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ADVANCES IN MORPHODYNAMICS OF COASTS AND LAGOONS

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Abstract: Coastal evolution due to natural and human-induced causes or factors can be rather variable on quite different temporal and/or spatial scales. Our capability to understand and especially predict this variability on the longer time scales is still limited. This can lead to misinterpretation of coastal change information, which hampers informed decision-making and the subsequent functional design process related to (soft) engineering interventions. Research progress aimed at understanding observations of long-term coastal evolution is important to support and improve the decision-making and functional design process. In this contribution an attempt is made to review recent research methodological advances and to illustrate the concept with one exemplary result.

Key words: Coastal tract, Coastal morphology, Hierarchical modelling, Tidal basin, Sea-level rise

1. INTRODUCTION

Evolution of coastal morphology over centuries to millennia (low-order coastal change) is relevant to chronic problems in coastal management (e.g., systematic shoreline erosion). 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 (where present). 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. After a description of the coastal tract cascade the underlying principles are illustrated by presenting a model approach that describes the impact of relative sea-level rise on tidal basins and their adjacent coasts.

2. COASTAL TRACT AND CASCADE HIERARCHY

2.1 THE PRACTICAL IMPERATIVE

Low-order coastal change involves morphological evolution on a geological time scale (order 10^3 years) that has significance on coastal management time scales (10^0 to 10^2 years). Issues of morphological stability and change in coastal management largely involve the need to predict and control the position of the shoreline. At any location along the coast, the shoreline position is governed by gains and losses of sediments in the alongshore and across-shore directions (i.e., the local sediment budget), and by tendencies toward flooding or emergence of the backshore due to changes in sea level. Sea-level change also mediates across-shore sediment displacements, and can influence alongshore sediment budgets through effects on the hydrodynamic conditions caused by changes in the effective bathymetry experienced by nearshore wave and current fields.

Prediction of shoreline change adopts different approaches, depending on the space and time scale over which predictions are required. For short-term (sub-decadal) coastal change (event and synoptic-scale changes occurring over hours through seasons to years), the focus is generally on the local sediment dynamics. These affect the shoreline planform and the across-shore profile (e.g., shoreline and profile models) in response to fluctuations in environmental conditions (i.e., the wave climate, littoral sediment budgets, sea level and the effects of anthropogenic activities). Theoretical and empirical approaches to these sub-decadal time scales generally focus on changes to the upper shoreface (defined loosely as the active zone; cf. STIVE and DE VRIEND, 1995), which correlate with shoreline movements. These changes are moderated by littoral sediment budgets and by sediment 'production' via shoreline erosion cutting into onshore sand reserves (e.g., eroding dunes or cliffs), or through artificial nourishment of beaches.

The practical imperative for long-term prediction (decades or longer), requires an expanded scope that also includes the lower shoreface and the interaction between the shoreface and backshore environments (Fig. 1). The upper shoreface has cross-shore length scales that are typically two to three orders of magnitude less than for the lower shoreface (depicted in Fig.1). This scale difference means that changes on the lower shoreface are associated with disproportionately larger changes on the upper shoreface, due to mass continuity for sediment exchanges between the two zones (ROY et al., 1994; COWELL et al., 1999a). The upper shoreface is subject to a similar interaction with the backshore, which comprises a morphologically active zone located between the upper shoreface (ocean beach) and the mainland. This zone may variously include dunes, washover surfaces, flood-tide deltas, lagoonal basin, tidal flats (Fig. 1A), mainland beaches (Fig. 1B) and fluvial deltas (Fig. 1C). Each of these may be present or absent, depending on local conditions, especially the regional substrate slope (ROY et al., 1994; COWELL et al., 1995).

The sediment exchanges depicted by the arrows in Fig. 1 occur in principle during any average year and on all time scales longer than this. These exchanges are summarised schematically in Fig. 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 (HALLERMEIER, 1981; NICHOLLS et al., 1998). 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 Fig. 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.

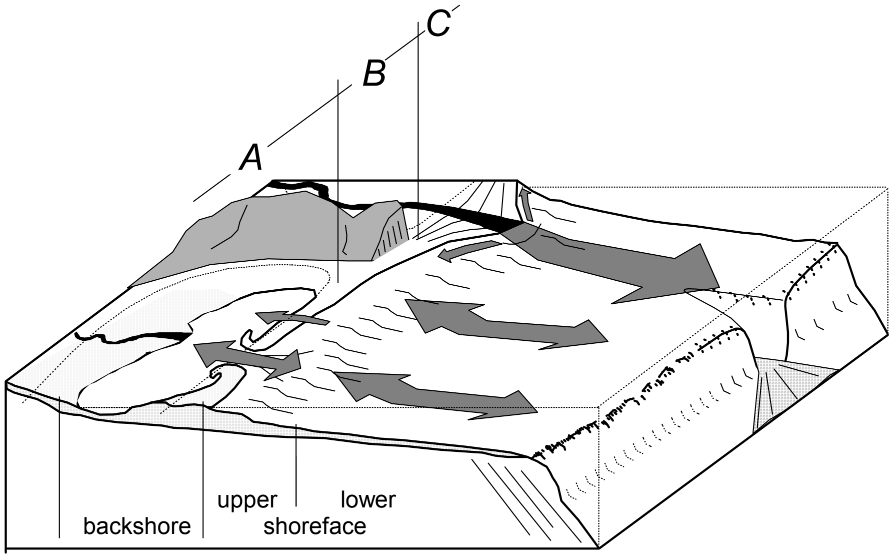

Fig. 1 Physical morphology encompassed by the coastal tract (after COWELL et al., 2003)

