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TRANSPORT OF FINE SEDIMENTS IN A NARROW CONVERGENGING ESTUARY – THE SEA SCHELDT

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ABSTRACT

In this paper we study the transport of fine sediments by river-induced flushing, estuarine circulation and tidal asymmetry in the Sea Scheldt, Belgium. This study is carried out with an idealized schematization of the river, modeled as an exponentially converging river with constant depth and rectangular cross section, using Delft3D. Sediment is only imported from the lower sea boundary. Values for tidal amplitude and river flow, prescribed at the models' open boundaries are comparable to those in the Scheldt. We show that a turbidity maximum is formed at the head of the salinity intrusion, driven by estuarine circulation, and in balance with river-induced flushing. For the given conditions, the model does not predict any fine sediment transport beyond that turbidity maximum.

Keywords: Estuary, fine sediments, estuarine circulation, tidal asymmetry

1. INTRODUCTION

In three previous papers (Winterwerp, 2011; Winterwerp and Wang, 2013; Winterwerp et al., 2013), we have argue that the Loire and Ems River have evolved into a hyper-turbid state with suspended sediment (SPM) concentrations of many 10s g/l. This evolution is the response to large-scale engineering works, such as narrowing, rectification and deepening, sometimes accompanied by redirection of fresh water as well. Not so long ago, these rivers depicted more classical estuarine features, with a pronounced estuarine turbidity maximum (ETM) at the head of the salinity intrusion, with concentrations of a few 100 mg/l. Hence, the question arose under which conditions this transition towards hyper-turbid conditions may occur, and whether other tidal rivers, such as the Sea Scheldt, may evolve towards such a state. In this paper, we study the role of the river flow in conjunction with tidal asymmetry on such a transition.

The distribution of fine sediments in an estuary is governed by a balance between the down-estuary transport by river-induced flushing and the up-estuary transport by estuarine circulation and tidal asymmetry. The three tidal asymmetries that may affect fine sediment transport are:

1. **Asymmetry in peak velocities**. As sediment transport in alluvial systems is proportional to U^n, with $n > 1$, peak flood velocities larger than peak ebb velocities induce net up-estuary sediment transport (e.g. Friedrichs and Aubrey, 1988).
2. **Internal asymmetry**. As vertical mixing scales with U^2, peak flood velocities larger than peak ebb velocities induce more mixing during flood than during ebb. Sediment is then better mixed over the water column during flood than during ebb, yielding a net up-estuary transport owing to the larger flow velocities higher in the water column (e.g. Jay and Musiak, 1994).
3. **Asymmetry in slack water duration**. We have to distinguish between Eulerian and Lagrangean conditions. In the latter case, flood dominant conditions prevail when the flow velocity decreases towards the head of the estuary, as sediment resides then longer on the bed at high water slack (HWS) than at LWS, e.g. Van Straaten and Kuenen (1958) and Postma (1961). For an Eulerian analysis, the temporal velocity gradient is relevant. If the gradient at high water slack (HWS) is smaller than at LWS, the duration of flow velocities below a critical velocity for erosion is longer during HWS than during LWS. Hence sediment resides on the bed longer during HWS than during LWS, resulting in a net-up-estuary transport (e.g. Dronkers, 1986).

It is important to realize that asymmetries in tidal velocities are not only governed by the higher harmonics of the tide, but also depends on the river flow. At large river flows, peak ebb velocities may exceed the otherwise dominant peak flood velocity. This obvious observation also implies that in up-estuary direction, the net effect of tidal asymmetry ultimately becomes ebb dominant in a converging estuary. Also, the salinity distribution, driving the estuarine circulation, interacts with the tidal asymmetry, in particular with the internal asymmetry. In general, the water column in the area of salinity intrusion is more stratified during ebb than during flood, owing to tidal straining, which adds to the effects of internal tidal asymmetry.

Further, we should appreciate how the various transport processes interact with the state of the sedimentary system. River-induced flushing, estuarine circulation and internal tidal asymmetry work on the sediment in suspension. However, estuarine circulation can only induce net (up-estuary) transport when the suspended sediment depicts a gradient over the water column – estuarine circulation does not induce net transport on homogeneously mixed matter.

Net sediment transport by tidal asymmetry in peak velocities can only occur over an alluvial bed, i.e. the sedimentary bed contains an abundance of fine sediments so that the amount of sediments in the water column scales with the flow velocity through enhanced erosion (mobility) and vertical mixing. In case that only limited amounts of fines are available in/on the sedimentary bed, the timing of mobilization of those sediments into the water column becomes relevant, which is governed by the length of the slack water period. We refer to starved bed conditions.

These starved bed conditions are encountered in most low-concentration estuaries, such as the current Upper Sea Scheldt, whereas alluvial conditions are met in hyper-concentrated rivers, such as the Ems and Loire.

In this study, we investigate whether the three processes of river-induced flushing, estuarine circulation and tidal asymmetry can induce a transition from starved bed conditions to alluvial conditions. For fine sediments, this implies a transition from a "normal" low-concentration estuary to hyper-concentration conditions.

