current on the ebb. In this case coarse sediment moves landward (in the direction of peak velocities [see, e.g., Allen, 1985, section 13.8, p. 255]), and vice versa in the ebb-dominated tidal regime. Some sedimentological and morphodynamic studies, on the other hand, refer to a flood-dominated system as to one in which there is a net floodward transport of sediment. However, this can also be due to temporal asymmetries (e.g., flood currents need shorter time to reach maximum speeds than ebb currents [Groen, 1967]). Thus shorter-falling (rising)
and peak-current asymmetries should be distinguished carefully because they affect different portions of the sediment load differently, enhancing sorting capabilities of sedimentary systems (it is possible for a tidal regime to have larger peak velocities on the flood but a shorter high water slack: this would tend to move coarse sediment floodward but finer sediment ebbward). In the following we will rather consider effects of asymmetry in the duration of slack water, ('slack water asymmetry') because they are the dominating ones in our area of interest (SFW). This asymmetry measures the duration of the low-velocity period around slack water during which settling dominates.

With increasing distance from the inlet the temporal asymmetry of transport changes. In the tidal inlets (location 2) the interval between times of appearance of velocity maxima are shorter around the time of high water (the duration of the low-velocity period around high water is shorter). This interval substantially increases during low water (shorter-falling asymmetry or ebb domination). In the other two locations (1 and 3) the corresponding maxima come closer to each other at the time of low water, and go apart from each other during high water (shorter-rising asymmetry or flood domination). This reveals that the response of transport to tidal forcing is not only temporally asymmetric, but also that the asymmetry has a pronounced spatial dependence. This presents another potentially important mechanism controlling sediment transport.

Figure 4. (a, c, e) Evolution of the vertical profiles of zonal currents (in m s$^{-1}$) and (b, d, f) friction velocity (in cm s$^{-1}$) in the entrance to the Spiekeroog basin. From top to bottom, the plots correspond to locations 1, 2, and 3 in Figure 1. The stepwise shape of the sea level is a mere result of the mapping from $\sigma$ coordinate system to $z$ coordinate system.
The temporal evolution of the level of turbulence (Figures 4b and 4d) follows the velocity (the time of appearance of maximum velocity almost coincides with the time of appearance of maximum friction velocity). The important difference between the two patterns (velocity and friction velocity) is that maximum velocities are at the sea surface (the model is tidally driven), whereas the maximum friction velocity is at the bottom where the turbulence is generated.

4. Sediment Dynamics

4.1. Numerical Experiments

Appendix A gives an idea about controls exerted by turbulence and currents on the sediment dynamics. Using a simple analytical approach we show that currents explain the “displacement mechanism” and result in a one-modal response of a sedimentary system to tidal forcing. The turbulence is responsible for the higher frequency response. In quasi-linear systems the individual contributions of each of the two mechanisms are superimposed, and depending on ϵ and λ (see Appendix A) the sedimentary system is dominated either by effects of turbulence or transport. However, it is not only the change of the level of turbulence and advection, but also the correlation between the two which tends to change the concentration of SPM. This idea illustrates one possible nonlinear coupling mechanism between advection and turbulence. This possible process cannot easily be addressed analytically; therefore we compare below results of three numerical experiments aimed at revealing the role of advection and turbulence in a more realistic sedimentary system described by a 3-D numerical model (see Table 1). The difference between the three experiments is in the different form of equation (14). The first experiment (control run, CR) uses the full equation (14). In the second experiment, which is called “no advection” (NA), we take (u,v) = 0 in equation (14). Thus we assume that the sediment does not move horizontally, but only up and down, which is caused by the vertical diffusion and gravitational sinking. The last two terms in equation (14) are accounted for in experiment NA providing thus the fluxes at the bottom.

