CT5303

# Coastal inlets and Tidal basins

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Subfaculty of Civil Engineering Hydraulic and Offshore Engineering Division



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## **Coastal inlets and Tidal basins**

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"In Greek mythology Ocean, the personification of the waters flowing along the coasts of the Earth, was a powerful and good man. Daily he gave birth and shelter to the sun and the moon and all the gods of the sea called him "father". His wife was Tethys, personifying the fertility of the sea. She gave him 3000 children, all the waves of the ocean."

These lecture notes intend to highlight the importance of some particular entities of the coastal zone, viz. coastal inlets and tidal basins, which may only be on "the periphery of the periphery" of the ocean, but which -due to their fertile conditions- play an important role in both the global function of the ocean and in the -hopefully sustainable- human exploitation of the earth's space.

These notes serve to support and supplement the lecture course on "COASTAL INLETS AND TIDAL BASINS". This is a new (per the academic course year 1998-1999) fifth-year course for MSc graduate students following the Hydraulics Engineering direction of the Faculty of Civil Engineering and Geosciences of Delft University of Technology. This course focuses on the interrelation between physical flow- and transport-phenomena and the morphodynamics of coastal inlet and tidal basin systems. The objective is to provide both insights into the phenomenological, theoretical and applied aspects for civil engineering MSc students. It builds upon the MSc courses treating flow, wave and transport dynamics, and is complimentary to the course on "COASTAL MORPHOLOGY". Some of the course elements are derived from the MSc graduate lecture notes on "PHYSICS OF COASTAL SYSTEMS" given at the Faculty of Physics and Astronomy of the University of Utrecht by Job Dronkers, assisted by Marcel Stive.

The course responsibility rests with Huib De Vriend and Marcel Stive. The course is given by the following lecturers, who are all active members of the Netherlands Centre for Coastal Research: Huib De Vriend (University of Twente and Delft University of Technology), Marcel Stive (WL | Delft Hydraulics and Delft University of Technology), Job Dronkers (Rijkswaterstaat | National Institute for Marine and Coastal Management and University of Utrecht), Zheng-Bin Wang (WL | Delft Hydraulics and Delft University of Technology). These lecture notes were composed with the valuable assistance of Birgit Cloin (WL | Delft Hydraulics and Delft University of Technology) and Ap Utrecht (University of Technology), Edwin Elias (Delft University of Technology) and Anneke Hibma (WL | Delft Hydraulics and Delft University of Technology).

Although there exist several valuable course monographs and books on the topic of coastal inlets, tidal basins and estuaries we feel that there are no complete standard teaching books for the purposes that we have in mind, as yet. It is our expectation that the present notes may develop towards that end. The present form and format of these notes is not final and far from perfect, both in language and contents. We feel that this is a matter of evolution, strongly steered by the experience of both teaching and examination.

Deift, October 1998.

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#### 1. Introduction

#### **1.1 Problem definition**

Tidal inlets and their associated basins (lagoons) are a common feature of lowland coasts all around the world. A significant part of the world's coastlines is formed by barrier island coasts, and most other tidal coasts are interrupted by estuaries and lagoon inlets. These tidal systems play a crucial role in the sediment budget of the coastal zone and thus influence the long-term coastal evolution.

From a morphological point of view, tidal inlets form highly dynamical systems, which are interlinked with the adjacent coast and the tidal basin or backbarrier area to which they give access. Often, the natural morphodynamic behaviour interferes with unnatural constraints (e.g. coastal defence works) and with the effects of human utilisation (e.g. sand mining).

Estuaries and tidal lagoons attract a variety of human activities, such as navigation, recreation, fishing and aquaculture, economical activity in the border zone, sand mining, land reclamation and in some cases hydrocarbon mining. On the other hand, many estuaries and lagoons form the basis of highly valuable and sometimes unique ecosystems. They function as nursery grounds for many species and as resting and feeding grounds for many others. Hence human activities which affect the properties of such a system, or put the environmental functions otherwise at risk, may have important environmental implications. For the proper management of these systems, it is therefore most important to be able to predict the impacts of such activities.

At a larger scale, a deficit of sediment in the backbarrier area, due to sea level rise, for instance, can have major effects on the sediment budget of the coastal zone. The outer deltas of the inlets seem to act as sediment buffers, but the ultimate source of the sediment which goes to the backbarrier area is the coast. Since long-term coastal zone management should include sediment management, the capability to predict the large-scale exchange of sediment is of great importance to ICZM (Integrated Coastal Zone Management). The inlets and their outer deltas play a key role in this exchange.

These lectures intend to develop an insight into the physical functioning of coastal inlets and tidal basins, such that possible engineering interventions are executed from a sustainable, holistic and integrated management perspective. The emphasis will be on inlet and basin systems which are subject to a mixed tidal and wave forcing, with negligible fresh water run-off, typical examples being the Wadden Sea and the Zeeland inlets and basins. Foreign examples are typically barrier-inlet coasts, such as found abundantly along the east-coast of the USA. This implies that fresh and salt water dynamics are not of strong relevance, and that the morphodynamics of these systems are largely determined by the interaction between the coarser sediment (fine to medium sands) and the tide and wave induced water motions.