2.2 DEFINITION OF THE COASTAL TRACT

We introduce the coastal tract as a composite morphology that is a physically identifiable feature. The composite form however also underpins a more abstract framework for aggregation methods (i.e., the coastal-tract cascade, outlined below). Identification of the coastal tract provides a) the rationale for spatial extension of coastal-change models (i.e., to include the lower shoreface and backbarrier as intrinsic components), and b) an explanation for end users of why these broader considerations are essential ingredients to long-term coastal management.

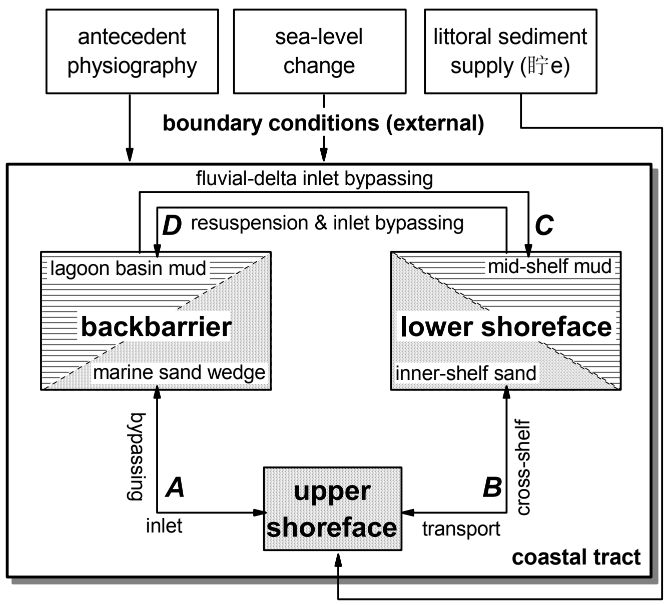

Fig. 2 Mechanisms steering the location of the upper shoreface (after COWELL et al., 2003)

We formally define the coastal tract in both physical and abstract terms. Under the physical definition, the coastal tract is a spatially contiguous set of morphological units representative of a sediment-sharing coastal cell.

Although we propose the tract as a natural physiographic feature, its composite nature means that its actual form can vary geographically in terms of its constituents (cf. regions A, B, and C in Fig. 1). Thus, an individual coastal tract has meaning only in the context of a specific engineering, management or research problem. That is, the tract also is an abstract entity (or meta-morphology) constructed (or templated) for analysis and prediction of a specific site or region in nature, on an associated time-scale.

The physical definition contains three key terms: a) the morphological units, which are constituents within our formal framework (that we term the coastal-tract cascade) for partitioning and aggregation of processes within the tract on the basis of scale; b) sediment-sharing systems, which form the scale-related defining entities of the coastal-tract cascade, and c) the coastal cell, which defines the coastal tract in relation to alongshore homogeneity of morphology and processes. We elaborate on each of these three concepts in the following.

2.3 COASTAL-TRACT CASCADE

We introduce the coastal-tract cascade to manage process aggregation. The tract cascade is thereby the means of separating out low-order coastal change from morphodynamics on smaller space and time scales. The contiguous morphological units referred to in the coastal-tract definition are associated with an intermediate morphodynamic scale in the cascade hierarchy. The contiguity relates to the coupling of adjacent morphologies within the coastal tract. The coupling manifests as coastal change (i.e., movements in the shoreline and changes in elevation of the bed).

2.3.1. Physical Tract Constituents

In terms of the simple physical definition of the tract as a composite morphology, its constituent morphological units are arranged perpendicular to the shoreline within a coastal cell. The across-shore sequence is: mainland beach (or fluvial delta), estuary-lagoon, barrier-beach-dune complex, upper shoreface, lower shoreface and continental slope (Fig. 1A). The fluvial delta and estuary-lagoon may be absent: e.g. in the case of a steep continental margin where the mainland beach (with or without dunes) fronts directly onto the shoreface (Fig. 1B).