In the third experiment, which is called “no shear” (NS) we replace the velocity in equation (14) with the vertically averaged velocity from the 3-D simulations. The comparison between results from this experiment and CR (both with identical hydrodynamics) aims at demonstrating the contribution of the vertical velocity gradient in equation (14) (without this gradient there would be no shear diffusion). We remind that effects of 3-D tidal dispersion are considered earlier by Dronkers [1982], Zimmerman [1986], and Geyer and Signell [1992]. It is noteworthy that in NS we do not completely avoid vertical shear in the velocity. Without this shear there would be no generation of turbulence in the boundary layers. Thus we have to keep in mind that the vertical diffusion, sedimentation rates and bottom erosion in experiment NS are dependent on the velocity shear (turbulent kinetic energy), the latter being computed in the dynamical model and used in the sediment model. Another way to express the basic assumptions in experiment NS would be to assume that we use a shallow water model output (2-D) and externally prescribed turbulence (in our case simulated in the 3-D hydrodynamic model) to solve the SPM equations in the 3-D space. Similar approaches are known from the literature on sediment dynamics where only the SPM equations are solved in 3-D, but the dynamics are based on 2-D simulations and some simple assumptions are made about vertical profiles of sediment [see Teisson and Latteux, 1986; Brenon and Le Hir, 1999, and references therein].

4.2. Horizontal Patterns

The vertically integrated mass of fine sediment (Figure 5) reveals two distinct areas with high values: in the tidal inlets and north of the back barrier islands. In both areas the depth is relatively large (see Figure 1), which could partially explain the large mass of suspended fine sediment fractions there. However, the temporal variability in the area north of the islands, which is different from what we observe in the tidal channels, suggests another (complementary) mechanism. In this area the amount of fine SPM reaches its maximum at low water (compare Figures 5c and 2c). This maximum could indicate that the delivery of sediment by the ebb current is constantly increased. This effect is similar to the well known “displacement of gradients of SPM” established in other studies [e.g., Weeks et al., 1993]. The results of the experiment with no advection (NA) do not reveal such a maximum.

During flood the “sediment clouds” north of the back-barrier islands are displaced slightly southward. However, instead of “entering” the back barrier basin of its origin through the tidal inlet, part of the sediment moves eastward. The “sediment clouds” reach their easternmost displacement during high water (Figures 5a and 2a), which is in the middle of the northern coast of the islands. This area could be assumed to act as a depository of sediment during high water. The east-west displacement of “sediment clouds” allows fine SPM to propagate laterally in front of the barrier islands. SPM can thus leave one tidal basin and enter a neighboring (eastward) one via the open ocean. This tendency is revealed by the correlation between velocity (Figures 2 and 4) and concentration of SPM (Figure 5) in front of the tidal inlets. The eastward direction is mostly due to the fact that SPM concentrations are much higher shortly after low water because of the SPM rich water originating from the tidal inlets (see further in text), and at the same time velocities are eastward. These balances are region-dependent and it can be expected that variability in the tidal cycle (e.g., neap-spring) and meteorological forcing shape the direct and indirect (via the open ocean) exchange between individual basins.

These results indicate that the strongest signal in the sediment dynamics is associated with the oscillation of sediment pools in east-west direction. In this signal the dominating period is the tidal one, not the semitidal one, the latter being associated with vigorous erosion/deposition. By

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Description</th>
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<tbody>
<tr>
<td>CR (control run)</td>
<td>Full dynamics coupled with sediment dynamics</td>
</tr>
<tr>
<td>NA (no-advection)</td>
<td>Velocity set to zero in equation (14)</td>
</tr>
<tr>
<td>NS (no velocity shear)</td>
<td>Vertically averaged velocity used in equation (14)</td>
</tr>
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circulating around the islands fine SPM undergoes complicated transformations resulting from erosion and deposition in quite different areas during different parts of the tidal cycle. It is not very plausible that these transformations could be accurately described supposing that sediment particles move along rectilinear trajectories, as in some theoretical models [e.g., Pritchard and Hogg, 2003].

The concentration of fine SPM in experiment NA reveals the role of turbulence for the sediment dynamics. In this experiment the patterns of sediment in some areas resemble the ones of TKE, the latter showing their largest magnitudes in the tidal channels. However, it is also clear from the numerical simulations that there are important differences between TKE and SPM concentration. This result demonstrates the problem in deducing the properties of sedimentation using only the TKE as input from dynamics. The results of experiment NA demonstrate also that without advection the vertically integrated sediment concentration is higher in tidal channels. The negative anomalies (Figure 6) north of the tidal inlets give a proof that without advection the dynamics of sediment pools is poorly represented.