#### **1.2 Predictability and scales**

Tidal inlet and basin morphology is the result of a stochastically forced, non-linear interaction between the water and sediment motion and the bed topography. Understanding and predicting their functioning is one of the major scientific challenges of coastal physics. It implies dealing with a wide range of space and time scales, with complex multi-scale interactions of the constituent processes, and with strong, partly stochastic variations of the forcing. Besides complexity, the possibility of limited predictability has to be taken into account, because these systems seem to satisfy all conditions for inherently unpredictable behaviour (cf. De Vriend, 1998). This would imply that large-scale behaviour cannot be derived in a deterministic way from small-scale processes.

At the moment, the hypothesis of the existence of inherent predictability limits in tidal inlets, other than those associated with turbulence and individual grain motion, has neither been verified, nor falsified. Practical limits, associated with computer time, for instance, have been encountered. One example is the simulation of a two-day storm event around Thyborøn Inlet, Denmark, which took about one day of computer time (Brøker et al., 1996).

Moreover, recent experience with a number of 2-D and 3-D morphological simulations of inlet systems in The Netherlands has shown, that we are not yet in full control of these models. Including new small-scale mechanisms, such as curvature-induced secondary flow, sometimes improves the results, sometimes it does the opposite, but we are unable to explain why. Apparently, the physical understanding which has to underpin the modelling is still insufficient.

In summary, we must conclude that our capability to predict the morphological behaviour of tidal inlets and tidal basins is still unsatisfactory for practical use. This is for instance true for coastal inlet systems, with its complex, multi-scale dynamical interaction of waves, currents and bed topography. Since the deltas play a key role in the large-scale behaviour of barrier island coasts, as a sediment buffer and as a link between the islands, the lack of sufficient knowledge on how they work is a major obstacle to a better capability to predict the behaviour of inlets and interrupted coasts.

In order to tackle the problem of limited predictability we introduce the concept of dealing with our interest on a *cascade of scales*:

the *micro-scale* level, i.e. the level inherent to the underlying processes<sup>1</sup> and the smallestscale morphological phenomena (ripple and dune formation); the principal forcings are the diurnal tide and the weather,

the *meso-scale* level, i.e. the level of the principal morphological features, such as channels and shoals (hundreds of metres in space, years in time); the principal forcings are seasonal and interannual variations in the tide and the weather conditions, and human activities such as sand mining; a special category of phenomena at this level is the response to extreme events,

the *macro-scale* level, i.e. the level at which these features interact (e.g. the outer delta in space, decades in time); the principal forcings are the longer-term cycles in the tide, decadal-scale variations in the wave climate, consistently repeated human interference activities, etc., the *mega-scale* level, i.e. the level at which the principal elements of the entire system (barrier islands, outer deltas, inlets, lagoon) interact, so generally many kilometres in space and centuries in time; the principal forcings are mean sea level rise, climatic change, long-term tidal variation, subsidence, etc.

The highly dynamic nature of the non-linear, stochastically forced systems we are dealing with makes it not very likely that one model will be able to cover all these scale levels at one time. At various points, one must expect to run into intrinsic or practical limits of predictability

<sup>&</sup>lt;sup>1</sup> The term *processes* is used here for the water and sediment motion at its own appropriate scale level (e.g. for wind waves: wave length in space, wave period in time)

(cf. De Vriend, 1998; also see Fig. 1.1). In other words: brute-force computing is probably not a viable approach to predicting the macro- and mega-scale behaviour of these systems.



In general, predictability limits can be overcome by aggregation. Based on what is known of the system's behaviour at scales below and above this limit, another model is formulated at the higher scale level, without attempting to describe every detail of what happens at the lower scale level. Hence, we should aim at a cascade of models at different levels of aggregation.

According to the present insights, the transitions in the above scale cascade approximately correspond with the points where predictability limits are to be expected. Hence this cascade should be reflected in the cascade of models. This leads to the system diagram shown in Fig.1.2.

The scales typically addressed in these lectures are the meso- and macro-scale. On the macro-scale one typically reverts to more aggregated modelling approaches, while the process-based modelling is typically relevant on the meso-scale. It is noted that in these lectures the use of the term coastal system is sometimes applied to the natural system and sometimes to the model system, i.e. the representation of the natural system in measurable characteristics or in mathematical variables. It is however important to make a distinction between the natural and the model system.



Fig.1.2: System diagram

In the context of coastal inlets and tidal basins one may think of the following processes on each of these scales:

- micro-scale: bottom ripple and dune structures due to flow and waves;
- meso-scale: ebb- and flood chutes; channel-flat exchange;
- macro-scale: ebb-tidal delta functioning;
- mega-scale: morphologic interaction between coastal inlet, tidal basin and adjacent coast.

