Morphodynamics of Coastal Inlets and Tidal Lagoons

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ABSTRACT

The impact of tidal lagoons on low-order evolution of adjacent coasts is large. It is in this context that meso-scale dynamic process and macro-scale aggregated process knowledge of tidal lagoons and estuaries are reviewed. Despite promising progress in meso-scale dynamic process knowledge we conclude that this knowledge is insufficient to predict the evolution of tidal basins and their interaction with the coast on time scales much longer than the process scale. We subsequently discuss the morphodynamic interaction between tidal wave propagation and basin geometry, through which we establish a macro-scale relation between tidal propagation and geometry that leads to morphodynamic equilibrium. These findings are used in a macro-scale aggregated model approach, which is based on the concept of looking for morphodynamic response in case macro-scale elements are forced out-of-equilibrium. It is shown that the process-aggregated approach enables us to at least hindcast and possibly forecast the morphodynamic behavior of tidal lagoons and coastal inlets under accelerated rates of sea level rise.

ADDITIONAL INDEX WORDS: Tidal lagoons, coastal inlets, geomorphology, modeling, morphodynamics, sea-level rise, Waddensea.

INTRODUCTION

Tidal lagoons and estuaries, collectively named tidal basins hereafter, interrupt a large part of the world’s shorelines with their associated coastal inlets. Compared to the morphodynamic behavior of uninterrupted coastlines and of rivers, the morphodynamic behavior of tidal basins is a degree more complex and less well understood. These systems are important for both ecological (e.g. for marine life, birds) and for socio-economic reasons (harbors, inland waterways, recreation, resource exploitation, etc.). Also, there are strong indications (Stive and Wang, 2003) that the morphological response of tidal basins to natural and human interventions has an impact on the coastal sediment budget, which is large compared to that of interventions along an uninterrupted coast.

This type of coastal change involves parts of the coast normally ignored in predictions required for management of coastal morphology: i.e., shoreline evolution linked to behaviour of the continental shelf and coastal plain. Cowell et al. (2003) therefore introduce a meta-morphology, the coastal tract, defined as the morphological composite comprising the lower shoreface, upper shoreface and backbarrier (Figure 1). It is the first order-system within a cascade hierarchy that provides a framework for aggregation of processes in modelling low-order coastal change. This framework is used in defining boundary conditions and internal dynamics to separate low-order from higher-order coastal behaviour for site-specific cases. This procedure involves preparation of a data-model by templating site data into a structure that complies with scale-specific properties of any given predictive models. Each level of the coastal-tract cascade is distinguished as a system that shares sediments internally. This sediment sharing constrains morphological responses of the system on a given scale. The internal dynamics of these responses involve morphological coupling of the upper shoreface to the backbarrier and to the lower shoreface. The coupling mechanisms govern systematic lateral displacements of the shoreface, and therefore determine trends in shoreline advance and retreat. These changes manifest as the most fundamental modes of coastal evolution upon which higher-order (shorter-term, i.e. subdecadal scale) changes are superimposed.

The sediment exchanges depicted by the arrows in Figure 1 occur in principle during any average year and on all time scales longer than this. These exchanges are summarised schematically in Figure 2, which differentiates sediment fluxes into sand and mud fractions. For coastal change on any scale,
antecedent morphology, sea-level change and littoral sediment budgets can be regarded as boundary conditions for the coastal area of interest.

For sub-decadal prediction of horizontal movements in the upper shoreface, sand exchanges with the lower shoreface (Fig. 2B) are usually ignored because these fluxes are so small that resulting morphological change is negligible: i.e. the annual closure-depth concept. The fluxes of fine sediments (Fig. 2, C and D) are not directly relevant to the upper-shoreface sediment budget because mud deposition there is negligible. For long-term predictions however, none of the internal sediment exchanges depicted in Figure 2 can be ignored. This is because systematic residual fluxes, that are small on the sub-decadal time scale, eventually cumulate through time enough to produce non-negligible (i.e., measurable) morphological changes. Moreover, the changes in morphology of the backbarrier, lower shoreface and upper shoreface cause these three zones to interact dynamically: i.e., the sediment exchanges themselves become influenced by the morphological changes.

