Research papers

The role of flow asymmetry and mud properties on tidal flat sedimentation

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\textbf{A B S T R A C T}

It is well known that landward transport of fine sediments on tidal flats is caused by lag effects, of which the scour lag and the settling lag are the best known. These lag effects result from a combination of sediment properties and hydrodynamic asymmetries. However, it is not well-understood how, in a quantitative way, these lag effects depend on the sediment properties and the hydrodynamic forcing, and what the relative importance of these sediment properties and hydrodynamics is. As a result, it is not known which lag effect is more important under which conditions. We therefore set out to explore the relative importance of hydrodynamics and sediment properties on tidal-flat sedimentation using a schematized 2DV cross-shore profile model with a tidal range of 6 m and an intertidal width of \( \sim 5 \) km. The effect of hydrodynamics is parameterized through a varying offshore tidal asymmetry and four different cross-shore profiles (a horizontal, convex, linear and concave profile). The effect of sediment properties is examined by evaluating a range of settling velocities \( w_s \) and critical bed shear stresses for erosion \( \tau_{cr} \). The main findings of this work are that (1) conditions of maximum sediment deposition rates exist for \( w_s \sim 0.5 \) mm/s and \( \tau_{cr} \sim 0.1 \) Pa, (2) deposition rates due to slack tide asymmetry are comparable to symmetric tides, while peak flow asymmetry produces the greatest deposition rates, (3) tidal flat deposition rates are greatest for concave profiles and least for convex profiles mainly due to the horizontal velocity gradient, and (4) the type of lag effect dominating landward transport varies spatially.

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1. Introduction

Many tidal embayments and estuaries harbor extensive tidal flats where fine-grained sediments accumulate. These tidal flats play an important role in the ecological functioning of such marine systems, and in protecting the low-lying hinterland against flooding. However, mudshores are still poorly understood compared to their sandy equivalents, partly because of the complexity of behavior of cohesive sediments, which is influenced by chemical, physical, and biological processes (e.g., Pethick, 1996; Kirby, 2000; Widdows et al., 2004; Friedrichs, 2012). The transport of fine-grained sediments towards these flats has therefore been extensively studied in the past, starting with the pioneering work of Van Straaten and Kuenen (1957, 1958) and Postma (1954, 1961, 1967). Over long timescales, tidal flats generally keep pace with rising sealevel owing to a variety of landward sediment transport processes related to lag effects (notably the settling lag introduced by Postma (1954), and the scour lag introduced by van Straaten and Kuenen (1957)).

Bartholdy (2000) concluded that lag effects are most important for coarse silt and fine sand, while Ridderinkhof and Eisma (in Eisma et al., 1998) state that settling and scour lag are only pronounced when suspended sediment concentrations are high, as then flocculation processes become important. Hence, the response of tidal flats to a change in sediment supply and grain size is strongly non-linear. This response may become even more complicated because of differences in Eulerian tidal asymmetries (slack tide asymmetry and peak flow asymmetry) and Lagrangian asymmetries in flow velocity or water depth. Due to the interaction between sediment properties, sediment concentration, hydrodynamics and bathymetry, the various contributions of these processes to tidal-flat sedimentation are not easily determined.

Previous analyses of tidal-flat transport processes using analytical (Friedrichs and Aubrey, 1996; Pritchard and Hogg, 2003) and numerical models (Roberts et al., 2000; Pritchard et al., 2002; Le Hir et al., 2000) of cross-shore profile evolution have lead to improved understanding of the long-term behavior of tidal flats. Roberts et al. (2000) and Pritchard et al. (2002) determined equilibrium bed profiles for tidal flats with a variation in tidal range, sediment supply and tidal asymmetry. An increase in tidal...
range and a decrease in sediment concentration both lead to a steeper tidal-flat profile. An asymmetrical tide with greater peak flood currents leads to accretion, while ebb-dominance leads to erosion; the bed profile shaped by ebb-dominant tidal asymmetry is steeper than for flood-dominant tidal asymmetry. Inclusion of wave effects by Roberts et al. (2000) leads to a transition from convex to concave equilibrium profiles, in agreement with observations (i.e., Kirby, 2000; Pritchard and Hogg, 2003) compute the equilibrium state of tidal flats under constant forcing conditions by applying an analytical model. They concluded that the gross morphological evolution of tidal flats (erosion/accretion rates as well as shape) depends on the offshore sediment supply. Although tidal asymmetries were related to equilibrium tidal-flat morphology, the actual transport rates and processes were not analyzed in detail.

A fundamental aspect not covered by previous modeling studies on tidal-flat sedimentation is the relation between sediment characteristics (settling velocity, critical shear stress for erosion), tidal asymmetry, and settling lag. The relative contribution of lag effects and tidal asymmetry on residual transport has mainly been explored on larger scales, e.g., for bays (Pritchard, 2005) or estuaries (Chernetksy et al., 2010). Only recently Hsu et al. (2013) developed a detailed intertidal flat model simulating the effect of tides and thereby the settling lag on residual fine sediment transport. Therefore, the aim of this paper is to analyze and quantify the effects of tidal asymmetry and sediment characteristics on transport rates and mechanisms for various tidal-flat geometries using a numerical model. The structure of this paper is as follows. First, we review tidally-driven fine sediment accumulation processes in intertidal areas. This is followed by a model description and model results. These are further interpreted in the discussion, followed by conclusions.