Figure 5. Vertically integrated mass of fine SPM in the control run (CR).
The spatial patterns in experiment NS are similar to the ones in CR (Figure 5). Some differences (Figure 7) are observed north of the barrier islands where the sediment pools are much more diffuse in experiment NS. The most pronounced differences are observed in the tidal channels where the absence of velocity shear enables larger accumulation of sediment in the water column (in particular during high water). This result indicates that the velocity shear is important not only for establishing the level of turbulence, the latter controlling the erosion and deposition, but also the redistribution of fine SPM in the vertical because SPM is subject to different advection at different depths (and thus shear diffusion).

The NA experiment overpredicts or underpredicts the vertically integrated SPM in CR by as much as 4 kg/m² (the maximum in CR is \( \sim 3 \) kg/m²). This error is large, but is not much greater than the expected error in any predictive model of sediment transport [Eidsvik, 2004]. The NS experiment, on the other hand, overpredicts or underpredicts by no more than 1.2 kg/m².

The contribution of different dynamics to the formation of the transport of sediment becomes clear if we analyze \( D - E \), that is, the bottom flux. This flux indicates

**Figure 6.** Difference between vertically integrated mass of fine SPM in experiments NA and CR.
potential accumulation or erosion trends (Figures 8 and 9). In the absence of vertical shear of velocity in NS, fine SPM tends to propagate from the open sea into the tidal flats. On the contrary, sand in NS does not show a pronounced difference to CR, in particular over larger scales. Instead, there is a stronger trend of sand to be eroded from the tidal channels and to be deposited just in front of them. Furthermore, the difference between $D - E$ in NA and CR shows similar patterns both for sand and fine sediment (erosion in channels and other deep regions and deposition on the tidal flats). The major difference in the deposition-erosion patterns of sand and fine SPM is on the shallows, just in front of tidal channels where the erosion of sand is very large. From the above results one could expect that the morphodynamic response to tidal forcing is very sensitive to the specific combination between dynamics and properties of sediment. It is thus not plausible that the full variety of possible trends in the real systems could be predicted with simple models.

4.3. Temporal Variability and Sediment Sorting

[53] Here we will first compare numerical simulations with observations carried out by Jördel et al. [2004] in the Otzumer Balje (Figure 1). Currents were recorded over the
entire water column using an acoustic Doppler current profiler (ADCP). A laser in situ scattering and transmissometry (LISST) system was used to determine particle size distributions of suspended sediments in the water column. In situ measurements obtained by means of a flow volume–controlled pump centrifuge were used to calibrate the signal from the acoustic and optical instruments. As demonstrated in Figure 10 the phase of signals from observations and numerical simulations agree quite well. The amplitudes are also similar. This agreement between observations and numerical simulations can be regarded as an encouraging support of the model performance. Therefore we expect that the results presented in this paper are relevant to the real sedimentary system. Further support of this relevance is presented by Gemein et al. [2006] where satellite data with high spatial resolution are compared against simulation with the model presented here.

The patterns of temporal variability are very different for sand and fine SPM (Figure 11) and this difference shows a pronounced spatial dependence. The signal associated with sand penetrates to very small heights above the bottom (~3 m). In order to facilitate the comparison between temporal variability of simulated sand and fine SPM concentrations we show in Figure 11 the normalized concentrations in the bottom layer. The mean values are also given in the plots.

The response of sand to the tidal flow occurs during very short times and shows a clear coherence with the signal in the sea level (the phase difference between the two is almost constant). However, the magnitude of the response (sand) is quite variable in space. North of the barrier islands the ratio between magnitudes of flood and ebb maxima varies throughout the simulations. One possible explanation of this feature is the very low concentration of sand during all times making the response noisy.

[56] The clearest two-modal response of sand is simulated in the tidal channels. In these areas the dynamics depend on the turbulent kinetic energy, which shows a two-modal behavior (Figure 4). Because sand is observed in suspension only in the near-bottom levels, and because in these levels the currents are relatively weaker, its dynamics are more local than the ones of fine SPM.

[57] The simulated sediment concentration in the tidal flats is characterized with one-modal variability pattern. The domination of the flood signal (higher maximum during flood) in these areas is even better seen in the transport curves (Figure 11f), which supports the idea of Staney et al. [2007] about changing tidal response, i.e., from ebb dominated response in tidal channels to flood dominated one in the tidal flats.