Only the lower shoreface departs from the conventional form, in that it is generalized to extend seaward to the edge of the continental shelf (Fig. 3). This extended concept encapsulates the regions traditionally (albeit inconsistently) termed the shoreface, as well as the inner, mid, and outer continental shelf (COWELL et al., 1999a). A lumped definition of the lower shoreface is essential if coastal change is to be understood across a sufficiently large range of time scales to enable practical prediction of low-order coastal behavior, and hence higher-order behaviors within a process cascade. Time scales for morphological change (Fig. 3) decrease by several orders of magnitude between the upper shoreface and the shelf edge (NIEDORODA et al., 1995). The extended lower-shoreface concept also admits the inclusion of fine sediments (i.e., mud) into the problem through maximum aggregation of sedimentation processes: i.e., across the entire contental-shelf surface. The physical representation of the coastal tract as a composite feature is simple because the constituents are traditional text-book morphodynamic systems. These systems however collectively contain a large amount of complexity and are also numerous. Paradoxically therefore, the simple physical representation of the coastal tract is too complex as a basis for robust and transparent models of low-order change. Greater aggregation of the constituents is required to achieve this, for which reason we turn to the more abstract concept of the tract as a meta-morphology.

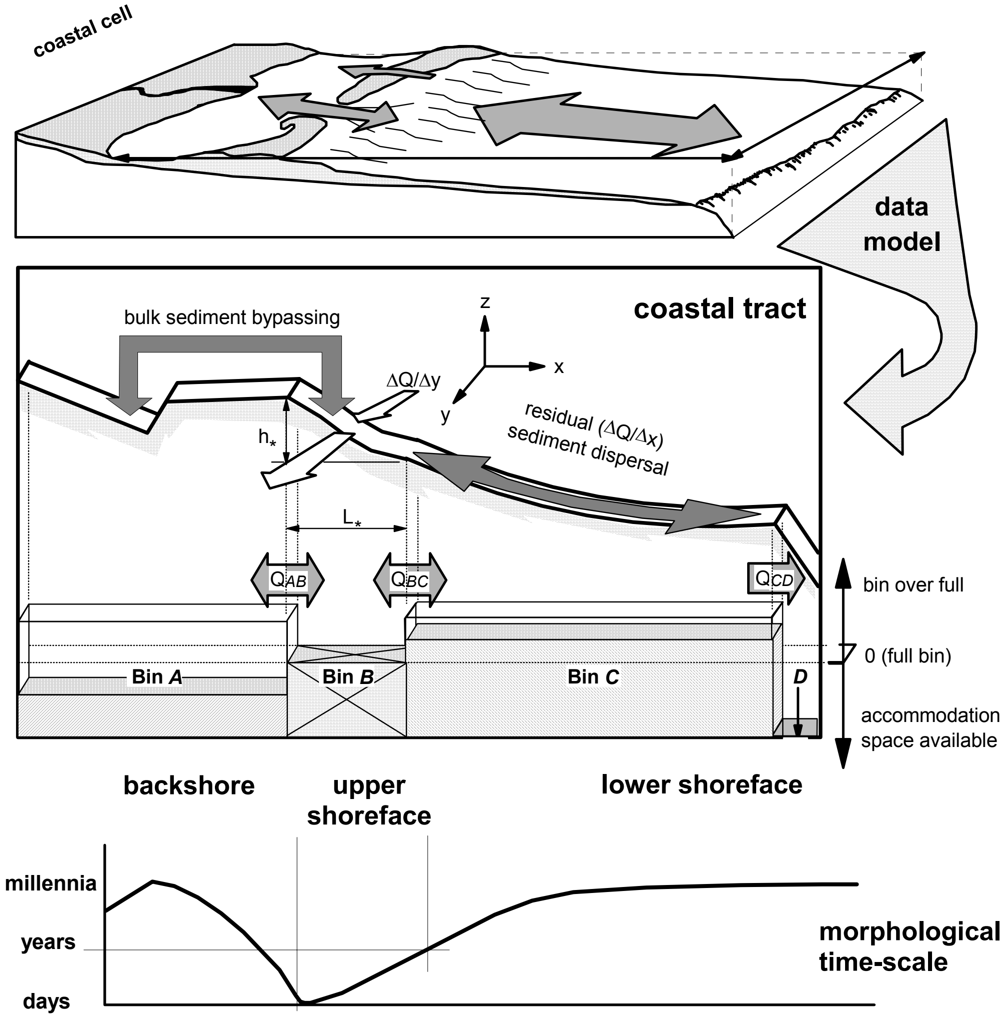

Fig. 3 Definition sketch of the coastal tract and sediment-accommodation space as bins in a sediment-sharing system (COWELL et al., 2003)

2.3.2. Tract Cascade Hierarchy

The coastal tract is a metamorphology forming the lowest-order level in a hierarchy of processes and morphologies (Fig. 4). The hierarchy involves a process cascade in which coastal behavior at any intermediate level results from the residual effects of higher-order processes, while constrained by the effects of lower-order systems in the cascade (CHORLEY et al., 1984). These constraints constitute internal boundary conditions that operate in addition to the external boundary conditions. The coastal tract contains and integrates the effects of all higher order morphodynamic systems in the cascade.