As an example consider the impact of sea-level rise on a coast adjacent to a tidal basin compared to that on an uninterrupted coast. Under the assumption that a tidal basin establishes morphodynamic equilibrium in following the rate of sea-level rise (see further) shoreline recession of the adjacent coast can be expressed as follows (STIVE and WANG, 2003):

\[ c_{pr} = \frac{\partial \zeta}{\partial t} \frac{L_p}{H_p} + \frac{\partial \zeta}{\partial t} \frac{A_b}{H_p L_{ac}} \]  \hspace{1cm} (01)

Where:
- \( c_{pr} \) is the rate of profile recession;
- \( \frac{\partial \zeta}{\partial t} \) is the rate of sea-level rise;
- \( L_p \) is the active cross-shore profile length;
- \( H_p \) is the depth of the active cross-shore profile;
- \( A_b \) is the tidal basin area;
- \( L_{ac} \) is the length of the adjacent coast impacted.

In the above equation the first term on the right-handside expresses the Bruun effect (BRUUN, 1962) and the second term expresses the basin effect. The Bruun effect is exceeded by the basin effect as soon as:

\[ A_b > L_p L_{ac} \]  \hspace{1cm} (02)

Typical orders of magnitude for \( L_p \) and \( L_{ac} \) are 1 km and 10 km respectively, so that basin areas larger than \( O \) (10 km²) cause an impact on shoreline recession rates which exceeds the direct impact due to the Bruun effect.

We would therefore argue that the interest into the morphodynamic behavior of tidal basins is not only motivated by their intrinsic characteristics but also by their extrinsic impacts on adjacent coasts. Hindcasting and forecasting the impact of natural and human interventions in tidal basins, based on physical insight, is therefore of utmost importance.

In our considerations of morphodynamic processes it is essential to distinguish spatial and associated temporal scales (cf. DE VRIEND, 1998). On the micro-scale we encounter bed forms, which are a result of interaction between local bed properties and local flow conditions. On the meso-scale we focus on alternating bars or flats and curved channels, which may be the result of a feedback between secondary currents due to centrifugal and Coriolis effects and basin geometry. On the macro-scale we focus on the macro-inlet elements, such as flats, channels, inlet gorge and ebb-tidal delta, the aggregated properties of which may be the result of an interaction between sea, swell and tidal wave properties and the elements’ geometric properties. The mega-scale concerns the interaction between adjacent basins and adjacent coasts including the shoreface on time-scales of centuries to millennia, which bears importance on geological scale (BEETS et al., 1992).

In this contribution we limit ourselves to the meso- and the macro-scales, which play a role on human and/or engineering time-scales (touching the low-order coastal change domain). Micro-scale processes concern the interaction between flow and bed resulting in a variety of bed forms which in turn influence the flow. We assume though that these are of secondary importance in the context of our work. The mega-scale interactions (well into the low-order coastal change domain) play a role on geological scales and are considered out-of-scope for this contribution.

**CHARACTERISTIC PROCESSES OF TIDAL LAGOONS AND COASTAL INLETS**

In this section we introduce and discuss a number of observations and interpretations of characteristic processes in tidal basins and coastal inlets. We distinguish two typical, morphologically near-equilibrium modes of Dutch tidal basins, viz. near square, rectangular tidal lagoons, characteristic for the Waddensea basins, and funnel-shaped estuaries, characteristic for the Scheldt basins. Obviously these two modes are not
representative for the whole variety of basin modes that may be encountered worldwide, but they should be considered representative for two important end-modes.

