On the superimposition of bedforms in a tidal channel

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ABSTRACT: High resolution bathymetric measurements reveal the super-imposition of bedforms in the Grådyb tidal inlet in the Danish Wadden Sea. Preliminary results of numerical model simulations are discussed: A linear stability model was tested to explain the large bedforms as being caused by tidal system instabilities. Results show comparable dimensions and migration rates. A three-dimensional morphodynamic model is shown to reproduce small scale transport rates but lacks realistic trends of morphodynamic evolution.

1 INTRODUCTION

Bed forms as a product of the interaction of a deformable bottom and forcing currents have been a phenomenon fascinating to the casual observer as well as to the scientific community for more than a century (e.g Darwin, 1883). Despite the simplicity in form, their ubiquitous nature and a multitude of applied methods and approaches, a satisfactory formulation of the underlying physical principles probably still needs several more years of dedicated research (cf. Kennedy, 1963). However, in the recent past new sensors and positioning technology in the field and laboratories on the one hand as well as enhanced mathematical schemes on the other hand have led to new opportunities in exploring the natural system and modelling its behaviour. This contribution presents preliminary findings from two different model approaches and compares these with high-resolution morphology of compound bedforms in a tidal channel. As mistakable nomenclature might obscure the scientific progress, we avoid the terms “sandwave” and “dune” - which both have been understood purely descriptive by many (cf. Ashley, 1990) but also process related by others - in favour of the innocent term “bedform” for all morphological features observed in the study area of the Grådyb tidal channel in the Danish Wadden Sea (Fig. 2).

For decades echo soundings have shown the superimposition of bedforms of different sizes in tidal environments (Allen & Collinson, 1974; Rubin & Mcculloch, 1980; Ten Brinke et al., 1999), but only recently high-resolution multibeam sonars provide the opportunity to show the complex three-dimensional patterns and to assess temporal variations in morphology along one direction and the same channel reach with high precision. In combination with high-accuracy positioning, horizontal and vertical changes at sub-decimetre scale can be quantified (Ernstsen et al., 2006).

The Grådyb tidal channel displays well developed compound bedforms that have been thoroughly investigated in the last years (Bartholdy et al., 2002; Bartholdy et al., 2005; Bartholomä et al., 2004; Ernstsen et al., 2006; Ernstsen et al., 2005; Winter & Ernstsen, 2007)Ernstsen et al., 2007. Successive inter-tidal as well as repeated annual measurements have formed a sound data-base on the characteristics of this domain. Generally the compound bedforms of the central Grådyb tidal channel can be characterised by large asymmetric ebb-directed features with lengths of about one hundred meters and heights of about three meters, which migrate slowly in the order of ten meters per year towards the open sea. Smaller superimposed bodies with lengths of a few meters and heights in the order of several decimetres are much more dynamic. These very active bedforms are asymmetric and reverse direction within each semi-tidal cycle, the crests moving in the order of two meters. These super-imposed bedforms tend to grow in size from the trough towards the crest of the underlying large feature. Although the smaller bedforms are expected to have an ebb directed net migration, and thus contribute to the dynamics of the large ones in the course of time (Bartholdy et al., 2002), some observed tidal cycles also reveal a flood oriented net migration (Bartholomä et al., 2004; Ernstsen et al., 2006).

Several empirical bedform stability relations have been formulated to describe and predict the interdependency of forcing and morphological reaction for equilibrium conditions (e.g. Allen, 1968; Southard & Boguchwal, 1990; Van Rijn, 1993). However, the occurrence of compound bedforms in the continuously changing natural environments requires a more detailed investigation. Two possible explanations for the superposition of different bedform hierarchies are frequently cited (Dalrymple & Rhodes, 1995) "Disequilibrium superposition" of smaller forms on top of larger ones is related to fluctuations in forcing in such way that large forms are generated by extreme events (e.g. floods) and remain because of their large lag time followed by the formation of smaller features on their back, which are related to the currents (Allen & Collinson, 1974). On the other hand "equilibrium superposition" describes the formation and existence of small bedforms within an internal boundary layer on the back of the large features (Rubin & Mcculloch, 1980; Dalrymple, 1984). Whereas the former seems not to apply for the tidal Grådyb area where both bedform generations are active and lasting and extreme flood events are rare, the latter equilibrium concept of an internal boundary layer may hold. The observed increase of height and length of the superimposed bedforms towards the crest of the underlying structure could point towards a growing internal boundary layer. However, the origin and dynamics of the large features are not explained by this concept.