2. SETUP OF THE STUDY

We study this transition with Delft3D simulations of the sediment transport and fate in an idealized schematization of a converging estuary. Its plan form is depicted in Fig. 1, and its dimensions in Table 1. This idealized schematization has similar features as the Scheldt estuary between Bath and Ghent.

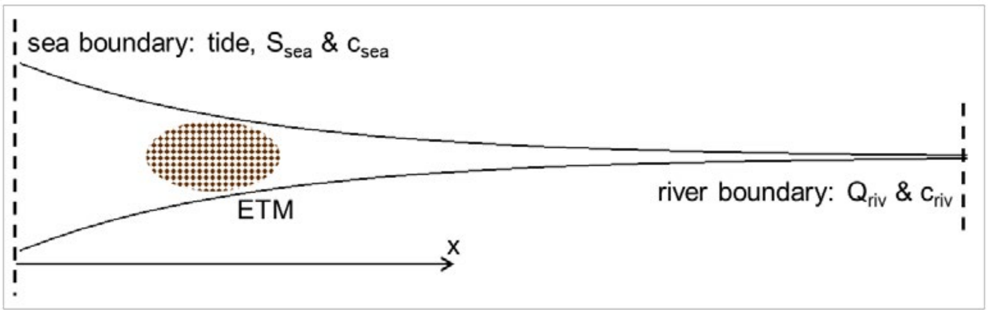

Fig. 1: Plan form of idealized, converging estuary.

Table 1: Reference settings Delft3D simulations; buoyancy coupling means that SPM affects the bulk density of the water-soil mixture, modifying vertical turbulent mixing.

L	estuary length	104 km	S_{sea}	salinity at mouth	20 ppt
h	water depth	6 m	c_{sea}	SPM concentration at mouth	100 mg/l
b_0	width at mouth of estuary	3 km	c_{riv}	SPM concentration at river	0
L_b	convergence length	30 km	W_s	settling velocity	0.5 mm/s
k_s	roughness height	3 cm		water-bed exchange	yes/no
Q_{riv}	river flow	30 m^3/s		buoyancy coupling	yes/no
M_2	amplitude semi-diurnal tide	2.13 m	$\tau_{c,e}$	critical shear stress for erosion	0.3 Pa
ϕ_2	phase semi-diurnal tide	91 deg	M	erosion parameter	0.1 g/m^2/s
M_4	amplitude first overtide	0.12 m	c_{gel}	gelling concentration	100 g/l
ϕ_4	phase first overtide	179 deg			

The width of the estuary follows an exponential function $b(x) = b_0 \exp\{-x/L_b\}$. The depth is constant and we do not include any intertidal area – the tide is therefore profoundly flood-dominant. Only the primary semi-diurnal tidal component (M_2) and its first overtide (M_4) are prescribed at the seaward boundary, while their values are equal to the measured values at Bath in the Western Scheldt.

This basically one-dimensional topography is implemented in Delft3D, using curvi-linear coordinates with grid size in longitudinal direction of Δx = 140 m and at the mouth a grid cell Δy = 140 m, decreasing down to 5 m at the river boundary. Over the water column, we use 23 sigma layers; in the lower half mater, the mean layer thickness measures 5 cm, while increasing by a factor 1.3 higher in the water column. We use a time step of 0.5 min, and run the simulations for many months (0.5 – 1 year). In almost all simulations, we start with an "empty" estuary, i.e. all sediment has to enter the system through its sea boundary.

In this study, the following simulations are analysed, varying the model parameters around the reference conditions given in Table 1:

Table 2: Delft3D simulations.

run	description	study objective
1	salinity, SPM, buoyancy coupling, no water-bed exchange	reference
2	same as 1, but no sediment	development overtides
3	same as 1, but no buoyancy coupling	effect SPM-induced stratification
4	same as 1, but with water-bed exchange	effect slack water asymmetry
5	same as 1, but no water-bed exchange, continuing on run 4	effect peak velocity asymmetry
6	same as 1, but no water-bed exchange and initial sediment on bed	effect peak velocity asymmetry
7	same as 1, but larger M_4 amplitude at boundary	effect external overtide
8	same as 1, but smaller river flow	effect river-induced flushing

3. FIRST RESULTS

When this document was drafted, the Delft3D simulations were not yet in equilibrium – the long simulation times required for attaining equilibrium, in conjunction with the high spatial resolution, yields long computational times.

Fig. 2 shows how the tidal asymmetry develops along the idealized estuary. In concordance with theory, a pronounced asymmetry is developed. At the estuary mouth, the peak ebb and peak flood velocity almost are equal (see Table 1: $2\phi_2 - \phi_4 = 3^0$, i.e. at the edge of a transition between ebb to flood-dominant conditions, e.g. Friedrichs and Aubrey, 1988). However, further up-estuary, peak flood velocities are almost twice the peak ebb velocities, as observed in the Upper Sea Scheldt. Hence, the estuary becomes progressively more flood dominant with respect to the peak tidal velocities in up-estuary direction.