#### **1.3 Mathematical-physical modelling**

As explained in Section 1.2, the physical laws that are described by a mathematical-physical model represent a limited class of processes, the characteristic scales of which are limited, temporally and spatially. In a certain sense, a mathematical-physical model is a filtered description of the natural system: only for a restricted window of spatial and temporal scales a description of the system's dynamics is represented. For smaller space and time scales certain parameterizations are introduced; in the matter balance equations these are for instance the diffusion and dispersion terms. For time and space scales external to the boundaries of the model or system forcing terms or boundary conditions are adopted.

Some models only describe the response of the coastal system due to a small external forcing, such that the system is only slightly out of its equilibrium state. These are linear models; the response does not exhibit any other frequencies than that of the forcing. The spatial structures are coupled to the geometrical scales which are adopted in the model. The model doesn't display "internal" dynamics; as soon as the forcing is released the system will return to its original equilibrium state.

When describing watermotion and transport processes in coastal ocean systems one commonly applies (geo)morphostatic models. In these models the bottom is assumed to be fixed from the consideration that significant changes in the bottom morphology occur on much larger time-scales than the time-scales involved with the adaptation of the flow and matter transport patterns. When the model includes non-linear processes scale-interactions may occur, not due to the boundary forcing, such as longshore flow instabilities (meandering longshore flow). The morphostatic assumption causes the model to return to its original state as soon as the forcing is released; evolution of the coastal system on larger scales is not described.

In (geo)morphodynamic models the bottom morphology is dynamically coupled into the system. Also due to a periodic and symmetric oscillating forcing residual transports may occur due to non-linear interactions, such that the bottom undergoes a non-periodic change. In this case bottom structures may develop on a wide range of time and space scales. Yet it must be realised that processes on the smaller time and space scales are not described dynamically correct by the model. In case of discrete numerical models the timestep and the gridsize determine the lower scale boundary on which processes can be described. In a morphodynamic model the system will generally not return to its original state when the boundary forcing is turned off.

The physical laws described by mathematical models are generally balance equations, that are derived from conservation laws for the balance variable; for instance conservation of mass, momentum, vorticity, energy or energy flux, suspended matter or sediment. In the 3-dimensional space these balance equations take the form:

$$\partial f / \partial t = -\nabla \cdot \Phi - \nabla \cdot \Psi + P \tag{1.1}$$

The various terms have the following meanings:

- left-hand term: temporal change of the balance variable f in a unit volume element, averaged over the time and space scales of small scale processes;
- first right-hand term: spatial gradient in the transport of *f* (difference between influx and outflux of *f*); often φ is given by φ= u.f.
- third right-hand term: production or destruction of f,
- second right-hand term: transport due to fluctuations in the flowfield caused by small scale processes.

Often  $\Psi$  is given by:

$$\psi = \langle u' f' \rangle$$
 (= diffusion or dispersion)

(1.2)

In case of sufficient small scale processes state in time and space the following is approximately valid:

$$\psi \approx -N_1(u)f_x - N_2(u)f_y - N_3f_z$$
(1.3)

The nonlinearity is especially due to the transport term  $\Phi = u.f$ . We may make a distinction in cases where f = f(u), e.g.

- momentum balance f = u, v, w;
- bottom or sand transport  $f \propto |u|^{b-1} u$ , in which he coefficient b ranges from 3 to 4;

and cases in which u = u (f), e.g.

- density driven flow (f = density);
- bottom form.

Balance equations may also be derived in two-dimensional space (either after depth- or width averaging) or in one-dimensional space (after cross-sectional averaging). In the latter case the balance equation takes the form:

$$(\overline{Af})_{t} + (A.\overline{u}.\overline{f})_{x} = A\overline{P} - \psi_{x}$$
(1.4)

in which  $\Psi$  is the dispersive transport:

$$\Psi = A(\overline{u-u}).(f-\overline{f})$$
(1.5)

In some cases this may be approximated by a gradient-type transport, with dispersion coefficient D:

$$\Psi \approx -AD\overline{f_x} \tag{1.6}$$

see for example the sections on "Physics of tidal inlets and of basins".

When the balance variable *f* is depending on a periodically varying function, it is attractive to attempt a formal averaging over the periodic motion. This is e.g. the case as when *f* represents the sand transport due to an orbital wave motion or due to a tidal motion. Let us consider the general case of the periodic temporal dependence of *f* on u(t) as suggested above, while u(t) consists of a small residual component and of a first and second harmonic component:

$$f(t) \propto |u(t)|^2 u(t), \text{ in which } u(t) = \langle u \rangle + u_1 \cos(\omega t) + u_2 \cos(2\omega t + \varphi_2)$$
 (1.7)

In the case of

- orbital wave motion:
  - the residual component <u> may be due wave drift or undertow;
  - the first order component is the primary wave frequency, viz. the first order Stokes component
  - the second harmonic or second order Stokes component;
- tidal motion:
  - the residual component *<u>* may be due a river run-off, an internal ebb-flood channel circulation and/or a HW-LW induced channel-flat flow difference;
  - the first order component is the primary tidal wave frequency, viz. diurnal M<sub>2</sub> tide;
  - the second harmonic or the M4 tide.