2. Net fine sediment transport by oscillating tidal currents

Net transport of (fine) sediments is always related to asymmetries in the hydro-sedimentological conditions. Basically, two types of asymmetries can be distinguished, which often occur simultaneously: (1) asymmetries in the water movement, and (2) asymmetries in the sediment properties/behavior. Therefore we will discuss the relevant hydrodynamic and sedimentary processes, limited to abiotic processes of tidal origin.

2.1. Asymmetries in the sediment properties/behavior

The energy required to mobilize fine cohesive sediment particles from the bed is much greater than the energy to keep these particles into suspension. Also, because cohesive sediment particles (flocs) are very small (with settling velocities between \(\sim 0.1 \text{ mm/s}\) and \(\sim 1 \text{ mm/s}\)), the time scale to mix the particles over the water column is much smaller (\(\sim \text{minutes}\)) than the settling time (\(\sim \text{hours or more}\)). These asymmetries in mobilization, vertical mixing and settling can contribute to processes known as settling lag and scour lag. The term settling lag was introduced by Postma (1954), and further elaborated by van Straaten and Kuenen (1957, 1958), and Postma (1961, 1967). The settling lag is the period or distance a particle can travel when the flow velocity has fallen below the critical shear stress for erosion, before settling on the bed. Combined with a spatial asymmetry in the water depth or flow velocity and a certain critical shear stress for erosion, this leads to transport in a preferential (usually landward) direction. The term scour lag was introduced by Van Straaten and Kuenen (1957), describing another form of landward sediment transport. They defined scour lag as the net transport arising from the greater critical shear stress needed for erosion than for maintaining particles in suspension. Combined with an asymmetry in the flow, this may lead to a net transport of sediment. Physical mechanisms for this increase in critical shear stress with time are self-weight consolidation, bed strengthening by diatom mats, and/or drying of intertidal areas. Nicholls (1986) and Dyer (1994) defined scour lag as the time required to vertically disperse sediment, thereby focusing on processes in the water column rather than at the water–bed interface. They defined the net transport arising from an increase of shear stress within the bed (comparable to van Straaten and Kuenen’s scour lag) as erosion lag. We will use both the Dyer and van Straaten and Kuenen terminology, defining scour lag as the net transport of sediment resulting from time-lags associated with the vertical mixing of sediment and the mobilisation of sediment from the bed.

2.2. Asymmetries in the water movement

Asymmetries in water movement may occur over time, over horizontal space, and/or over the water column.

2.2.1. Time asymmetries

In semi-diurnal tidal regimes, tidal asymmetry results from the generation of the \(M_4\) overtide, and asymmetry is often measured in terms of phase differences between the \(M_2\) and \(M_4\) tidal component. In mixed tidal regimes, other components may play a role as well (Hoitink et al., 2003; Niedziko, 2010; Song et al., 2011; van Maren and Gerritsen, 2012). For instance, when \(0^\circ < (2\phi_{M2}-\phi_{M4}) < 180^\circ\) (where \(\phi_{M2}\) and \(\phi_{M4}\) are the phase angle of the \(M_2\) and the \(M_4\) component, respectively) peak flood velocities exceed peak ebb velocities, resulting in a net flood-directed sediment transport (as the sediment transport rate generally scales with the flow velocity to the power \(n\), with \(n > 1\)). This process has been extensively elaborated in the literature for granular sandy material (e.g., Friedrichs and Aubrey, 1988; Van de Kreeke and Robaczewska, 1993), but it is relevant for alluvial systems of fine sediment as well (Dronkers, 1986). Asymmetry in the slack tidal period is defined as the difference between ebb and flood slack tide period at which the flow velocity is below a critical velocity threshold. Landward transport prevails if the duration of HWS (high water slack) exceeds the duration of LWS (low water slack); this occurs for \(90^\circ < (2\phi_{M2}-\phi_{M4}) < 270^\circ\). In general, coarser sediment is more sensitive to local asymmetries in maximum velocity, while fine sediment is more sensitive to local asymmetries in the duration of slack (Friedrichs, 2012), as long as suspended sediment concentrations are fairly low.

2.2.2. Vertical asymmetries

Asymmetry in vertical mixing (often referred to as internal asymmetry; Jay and Musiak, 1994), results from asymmetries in peak flow velocities (with vertical mixing scaling non-linearly with flow velocity). If vertical mixing during flood is more intense than during ebb, particles will travel longer distances during flood owing to the larger flow velocities higher in the water column. Tidal asymmetries in vertical velocity distribution are associated with accelerating and/or decelerating flows, and density currents. Combined with the vertical distribution in suspended sediment concentration, an asymmetry in the flow velocity profile may induce net sediment transport by differential advection. However, the asymmetries in vertical suspended concentration often occur in conjunction with asymmetries in vertical mixing, in particular in the case of gravitational circulation (e.g., Dyer, 1994).
2.2.3. Horizontal asymmetries

Asymmetries in longitudinal velocities are generally important in short basins and on tidal flats (where substantial velocity gradients exist within the tidal excursion), where flow velocities decrease towards the head of the basin or towards the top of the tidal flat. For water particles traveling with the flow, this longitudinal asymmetry introduces an asymmetry in slack water period, even though Eulerian flow velocities may be symmetric. This effect, noted by Postma (1961), is visualized in Fig. 1. The Eulerian flow velocity at two locations (x1 and x2) is fully symmetric. However, a water particle departing at the onset of flood at location x1 arrives at x2 at the end of flood. Since the flow velocity here is smaller than at x1, the duration of slack tide (at which the velocity is below a critical threshold velocity) experienced by the water particle at x2 is longer than at x1.