The concentration of sand during the ebb phase is very low (Figure 11e), particularly in the tidal basins. This explains why the transport of sand (product of concentration and velocity) in the bottom level is directed toward the coast. However, the transport of fine SPM is much better balanced (the integrals of curves in Figure 11f almost equal). Obviously, the difference between transports of sand and fine SPM which we observe in the model, from one side and the difference between them and water transport from another, demonstrates the sensitivity of sediment transport patterns on properties of transported matter. These individual properties can be roughly described by the settling velocities \( w_s \), which are different for fine SPM and sand. The corresponding response times measured by \( H/w_s \) also

---

**Figure 8.** Difference between deposition minus erosion in experiments NA and CR accumulated for one tidal period: (top) fine SPM and (bottom) sand.
differ substantially. Because sand spends a much shorter
time in suspension compared to SPM its transport patterns
differ more from the ones of water (formally the response
time of water is infinite) than the transport pattern of SPM.
Furthermore, while the transports of fine SPM and water
almost coincide in the deep tidal channels, large differences
are observed elsewhere.

These results are indicative for different sorting
processes which are associated with the different transports
of different types of sediment. The spatial dependence of
sediment sorting in the area of our interest is further
supported by the different sensitivity of patterns of
\( D - E \) (Figures 8 and 9) to a variety of dynamics.

One further example of different responses of sand
and SPM is given below. There is a pronounced phase lag
between concentrations of sand and fine SPM in the tidal
channels and north of the barrier islands (Figures 11a and
11c), which reveals the different exposure of the two
materials to advection. Differently from the sand, which is
less sensitive to advection showing thus two modal curves,
the fine SPM exhibits a clear one-modal variation north of
the barrier islands. The phase shift between times of maxi-
mum concentrations of sand and fine SPM is \( \sim \pi/2 \) north of
the barrier islands, which indicates that an increase of
concentration of fine SPM is due to the upstream current
carrying eroded material from the tidal channels.

### 4.4. Evolution of Sediment Stratification During the
Tidal Cycle

The above discussion about the different response of
sand and fine SPM does not reveal the full variety of
processes, most of which are characterized by strong
gradients in the vertical. To more clearly describe the
response of fine SPM concentration to the variability of
the tidal flow we use the diagrams in Figure 12 displaying
the evolution of the vertical profiles for the three experi-
ments. In the experiment CR the flood signal in the tidal
channels is narrower in time, in particular in the deeper
layers, and the concentration of fine SPM reaches slightly
lower levels than during ebb (Figure 12d). Most of the
sedimentation occurs during a short period after high water.

The fine SPM concentrations in the surface layers of
tidal channels are about two times smaller during slack
water compared to the ones during flood and ebb. The SPM
minimum during low water is less pronounced than during
high water and is not observed down to the bottom. This
difference is presumably due to the fact that during flood the
waters in the channels are replaced by relatively sediment-
poor open ocean water.

One important result from Figure 12 is that the
number of maxima and minima per tidal period differs, as
well as their time of occurrence. The two-modal (two
maxima and minima per tidal period) behavior of fine
SPM concentration occurs only in the deep channels. In
contrast to this, the time dependence in the open North Sea
and in the shallow extensions of channels (toward the tidal
flats) is one modal. The maximum of fine SPM concen-
tration in the North Sea is reached during low water, the
minimum of \( \sim 0 \) mg/l is reached during high water. Near the
tidal flats almost a phase opposition to the case in the open
sea dominates the entire water column: the maximum
concentration is reached shortly before high water. The
explanation of this phenomenon has been given above
and assumes that the source of the sediment in suspension
is in the deep channels. Then, the high concentrations are

![Figure 9](image_url)

**Figure 9.** Difference between deposition minus erosion in experiments NS and CR accumulated for one
tidal period: (top) fine SPM and (bottom) sand.
advection toward the north and south in phase opposition (once per tidal period). This theory is in agreement with the "displacement mechanism" and is supported by the fact that the two-modal sediment concentration recovers north of the back barrier islands in experiment NA. In this experiment the advection of sediment is switched off and the control is taken by turbulence which is two-modal (see Figures 4b, 4d, and 4f).

Although one could expect that without advection (experiment NA) the variability of stratification of fine SPM would tend to follow the one of TKE (Figure 4), this is not always the case; that is, the sensitivity of fine SPM to the level of turbulence is not so simple. One feature which is robust in the three locations and correlates well with the TKE variability is the minimum of fine SPM at the sea surface occurring just after the low and high water (Figures 12b, 12e, and 12h). The two minima are most pronounced in the deep tidal channels, where the signature of the one occurring during low water is observed until the time of high water (Figure 12e). Unlike in experiment CR, the fine SPM maxima in experiment NA are much "wider," in particular the flood maximum.