Note that when the second order component is involved in the case of water waves one speaks of asymmetric waves, while in the case of tidal waves one speaks of "overtides".

Let us now first perform a formal time-averaging of f(t) under the assumption that the residual component is small compared to the oscillating component, which yields the following result:

$$< f(t) > \propto < u(t) |u(t)|^{2} > \cong < \widetilde{u} |\widetilde{u}|^{2} > + 3\overline{u} < |\widetilde{u}|^{2}$$
(1.8)

This result indicates that the sediment transport is determined by two components, one due to the oscillating motion, which is zero in case the oscillating component is purely due to a first harmonic motion, and another one due to the transport by the residual component of the periodic stirring of the sediment, yielding the assumption that the concentration is often assumed to be proportional to b-1. The latter case is usually the case when there exists little or no asymmetry in the periodic motion. The oscillatory motion stirrs up the sediment and the residual motion advects the sediment (cf. the Bijker transport formula).

Let us now consider the transport due to the oscillating motion in case the oscillating component is due to both a first and a second harmonic, including a possible phase shift between these components. In that case we may derive:

$$\langle \widetilde{u} | \widetilde{u} |^2 \rangle \cong 3/4 u_1^2 u_2 \cos \varphi_2 \tag{1.9}$$

Note that in this result we may conclude that in case the phase shift is zero an asymmetry in the periodic motion may induce an important contribution, since both  $\langle u \rangle$  and  $u_2$  may be of the same order, viz. some 10% of the primary harmonic amplitude. Also note that the phase shift  $\varphi_2$  is fairly important, viz. if it approaches  $\pi/2$  the asymmetry contribution goes to zero and no contribution due to asymmetry is exerted.

In the sections to follow, the dynamical model descriptions are often derived from the balance equations. The variables defined from the balance equations and associated boundary conditions are mathematical objects. These should be separated from the physical reality. We may only demand a limited agreement with physical reality, concerning those processes which are represented correctly on those time and space scales which are represented by the balance equations.

#### 2. The coastal ocean in global perspective

#### 2.1 The importance of the coastal ocean

Along the edges of the continents there exists an intensive interaction between land and sea. At this interface a large variety of coastal ocean systems has developed. These coastal ocean systems may be characterized as relative shallow areas (i.e. depths less than approximately 100 m) bordered by or partly enclosed by land, influenced by the sea and connected to the oceans, and in which ocean disturbances propagate. Relatively shallow may be defined as those depths at which the bottom is disturbed by the water motions, such that the bottom affects the water motion and whereby the bottom itself is changed. So, there exists an interaction between bottom morphology and water motion, where there develops a mutual balance. The transitional zone between the ocean and the continents or the ocean's peripheri has a characteristic form (see Figures 2.1 and 2.2), along which the dominant water motions change. This zone is named the continental shelf, while the peripheri of the continental shelf is referred to as the shoreface. This shoreface may be defined as -on its landward side- to include the partially enclosed basins, such as tidal rivers, estuaries, tidal basins (lagoons), and coastal inlets, of which the latter two are central in these lectures.



Figure 2.1: The shelf and dominant processes



Figure 2.2: The shoreface (after Cowell et al, 1998)

Coastal ocean systems have developed due to the deposition of erosional products and organically produced particles towards the land ocean interface. These erosional products are primarily transferred by rivers, but also transfer by wind and local erosion, and local production may play a role. These products determine the coastal ocean bottom sediment, that is moved by the local water motions. These water motions are driven by several forces, primarily tides and wind. Tidal energy is confined to some small frequency bands, while the influence of wind and atmospheric variations is manifest over a wider frequency spectrum, varying from windwaves to storm surges (see Figure 2.3).



Figure 2.3: Energy spectrum in the coastal sea

1947 -	THE COASTAL OCEAN'S PASSPORT
	COASTAL OCEAN DATA
•	occupies 18 % of the globe's surface
•	accounts for 25 % of the global primary production
•	is the area where around 60% of the human population lives
	is the area where between 60% and 70% of the world's cities with populations of over
	1.6 million people are located
	number approximately 90% of world fish catch
•	
<u>17.399</u> 14.5865	THE COASTAL OCEAN ACCOUNTSTON
•	8% of the ocean sunace
•	< 0.5% of the ocean volume
•	between 18% and 33% of global ocean production
•	80% of the global organic matter burial
•	90% of the global sedimentary mineralisation
•	75% to 90% of the global sink of suspended river load and its associated
	oloments/pollutants

in excess of 50% of pesent day global carbonate deposition

Table 2.1: Characteristics and importance of the coastal zone

The importance of the coastal ocean systems both in physical, biological and human terms is highlighted in Table 2.1. This data explains why globally coastal areas are the areas with the largest growth in population and what is the importance of the coastal ocean for the ocean itself. In its turn, the importance of the oceans both for the global climate and its resources is indicated in Table 2.2. Clearly, pressures on the coastal ocean are of global impact. It is expected that these pressures increase in future due to:

- trends in tourism, population migration and transport, modifying physically large coastal areas, and threathening habitats and landscape,
- changes in land use and ineffective catchment management, causing coastal and marine pollution, in particular by diffuse pollution sources,
- overexploitation of resources, related to a lack of consistent institutional and legal frameworks, causing depletion of fish stocks and destruction of the coastal zone,
- climate change and sea level rise, causing coastal erosion and threathening safety of life and investments.

THE OCEAN'S PASSPORT	
Sex;	male
Colour:	varying from greys and browns to deep-blue
Mean depth:	3,750 m
Largest depth:	11,034 m Marianen abyss, Phillippines
Water volume:	13,988,983 * 1000 km <sup>3</sup> (98 % of all water on earth)
Surface extent:	360,000,000 km <sup>2</sup> (70.8 % of the earth's surface)
Mean	varying from 30 degrees Celsius on the surface to 2
temperature:	degrees celsius at the deepest location
Salt content:	34 to 37 grams per liter
Species:	1.7 million known, between 5 and 100 million estimated, representing about 80 % of the earth's biodiversity
Married to:	the atmosphere
Children:	the world climate and the ocean circulation
Profession:	climate regulator, resource provider, fish producer,
	transport facilitator, garbage processor, therapist and inspirator
Coastal length:	504,000 km (a little more than 12 times the equator's circumvention)

Table 2.2: Characteristics and importance of the ocean

In conclusion, it is obvious that knowledge of the physical dynamics of coastal inlets and tidal basins as important elements of the coastal ocean systems is a necessary prerequisite to allow for an efficient and sustainable development of these systems. This knowledge however needs to be translated into quantitative predictive abilities, for which these lectures intend to provide the skills.

## 2.2 Coastal classifications

(Source: P.D. Komar; Beach Processes and sedimentation.)

A variety of coastal features have been considered and generally referred to as erosional or accretional. It is worthwhile to further categorize coastal morphology in order to bring about a better understanding of the factors that are significant in controlling the morphology. This has led to the development of coastal-classification schemes. Coastal morphology reflects the complex imprint of the tectonic setting, modified by the combined actions of more local agents and processes. The resulting classifications have included purely descriptive schemes, as well as systems that relate to physical processes important to the morphology. The classification scheme developed by Inman and Nordstrom (1971) is modern in its outlook in that it is based on the concepts of global plate tectonics discovered in the 1960s. Inman and Nordstrom recognized that the gross aspects of the topography of a coast are related to its position on the moving tectonic plates. For example, this can explain the observed contrast in the west versus the east coasts of the United States, shorelines dominated, respectively, by erosion (i.e., sea cliffs, rocky headlands, etc.) and deposition (barrier islands). Along the northwest coast of the United States, portions of the tectonic plate in the ocean basin are moving eastward and colliding with the continental plate of North America. This collision has resulted in earthquakes, coastal uplift, and the formation of mountains immediately inland from the coast. In contrast, the east and Gulf coasts of the United States lie within the interior of the North American plate, and therefore experience little tectonic activity and are either stable or subside rather than being uplifted. Furthermore, larger quantities of sediments are delivered by rivers to the east and Gulf coasts due to the overall asymmetry of North America, with the largest mountains and divides located closer to the west coast. Inman and Nordstrom (1971) have classified coasts into three categories according to their positions within the moving tectonic plates:

- 1. Collision coasts (convergent margins)
  - Continental collision coast: the margin of a thick continental plate colliding with a thin oceanic plate (e.g., west coasts of North and South America)
  - Island arc collision coasts: along island arcs where thin oceanic plates collide (e.g., the Aleutian island arc)
- 2. Trailing-edge coasts (divergent margins)
  - Neo-trailing-edge coasts: new trailing-edge coasts formed near beginning spreading centers and rifts (e.g., the Red Sea and Gulf of California)
  - Amero-trailing-edge coasts: the trailing edge of a continent having a collision coast on its opposite side (e.g., east coasts of the Americas)
  - Afro-trailing-edge coasts: the coast on the opposite side of the continent is also trailing (e.g., the east and west coasts of Africa)
- 3. Marginal sea coasts: coasts fronting on marginal seas and protected from the open ocean by island arcs (e.g., Korea)

Figure 2-4 shows the worldwide distribution of coastal and shelf types as determined by Inman and Nordstrom. They distinguish between Afro-trailing-edge coasts and Amero-trailingedge coasts due to differences in the resulting sediment supplies. As was pointed out previously, the east and Gulf coasts of the United States receive large quantities of sediments since many of the rivers originate in mountainous regions and have very large drainage basins.