In principle, tidal lagoons (i.e. the idealized mode of a rectangular basin which is short relative to the tidal wave length) and their associated coastal inlets consist of two major macro-scale morphological entities, viz. the ebb-tidal delta and the flood-tidal delta, connected by the inlet entrance. A large number of different modes of tidal basins are determined by the extent in which the flood-tidal delta is developed over the basin. In the Dutch case the current basins all display a well-developed, equilibrium flood-tidal delta. This is believed to be due to the combination of moderate sea-level rise and availability of sediment from the adjacent barrier coasts (Van Groor et al., 2003). There are many examples of underdeveloped flood-tidal deltas. One example is Red Fish Pass, Florida. Both deltas are characterized by channels and adjacent flats, which are counterparts in the same system.

An analysis of the well-developed flood-tidal deltas of the Waddensea lagoons indicates that the branching behavior is of fractal nature (Cleversinga and Oost, 1999). This is very probably due to an instability mechanism leading to self-organization. Based on a strongly simplified linear stability analysis, using a linear bottom (with decreasing depth from sea to land) as equilibrium (zero order solution), Wang (1992) derived a relation for the most stable mode of channels for tidal lagoons like the Wadden Sea, which suggests that:

- The larger the slope the smaller and the closer spaced the channels;
- The more active (morphologically) the region, the smaller and closer spaced the channels;
- The larger the water depth, the larger and the wider spaced the channels.
- A channel after a bifurcation will have half the length of the original channel.

With this latter rule a fractal structure (Cantor's tree) of the channels can be constructed. Field data of the Dutch Wadden Sea appear to confirm this rule. Especially the rule that channels after a bifurcation become half as deep is well verified by data (Fokkink, 1993).

For estuaries the above mentioned self-organization concept is becoming of better understanding. In estuaries we observe the existence of both channels and bars or flats in case an estuary exceeds a certain width. Observations in the Western Scheldt Estuary suggest that there seems to be a maximum size of the channels, above which more a channel will bifurcate into more channels (Allersma, 1992). However, by analyzing data of more estuaries in the world, Allersma (1994) later concluded that this maximum size is not a universal constant, but seems to be dependent on the size of the estuary and type of sediment in the estuary. The larger the size of the estuary and the finer the sediment the larger the maximum size of the channels, and the number of channels in a cross-section is determined by the width to depth ratio of the estuary.

Recent analytical stability studies such as those by Schuttelalaars and De Swart (1999, 2000) and Seminara and Tubino (1998, 2001) seem to explain the above observations to a certain extent. These studies indicate that when the width scale exceeds a certain tidal length scale, instability mechanisms trigger the existence of alternating bars and channels. Hibma et al. (2003) has used a process-based approach to investigate whether the full non-linear equations confirm such findings. Indeed it is found that process-based models of the 2DH long wave equations, when seeded with small 2-D periodic or random perturbations, predict the development of alternating channels and bars (Figure 3).

Feedback between secondary flow effects and basin geometry is found to explain the existence of both channels and flats as counterparts in many tidal basins and ebb-tidal deltas (Coeveld et al., 2003).

These process-based studies, however, have shown their relevance on smaller time scales but have not proven to be of importance in longer-term development and prediction of tidal basins (see also De Vriend and Ribberink, 1996). In conclusion the author would state as follows. Although we have a rich description of processes in tidal basins and an initial insight into the complex feedback between local processes and morphology, we lack sufficient knowledge to predict the evolution of tidal basins and their interaction with the coast on time scales much longer than the process scale.