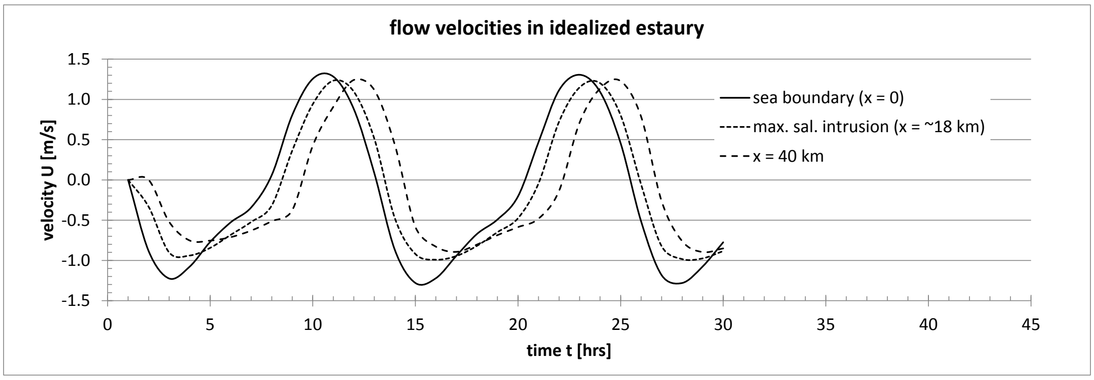

Fig. 2: Development of tidal asymmetry along idealized estuary – positive velocities are up-estuary..

Fig. 2 also shows that the temporal velocity gradient at LWS is larger than at HWS – if we assume a critical velocity U_{cr} = 0.2 m/s, the duration of HWS and LWS are given in Table 3, from which we conclude that with respect to slack water conditions, the idealized estuary is ebb-dominant.

Table 3: Computed duration of slack water period, assuming 02. m/s threshold erosion velocity (run 3).

mouth *x* = 0		maximum salinity distribution		*x* = 40 km	
LWS	HWS	LWS	HWS	LWS	HWS
45 min	24 min	33 min	24 min	24 min	21

Fig. 3 shows an example of the longitudinal and vertical distribution of salinity and SPM in the idealized estuary, with a pronounced estuary turbidity maximum (ETM). Note that this ETM migrates with the tide, the results of which are not shown here. Though salinity is at equilibrium, the SPM distribution is not – sediment continues to be pumped into the estuary. Note the accumulation of fines around the location of maximum salinity intrusion, well exceeding the SPM concentration at the model's boundary. The SPM concentration is well mixed over the water column owing to the absence of sediment-induced buoyancy destruction.

Fig. 3 also shows that no sediment can escape beyond the ETM in up-estuary direction. Apparently, up-estuary transport by estuarine circulation and down-estuary transport by river-induced flushing converge at this ETM-location. As sediment interaction with the bed (erosion and deposition) is not modelled, slack water asymmetry, nor asymmetry in peak velocities contribute to the net fine sediment transport in the estuary.

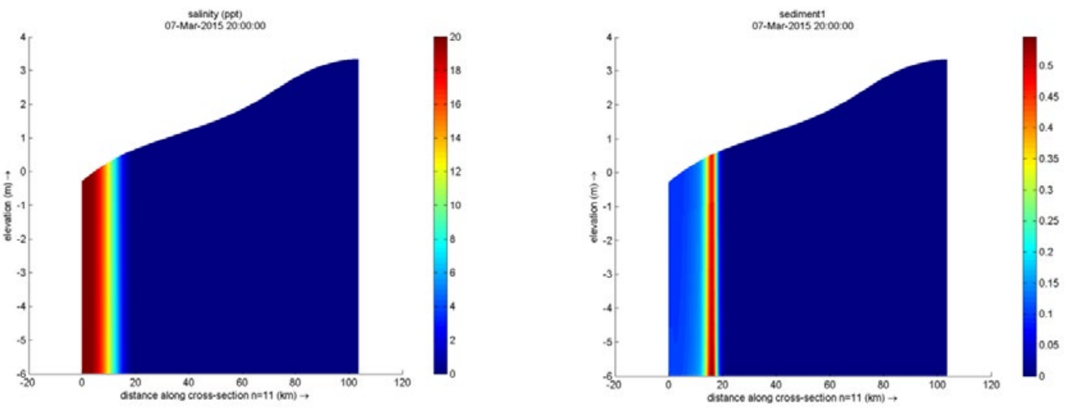

Fig. 3: Longitudinal salinity (left panel) and SPM (right panel) distribution around maximal salinity intrusion, run 3; 7 month simulation time. Colour bars represent salinity [ppt] and SPM [g/l].

These results bear the remarkable conclusion that for starved bed systems in an idealized schematization of a converging estuary, sediment cannot be transported beyond the ETM induced by the longitudinal salinity gradient. This conclusion holds for the boundary conditions prescribed – at this boundary, $t_{LWS} > t_{HWS}$.

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