Figure 2.4: A tectonic-based classification of coasts.

This contrasts with Africa, which lacks significant mountains due to the absence of plate collisions, and accordingly smaller quantities of sediment are delivered to the coastal zone. As noted by Inman and Nordstrom, the categories within their classification correspond at least grossly to the physical nature of the coast. The collision coasts are all relatively straight and mountainous and generally are characterized by sea cliffs, raised terraces, and narrow continental shelves. The trailing-edge types of coasts are more variable. The 'Amero'' types have low-lying depositional coastal forms such as barrier islands and the widest continental shelves. The neo-trailing-edge coasts only recently have come into being, as spreading between two newly formed plates has split the continental crust apart. The coast is typically steep with beaches backed by sea cliffs, so in many respects these new-trailing-edge coasts are similar to collision coasts. The marginal-sea coasts have the greatest diversity of form. The land may be low lying or hilly, and the form of the coast can be dominated by local processes such as the formation of river deltas.

Inman and Nordstrom refer to the above-mentioned tectonic classification as the control on first-order features, the gross morphology of the coast. They further recognize the importance of second-order features superimposed on the first-order tectonic control. They have in mind the formation of river deltas, the development of barrier islands, the erosion of sea cliffs, special processes such as glaciation that produces fjords, and organic action that forms coral reefs and mangrove coastlines. This level of smaller-scale coastal features and processes has been the primary focus of most classifications. For example, Figure 2-5 is derived from the classification of Shepard (1976), which relates the groupings of such features to the level of coastal stability.

Such classifications can be useful in helping one to focus on the tectonic controls, geologic history, and the many processes that have shaped a specific area of shore or have given rise to a certain coastal landform. For example, Glaeser (1978) has examined the global distribution of barrier islands in terms of the classification of Inman and Nordstrom (1971), leading to a better understanding of the physical setting required for the formation of that coastal landform.



Figure 2.5: A coastal classification based on the second-order level of features and processes, the tectonic settling being the first-order control.

#### 2.3 Geological sea-level changes

The Quaternary, a period of the Cenozoic, the youngest geological area, is characterized by the occurrence of an alternatory interglacial and glacial climate at the higher latitudes of the northern hemisphere. These glacial-interglacialcycles occurred approximately every 100.000 year during the last one million years. The forcing factor is likely to be changes in the earth's radiation budget due to orbital parameter effects.

An important phenomenon of the Quaternary are changes in sea-level. During the cold periods -the glacials- enormous icecaps were developed, resulting in an important lowering of the oceanlevel (more than 100 m) and large parts of the Continental Shelf became dry land or coastal swamp. In Southeast Asia large parts of the Sunda Shelf became dry land or coastal swamp, accordingly fauna and flora could migrate from continental Asia to the present islands of the archipel (Indonesia and Malaysia). The melting down of the icecaps during the warmer periods -the interglacials-resulted in the restoration of ocean level. These changes in sea-level are called "eustatic".

There are local and regional factors that influence the eustatic sea-level changes, such as upward and downward tectonic movements and compaction of sediments. Since these factors are difficult to separate from each other we use the terminology relative sea-level changes, thus indicating the combined result of eustatic, tectonic and compaction factors.

#### **Old Concept**

Until 1971 relative sea-level changes were thought to be caused by eustatic and tectonic movement. Thus, in so-named "stable areas", where no tectonic processes occurred, the observed sea-level changes were assumed to be due to eustatic effects only. The eustatic sea-level changes in the stable areas during the last 6,000 years, as reported by several authors, were rather divergent. There exist three groups of results:

- the oscillating sea-level (e.g. Fairbridge)
- the standing sea-level (e.g Fisk)
- the continuous rising sea-level (e.g. Shepard)

#### New Concept



Fig 2.6: Sketch illustrating the sea-level movement, connected to gravity change. The deformed geoid is the sea-level position prior to the melting. The undeformed geoid is the spherical surface, the sea-level position after ice melting.

The controversial findings of the above mentioned three groups could not be explained until a new concept of sea-level changes came into being due to a new geophysical concept. The definition of eustasy in the sense of worldwide simultaneous changes in sea-level is not correct. The sea-level is an equipotential surface of the earth 's gravity field (Fig 2.6). Sea-level can be measured only over the oceans but can also be defined at the land area, as the corresponding equipotential surface is a completely closed surface. Unlike ice, water can flow rapidly to create an equilibrium ocean surface on a gravitational equipotential surface,

the geoid. The redistribution of water causes differential loading of the ocean basins and additional deformation of the ocean floor. This feed back process continues until equilibrium is attained.

The above factors controlling sea-level are interrelated because both the ice sheets and the water load deform the earth's surface and perturb the geoid. Geophysicists have made a model of the deformation of the geoid and the earth's surface in the last 6000 years as a result of viscoelastic isostatic responses to water and ice loads (Clark et al, 1978). This model results in six zones (Fig 2.7) with typical relative sea-level change curves.