Van Straaten and Kuenen (1957) stressed the importance of spatial variations in water depth. For convex profiles (typically channel-shoal systems), the average water depth is less during HWS (when the flats are inundated) than during LWS (when water is confined to the deep channels). More sediment can therefore be deposited on the bed at HWS than at LWS, inducing landward sediment transport.

2.3. Interactions of hydrodynamics and sediment properties

Net sediment transport is generated by a combination of hydrodynamic asymmetries and sediment properties. A neutrally buoyant particle can only be transported in a tidally averaged preferential direction by residual flow, but not by the other hydrodynamic asymmetries discussed above. The latter require sediment with a certain settling velocity and/or critical shear stress for erosion to transport the cross-shore tidal flow, and the relative importance of sediment properties and tidal flow asymmetries, we undertake experiments with a numerical model in which flow asymmetries and sediment properties are systematically varied.

3. Methods

3.1. Modeling approach

The present study is restricted to sedimentation from tidal flow, and the relative importance of sediment properties and tidal asymmetry therein. We only account for cross-shore tidal flow, since alongshore flows mainly dominate the sediment transport in deeper water (Le Hir et al., 2000). For conditions where cross-shore flow velocities are generated by tidal filling and emptying, and when the hydraulic roughness is spatially uniform, the cross-shore distribution of flow velocity is determined by the cross-sectional profile. This can be illustrated with a simple mass balance, providing the variation of the depth-averaged flow velocity $u$ over distance $x$:

$$u_x = \frac{\ell_x}{h_x} \frac{dh}{dt}$$

where $h_x$ is the water depth and $\ell_x$ is the length of the intertidal tidal flat at location $x$, in the flow direction. For linear profiles, $\ell_x/h_x$ is spatially constant, and therefore the flow velocity is independent of $x$, i.e., spatially uniform. Convex profiles have a landward increasing $\ell_x/h_x$ ratio, thus landward increasing flow velocities; for concave profile the opposite holds. Note that bed friction may generate a spatial variation in $dh/dt$ through deformation of the tidal wave, leading to non-uniform $u_x$ in case of constant $\ell_x/h_x$. In order to evaluate the relative contribution of the cross-shore variation in depth and flow velocity on the effectiveness of landward transport, we use four tidal-flat schematizations: a uniform, linearly increasing, concave-upward, and convex-upward profile.

3.2. Model application

We simulate tide-induced fine sediment transport and bed level changes using Delft3D, solving the unsteady shallow water equations under the hydrostatic pressure assumption, and computing sediment transport and morphological update simultaneously with the flow (see Lesser et al., 2004). This model can accurately reproduce complex sediment transport and deposition patterns in high-concentration fine sediment environments: see e.g., van Maren, 2007; van Maren et al., 2009; and Hu et al., 2009).

The schematized single tidal flat of infinite length has a width of 10 km (with an intertidal region of ~5 km), a horizontal resolution of 50 m, and 20 evenly spaced vertical sigma layers. The model is forced with a semidiurnal tide (12 h period for convenience, called $D_2$) with amplitude $A_2=2.7$ m, and a quarter-diurnal over tide with $A_4=0.3$ m or with a symmetric tide with $A_2=A_4=3.0$ m (and $A_4=0$ m). The asymmetry of the tide is determined by the relative phase lag ($2\phi_2-\phi_4$) of the semi-diurnal and the quarter-diurnal constituents. The model is forced with tidal asymmetry at its open boundary, where ($2\phi_2-\phi_4$) is set at $0$, $90^\circ$, $180^\circ$, and $270^\circ$. In order to have similar tidal prisms for simulations with and without tidal asymmetry, the amplitude of $A_2$ is increased from 2.7 m to 3 m for simulations with $A_4=0$. Generally, most sediment transport occurs during spring tides; neap tide transport rates are usually smaller but not differing in pattern or magnitude. To keep our analyses simple, the possible effects of a spring-neap cycle are therefore not included. For the roughness a constant Chezy coefficient of 70 m$^{1/2}$/s is used.

Four tidal-flat configurations are defined with horizontal, linear, concave, and convex cross sections (Fig. 2). The origin of the vertical co-ordinate system ($Z=0$) is set at mean sea level (MSL), and the top of the tidal flat is 4 m above MSL (except for the case with horizontal bed level). The length of the model was set at 10 km as a trade-off, to minimize effects of boundary conditions (requiring a long cross-section) and overide production within the model domain (requiring a short cross-section). Internally produced overides should be minimized since this complicates quantification of external tidal asymmetry effects. The slopes (i) of the convex, concave, and linear profiles are given in Table 1. These cross-sectional profiles are typical for tidal flats in mesotidal to macrotidal conditions. Generally, cross-sectional...
profiles tend to become more convex-upward when the relative importance of tidal effects (relative to wave effects) increases. Therefore the convexity of cross-shore profiles is often observed to increase with tidal range (Dieckmann et al., 1987; Kirby, 2000; Friedrichs, 2012). In the Dutch–German part of the Wadden Sea, for example, the shape of the tidal flats changes from concave in the mesotidal part to convex in the macrotidal sections (Dieckmann et al., 1987). However, for a given tidal range, the profile shape may also change over short distances in response to wave exposure. As a result, concave-upward profiles may even be found in hypertidal estuaries such as the Severn (see Allen and Duffy, 1998).