This contribution compiles results of high resolution bathymetric data and simulations of two different numerical models. It is discussed if the observed dynamic compound dunes in the Grådyb tidal inlet can be genetically separated into large scale features which are explained to develop as instabilities of the tidal system, and into the smaller superimposed features that are in dynamic equilibrium with the quasi-steady tidal currents on the back of the larger ones.

2 METHODS

2.1 *Field Data*

Ship based bathymetric profiling was conducted in the Grådyb tidal inlet at about N55.46/E8.33. For this study, a 500 m section of the channel measured on July 15, 2003 is considered. The bathymetric data were recorded from RV Senckenberg using a Seabat™ 8125 (RESON) multibeam echo sounder (MBES) system operating at 455 kHz. The MBES system was coupled with an AQUARIUS 5002 (THALES) dual frequency (L1/L2) Long Range Kinematic (LRK) Global Positioning System (GPS). The horizontal and vertical precision of the MBES system is ±20 and ±2 cm, respectively, at a 95% confidence level (Ernstsen et al., 2006). Data was interpolated by linear interpolation to a line along the central main channel.

2.2 *Linear Stability Model*

Stability models are based on the assumption that the generation of large scale bedforms (in this context most often termed "sandwaves") is driven by tidal averaged circulation cells produced by the interaction of the oscillatory tidal current with bottom perturbations (Besio et al., 2008; Hulscher, 1996). Depending on the value of the parameters, the net transport of these residual currents can be directed towards the crests of the initial perturbations, causing further local accumulations which lead to the growth of bedforms. Linear stability analyses comprise the detection of stable features of distinct wavelength and evolution timescale.

A linear stability model (Blondeaux & Vittori, 2005a) has been applied to the system characteristics in the Grådyb tidal inlet. The model is describing a reduced system in that the flow generated by a M2 tidal wave propagating over a flat bottom of infinite extent composed of non cohesive sediment is considered and the time development of the bottom profile is investigated.

The hydrodynamic module considers a three-dimensional turbulent current which is modelled by means of a Boussineq-type closure with an eddy viscosity ν_T assumed to depend on the distance from the bottom (see Besio et al., 2006). Hereinafter a star denotes dimensional quantities. Viscous effects are neglected. The flow field is determined by solving continuity and momentum equations which are more conveniently written in non dimensional form. The mean water depth h_0^* is used as length scale, the maximum value U_0^* of the depth averaged fluid velocity during the tidal cycle is used as velocity scale and the inverse of the angular frequency ω^* of the tide is used as time scale. Hence the hydrodynamic problem reads:

$$\nabla \cdot \mathbf{u} = 0 \tag{1}$$

$$\frac{\partial \mathbf{u}}{\partial t} + \hat{r}(\mathbf{u} \cdot \nabla)\mathbf{u} = -\nabla p + \mathbf{g} + \hat{\delta}^2 \nabla \cdot \mathbf{T} \tag{2}$$

In which u = (u, v, w) is the velocity vector and p is the pressure. The operator ∇ is defined by ($\partial/\partial x$, $\partial/\partial y$, $\partial/\partial z$), where x and y are two horizontal axes lying on the free surface and z is the vertical coordinate pointing upward. g is the dimensionless gravity acceleration (g=(0,0,−g)) and T is the turbulent stress tensor defined as:

$$T_{i,j} = \nu_T \left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) \tag{3}$$

At the free surface the dynamic boundary conditions, which force the vanishing of the stresses, as well as the kinematic boundary conditions are imposed. The vanishing of velocity is forced at a distance from the bottom related to bottom roughness.