These ideas follow hierarchy theory, according to which nature can be partitioned into 'naturally occurring' levels that share similar time and space scales, and that interact with higher and lower levels in systematic ways (HAIGH, 1987; CAPOBIANCO, et al., 1998). Each level in the hierarchy sees the lower levels as extrinsic constraints or boundary conditions, and the higher levels as intrinsic (sub-scale or 'sub-grid') processes. At successively higher levels in the hierarchy, these intrinsic processes may lose their relevance for lower levels, turning them effectively into some combination of unimportant variations ('noise') and sub-scale processes that must be generalized for representation at the scale of interest (Fig. 4). The criterion by which we partition the cascade is that each level forms an internally sediment-sharing system.

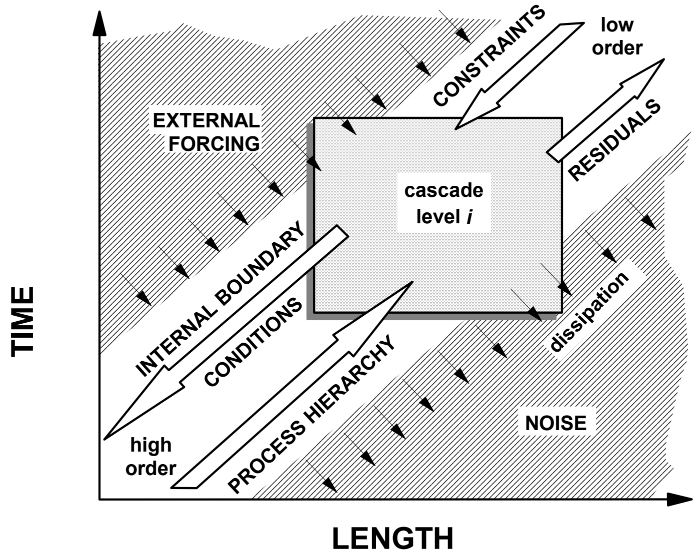

Fig. 4 Coastal-tract cascade. Sub systems in the process hierarchy and internal and external constraints. Lower order systems have the largest length and time scales.

Based on observations of large-scale coastal behaviour, for example throughout the Holocene (BEETS et al, 1992; ROY et al, 1994; COWELL et al., 2001), we consider a coastal-tract system to form the first-order level in the hierarchy (Table 1). At this level, the coastal tract behaves as a single unit that adjusts internally in response to a) gross environmental factors, such as relative sea-level rise, coastal ocean climatology and external sediment sources (such as river input) or sediment sinks (such as submarine canyons), and b) lower level constraints, such as a geologically inherited substrate and tectonic movements (the zero-order system).

Table 1 Coastal-tract systems and system scales (Cowell et al., 2003).

system	system-scale	time scale	space scale
zero order	meta-scale	Quaternary Period ($\geq 10^4$ yrs)	tract environment
first order		Holocene Epoch (10^2 -10^3 yrs)	coastal tract
second order	macro-scale	late Holocene Age (10^1 - 10^3 yrs)	morphological complex
third order		years to decades	morphological unit
fourth order	meso-scale	seasons to years	morphological element
fifth order		days to seasons	
≥fifth order	micro-scale	seconds to days	sub-grid phenomena

Second-order systems we term morphological complexes because they comprise aggregations of the various text-book morphologies that constitute the third-order systems (morphological units in Table 1). We introduce second-order systems to minimise aggregation errors and loss of model transparency (COWELL and THOM, 1994) that might result from integrating too much spatial and functional complexity across a single level in the cascade. The backbarrier is a morphological complex that may include dunes, estuaries (or tidal lagoons) together with the fluvial deltas entering them (Fig. 1A), and the coastal lowlands formed as lagoons become sediment filled (Fig. 1C). The upper shoreface is a morphological complex that may incorporate river and ebb-tide deltas as well as surf zone morphologies. We say "may", because the way we define morphological complexes depends on the site-specific problem and is part of the templating process involved in development of the data-model.

Similarly, we introduce a fourth-order level (morphological elements in Table 1), on even smaller temporal and spatial scales, with which we account for the spatial and functional complexity of the more traditional morphological units by distinguishing their main internal elements. Examples include,

- for beaches – the beach face and surfzone bars;
- for a tidal inlet – the gorge plus the bars and channels on the ebb- and flood-tide delta;and
- for the inner tidal basin – the channels, lower and higher tidal flats, plus fringing salt marshes or mangroves.

Whereas here we consider first, second and third order systems with decades as the smallest time scale, the fourth-order system displays relevant behaviour on the sub-decadal time scale. On smaller time scales (days to seasons), fifth-order systems may be distinguished, such as beach states on the upper shoreface. For the low-order processes (first to third), these fifth-order systems can be considered as noise (Table 1). These distinctions provide a formal basis for aggregation methods used in development of models of coastal change in general, and for low-order change models in particular, as illustrated by the example presented in the following section.