Characteristic concave tidal-flat systems are found in the Yangtze estuary (e.g., Yang et al., 2008 and Fan et al., 2002). The concave profile modeled here is based upon profiles presented by Yang et al. (2008) and Fan et al. (2002), even though spring tides in the Yangtze Estuary are between 4 m and 5 m. Sediment on the Yangtze tidal flats primarily consists of mud with a relatively large silt content. Our convex profile is modeled after the Skeffling flat in the Humber (6 m; Bassoullet et al., 2000). The sediment dynamics are computed with a morphodynamic model based on the Partheniades equation for erosion $E$, and a permanent deposition flux $D$:

$$E = M \left( \frac{\tau}{\tau_c} - 1 \right)$$

$$D = \omega_s c$$

where $M$ is the erosion parameter (kg/m²/s), $\tau$ is the bed shear stress, $\tau_c$ is the critical bed shear stress for erosion, $\omega_s$ is the settling velocity (m/s) and $c$ is the sediment concentration (kg/m³). The mud properties varied here are $\tau_c$ and $\omega_s$. Mainly fine sediments (mud) are transported towards tidal flats by cross-shore tidal currents (in the absence of additional wave-induced bed shear stresses); the flow velocity is usually too small ($< 0.4$ m/s; see Fig. 2) to transport large amounts of sand in suspension. Mud consists of a flocculating clay fraction and non-cohesive silt. The settling velocity of flocculated mud and fine to coarse silt is typically between 0.1 mm/s and 2.0 mm/s; these values are much larger than the settling velocity of the primary particles, the building stones of the flocs. Although the degree of flocculation (and therefore the settling velocity) may vary throughout the tidal cycle (e.g., Winterwerp, 2002), we use a constant settling velocity per simulation. The critical shear stress for erosion of cohesive mud may range from around several 0.01 Pa to several Pa, depending on the degree of compaction. We use values in the range of 0.01 to 0.2 Pa, typical for non-consolidated to loosely consolidated clay (see e.g., Widdows et al., 2007; Dickhudt et al., 2011).
The sediment concentration at the offshore boundary is set at a constant value of 100 mg/l (which is a simplification since in reality the offshore sediment concentration varies throughout the tide). Initially, no sediment is available for erosion (starved bed conditions); only sediment entering the model domain through the offshore boundary becomes available for erosion and deposition.

4. Results

4.1. Hydrodynamics

The flow velocities computed at the offshore boundary vary for the different cross-sectional profiles, since these depend on the local depth and the tidal prism (which are different for all cross-sections). Velocities at the offshore boundaries are larger for the concave and the horizontal profiles than for the convex and the linear profile (Fig. 2). For both the concave and the horizontal profiles, this is caused by the larger tidal prism, while for the concave profile, this is additionally caused by the shallower depth. From deep water to MSL, the flow velocity for the linear profile is spatially uniform while computed flow velocities for the convex profile increase in the landward direction (Fig. 2). For all profiles, the magnitude of the peak flood flow velocity is equal to the peak ebb flow velocity in case of symmetric hydrodynamic forcing (not shown).

4.2. Net transport induced by symmetric tides

Forced by a symmetric semi-diurnal tide only, fine sediment is transported landward. The amount and the location of the deposited sediment vary with the critical bed shear stress for erosion \(\tau_{cr}\) and the settling velocity \(w_s\). For a horizontal profile, the net deposition rate is greater for a large settling velocity than for a small settling velocity (Fig. 3). This is probably related to the settling lag: sediment with a low settling velocity remains suspended throughout slack water and is transported back seaward while more sediment with a larger settling velocity is deposited. The cross-shore position of sediment deposition appears to be independent on the settling velocity: peaks in deposition rate occur at approximately the same locations for large and small settling velocities. The effect of \(\tau_{cr}\) is different: sediment is transported farther landward for a smaller \(\tau_{cr}\), in line with the linear decrease in flow velocity (Fig. 2), but the deposition rate varies little with \(\tau_{cr}\).

The effect of \(w_s\) and especially \(\tau_{cr}\) is different for the linear profile, and more complex (Fig. 4). Sediment with small \(w_s\) and \(\tau_{cr}\)
is transported farther landward than sediment with large \(w_s\) and \(\tau_{cr}\), while the deposition rate decreases. This landward transport of sediment with small \(w_s\) is explained by the landward decreasing flow velocity. Although the flow velocity \(u\) is spatially uniform at a given point in time for a linear profile, the peak flow velocity within a tidal cycle (\(\ddot{u}\)) does spatially vary. This is caused by the relation between \(dh/dt\) and bed level \(Z\). The water level gradient \(dh/dt\) is greatest halfway between flood and ebb (thus at MSL), and consequently \(u=\ddot{u}\) at MSL. In the area below MSL, \(\ddot{u}\) is therefore spatially uniform. However, \(\ddot{u}\) decreases from MSL to HW (see Fig. 2a), because flooding and drying occurs at progressively smaller \(dh/dt\). As a result, \(\ddot{u}\) is constant from deep water to \(Z=\text{MSL}\), and then decreases to 0 at \(Z=\text{HW}\). With landward decreasing maximum bed shear stress, the sediment transported in suspension becomes increasingly finer in that direction. The deposition decreases with \(w_s\) because relatively more sediment remains suspended, but also because lag effects become less important: the settling lag leads to residual transport if time lags are longer on either flood or ebb. When sediment remains in suspension during both ebb and flood, there is no lag effect.