The dimensionless hydrodynamic problem is characterized by the dimensionless parameters:

$$\hat{r} = \frac{U_0^*}{\omega^* h_0^*}, \quad \hat{\delta} = \frac{\sqrt{\nu_{T0}^*/\omega^*}}{h_0^*}. \tag{4}$$

The parameter $\hat{r}$ is the ratio between the amplitude of horizontal fluid displacement oscillations and the local depth and assumes typical values of $O(10^2)$. The parameter $\hat{\delta}$ is the ratio between the thickness of the turbulent bottom boundary layer and the local depth. A rough estimate shows that $\hat{\delta}$ is of order one. In the definition of $\hat{\delta}$, ν_{T0}^* is a dimensional constant and provides the order of magnitude of the eddy viscosity and is computed as:

$$\nu_{T0}^* = k\frac{U_0^* h_0^*}{C} F(\xi) \tag{5}$$

Where k is the Von Karman constant, $F(\xi)$ an appropriate function which describes the vertical distribution of the eddy viscosity and C is the friction factor, which only depends on the roughness size and is computed by means of the standard formula used for steady currents (Fredsoe & Deigaard, 1992). Since ν_{T0}^* is proportional to U_0^* it is convenient to introduce the new viscous parameter $\hat{\mu} = \hat{r}/\hat{\delta}^2$ unrelated to the strength of the tidal current.

The morphodynamics module of the model is governed by the sediment continuity equation which, in dimensionless form, reads:

$$\frac{\partial h}{\partial T} = \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} \tag{6}$$

In which T now is a slow time scale which will be defined below (12). The dimensionless parameters which characterize the morphodynamic problem are the sediment porosity p_{or}, the dimensionless sediment size d, the mobility number ψ_d and the particle Reynolds number R_p

$$d = \frac{d^*}{h_0^*}; \quad \hat{\psi}_d = \frac{\left(\omega^* h_0^*\right)^2}{(\rho_s/\rho - 1) g^* d^*}; \tag{7}$$

$$R_p = \frac{\sqrt{(\rho_s/\rho - 1) g^* d^{*3}}}{\nu} \tag{8}$$

In which ρ and ρ_s are the water and sediment density respectively.

The sediment flux is computed considering both bed load and suspended load. The dimensionless bedload flux $\left(q_{Bx}, q_{By}\right) = \left(q_{Bx}^*, q_{By}^*\right) / \sqrt{(\rho_s/\rho - 1) g^* \left(d^*\right)^3}$ due to the tidal current is computed by means of the relationship proposed by (Van Rijn, 1991), which relates (q_{Bx} , q_{By}) to the dimensionless bottom shear stress $\left(\vartheta_x, \vartheta_y\right) = \left(\tau_x^*, \tau_y^*\right) / \left(\rho_s^* - \rho^*\right) g^* d^*$.

The dimensional shear stress components $\left(\tau_x^*, \tau_y^*\right)$ can be easily evaluated by means of the constitutive law. Moreover, as suggested by Colombini (2004) and assumed by Cherlet et al. (2007), to compute the sediment transport rate, the shear stress is evaluated at the top of the bed-load layer. If the bed is covered by ripples the thickness of the bed load layer scales with the ripple height.

To complete the evaluation of the bedload sediment transport, the model takes into account the weak effects due to a slow spatial variation of the bottom topography. If the bottom slope ∇h is small, simple dimensional arguments and linearization lead to

$$\left(q_{Px}, q_{Py}\right) = -q_B \mathbf{G} \nabla h \tag{9}$$

where G is a dimensionless second order 2-D tensor (Seminara, 1998).

The suspended sediment transport (q_{Sx}, q_{Sy}) is computed as the flux of sediment concentration in the water column. The sediment concentration $c = c\left(x^*, y^*, z^*, t^*\right)$ is computed by solving a standard convection-diffusion equation.