MACRO-SCALE INTERACTION BETWEEN TIDAL PROPAGATION AND BASIN GEOMETRY

Because of the insufficient capabilities of process-based modeling to predict behavior of longer-term development of tidal basins and coastal inlets, one often relies on empirical observations. As examples consider the relation between the tidal prism and channel cross-section (cf. O'Brien, 1969 and Einsink, 1990) and between tidal prism and ebb-tidal delta volume (cf. Dean and Walton, 1975). De Vriend and Vinck (1996) discusses the use of such empirical observations in process-aggregated models, which may be used to hindcast and forecast morphodynamic evolution of near-equilibrium basins when forced out-of-equilibrium. A recent, comprehensive model based on this approach and applied to macro-scale elements of tidal basins is central in the last section. Before presenting this, we first introduce the work of Dronkers (1998) who derived a relation between basin geometry and tidal propagation properties which yields zero asymmetry-induced transport for sand. The role of the intertidal areas in the tidal basin appears to be essential here. This is also found in other studies dealing with the interaction between tidal asymmetry and morphology of tidal basins (Friedrichs and Aubrey, 1988; Van de Kreeke and Dunsbergen, 1998).

Dronkers (1998) considers the one-dimensional tidal equations for rectangular, uniform basins and for funnel-shaped...
basins. It is assumed that the tide propagates along the main channel(s) of the basin, such that the large-scale geometry rather than local variations determine its properties. The basin's cross-section is schematized as in a tidal river with an approximate uniform velocity, \( u \), in the flow dominant cross-section, \( A_s \), and approximate zero velocity, \( u \), in the adjacent intertidal storage cross-section. In contrast to tidal rivers the flow-dominant cross-sectional width of tidal basins is generally much smaller than the total cross-sectional storage width, \( b_k \), certainly at high water. The apparent depth \( H_H \) \( A/b_k \) \( H_b/b_k \) is generally smaller at \( H_W \) than at \( L_W \). DRONKERS (1998) derives expressions for ebb and flood duration in terms of cross-sectional variables. It happens to turn out that in both these cases the difference in ebb and flood duration, \( t_e - t_f \), reads:

\[
t_e - t_f \approx H_H^+ H^- - H_b^+ H^-
\]  

(03)

where the superscripts + and - indicate \( H_W \) and \( L_W \) level respectively.

In case ebb and flood duration are equal we may aspect virtual zero asymmetry-induced transport when assuming sand transport to be proportional to a power higher than one of the tidal current speed. This requires that the basin geometry satisfies:

\[
H_H^+ H^- \approx H_b^+ H^-
\]  

(04)

Or

\[
H^+ / H^- \approx \sqrt{b_k^+ / b_k^-} \sqrt{b_e^+ / b_e^-}
\]  

(05)

In many tidal basins the flow-dominant cross-section at \( H_W \) does not differ much from the flow-dominant cross-section at \( L_W \), implying that

\[
\alpha = H^+ / H^- \approx 1
\]  

(06)

In case \( \alpha \) is larger than one the basin is importing, since the basin is shallower than would be necessary for morphologic equilibrium. In contrast, when \( \alpha \) is smaller than one, the basin is exporting. Figure 4 indicates that for all Dutch tidal basins, be it funnel shaped estuaries or rectangular lagoons, the mean cross-sectional geometry satisfies Equation (6) approximately.

MACRO-SCALE AGGREGATED MODELLING OF TIDAL BASINS

An important objective of the present section is to discuss a process-aggregated approach to deal with macro-scale morphodynamic evolution, and explore the application of such an approach to hindcast and forecast the morphodynamic evolution of tidal basins in the light of accelerated sea level rise. This approach, embedded in the ASMITA (Aggregate Scale Morphological Interaction between a Tidal basin and the Adjacent coast) model, has recently been introduced by STIVE et al. (1998, see also VAN GOOR et al., 2003). The approach heavily depends on the concept of the existence of morphodynamic equilibrium in case of constant forcing based on the above-mentioned and to a certain degree explained observations. These findings are used to study morphodynamic evolution of near-equilibrium basins when forced out-of-equilibrium either due to natural or human causes.