The sedimentation patterns in relation to \(w_s\) and \(\tau_{cr}\) are qualitatively similar for the linear, concave, and convex profiles, and therefore the longitudinal distributions are not further discussed. Instead, the total deposition rate is averaged over the upper tidal flat (from MSL to HW) and the lower tidal flat (from LW to MSL). This allows for a more detailed analysis of the effect of \(w_s\) and \(\tau_{cr}\) on deposition rates. We compare model results with a starved bed (analyzing model results after one-week simulation) with the results for an alluvial bed (model results after 12-week simulation). For both starved bed and alluvial bed conditions, deposition rates on the upper tidal flat (UTF) are greatest for the concave profile (up to 1 mm/day), followed by the linear profile (up to 0.5 mm/day) and the convex profile (up to 0.2 mm/day). These results suggest that the model tends to evolve towards a convex profile. For all model scenarios, a condition of maximum deposition rates exists, depending on \(w_s\) and \(\tau_{cr}\) (Figs. 5 and 6). For the convex, the concave, and the linear profile, deposition rates on the tidal flat (HW to MSL) are large for sediment with \(w_s=0.5 \text{ mm/s to 1 mm/s}\) (Fig. 5a, c and e). For the concave and the linear profile, deposition rates peak at \(\tau_{cr}\approx 0.1 \text{ Pa}\) (Fig. 5c, e), while sediment deposition rates peak at \(\tau_{cr}\) around 0.05 Pa for the convex profile (Fig. 5a).

Of particular interest is the change in the condition of maximum deposition rate in time. After one week simulation (Fig. 5), the amount of sediment available for erosion from the initial bare tidal flat is limited. After 12 weeks simulation, an abundant amount of

**Fig. 5.** Deposition rates (cm/day) after one week, using a symmetric tide, averaged over the upper tidal flat (left; a, c, and e), and the lower tidal flat (right; b, d, and f) for the convex profile (a and b), linear profile (c and d), and concave profile (e and f). The upper tidal flat is the area between HW and MSL; the lower tidal flat between LW and MSL. The length of the upper tidal flat is 2400 m, 2400 m, and 1900 m whereas the length of the lower tidal flat is 1600 m, 2400 m, and 3200 m (for the convex, linear, and concave profile, resp.).
sediment is available throughout the model domain, and deposition rates become constant (Fig. 6). Note that this condition of constant deposition rate does not imply morphodynamic equilibrium: at morphodynamic equilibrium net deposition rates would be zero. Constant deposition rates over an equilibrium profile (i.e., seaward progradation of the profile) was also found by Pritchard et al. (2002), which they attributed to persistent settling lag processes. Friedrichs (2012) presumed that Pritchard’s constant deposition rates at equilibrium to result from tidal bores (i.e., flood-dominant peak flow asymmetry generated within the model). The convex profile is closest to morphodynamic equilibrium, while the concave profile is furthest from equilibrium, in line with expectations for a tide-dominated environment (e.g., Kirby, 2000). After 12 weeks, the peak in deposition rates is at larger \( \tau_{cr} \) and larger \( w_s \) than after 1-week simulation. Also note that deposition rates on the UTF is greater after 12 weeks than after one week, while deposition rates have decreased on the lower tidal flat (LTF). This is explained by the time required for sediment to be transported from the subtidal zone to the UTF.

The transition from starved bed (Fig. 5) to alluvial bed (Fig. 6) is of the order of eight weeks (Fig. 7). Deposition rates are higher on the LTF than on UTF in the first 2–4 weeks (Fig. 7). Deposition rates on the UTF progressively increase, whereas the deposition rates on the LTF decrease. In order to better understand the relation of the peaks in deposition rate with \( w_s \) and \( \tau_{cr} \) (as in Fig. 5 and 6), we first analyze the processes responsible for the computed landward sediment transport.

### 4.3. Transport mechanisms

The hydrodynamics and sediment transport for a symmetrical tide, with average sediment parameter settings (\( \tau_{cr}=0.1 \) Pa, \( w_s=1 \) mm/s, linear profile) are visualized in Fig. 8 (5 cycles, depth-averaged values) and Fig. 9 (single cycle, vertical distributions). The computed sediment concentration increases from 0.1 kg/m³ at the seaward model boundary to slightly over 2 kg/m³ (values up to 1 kg/m³ shown in Fig. 8c) on the landward side. During flood, sediment is transported in the landward direction, while less sediment is transported in seaward direction during the subsequent ebb (Fig. 8d). This results in a net transport in flood direction (Fig. 8e), and therefore deposition (Fig. 8f). The mechanisms responsible for this landward transport differ for the subtidal zone and the intertidal zone, and will therefore be discussed separately below.

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**Fig. 6.** Deposition rates (cm/day) after 12 weeks, using a symmetric tide, averaged over the upper tidal flat (left: a, c, and e), and the lower tidal flat (right: b, d, and f) for the convex profile (a and b), linear profile (c and d), and concave profile (e and f). The upper tidal flat is the area between HW and MSL, the lower tidal flat between LW and MSL. The length of the upper tidal flat is 2400 m, 2400 m, and 1900 m whereas the length of the lower tidal flat is 1600 m, 2400 m, and 3200 m (for the convex, linear, and concave profile, resp.).
4.3.1. Subtidal zone

In the subtidal zone (X=9000 m, black dashed lines), landward transport occurs (Fig. 8d and e) despite \( \bar{u}_{flood} = \bar{u}_{ebb} \) (close to 200 mg/l; see Fig. 8c) and \( \bar{u}_{flood} = \bar{u}_{ebb} \) (0.35 cm/s, see Fig. 8b), where \( \bar{u} \) and \( \bar{c} \) are maxima in flow velocity and sediment concentration. This residual landward transport is induced by phase lags between concentration. This residual landward transport is induced by phase lags between concentration. This residual landward transport is induced by phase lags between concentration. This residual landward transport is induced by phase lags between concentration. This residual landward transport is induced by phase lags between concentration. This residual landward transport is induced by phase lags between concentration. This residual landward transport is induced by phase lags between concentration. This residual landward transport is induced by phase lags between concentration. This residual landward transport is induced by phase lags between concentration. This residual landward transport is induced by phase lags between concentration.