In the study of the stability of the flat bottom configuration of a shallow tidal sea, small perturbations of the bottom are considered so that the bottom configuration differs from the flat one by a small (strictly infinitesimal) amount proportional to the non-dimensional order of magnitude of the height of the perturbations (ε). Hence, due to the small values attained by non-linear terms, the bottom profile can be thought to be given by the superposition of different spatial components which evolve one independently of the other. A normal mode analysis can be performed and the problem can be solved for the generic spatial component of the perturbation of the bed which is periodic in the x- and y-directions with dimensionless wave numbers $\alpha_x = \alpha_x^* h_0^*$ and $\alpha_y = \alpha_y^* h_0^*$:

$$\eta\left(\alpha_x, \alpha_y, t\right) = \varepsilon A(t) e^{i\left(\alpha_x x + \alpha_y y\right)} + c.c. + O\left(\varepsilon^2\right) \tag{10}$$

Where $\varepsilon A(t)$ is the amplitude. The small value of ε allows the solution to be expanded in terms of ε.

At the leading order of approximation, i.e. $O(\varepsilon^0)$, the problem is reduced to that of determining the flow and sediment transport induced by tide propagation over a flat seabed. Then, the hydrodynamic problem for the flow perturbations is obtained at $O(\varepsilon)$ and is solved following Blondeaux & Vittori (2005b).

The time development of the amplitude of the bottom perturbation, follows from the sediment continuity equation at order ε:

$$\frac{dA(T)}{dT} = \Gamma(t)A(T) \quad (11)$$

In which Γ is a periodic, complex function of t which depends on the parameters of the problem and T is the slow morphodynamic time scale

$$(\, T = td / \left|(1 - p_{or})\sqrt{\hat{\psi}_d}\right|). \quad (12)$$

The solution of (11) shows that the growth or the decay of the bottom perturbations is controlled by the real part $\overline{\Gamma}_r$ of the time average of $\overline{\Gamma}$, thus leading to the wavelengths, while the imaginary part $\overline{\Gamma}_i$ is related to the migration speed of the perturbations.

2.3 *Morphodynamic Model*

Based on the process based modelling system Delft3d-FLOW (Lesser et al., 2004) a model of the Grådyb tidal channel system has been set-up to calculate highly resolved long term hydrodynamics and short-term morphodynamics of the domain. The finite-differences modelling system solves the non-linear shallow water equations in three dimensions on curvilinear grids. The vertical momentum equation is reduced to the hydrostatic pressure relation. Thus, vertical accelerations are assumed to be small compared to the gravitational acceleration and are therefore not taken into account. A full coupling between computational modules allows the simultaneous calculation of currents, sediment transports and bed evolution at every computational time step. The suspended sediment transport is based on the numerical solution of an advection-diffusion equation:

$$\begin{aligned} &\frac{\partial c}{\partial t} + \frac{\partial uc}{\partial x} + \frac{\partial vc}{\partial y} + \frac{\partial (w - w_s)c}{\partial z} + \\ &- \frac{\partial}{\partial t}\left(\varepsilon_{s,x}\frac{\partial c}{\partial x} + \varepsilon_{s,y}\frac{\partial c}{\partial y} + \varepsilon_{s,z}\frac{\partial c}{\partial z}\right) = 0 \end{aligned} \quad (13)$$

In which c is the mass concentration of sediment, u,v,w the flow velocity components, $\varepsilon_{s,x}$, $\varepsilon_{s,y}$, $\varepsilon_{s,z}$, are the eddy diffusivities and w_s the settling velocity of the median grain size, calculated by the Van Rijn (1993) method. Whereas the vertical diffusive flux through the free surface is set to zero, the bed boundary condition is given by:

$$- w_s c - \varepsilon_{s,z}\frac{\partial c}{\partial z} = D - E \quad (14)$$

In which $D=w_s c_a$ is the deposition rate and $E=\varepsilon_s dc/dz$ the erosion rate. The reference concentration c_a at reference level a is calculated directly by the sediment transport formula of Van Rijn (1993) as $c_a= S_s/h\ u$. Here S_s is the derived suspended sediment transport rate and h the water depth. The elevation of the bed is dynamically updated at each computational time step as a result of the sediment sink and source terms. For details on the modelling system and the underlying numerics the interested reader is referred to (Roelvink & Van Banning, 1994; Lesser et al., 2004), (Tonnon et al., 2007) or the Delft3D-FLOW manual (Wl|Delft-Hydraulics, 2007).