Model Concept

The basic idea of the approach is that a tidal inlet can be schematized into a number of morphological elements. For each element a volume can be defined acting as integral state variable. The level of schematization is similar to that of the ebb-tidal delta by DEAN and WALTON (1975). A tidal inlet is thus schematized into three sequentially connected elements:

- the ebb-tidal delta (state variable = integral excess sediment volume relative to an undisturbed coastal bed profile, \( V_e \));
- the total channel volume in the tidal basin (state variable = integral water volume below MLW, \( V_c \));
- the inter-tidal flat area in the tidal basin (state variable = integral sediment volume between MLW and MHW, \( V_f \)).

Following this schematization the adjacent coastal stretches are considered as an external boundary - "the outside world" - which can exchange sediment with the considered inlet system.

As mentioned before, the important hypothesis used in the model concept is that a morphological equilibrium can be defined for each element depending on the hydrodynamic conditions (e.g. tidal prism, tidal range) and morphometric conditions (e.g. basin area). Theoretical arguments for the existence of such equilibrium were given by DRONKERS (1998), but is also supported by various field investigations, which have resulted in empirical relations between state variables and parameters of the governing hydrodynamic and morphometric conditions (cf. EYSSINK 1990). In general, the dry or wet volume \( V_c \) of an arbitrary element \( n \) in a state of morphodynamic equilibrium, has appeared to be highly correlated to the tidal range \( H_{om} \), the tidal prism \( P \) and the basin area \( A_b \):

\[
V_n = V_n(P, H_{om}, A_b)^{0.7}
\]  

(07)

According to this hypothesis no morphological change takes place when all elements in the system are in equilibrium. When one or more elements are out of equilibrium morphological changes will take place tending to restore the system to (a possibly new) equilibrium.
Obviously, sediment transport must accompany morphological changes. It is assumed that suspended load is representative for the transport mode (DI SILVIO, 1989). The sediment transport formulation is basically the same as for any other suspended sediment transport model. However, unlike process-based models describing flow and sediment transport within tidal cycles residual sediment transport $T$ is directly modeled here. This means that the long-term (time scale much larger than tidal period) mass-balance is considered for every morphological element:

$$\pm \frac{dV_n}{dt} = \sum T_{ni}$$  \hspace{1cm} (08)

The left-hand side of this equation represents the erosion rate within the element. Its sign is positive for a wet volume and negative for a dry volume. The right-hand side represents the sum of the transports leaving the element via all connections to other elements including the outside world. The erosion rate is assumed to be proportional to the difference between the local equilibrium concentration and the actual concentration like the process-based model for suspended sediment transport of GALAPPATTI and VREUGDENHIL (1985):

$$\pm \frac{dV_n}{dt} = w_i \cdot A_i \cdot (c_{eq} - c_n)$$  \hspace{1cm} (09)

Herein $w_i$ [m/s] is the vertical exchange rate and $A_i$ [m$^2$] is the horizontal area of the element. Erosion occurs when the actual sediment concentration $c_n$ is smaller than the equilibrium concentration $c_{eq}$ sedimentation occurs when the actual sediment concentration is larger than the equilibrium concentration. Also like any suspended sediment transport model the (long-term residual) sediment transport between two elements is assumed to be of the advective-diffusive type:

$$T_{ni} = Q_{ni} \cdot \frac{c_n + c_i}{2} + \delta_{ni} \cdot (c_n - c_i)$$  \hspace{1cm} (10)

Herein $Q$ [m$^3$/s] is the residual flow rate, $\delta$ [m$^3$/s] the diffusion exchange rate between the two elements and $c_i$ the sediment concentration in the adjacent element.