4.3.2. Intertidal zone

In the intertidal zone (X=2000 m and 5000 m), maximum concentrations and maximum transports are greater during ebb than during flood (Fig. 8c and Fig. 9h). However, the durations of the high concentration events are longer during flood (4 h) than during ebb (3 h). The concentration during flood is high from the beginning (at \( T=2 \) h at X=5000 m; see Fig. 9h), until sediment settles shortly before HW. During ebb, the concentration remains low for the first 2 h. As a result, even though the maximum transport rate may be greater during ebb than during flood (as in Fig. 9h), the longer duration of flood transport results in residual transport in the flood direction. This process can be classified as settling lag, which will be further explained hereafter.

An asymmetry in transport duration can be caused by a longitudinal gradient in flow velocity. At the end of the flood, sediment particles will start settling from suspension when \( \tau_s < \tau_r \). In case of a longitudinal gradient in flow velocity, these particles are transported in flood direction by the waning current towards areas with small bed shear stress. Due to the smaller bed shear stress, it takes time before \( \tau_s \) exceeds \( \tau_r \) during the following ebb, and as a result, transport in ebb direction lasts shorter than in flood direction. This horizontal asymmetry generates the difference in flood and ebb transport duration in Fig. 9. In addition to this horizontal flow velocity gradient, landward transport may be strengthened by time-asymmetries, which will be elaborated hereafter.

4.4. Effect of tidal asymmetry

So far, we have focused on symmetric tides forced with a sinusoidal \( D_2 \) signal. Next, we introduce asymmetry in the tide, by adding a \( D_4 \) overtide (with amplitude ratio \( D_4/(D_2+D_4)=0.1 \)). The phase difference between \( D_2 \) and \( D_4 \) is varied such that \( (2\phi_2-\phi_4) \) is 0°, 90°, 180°, and 270°. A phase difference of 0° and 180° induces a slack tide asymmetry (with a longer LWS or HWS period, respectively) but equal ebb and flood peak flow velocities. A phase difference of 90° and 270° implies peak flow asymmetry (flood or ebb dominant, respectively), with identical slacks. For \( 2\phi_2-\phi_4=90° \), the flow velocity during flood is larger than during ebb, and since sediment transport scales approximately with the cubed flow velocity, net transports should be greater than for symmetric tides. For \( 2\phi_2-\phi_4=180° \), the period around HWS lasts longer than around LWS. Hence, particles have a longer time to settle at the end of flood than at the end of ebb, and residual landward transport is probably greater than for symmetric tides.
The simulations show that deposition rates in the upper intertidal zone are greatest during flood-dominant peak flow asymmetry \(2\phi_2-\phi_4=90^\circ\), see Fig. 10. Flood-dominant slack tide asymmetry \(2\phi_2-\phi_4=180^\circ\) also results in deposition, but at a lesser rate. Similar to symmetric tides, conditions of maximum deposition rates are found for peak flow asymmetry and slack tide asymmetry (Figs. 11 and 13). Deposition rates for peak flow asymmetry (left panel in Fig. 11) are more than six times greater (around 2 mm/day) than for symmetric tides or slack tide asymmetry (both 0.3 mm/day; Fig. 5c and the right panel in Fig. 11). The critical shear stress at which maximum deposition rates occur is equal \(0.1 \text{ Pa}\) for these three cases (symmetric tide, peak flow asymmetry and slack tide asymmetry) but the settling velocity at which maximum deposition rates occurs differ. Deposition rates are maximal at the smallest settling velocity for slack tide asymmetry \((w_s=0.5 \text{ mm/s})\) followed by symmetric tides \((w_s=0.5–1 \text{ mm/s})\) and peak flow asymmetry \((w_s=1 \text{ mm/s})\).

Deposition rates are larger for alluvial bed conditions (Figs. 12 and 13) than for starved bed conditions (Figs. 10 and 11), with maximum deposition rates due to peak flow asymmetry exceeding 2.5 mm/day, and maximum deposition rates of 0.45 mm/day due to both symmetric tides and slack tide asymmetry. Conditions of maximum deposition rates are at a slightly larger settling velocity for alluvial bed conditions than for starved bed conditions: \(w_s=0.5–1 \text{ mm/s}\) for slack tide asymmetry, 1 mm/s for symmetric tides, and 1–1.5 mm/s for tidal asymmetry. For both types of tidal asymmetry, the critical shear stress at which the deposition rate is maximal is nearly equal for starved bed and alluvial bed conditions.

Summarizing, the model results suggest that peak flow asymmetry is much more important than slack tide asymmetry for landward sediment transport, both for starved bed and for alluvial bed conditions. Also, slack tide asymmetry seems to increase deposition rates only marginally compared to symmetric tides.
Conditions of maximum deposition rates are found at larger $w_s$ and $t_{cr}$ for peak flow asymmetry than for symmetric tides and slack tide asymmetry; the peak occurs at smallest $w_s$ and $t_{cr}$ for slack tide asymmetry. Since larger $w_s$ and $t_{cr}$ are representative for increasing grain size or floc size, peak flow asymmetry favors transport of coarser sediment more than slack tide asymmetry and symmetric tides, while slack tide asymmetry promotes transport of the finest fraction (in the landward direction in case of a longer HW slack period; seaward for a longer LW slack period).

5. Discussion

5.1. Model limitations

Our 2DV modeling approach is an oversimplification of the complex three-dimensional environment. We first discuss processes shaping tidal flats which are not included in the model.