According to the morphology of the Grådyb tidal channel in the Danish Wadden Sea between the barrier spit Skallingen and the barrier island Fanø, two models have been set-up: An overall model grid covers the whole tidal flat areas and the main tidal channel with varying resolution of about 20 to 500 m. (Figure 1). In the vertical direction, a terrain-following sigma grid uses ten logarithmically scaled layers. At the open sea boundary water level conditions are described by 94 harmonic astronomical constituents, based on long term records of a tidal gauge on the ebb tidal delta. This model was used to calculate boundary conditions for a highly resolved morphodynamic model and the linear stability model explained in section 2.3. To account for the discretisation of the small superimposed bedforms a second model of higher resolution was used. This model domain covers a 500 m long stretch (1DV) in 0.5 m grid cells in the longitudinal direction, and 20 equally spaced layers in the vertical. Despite the limited coverage of the latter model, computational time step restrictions limit the application of this model to a few tidal cycles. Model reduction techniques to speed up morphological timescales, as simple scaling factors or continuity updating have been avoided in this study to reflect the small scale morphodynamic capabilities of the system.

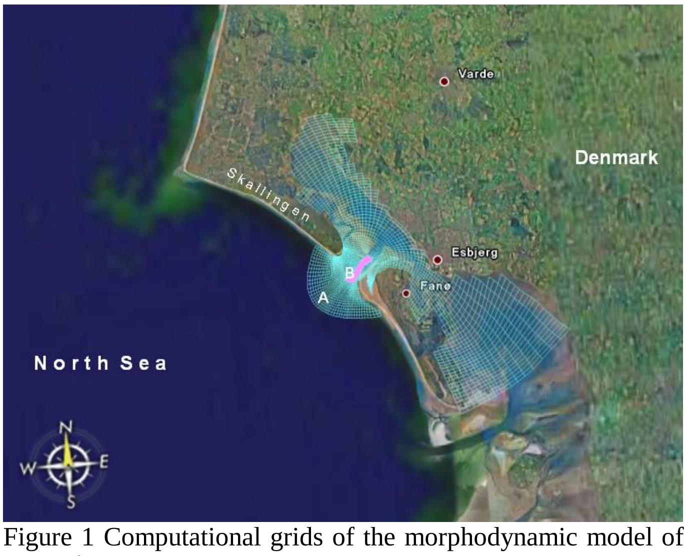

Figure 1 Computational grids of the morphodynamic model of the Grådyb tidal channel. A denotes the coarse overall grid, B the higher resolved detail grid.

3 RESULTS

3.1 *Field data*

Earlier studies on the Grådyb tidal channel have described the annual development of the well developed compound bedforms with lengths of about one hundred meters and heights of about three meters, which migrate slowly in the order of ten meters per year towards the open sea. Here data is shown from a stretch of compound bedforms in the central Grådyb tidal channel which were derived by successive measurements shortly after high water, at low water, and two hours before successive high tide. These profiles exemplarily reveal the morphological change during the ebb tide period (Figure 2, Figure 6a). It is observed that during this time mainly the superimposed bedforms change position, whereas the underlying large bedforms identified by the deeper troughs seem stable.

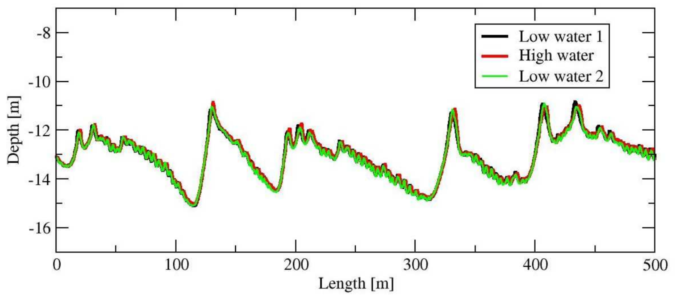

Figure 2: Bathymetric profiles showing the morphological effect of the flood and ebb current.