Substituting (9) and (10) into (8) yields an equation for the sediment concentration for each element. In this way a system of coupled equations for the sediment concentrations in all elements is established. It can readily be solved if the local equilibrium concentration is known. Equilibrium sediment concentration according to most sediment transport theories can be considered as proportional to a certain power of the flow velocity. In an aggregated scale model as considered here, flow velocity is not an available hydrodynamic parameter. However, the ratio between the equilibrium volume and the actual volume of e.g. the channel can be considered as the ratio between the flow velocity and that under equilibrium condition. Therefore the following formulation is used:

$$c_{eq} = c_i \cdot \left( \frac{V_{eq}}{V_n} \right)^r$$  \hspace{1cm} (11)

The magnitude of power $r$ is larger than one, commonly taken as 2 in compliance with a third power for the sediment transport as a non-linear function of the mean flow velocity. Its sign depends on the definition of the element volume, $V_{eq}$, positive for wet volume and negative for dry volume.

The parameter $c_i$ has the dimension of sediment concentration. When the whole system is in equilibrium the sediment concentration in all elements will be the same and equal to $c_i$. Therefore it is called the overall sediment concentration. It is usually prescribed at the outside world as boundary condition if the outside world can be considered as in equilibrium, which physically means that there is no limitation for supply or accommodation of sediment adjacent to the system under consideration. In the cases that we described this is proven to be valid. However, in case the adjacent coast is protected, constrained by headlands, or in general insufficiently dynamic this assumption may not hold. In this case the adjacent coast needs to be introduced as an intrinsic morphological element to the system.

**Response to Sea Level Rise**

Sea level rise induces a special kind of disturbance or forcing for the morphological development. The volume of a morphological element will then not only change due to sedimentation or erosion but also due to the change of the mean sea level. This can be implemented into the model by adding a term to equation (9)

$$\pm \frac{dV_n}{dt} = w_i \cdot A_i \cdot (c_{eq} - c_n) \cdot A_n \frac{d\zeta}{dt}$$  \hspace{1cm} (12)

Herein $\zeta$ is the mean sea level. For a model with a single (channel) element, $V_c$, the following equation for its morphological evolution can be derived:

$$\frac{dV_c}{dt} = w_i \cdot A_i \cdot \delta_{wc} \cdot c_{eq} - \frac{\left( \frac{V_{ce}}{V_c} \right)^r - 1}{\delta_{wc} w_i A_i} + A \frac{d\zeta}{dt}$$  \hspace{1cm} (13)

The volume of the channel $V_c$ will still be changing in time if it is equal to its equilibrium value. When sea-level rise rate is constant a dynamic equilibrium can be established and the equation becomes:

$$V_{ce} = V_c \left( \frac{1}{\delta_{wc} w_i A_i} - \frac{1}{\delta_{wc} w_i A_i} \right)$$  \hspace{1cm} (14)

Equation (14) gives the channel volume in case of a dynamic equilibrium under external forcing of a constant sea level rise. In case of no sea level rise (\(\frac{d\zeta}{dt} = 0\)) equation (14) reduces to a new dynamic equilibrium volume, which is larger than the original equilibrium volume ($V_{ce} > V_c$). Apparently there is a permanent difference between the equilibrium volume with SLR ($V_{ce}$) and the equilibrium volume without SLR ($V_c$). This difference in equilibrium volume is necessary to maintain the demand of sediment that drives sediment imports into the system to such extent that the system does not drown. Equation (14) also shows that the equilibrium volume $V_{ce}$ becomes infinitely large when:

$$\delta_{wc} = \frac{w_i \cdot \delta_{wc} \cdot c_{eq}}{w_i \cdot A_i}$$  \hspace{1cm} (15)

Apparently, there is a maximum SLR rate an inlet can keep pace with. It is referred to as critical sea level rise rate as it indicates the transition between preservation and degeneration of the inlet. The dynamic equilibrium volume (relative to the equilibrium value) as a function of the sea level rise rate (relative to the critical value) is shown in Figure 5.