The differences between the results of experiments CR and NA demonstrate that the advection is an important mechanism controlling the fine SPM, in particular north of the barrier islands. Without advection, the concentrations in this area are much lower (notice the different contour intervals). However, the comparison between results in CR and NA in the tidal basins, speaks against the "displacement mechanism." No decrease of the concentration is observed during the entire tidal cycle (the contour intervals are the same in Figures 12g and 12h), as it is in the case north of the barrier islands. On the contrary, the fine SPM concentrations in experiment NA are larger than in experiment CR. This indicates that the source of sediment in this area is local and results from the erosion of tidal flats. Without advection in experiment NA, this sediment does not leave the shallow area, thus the concentrations remain high during the entire tidal period, in particular in the deep layers.

In both experiments NA and CR the main maximum in fine SPM on tidal flats occurs shortly before high water. However, unlike in experiment CR where the secondary maximum of fine SPM occurring during the end of ebb phase is very small, the one in experiment NA is well pronounced. This again demonstrates the role of advection removing SPM from the tidal basins during ebb.

Without vertical gradients of velocity (experiment NS) the sedimentary system develops lower concentrations of sand (not shown here) and (in most locations) higher concentrations of fine SPM (Figure 12) than in CR. The sensitivity experiment NS reveals thus the role of physical controls on the sorting capabilities of the sedimentary system. This is important for further studies addressing morphodynamics and illustrates that simplifications in the models could lead to large artifacts. In this context we remind the reader that in a number of studies the sediment dynamics are simulated using vertically averaged velocity in the equation for sediment evolution. As seen from the comparison between experiments CR and NS the principal difference between the two is that the temporal variability in NS is described by one maximum and one minimum per tidal period. There is only in the upper layers of the tidal channels a small (secondary) decrease of fine SPM concentration shortly after high water in NS.

Overall, in experiment NS velocities in the surface layer are smaller than in experiment CR (and vice versa in the deep layers). As a result of that, north of the barrier islands, where the "displacement mechanism" is most relevant, the fine SPM maximum at low water reveals lower values of fine SPM concentration in NS than in CR. This is explained by the fact that a large part of the transport of fine SPM occurs in the upper layers where velocities are higher, as well as SPM concentrations because light particles are maintained at higher levels in the suspension.

The second important consequence of the missing velocity shear in experiment NS is the change in the temporal variability. The sediment minimum during high water is not as well pronounced in experiment NS as in CR. However, the minimum during low water is better pronounced in experiment NS. These changes are accompanied by much larger amplitudes of fine SPM concentrations in experiment NS (the amplitudes increase also in the tidal basins). Comparing the diagrams (Figures 12c, 12f, and 12i) displaying the evolution of the vertical profiles simulated in the three locations one finds a much simpler (than in experiment CR) response to the tidal flows because the second mode in the response is very small compared to the first one. Furthermore, the phases in locations 1 and 3 are just lagged against the phase in location 2 (locations 1 and 3 exhibit almost opposing phases). This gives a clear indication that in experiment NS the sediment dynamics are largely governed by the "displacement mechanism"; that is, the system is more linear when advection is described by vertically averaged fields.

5. Conclusions

This paper extends the simple conceptual model of Weeks et al. [1993] suggesting that tidal excursions explain the observed semidiurnal concentrations of fine SPM, while resuspension explains the quarter-diurnal variations. This
concept has been checked by analyzing results from numerical simulations focusing on the control of transport and turbulence on the sediment transport. It has been demonstrated with the help of a simple analytical model that a large variety of combinations between advection and turbulence could exist, which explains different appearances of SPM patterns in the ocean. One of them, which is known as “twin peaks” appears to be a solution of equations presented in Appendix A and corresponds to the intermediate case between the two extremes: one-modal oscillations (dominance of advection) and two-modal oscillations (dominance of turbulence).

The simulations with a 3-D hydrodynamic model coupled with a sediment transport model help understand the dominant patterns. It has been shown that “sediment clouds” north of the barrier islands present one of the most important elements of the sediment dynamics in the region. Their oscillation in the east-west direction is dominated by one-modal oscillations. By circulating around the islands fine SPM undergoes complicated transformations resulting from erosion and deposition in quite different areas during different parts of the tidal cycle.