3. AGGREGATED-SCALE MODELLING OF SEA-LEVEL RISE IMPACT ON TIDAL BASINS

3.1 INTRODUCTION

As explained above, coastal inlets and tidal basins provide an important sediment accommodation space for a situation in which the relative mean sea level is rising (the combined effect of sea level rise and bottom subsidence). Under present moderate rates of relative sea level rise a dynamic equilibrium between forcing and morphological evolution is commonly assumed, but whether this may continue to exist under accelerated relative sea level rise is uncertain. Obviously, morphological interaction between the adjacent coastal stretches, the lagoons and their submerged deltas takes place at various spatial and temporal scales, according to the intrinsic dynamics and external forcing factors.

To gain more insight in these morphological interactions the coastal/basin interaction model ASMITA (Aggregated Scale Morphological Interaction between a Tidal basin and the Adjacent coast; STIVE et al. 1998) is further developed. The main objective of this section therefore concerns a fundamental analysis of ASMITA in relation to external forcing due to relative sea level rise. This is done in compliance with the internal and external uncertainties and results in determining to what amount of external forcing tidal basins can maintain their dynamic equilibrium.

3.2 THE ASMITA APPROACH

Based on existing field information and on generic modelling experience regarding coastal inlets and tidal basins the ASMITA modelling approach has been developed, focusing on the 'residual' interaction occurring on long term scales. Because of the time-scales of interest, the geophysical elements of the coastal fringe are considered at an aggregated scale, e.g. single-inlet lagoons are schematised into two or at the most three spatial units such as ebb-tidal delta, channel area and (high and low) flats area. The approach is an aggregation and an extension of a model formulation for tidal basins (WANG et al., 1998). The aggregation concerns the fact that we characterise the system elements by only one state variable (viz. a total wet or dry volume).

The most important hypothesis used in the model concept is that an equilibrium state can be defined for each element depending on the hydrodynamic condition, for instance, tidal prism P and tidal range H. An empirical relation is required for each element to define the morphological equilibrium state as a function of such conditions in a situation of constant sea level rise rate. These relations are derived from literature (cf. EYSINK, 1990). To investigate the behaviour of a tidal inlet system under the external forcing of changing relative sea level rise rates we start to analyse a simple single element model and expand it to a more complicated three-element model.

3.3 SINGLE ELEMENT MODEL

As the channel element of the tidal inlet system is a central element physically, it is taken as the basis for the single element model. In this model we presume that channel evolution is dominated by diffusive transport between channel and the 'outside world' (see Fig. 5) and that this 'outside world' is always in a state of equilibrium. The external 'outside world' thus represents adjacent elements such as ebb-tidal delta, adjacent coasts and tidal flats.

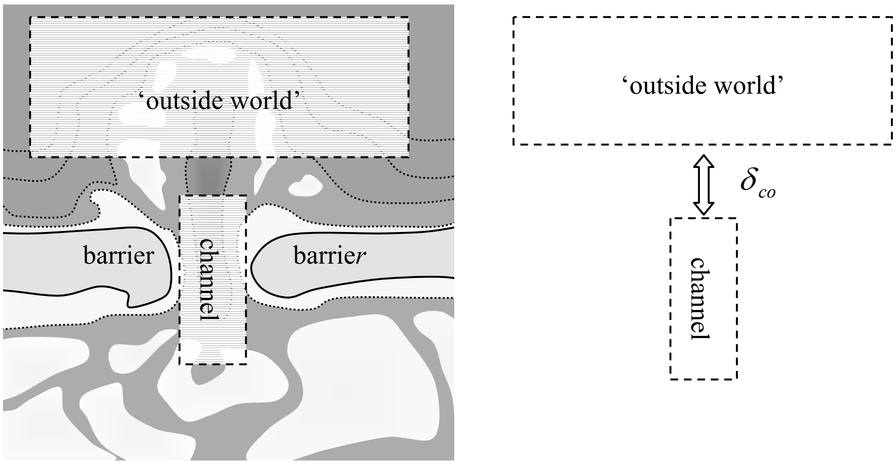

Fig. 5 Single element model

In this channel-outside world model, the equilibrium state for the channel is derived from the equilibrium relation for the channel volume which depends on the tidal prism:

$$V_{ce} = f(P) \tag{1}$$

where V_{ce} stands for the equilibrium volume for channel element (m^3) and P the tidal prism (m^3). In the model approach leading to stability limits, changes in tidal prism are considered as a secondary effect and therefore neglected.