However, also the local tectonic environment of individual coastal areas and islands must be understoood. Figure 2.8 shows causes and effects, affecting both land and sea levels and resulting in relative sea-level changes. for instance, there are areas with important tectonic movements like the earthquake zones in the western and southwestern pacific and in the mediterranean. In these areas shorelines can be uplifted, submerged and or tilted by these earth movements. Most deltas are situated in slowly subsiding sedimentary basins (tectonic subsidence between 2 and 10 cm per century).











Figure 2.8: Processes controlling the factors which give rise to the three different types of eustacy. Together with crustal movements these largely determine the relative sea-level changes.

In conclusion, during the last 6,000 years, when the landicecaps had already nearly vanished, sealevel changes are due to the above mentioned local and regional factors. Accordingly, the curves of relative sea-level rise versus time show large local and regional variations.



Finally, in stable areas sea-level changes are as indicated in Fig 2.9.

Figure 2.9: Indication sea-level changes in stable areas

#### 2.4 Role of sea-level rise in Holocene coastal evolution

Bird's (1985) worldwide inventory indicates that approximately 70% of the world's sandy coastlines have shown retreat over the past few decades; less than 10% have shown net progradation, while the remaining 20% to 25% have remained approximately stable. While shore retreat on a geological time scale is undoubtedly connected with eustatic sea-level rise, it is more or less generally assumed that even the relatively small rate of sea-level rise of the last century is driving this worldwide tendency of shore retreat (cf. Vellinga and Leatherman, 1989). It must be borne in mind though that there generally exists a complicated interaction of agents affecting shore retreat (as illustrated in Figure 2.10). For instance, both the sea-level rise relative to the land and its effects are rather site-specific. Besides the eustatic sea-level rise local contributions can for instance be due to glacial rebound, subsidence, compaction, and changes in ocean circulation. Additional (and often dominating) causes of erosion can be due to longshore or cross-shore losses, which in their turn can be due to a variety of causes such as the physical geometry (e.g. headlands, submarine canyons), hydraulic boundary conditions (e.g. related to waves, tides, wind) or human interference (e.g. harbours, erosion mitigating structures). At a very high level of aggregation these latter losses can be summarized to be represented by the terminology "sediment availability". The following three literature references illustrate and confirm this description.



Figure 2.10: Interaction of agents affecting shore retreat and erosion (after Morton, 1977)

With respect to the behaviour of barriers, Curray (1964) summarized the effects of both relative sea-level changes and sediment availability on the displacement of the coastline in a qualitative way as follows. In the case of absence of a net source or a net sink for sediment regression occurs in the case of a falling sea-level or emergence and transgression occurs in the case of a rising sea-level or subsidence. With a net source of sediment the transition between regression and transgression shifts towards low levels of rising relative sea-level, and alternatively with a net sink of sediment towards low levels of falling relative sea-level.

Also, Swift (1976) adopts this basic "model" and adds to it by giving valuable specifications of the shifts mentioned above. He relates the presence of either a sink or a source for the

upper shoreface to either longshore or cross-shore processes. In the longshore process sediment discharge by rivers is one of the more important factors, while for the cross-shore process the relative importance of storm and fair weather conditions is described.

For a larger variety of sedimentary environments the reasonably well-documented Holocene coastal evolution of the Netherlands (cf. Beets et al., 1990) gives the following qualitative picture of the role of sea-level rise (Van der Valk, 1990):

area	rate of sea-level rise fast	slow
river peat-bog tidal basin/estuary barrier	vertical architecture drowning drowning retreat	horizontal architecture spreading surface stabilization depending on sediment availability: nil : transgressive med : static high: regressive

Table 2.3Effects of sea-level rise on sedimentary environments in the western Netherlands<br/>(adapted from Van der Valk, 1990)

Thus, the influence of relative sea-level rise is generally acknowledged, but the more precise quantification is a matter of discussion. In this context it is noteworthy that there is a large amount of literature devoted to the description and subsequent discussion of concepts to quantify the effect. One of the most wellknown concepts is that of Bruun (1954, 1962), which basically states that the shore profile is vertically invariant in space and time relative to mean sea level. If no sediment sinks and sources are introduced this determines the profile shift in a simple way. Valuable extensions of the concept and discussions on the validity of the concept have been made by Dean (see Dean, 1990, for a review). Here, it will be argued and discussed in more detail furtheron that the Bruunconcept is not a valid approach in general. In this context it is also important to note that time is required for equilibrium to be established. Often this point is not considered by coastal scientists. Hands (1983) noted that (a) the Bruun rule requires a lag if it is to predict shoreline behaviour and (b) the depth of the active profile increases as one considers longer timescales. Both of these conclusions are of relevance when considering shore nourishment in relation to sea-level rise.