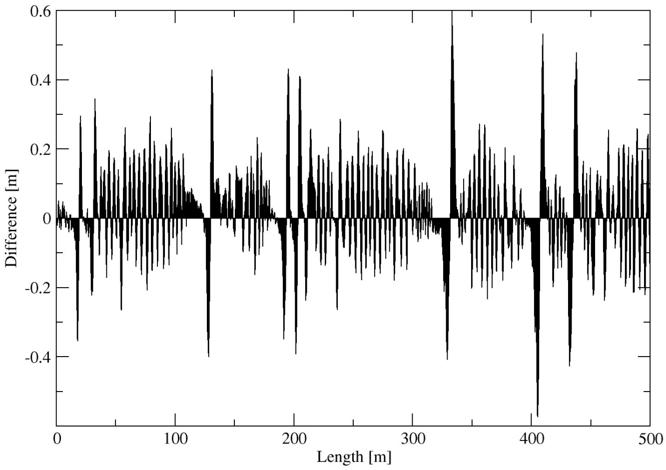

Figure 3: Bathymetric differences between measurements at low water and successive high water.

The superimposed smaller bedforms change orientation according to the direction of the current. Migration rates are in the order of three meters in a half tidal cycle for the highest bedforms, and about one meter in the deepest parts. It shall be stressed out that a net migration cannot be quantified on the basis of this dataset as the tidal cycle was not completely covered. However, a more quantitative assessment of bedform migration is possible by plotting the differences between two successive bathymetric states if bedform migration is less than one wavelength. As this is the case here, we exemplarily show the volumetric differences between low tide and high tide in Figure 3. The differences show characteristic erosion patterns at the stoss side and deposition at the lee side of the superimposed bedforms.

3.2 *Linear stability model*

The linear stability model aims towards a prediction of long time evolution of bedforms, driven by residual currents. Results of the application of the model for representative parameters of the central tidal channel are shown in terms of the growth rate $\overline{\Gamma}_r$ and $\overline{\Gamma}_i$ in Figure 4.

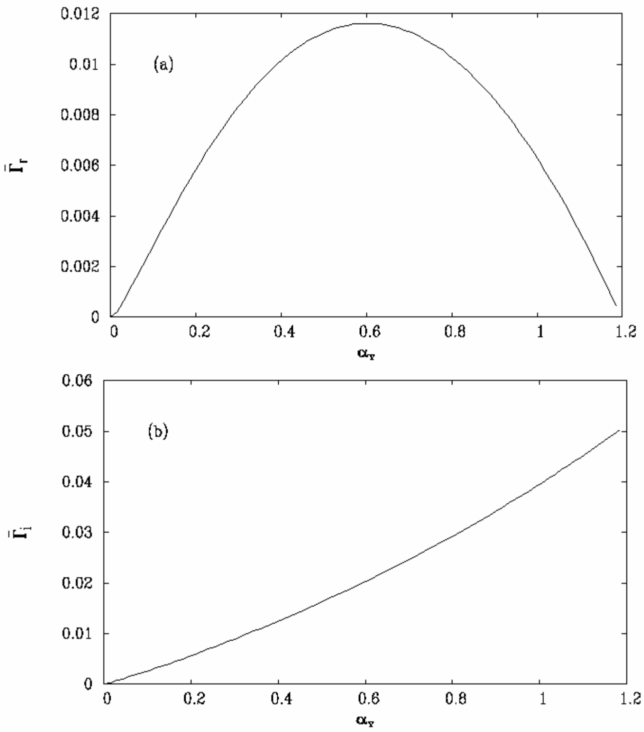

Figure 4: Results of the linear stability analysis: Real (a) and imaginary part (b) of the averaged growth rate Γ as a function of the wave-number (α_x).