The 3-D simulations identify the major physical controls. Without advection (NA) the vertically integrated sediment concentration is higher in tidal channels because the level of turbulence there is high. Without vertical gradients of horizontal velocities (NS), fine SPM concentration in the tidal channels is larger during ebb and smaller during flood than in CR. This result proves that velocity shear is important not only for establishing the level of turbulence, the latter controlling the erosion and deposition. It also controls the redistribution of fine SPM in the vertical direction, which is further propagated by the advection over large areas. Without velocity shear (experiment NS) the sedimentary system develops lower concentrations of sand and higher concentrations of fine SPM than in CR, illus-

![Figure 11](image_url)

Figure 11. Normalized to absolute maximum concentration of fine SPM and sand 1 m above the bottom: (a) north of the barrier islands, (c) in the tidal channels, and (e) in the tidal basin of Spiekeroog (the extension of the tidal channel). The locations are given in Figure 1. The oscillations of sea level (dotted line) are plotted in Figures 11a, 11c, and 11e to enable better understanding of the timing of different signals. (b, d, f) Normalized to absolute maximum transports in the 2 m bed layer (positive when directed toward tidal basins) in locations 1, 2, and 3, respectively. The dotted line represents the normalized water transport.
trating clearly that simplifications in the models (e.g., using 2-D dynamics) could lead to artifacts.

[73] The sorting capabilities of the sedimentary system are identified by the differences in the transports of different types of sediment. Because sand spends much shorter time in suspension and is very sensitive to extreme turbulence its transport patterns are very different from those of SPM and water. On the contrary, the dynamics of fine SPM are strongly affected by the tidal excursions and resembles (in particular in the tidal channels) the water transport. It has also been demonstrated that the differentiation mechanisms show a pronounced spatial dependence.

[74] The simulated phase lags between concentrations of sand and fine SPM reveal the different exposure of the two materials to advection. Contrary to the sand characterized mostly by two-modal variability, the fine SPM exhibits a clear one-modal variability north of the barrier islands and in the tidal basins.

[75] Most of the theories about the “displacement mechanism” assume that the gradient of fine SPM is due to the coastal source. We showed, however, that the vigorous dynamics in the tidal channels could also build gradients. While the behavior of fine SPM in the area north of the back barrier islands obeys the principles of “displacement,” the tidal basins do not support this “rule,” but show a very complicated temporal evolution. This indicates that the source of sediment in this area is local (associated with the large erosion at shallow depths).

[76] From the discussion above one result becomes evident: the temporal asymmetry of fine SPM concentration (and the occurrence of double peaks per tidal period) in the realistic systems is not fully described by a superposition between effects of turbulence and advection, as this is the case in the analytical and conceptual models. The velocity shear in the vertical direction is crucial for the adjustments between temporal and spatial variability.

Appendix A

[77] The aim of this appendix is to present an analytical model, which gives an explanation of the role of transport and turbulence in the sediment dynamics. We consider a simplified two-dimensional analogue of equation (14):

\[
\frac{\partial c}{\partial t} + u \frac{\partial c}{\partial x} = \frac{\partial}{\partial z} \left( A_V \frac{\partial c}{\partial z} + \frac{w_s}{C_1} (w_s - w)c \right)
\]

[78] Using equations (16)–(18) we can present the bottom flux in the right-hand side of equation (A1) as

\[
\left( A_V \frac{\partial c}{\partial z} + w_s c \right)_{z_{min}} = E - D \\
= aM_c \left( \frac{\tau_b}{\tau_c} - 1 \right) - w_s c_b \left( 1 - \frac{\tau_s}{\tau_d} \right)
\]
for the notations here and below, see section 2. When writing this equation we take into consideration that at the bottom \( w = 0 \). We integrate equation (A1) over the whole water depth \( H + \zeta \) and assume that \( \tau_d = \tau_e = \tau_{cr} \). The corresponding approximate equation for the vertically averaged SPM concentration then reads:

\[
\frac{\partial c}{\partial t} + u \frac{\partial c}{\partial x} = \frac{M_c}{H} f_c^e(\tau_b) - \frac{w_{cb}}{H} f_f^e(\tau_b) \tag{A3}
\]

where

\[
f_c^e(\tau_b) = \begin{cases} \frac{\tau_b}{\tau_{cr}} - 1 & \text{when } \tau_b > \tau_{cr} \\ 0 & \text{otherwise} \end{cases}
\]

and

\[
f_f^e(\tau_b) = \begin{cases} 1 - \frac{\tau_b}{\tau_{cr}} & \text{when } \tau_b < \tau_{cr} \\ 0 & \text{otherwise} \end{cases}
\]