A disturbance of the channel volume causes a change in the local equilibrium concentration of the channel (c_{ce}), which depends on the actual volume V_c (see Stive et al., 1998). The equation determining the local equilibrium concentration is equal to:

$$c_{ce} = c_E \cdot \left(\frac{V_{ce}}{V_c} \right)^n \tag{2}$$

The overall equilibrium concentration c_E, which is the long-term averaged concentration of the system in equilibrium, also applies to the concentration at the boundaries and is assumed time-invariant. The difference between the actual concentration c_c and the local equilibrium concentration c_{ce} results in changes in the morphology. Morphological changes are governed by:

$$\frac{\partial V_c}{\partial t} = w_s \cdot A_c \cdot (c_{ce} - c_c) \tag{3}$$

where w_s is the vertical exchange coefficient (m/s) and A_c the horizontal area (m^2) of the channel. As a result of the disturbance, sediment is exchanged between the various elements. Assuming the diffusive transport dominating and considering the 'outside world' as the only neighbour of the channel, the mass balance of sediment for the channel is expressed by:

$$\delta_{co} \cdot (c_c - c_E) = w_s \cdot A_c \cdot (c_{ce} - c_c) \tag{4}$$

where $\delta_{co} \cdot (c_c - c_E)$ represents the horizontal diffusive exchange between the channel and the 'outside world'.

From equation (4) we can derive:

$$c_c = \frac{\delta_{co} \cdot c_E + w_s \cdot A_c \cdot c_{ce}}{\delta_{co} + w_s \cdot A_c} \tag{5}$$

which gives:

$$c_{ce} - c_c = \frac{\delta_{co}}{\delta_{co} + w_s \cdot A_c} \cdot (c_{ce} - c_E) = \frac{\delta_{co} \cdot c_E}{\delta_{co} + w_s \cdot A_c} \left(\left(\frac{V_{ce}}{V_c} \right)^n - 1 \right) \tag{6}$$

Equation (6) in combination with equation (3) yields:

$$\frac{\partial V_c}{\partial t} = \frac{w_s \cdot A_c \cdot \delta_{co} \cdot c_E}{\delta_{co} + w_s \cdot A_c} \left(\left(\frac{V_{ce}}{V_c} \right)^n - 1 \right) \tag{7}$$

The imposed external forcing on our 'single element system' is considered as linear, which means the sea level rise has a constant rate. The following expression is taken for the external forcing:

$$VI = A_c \cdot \frac{\partial \zeta}{\partial t} \tag{8}$$

where VI = sea level rise induced *Volume Increase* (m^3/year) and $\partial \zeta / \partial t$ = relative sea level rise or *SLR* (m/year).

When we include the sea level rise in equation (7) we derive:

$$\frac{\partial V_c}{\partial t} = \frac{w_s \cdot A_c \cdot \delta_{co} \cdot c_E}{\delta_{cout} + w_s \cdot A_c} \left(\left(\frac{V_{ce}}{V_c} \right)^n - 1 \right) + A_c \cdot \frac{\partial \zeta}{\partial t} \tag{9}$$

Equation (9) describes the morphological behaviour of the system. Either the system erodes ($\partial V_c / \partial t > 0$) or the system accretes ($\partial V_c / \partial t < 0$) or the system is in dynamic equilibrium ($\partial V_c / \partial t = 0$). In case of dynamic equilibrium we assume that the considered element(s) can undergo a self-organisation leading to a (quasi-) equilibrium morphology for each element under relatively constant hydrodynamic forcing conditions. If we consider a situation of (quasi-) equilibrium the volume changes of the element in time at the equilibrium stage should be zero. This results in:

$$\frac{\partial V_c}{\partial t} = 0 = \frac{w_s \cdot A_c \cdot \delta_{co} \cdot c_E}{\delta_{cout} + w_s \cdot A_c} \left(\left(\frac{V_{ce}}{V_c} \right)^n - 1 \right) + A_c \cdot \frac{\partial \zeta}{\partial t} \tag{10}$$

which may be solved for V_c:

$$V_c = \frac{V_{ce}}{\left(1 - \frac{\partial \zeta}{\partial t} \cdot \frac{\delta_{co} + w_s \cdot A_c}{w_s \cdot \delta_{co} \cdot c_E} \right)^{\frac{1}{n}}} \tag{11}$$

Equation (11) gives the dynamic equilibrium under the external forcing of a constant sea level rise. Apparently, the element can adjust itself and change its volume to satisfy a new equilibrium state. However, limitations to this adaptive capacity of the element exist. This is shown in equation (12), because equation (11) only holds for:

$$\frac{\partial \zeta}{\partial t} < \frac{w_s \cdot \delta_{co} \cdot c_E}{\delta_{co} + w_s \cdot A_c} \tag{12}$$

If we increase the external forcing $\partial\zeta/\partial t$ up to this limit, the volume of the element goes to infinity and the dynamic equilibrium stage can not be reached any longer. For a fictitious case this is shown in Figure 6. In this case, the sediment entrapment in the channel element gradually becomes insufficient and the increase in wet volume is due to the rate of sea level rise. Furthermore, it can be seen that before the theoretical limit is reached a considerable increase of volume is needed to reach a new dynamic equilibrium state.