For a median sand size $d_{50}^* = 0.63$ mm, a constant and uniform water depth of 11.5 m, a sinusoidal forcing (dominant tidal component M2) of amplitude 0.84 m/s, and a residual current of 0.03 m/s it is found that $\overline{\Gamma}_r$ peaks for $\alpha_x = \alpha_{\max} = 6.04\,10^{-1}$ and the value of $\overline{\Gamma}_i$ corresponding to $\alpha_{max}\left(\overline{\Gamma}_i^{max}\right)$ is $2.02\,10^{-2}$. The dimensional values of the wavelength $\left(L^*\right)$ and migration speed $\left(C_w^*\right)$, associated to the maximum of $\overline{\Gamma}_r$, can then be computed as:

$$L^* = \frac{2\pi h_0^*}{\alpha_{max}}; \quad C_w^* = \frac{\overline{\Gamma}_i^{max} d_{50}^* \omega^*}{\alpha_{max}\sqrt{\hat{\psi}_d}\left(1-p_{or}\right)} \tag{15}$$

This results in a wavelength L^* equal to 120 m and a bedform migration celerity C_w^* equal to 9.8 m/yr.

3.3 *Morphodynamic models*

In contrast to the linear stability analysis which covers much larger timescales and addresses the large bedforms in the tidal channel, morphodynamic model simulations were applied for short timescales in the range of tidal cycles. The coarse Grådyb model has been applied for a neap-spring tidal cycle for currents only in order to derive the main harmonic constituents of the horizontal tide and a residual current. A tidal analysis of computed current velocities for a position in the main channel brings about the main semi-diurnal tidal components M2, S2, N2. M2 has an amplitude of 0.84 m/s and a phase of 343 degrees. S2 has an amplitude of 0.20 m/s and a phase of 52 degrees. N2 has an amplitude of 0.12 m/s and phase of 316 degrees. The residual current velocity was found to be 0.03 m/s. Due to computational restrictions the morphodynamic model has been applied to simulate a few tidal cycles only. Preliminary morphodynamic computations with the uncalibrated coarse model give a picture on sedimentation/erosion rates for a tidal cycle (

Figure 5).

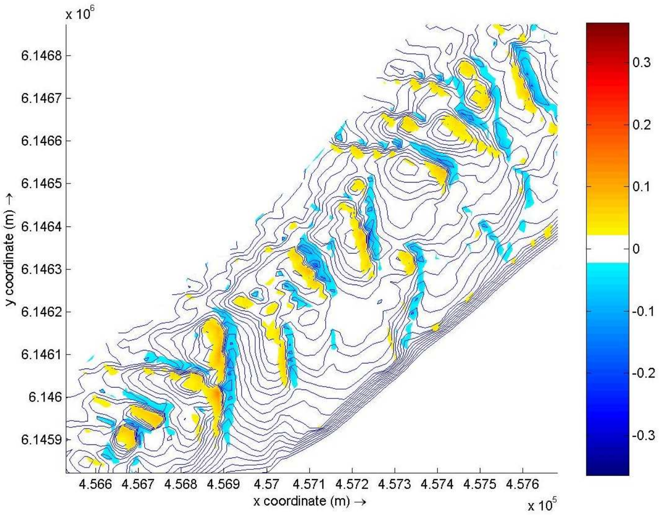

Figure 5: Coarse model: Computed sedimentation (red)/erosion (blue) rates in [m] after one tidal cycle for a part of the main Grådyb tidal channel. Isobaths depict 0.2 m intervals of the bathymetry. Coordinates are UTM (Zone 32).

Results of this simulation seem qualitatively reasonable, as computed transport rates at the crests of the large bedforms are comparable to the measured values. The simulations show erosion at the crests and deposition on the lee side of the large bedforms. However, the coarse resolution of this model does not allow a simulation of the smaller, superimposed bedforms. It shall be noted that a longer application of this morphodynamic simulation leads to unrealistic subsequent degradation of the bedforms towards a bed of minor undulations.