That there is a critical value of the sea level rise rate at which a tidal inlet will drown can also be found if the system is divided into more morphological elements. Van GOOR et al. (2003) applied the three-elements model as described above to investigate the impact of sea level rise on two tidal inlets in the Dutch Wadden Sea, the ‘Amelander Zeegat’ and the ‘Eierlandse Gat’ (see Figure 6a,b). In this figure the dynamic equilibrium volumes of the inlet elements are plotted as a function of sea level rise rates. The vertical line represents the current state of

![Figure 5. Dynamic equilibrium volume under influence of sea level rise (r=2).](image-url)
required sediment volumes for each of these units may be balance in case of e.g. a tidal basin closure in the sense that these relations one may determine the aggregated sediment tidal prism and ebb-tidal delta volume are well known. Using between tidal prism and channel cross-section and between of channel-flat development. For instance, the relations even before insights have emerged on process-based mimicking basins and coastal inlets on time-scales of decades or centuries. However, this knowledge has bottom evolution in an estuary is now capable of mimicking the phenomena. For instance, we have shown that process-based dominated by sand fractions rather than silt fractions. It is basins with little to no fresh water run-off, and that we assume process-scale. Note that our descriptions are limited to tidal and the associated impact on the adjacent coast justifies that with the highly dynamic nature of tidal basins and coastal inlets and rivers. Ecologically and socio-economically tidal basins and coastal inlets are of high importance. This in combination and coastal inlets finds it easier to adapt to a higher rate of sea level rise. Validation of these findings is difficult since morphological observations of tidal basins under increasing rates of sea level rise are to the authors' knowledge - non-existing. From geological reconstruction of the Holland coast (BEETS et al., 1992), however, validation in a reverse manner is to a certain extent confirmed. When sea level rise rates decreased to less than about 1 m/century the central Holland tidal basins changed from drowned systems to well-developed intertidal basins.

DISCUSSION AND CONCLUSIONS

Morphodynamic evolution of tidal basins and coastal inlets is a complex issue. Probably due to this complexity scientific attention has been less than that spent on uninterrupted coasts and rivers. Ecologically and socio-economically tidal basins and coastal inlets are of high importance. This in combination with the highly dynamic nature of tidal basins and coastal inlets and the associated impact on the adjacent coast justifies that scientific attention is focused on these systems. The present paper aims to make a contribution in this respect. We have started this chapter by describing observations and modeling capabilities of morphodynamic evolution on the process-scale. Note that our descriptions are limited to tidal basins with little to no fresh water run-off, and that we assume sediment transport as related to morphological evolution to be dominated by sand fractions rather than silt fractions. It is shown that we have a rich description of process-scale phenomena. For instance, we have shown that process-based modeling of the interaction between tidal wave propagation and bottom evolution in an estuary is now capable of mimicking the emergence of channels and flats. However, this knowledge has.

Not allowed us to hindcast or predict the evolution of tidal basins and coastal inlets on time-scales of decades or centuries. The value of empirical observations has been established even before insights have emerged on process-based mimicking of channel-flat development. For instance, the relations between tidal prism and channel cross-section and between tidal prism and ebb-tidal delta volume are well known. Using these relations one may determine the aggregated sediment balance in case of e.g., a tidal basin closure in the sense that required sediment volumes for each of these units may be assessed. As a result the sediment source or sink magnitude of the adjacent coast results. However, what lacks is the assessment of the time-scales involved with the adaptations of the macro-scale elements. Before resolving the latter imperfection in the main section of the present chapter we have introduced the elegant and validated derivation by DRONKERS (1998) explaining the relation between tidal propagation and basin geometry. The derivation explains that tidal basins with relatively shallow channels require relatively wide flats and that tidal basins with relatively deep channels require relatively narrow flats to create zero asymmetry-induced sand transport, thus promoting dynamic equilibrium.

Gratefully accepting these semi-empirical observations of dynamic equilibrium we have expanded on an aggregated-scale model with which we are able to assess the macro-scale response of tidal basins and ebb-tidal deltas to forcing by an accelerated rise of sea level. It is shown that strong changes in intertidal areas will result when sea level rise rates accelerate.

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LITERATURE CITED