(A4)

We make the following physical assumptions: \( c_b = c \), and \( \frac{d c}{d t} = \frac{d}{d t} \), the latter suggesting that the background concentration field varies exponentially in \( x \) with a characteristic length scale \( L \). We further nondimensionalize equation (A3), defining dimensionless variables (denoted by an asterisk) as \( \ast = \omega t \) and \( c^* = c w_c / M_c \), and we write the tidal velocity as \( u = U_0 u^*(\ast) \). We now have

\[
\frac{\partial c^*}{\partial \ast} + \lambda u^* c^* = \varepsilon \left[ f_c^e(u^*) - f_f^e(u^*) c^* \right],
\]

where

\[
\lambda = \frac{U_0}{L \omega} \quad \text{and} \quad \varepsilon = \frac{w_c}{H \omega}
\]

(A5)

The parameter \( \lambda \) represents the importance of advection of the background concentration, while \( \varepsilon \) represents the importance of local exchanges between the bed and the suspended sediment. This equation is of exactly the same form as those considered in Appendix A of Pritchard and Hogg [2003], where

\[
a(\ast) = \varepsilon f_c^e(u^* \ast^*),
\]

and

\[
b(\ast) = \lambda u^* + \varepsilon f_f^e(u^* \ast^*).
\]

(A6)

In order for the variation of \( c^* \) to be small compared to its mean value \( c^*_{0} \), we require that \( \varepsilon \ll 1 \) and \( \lambda \ll 1 \). If we further assume that \( \lambda \sim \varepsilon \), so the contributions from advection and erosion are of comparable magnitude, we can expand \( c^*(\ast) \) in a series, \( c^*(\ast) = c^*_{0} + \varepsilon c^*_{1}(\ast) + \ldots \), where the leading order (constant) term should satisfy

\[
c^*_{0} \int_{0}^{2\pi} \left( \lambda u^* + \varepsilon f_f^e(u^* \ast^*) \right) d\ast = \int_{0}^{2\pi} f_c^e(u^* \ast^*) d\ast
\]

(A7)

while the next-order term satisfies

\[
\varepsilon \frac{dc^*}{d\ast} = f_c^e(u^* \ast^*) - c^*_{0} \left[ \lambda u^* + \varepsilon f_f^e(u^* \ast^*) \right]
\]

(A8)

Consider now a sinusoidal tidal signal, \( u^* = \sin \ast \). Because we aim at finding analytical solutions, we will use a simple closure \( \tau_{cr} = \alpha u^* \). In this case the advective term provides a contribution \(-c^*_{0} \cos \ast \) to \( c^* \). The erosive and depositional terms, on the other hand, vary as \( \sin^2 \ast \), in other words with twice the periodicity of \( u \), and will contribute terms proportional to \( \varepsilon \cos 2\ast \) and \( \varepsilon \sin 2\ast \); the phase of this contribution will depend on the values of \( \tau_{cr} \) and of \( c^*_{0} \).

[79] The above conclusions support the results of Weeks et al. [1993]: the advection of sediment contributes terms with the same periodicity as that of the tidal signal, while the quadratic terms in the erosion and deposition functions contribute terms which vary at twice this frequency. This gives a simple theoretical explanation of “twin peaks” known from earlier experimental and conceptual studies of sediment transport. It is clear that the dominance of advection or erosion depends on the magnitude of \( \varepsilon \) and \( \lambda \) and a large variety of combinations could exist shaped by the dynamics and properties of the sediment. Depending upon the ratio \( \zeta \) the temporal evolution of the concentration of SPM could vary between the two extreme cases: one-modal and two-modal oscillations. We would expect higher harmonics to be generated by interactions between the forcing functions \( f_c^e, f_f^e \) and the concentration components \( c^*_{n} \). Obviously, the number of assumptions leading to equation (A8) is large and it is demonstrated in this paper that the numerical simulations show a much more complicated behavior.

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References


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