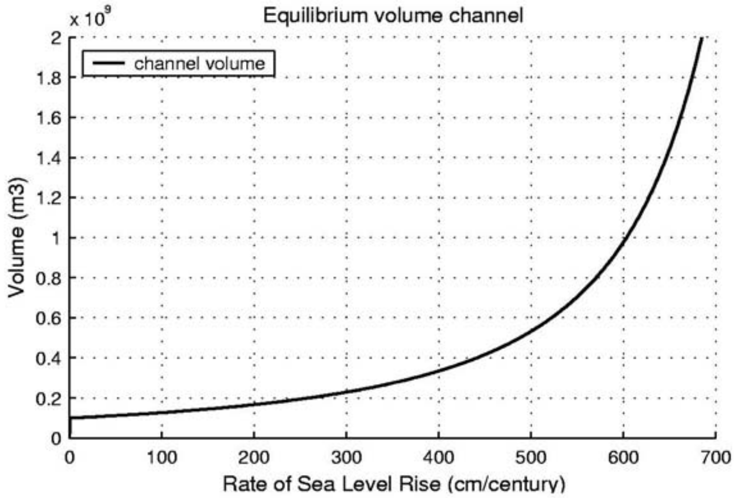

Fig. 6 Dynamic equilibrium volume channel element

3.4 THREE-ELEMENT MODEL

Here the ASMITA model is extended with two more elements, viz. the ebb-tidal delta and the flat elements. See Fig. 7.

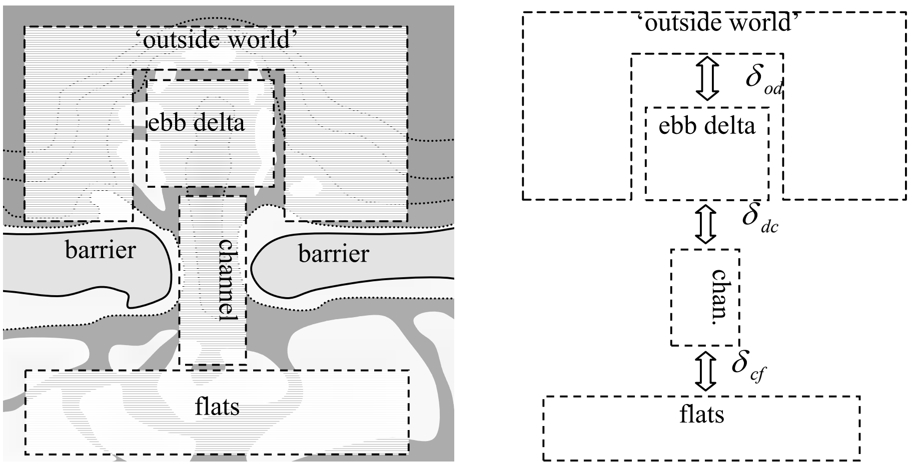

Fig. 7 Three-element model

The parameter governing the equilibrium sediment volume of the ebb-tidal delta is the tidal prism:

$$V_{de} = f(P) \tag{13}$$

The equilibrium concentration and the morphological change of the ebb-tidal delta area are given by:

$$c_{de} = c_E \cdot \left(\frac{V_{de}}{V_d} \right)^n \; ; \; \frac{\partial V_d}{\partial t} = w_s \cdot A_d \cdot (c_d - c_{de}) \tag{14,15}$$

The mass balance of sediment for the ebb-tidal delta is given by:

$$\delta_{od} \cdot (c_E - c_d) - \delta_{dc} \cdot (c_d - c_c) = w_s \cdot A_d \cdot (c_d - c_{de}) \tag{16}$$

where δ_{dc} represents the horizontal exchange rate between the delta and the channel and δ_{od} represents the horizontal exchange rate between the 'outside world' and the ebb-tidal delta. The sediment exchange between the channel and the flat is described similarly. Thus the mass balance of sediment for the channel is:

$$\delta_{cf} \cdot (c_c - c_f) - \delta_{dc} \cdot (c_d - c_c) = w_s \cdot A_c \cdot (c_{ce} - c_c) \tag{17}$$

When we use the equations for the ebb-tidal delta, the channel and the flat we may derive the following three coupled first-order differential equations for the ebb-tidal delta volume, the channel volume and the flat volume, respectively:

$$\frac{\partial V_d}{\partial t} = w_s \cdot A_d \cdot (c_d - c_{de}) - A_f \cdot \frac{\partial \zeta}{\partial t} \tag{18}$$

$$\frac{\partial V_c}{\partial t} = w_s \cdot A_c \cdot (c_{ce} - c_c) + A_c \cdot \frac{\partial \zeta}{\partial t} \tag{19}$$

$$\frac{\partial V_f}{\partial t} = w_s \cdot A_f \cdot (c_f - c_{fe}) - A_f \cdot \frac{\partial \zeta}{\partial t} \tag{20}$$

where c_{de}, c_d; c_{ce}, c_c and c_{fe}, c_f are the actual and equilibrium concentrations of the ebb-tidal delta, the channel and the flat. For a situation of dynamic equilibrium, the volume change in time of the delta, channel and flat should be zero ($\partial V_d / \partial t = 0$; $\partial V_c / \partial t = 0$; $\partial V_f / \partial t = 0$). For this three-element model we can derive three limits for the relative sea level rise ($\partial \zeta / \partial t$), respectively for ebb-tidal delta, channel and flat. The smallest of the given limits is determinative for the system.