To further explore the capabilities of the model in terms of small scale morphodynamics, the higher resolved 1DV model was applied for a tidal cycle. Resultant bathymetrical profiles after a half and a full tidal cycle are shown in Figure 6.

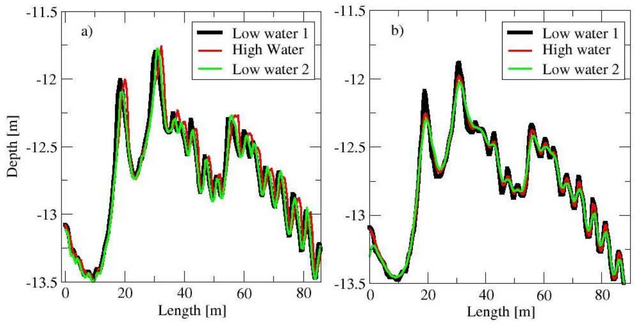

Figure 6: Inter-comparison of measured (a) and simulated (b) small scale morphological evolution of a compound bedform in the central main channel.

The simulation predicts the morphodynamic evolution of the bedforms in the same order of magnitude as the field data. However, similar to the coarse model simulations, these results lack reality in terms of the calculated migration of the single forms. Instead of a horizontal migration of the superimposed forms which is observed in the measured data, the simulation mainly results in a vertical adjustment of the profile. Bedform crests are predicted to erode while the troughs are smoothed out. Although this simulation could not be extended in time, a further evolution would probably lead towards a smooth bed.

4 DISCUSSION

Successive high-resolution bathymetric profiling in a tidal channel in the Danish Wadden Sea reveals the complex morphology and the dynamics of compound bedforms. Large features of about 80-120m length are stable in that they remain ebb oriented with a low angle stoss side and a steeper lee side facing the open sea. Repeated annual measurements have shown that these migrate in ebb direction in the order of ten meters per year. The superimposed smaller bedforms are more mobile, changing orientation within single tidal cycles. The common theory of “equilibrium superposition” presumes the formation and existence of small bedforms within an internal boundary layer on the back of the large features but does not explicitly describe the formation of the latter. Winter & Ernstsen (2007) have shown the different characteristics of the large and small bedforms and proposed a separate analysis of either group. Here preliminary results of two numerical models have been tested according to their suitability of application.

A linear stability method was shown to be able to predict the wavelength of the fastest growing mode, i.e. the large bedforms, for a set of parameters representative for the Grådyb domain. Also the predicted migration speed was practically similar to measured values. Information on the amplitude, form, and the time required to attain an equilibrium state are inherently not possible by a linear stability analyses.

In addition, a 3D morphodynamic model was applied to simulate the short-scale morphodynamic evolution in the horizontal domain, and with higher resolution along a bathymetric profile. The observed migration characteristics of the large forms cannot be resolved because of computational costs. The dynamics of the superimposed bedforms are only captured in a sense that transport rates are in the same order of magnitude as the measured data. Possible reasons, which form the basis for ongoing sensitivity studies, are a still too coarse discretisation of the domain, a transport formula (or the input data parameterisation) not suitable for bedform migration, or insufficient hydrodynamic forcing.

5 CONCLUSIONS

A linear stability analysis was shown to estimate a stable mode of bedform length in a computational domain characterised by a mean water depth, a median grain size and harmonic velocity forcing. Taking into account a residual current, also the celerity of the bedforms were derived. If the assumption is valid that residual tidal circulations are driving the large bedforms observed in the Grådyb tidal channel, they may be explained as tidal system instabilities similar to features which are mostly known in larger dimensions at open coasts - often called sandwaves (e.g. Knaapen, 2005). The small superimposed bedforms are assumed to be genetically different forms comparable to bedforms known from flume experiments and unidirectional currents. Preliminary morphodynamic computations shown here are obviously limited by computational costs and lack realistic bedform dynamics. However, high resolution data on the dynamics of the superimposed smaller bedforms suggest that these act autonomously according to the common concept of “equilibrium superposition”.

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