1. The modeled tidal-flat evolution patterns primarily represent tide-dominated environments, mostly having a convex-upward profile. The morphology of tidal channels and flats strongly depends on sediment resuspension by wind waves (Fagherazzi et al., 2007; Fagherazzi and Priestas, 2010), often leading to concave-upward tidal flats (Kirby, 2000). However, concave profiles may be tide-dominated for a certain period. The tidal flats in the Yangtze estuary, for instance, are shaped by erosion during the storm season, giving rise to an overall concave profile. The following non-storm season is characterized by a prolonged period of continuous tide-dominated deposition (e.g., Yang et al., 2008).

2. Fine-grained mud beds consolidate, resulting in a critical shear stress for erosion that increases with time after deposition (and therefore with depth). Since the critical shear stress for resuspension of consolidated sediments is greater than the shear stress at which unconsolidated sediments settle, this generates landward transport through scour lag.

3. Fine-grained sediments flocculate to form aggregates with a larger settling velocity, depending on the turbulent energy and sediment concentrations prevailing, resulting in spatially and temporarily varying settling velocities. Flocculation was postulated to be necessary to generate landward transport by...
settling and scour lags by Eisma et al. (1998). Indirectly, the model results demonstrate the importance of flocculation: the computed deposition rate is greatest for settling velocities around 1 mm/s while sediment often observed in intertidal areas has a single-particle settling velocity typically several orders of magnitude smaller.
4. Fine-grained sediments may, at even relatively moderate sediment concentrations (several 100 mg/l), generate vertical concentration gradients sufficiently strong to dampen turbulent mixing (Winterwerp, 2001). Van der Ham and Winterwerp (2001) concluded that this damping of turbulent mixing by sediment-induced density stratification, leading to rapid deposition during decelerating flows, is an essential aspect of tidal-flat deposition.

5. Resuspension and transport of sediment may be substantially influenced by the shallow water hydrodynamics, especially the tidal bore that occurs on rising tide and the hydraulic smooth flow that occurs during ebb is not accurately included. This requires a model with a spatial and vertical grid resolution much more detailed than applied here.

6. The model is forced with a constant tide without spring-neap variation. In reality, most sediment transport (and associated morphological change) occurs during spring tide conditions (see e.g., Fan et al., 2002). This implies that while in reality typically ~5 day out of the ~14-day tidal cycle are important for tidal-flat morphodynamics, our model is continuously morphologically dynamic. Resulting morphological change is therefore ~3 times larger than for tides with a similar tidal range (or representative for conditions with a larger tidal range). Furthermore, we presume that the precise shape of the equilibrium profile of a mud flat must be affected by such a spring-neap cycle, in conjunction with the sorting of sediment classes mentioned above.

In addition to these process formulation, the model also assumes along-shore uniformity through the 2DV approach. Most tidal-flat systems have complex 3D structures, with an intricate pattern of channels and gullies intersecting tidal flats. Accretion rates on tidal flats are strongly influenced by the transport of water and sediment in these channels and gullies, while the channels themselves strongly influence the development of tidal asymmetry (Speer and Aubrey, 1985). The small-scale gullies export sediment, compensated by flood-dominated transport on the flats (Mariotti and Fagherazzi, 2011). In addition, tidal flats may also be shaped by longshore currents. However, for the relatively broad flats considered here, cross-shore currents are generally more important (see Le Hir et al., 2000).

Summarizing, we underestimate landward transport by excluding consolidation, sediment-induced density effects, and detailed shallow water hydrodynamics. Simultaneously, we underestimate export by excluding waves and complex three-dimensional geometries. Extending the model with more processes or a more complex geometry would therefore make the model more realistic but increasingly difficult to analyze and interpret. Such complex 3D models are necessary to predict and understand sediment dynamics, but the simplified approach adopted here help in their interpretation.

5.2. Asymmetry in the flow

All cross-shore profiles have a Lagrangian hydrodynamic asymmetry (in water depth and/or flow velocity), of which the hydrodynamics of the linear profile are the least asymmetric. The flow velocity is spatially uniform in time, and for a particle advected by the tidal flow, the water depth at HW slack is equal to the water depth at LW slack. Flow asymmetries for the convex and concave profiles are more spatially variable, reflecting the cross-sectionally variable bed level gradient.

In the subtidal zone of the linear profile, there is no sediment transport resulting from horizontal or time asymmetries in the flow. Sediment is mainly transported towards the intertidal zone by the time lag between $u_t = \bar{u}$ and $c_t = \bar{c}$ (approximately 30 min to 60 min). This timelag is the result of hydrodynamic mixing time, the settling flux (introducing a relation with $w_s$), and $c_{tr}$ (resulting in a time difference between the onset of hydrodynamic mixing and mixing of sediment particles). Therefore this time lag increases with $c_{tr}$ and $w_s$, and residual transport increases accordingly.

The importance of the horizontal flow velocity variations on residual transport is illustrated by the difference between the linear profile and the concave profile. The main differences between these profiles are that for the concave profile (1) the instantaneous flow velocity decreases in the landward direction, and (2) for a particle transported by the oscillating tidal currents, the local water depth at HW is larger than at LW. The first would result in greater landward transports due to settling lag, while the second would promote seaward transport due to more sediment settling at LW than at HW. The predicted greater deposition rates on concave profiles than on linear profiles suggests that the spatial asymmetry in the flow velocity is a more important mechanism than asymmetry in the slack tide water depth.
5.3. Asymmetry in sediment properties

Settling lag is most effective for transporting sediment in the landward direction when sediment settles at the end of flood, but not at the end of ebb (the more sediment remains suspended at LW, the greater the net flood transports). For sediment with a very small settling velocity, which does not deposit on the bed, there is no residual sediment transport. Sediment with a large settling velocity is deposited at the end of flood, but also at the end of ebb. For these conditions, the lag effects are less effective for residual landward transport than for conditions at which most sediment remains in suspension at the end of ebb. Similar arguments apply for $\tau_{\text{cr}}$: sediment with a small critical shear stress for erosion does not remain on the bed but is continuously resuspended, and hence settling lags generating landward transport are less important. For very large $\tau_{\text{cr}}$, sediment is not remobilized and therefore transports are small.