3.5 CASE STUDY

In the light of the theoretical consideration above, an actual inlet is selected for a case study. The selected inlet, the 'Amelander Zeegat' situated in the Dutch Wadden Sea between the barrier islands Terschelling and Ameland, is assumed to be in dynamic equilibrium under present rate of sea level rise. For the analysis the following mean data is used (derived from BIEGEL, 1992).

Table 2 Element characteristics Amelander Zeegat

Element	Area (m^2)	Volume (m^3)	w_s (m/s)	δ_{od} (m^3/s)	δ_{dc} (m^3/s)	δ_{cf} (m^3/s)	c_E (-)
Ebb-tidal delta	7.47e7	1.31e8	1e-5				
Channel	9.83e7	3.02e8	5e-5	1500	1500	1000	0.0002
Flat	1.78e8	1.20e8	1e-4				

The values for the outside equilibrium concentration, horizontal exchange rate and vertical exchange rate are derived from BUIJSMAN (1997) for the calibration of the Friesche Zeegat and might need further research. Using these values for a stability analysis we obtain dynamic equilibrium volumes as presented in Fig. 8:

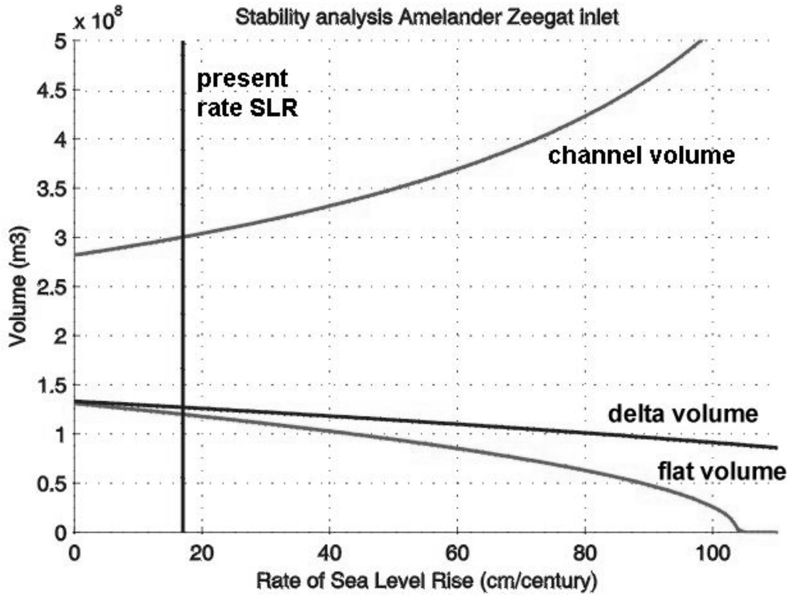

Fig. 8 Dynamic equilibrium volumes and stability limit

In Figure 8 the current state of dynamic equilibrium (under the external forcing of 17 cm sea level rise per century) is represented by the solid vertical line. With an accelerating rate of sea level rise (moving further to the right in the in plot) we see the dynamic equilibrium volume of the channel element increasing and the dynamic equilibrium volumes of the flat and ebb-tidal delta element decreasing.

Apparently for the modelled case it can be shown that an accelerating relative sea level rise results in the gradual drowning of the tidal basin. Nevertheless, the system may find a new dynamic equilibrium state. However, if the accelerated sea level rise exceeds the threshold value (the stability limit), the system will no longer regain a state of dynamic equilibrium. For the Amelander Zeegat case, the threshold value lies at a sea level rise of 105 cm per century. One has to keep in mind this limit is very sensitive to a change of area, horizontal exchange rate and outside world equilibrium concentration (sediment supply). Because the latter two variables represent large uncertainties it seems useful to accompany model results with an estimate of their matching reliability. A probabilistic approach might provide this reliability.

4. CONCLUSION

Evolution of coastal morphology over centuries to millennia (low-order coastal change) is relevant to chronic problems in coastal management (e.g., systematic shoreline erosion). 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 (where present). 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.

As an exemplary illustration of a data-model and the associated templating a prediction model is introduced for the long-term morphology of coastal inlets and tidal basins under the influence of accelerating relative sea level rise. The adopted model approach (ASMITA) considers the geophysical elements of a tidal inlet system at an aggregated scale, schematising the system into maximally three spatial units such as ebb-tidal delta, channel and flat. Analysis of model results shows that under the external forcing of a constant relative rate of sea level rise the system may tend towards a new dynamic equilibrium that depends on the sea level rise rates. However, if the rate of sea level rise exceeds a certain threshold value (stability limit), the system will no longer regain a state of dynamic equilibrium and the tidal basin will drown.

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