Our model results demonstrate that a combination of settling velocity and critical shear stress for erosion should exist, for which deposition rates are greatest. Postma (1961) already hypothesized that a settling velocity should exist for which deposition rates on tidal flats are maximal, but he could not confirm this with observations. However, although the existence of an optimal settling velocity has been identified for estuaries (e.g., Talke et al., 2009), it has not yet been quantified for tidal flats. An important consequence of the relation between settling velocity and deposition rate, as follows from our model, is that the settling velocity at which the deposition rate is large, should exceed $\sim 0.1 \text{ mm/s}$. This would imply that primary mud particles cannot be deposited on tidal flats (for which settling velocities are several orders of magnitude smaller), but need to be aggregated as flocs.

Following our model results, the dependency of the deposition rate on sediment parameters has implications for the horizontal distribution of sediment on tidal flats. The finer the sediment (low settling velocity), the farther it is transported in the landward direction. This suggests that the sediment size decreases in the landward direction, which is in line with observations (see e.g., Friedrichs, 2012). Additionally, the model indicates that the conditions of maximum deposition rate depends on the amount of sediment available on the bed. For low sediment availability (starved bed), the deposition rate is largest for the finest sediment fraction. With increasing sediment available for resuspension (dynamic equilibrium) the sediment deposited in the upper intertidal zone should become increasingly ‘coarser’: the settling velocity increases and the critical shear stress for erosion increases. Quantifying this transition requires the use of a multi-fraction multi-layer model. A transition from starved bed to equilibrium bed may occur, for instance, after a storm or during the winter to summer transition.

5.4. Lag effects

Even with a simple model, it is difficult to separate the relative contribution of settling lag and scour lag. Strictly speaking, the settling lag is only related to the settling velocity, combined with hydrodynamic asymmetries. However, the settling lag becomes insignificant in absence of a critical shear stress for erosion, because then sediment is permanently in suspension. And without a critical shear stress for erosion, residual transport can only occur due to asymmetries in vertical mixing, which is also scour lag (as defined in Section 2).

In our model results, the landward transport in the subtidal zone results primarily from asymmetries in vertical mixing (and thus scour lag). In the intertidal zone the residual transport is dominated by settling lag. This suggests a landward increasing importance of settling lag relative to scour lag. It should be noted we use a constant critical shear stress for erosion and no critical shear stress for deposition. Therefore, the net transport by scour lag resulting from a greater critical shear stress for erosion than for maintaining sediment in suspension (the definition used by Van Straaten and Kuenen (1957) is not accounted for in our model.

Postma (1954, 1961) and van Straaten and Kuenen (1957, 1958) both stress the importance of a landward decreasing flow velocity for landward transport. However, they disagreed on the relative importance of other processes. Postma related HW settling to asymmetry in the flow field (both Eulerian, generated by friction and Lagrangean, i.e., spatial flow velocity gradients). Van Straaten and Kuenen (1957) emphasize the importance of channel-shoal complexes: around HW all intertidal flats are inundated yielding a smaller average water depth than at LW, when water is confined to the deeper channels. This results in larger deposition rates around HW. Our model suggest a limited importance of slack tide asymmetry. Slack tide asymmetry was found to increase deposition rates, but much less than peak flow asymmetry.

6. Conclusions

Residual transport of fine sediment on intertidal tidal flats results from a combination of asymmetries in the flow and sediment properties. The flow on tidal flats is always asymmetric: even if the flow is spatially uniform through time, the peak flow velocity within a tidal cycle decreases spatially in the upper intertidal zone (from MSL to HW). More importantly, mixing and settling generate time lags which lead to landward transport. Sediment has a finite settling velocity and critical shear stress for erosion, and in combination with asymmetries in the flow velocity, a landward tide-induced residual transport of fine sediment onto the tidal flats always exists. Spatial variations in flow velocity are effective for landward transport of fine sediment. Peak flow asymmetries are found to be more important than other tidal asymmetries, such as slack tide asymmetry. Based on our model results, we conclude that the time lag needed to vertically mix sediment may be an important mechanism transporting sediment from the subtidal zone towards the upper tidal flats. Within the intertidal zone, settling lag contributes more to residual transport than scour lag.

Conditions for maximum deposition rates on tidal flats were found, depending on the critical shear stress for erosion and the settling velocity. This maximum is determined by asymmetries in flood and ebb transport (increasing with increasing settling velocity and critical shear stress) and gross transport rates (decreasing with increasing settling velocity and critical shear stress). Deposition rates were found to depend strongly on the cross-sectional profile, increasing with the concavity of the profile. Although the deposition rate decreases for very fine sediment (smallest settling velocity), this finer fraction is transported farther landward compared to sediment with a larger settling velocity. This explains the often-observed landward decreasing grain size of tidal-flat deposits. Deposition rates for slack tide asymmetry are comparable to those for symmetric tides, while peak flow asymmetry is found to generate the greatest landward transport.

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