Another important point that should be stressed here is that the timescale on which present sea-level rise influences coastal evolution in general and coastal erosion in particular, is relatively large. This is not generally true for coastal response in cases of larger rates of sea-level rise, since rapid sea-level rise induces rapid coastal response in general. Present coastal evolution on smaller time and space scales is dominated by other causes.

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### 2.5 Greenhouse effect, climate change and future sea-level changes

(Main Source: Commonwealth Secretariat, 1989, updated with IPCC FAR, 1990)

#### The Greenhouse Effect

It has long been realised that the atmosphere acts in a similar way to the glass walls and roof of a greenhouse in trapping heat from the sun: the effect was described by John Tyndall in 1861. The possibility that increasing concentrations of carbon dioxide due to the burning of fossil fuel could lead to global warming was raised by Arrhenius in 1896, and he calculated that a doubling of carbon dioxide could raise average temperatures by 5° Celcius (note the correlation between carbon dioxide and global temperature, Fig 2.11). Another scientist, G.S.Callendar, attributed an apparent rise in surface temperatures from the 1880s to the 1930s to industrial pollution by carbon dioxide. Interest in the subject waned when the earlier temperature rise was not sustained. But it has returned with unprecedented strength now that evidence has accumulated of the rise in atmospheric concentrations of carbon dioxide and other greenhouse gases and because there are now clear indications of an actual rise in global mean temperature.

There is no scientific doubt that the atmospheric concentrations of a number of "greenhouse gases" capable of global warming have increased. Such gases are transparent to short-wave radiation, but retain long-wave radiation that would, in their absence, pass from the Earth into space, thus warming the Earth's surface and lower atmosphere. These gases (excluding water vapour, also a greenhouse gas) are (data from Bolin et al, 1986, and Warrick et al, 1988):

- a) carbon dioxide (CO2), whose atmospheric concentration has increased from an estimated 275 parts per million by volume (ppmv) in "pre-industrial" times (mid-eighteenth centruy) to 315 ppmv in 1958, and to about 350 ppmv in 1988;
- b) methane (CH4) which has increased annualy by about 15 parts per billion by volume (ppbv) (equivalent to a growth rate of 1.0 percent averaged over the ten years between 1975 and 1985), and has now reached over double the pre-industrial concentration of 700 ppbv;
- c) nitrous oxide (N2O) which has been increasing at about 0.25 per cent per year to reach about 310 ppbv by 1988, as against 280 ppbv in preindustrial times;
- d) chlorofluorcarbons(CFC's), which increased rapidly (5 to 7 per cent per annum) throughout the 1970's.

In addition, tropospheric ozone, which is being produced near the earth's surface in industrialised areas as a result of chemical reactions involving hydrocarbons (largely from motor vehicle emissions) and nitrogen oxides, is also a greenhouse gas, but its contribution to the greenhouse effect is hard to estimate due in aprt to its high spatial and temporal variability.

The increasing concentration of greenhouse gases is the result of human activities related to energy use, agriculture and industrial expansion, especially the first. Some 65 to 90 per cent of the increase of carbon emissions in the atmosphere today is believed to come from the burning of "fossil fuels" (coal, oil, gas). The remainder presumably comes from the biosphere, and especially from the clearance of tropical forests.

It is generally accepted within the scientific community that the continuing increase in greenhouse gas concentrations will result in substantial global warming (see Figure 2.12 for an illustration of the principle system). There is, however, uncertainty over the magnitude and timing of the warming, the regional patterns of climate change, the

seasonal differences, the effects on climate variablity and extreme events and the extent of changes in the global sea-level.



Figure 2.11: From: S.H. Schneider, Scientific America, September 1989



Fig 2.12: Earth heat balance

Predictions of climate change depend on the validity of climate models and the accuracy with which trends in the greenhouse gas emissions and atmospheric concentrations are simulated. Many of those used sofar have grossly simplified the complex atmospheric system and, especially, have been unable to deal adequately with the crucial processes at the interface between atmosphere and ocean (e.g. Clark, 1982). Because of the resolution of global climate models, cloud processes are not adequately represented. Despite the uncertainties, however, the best models available today are consistent with one another in indicating a warming of global mean tempreature.

It is also important to model accurately the global carbon cycle. This is necessary in order to predict future CO2 concentrations, as well as to suggest ways of effectively preventing further increases. About half the carbon dioxide emitted today is taken up by the oceans, which provide the main "sink" for CO2, for example in the shape of limestone sediments. Carbon cycle modellers have difficulty, however, in balancing their carbon budgets, and some components of the carbon cycle are not well understood. For example, it is not clear how plant growth might be stimulated by higher CO2-concentrations and increase the global sink for CO2.

#### **Climate Change**

Given the best estimates of the increase of greenhouse gas concentrations due to anthropogenic emissions it is most likely that during the next century these emissions could increase global mean temperature by 0.2°C to 0.5°C per decade if no additional steps are taken to reduce them (Business As Usual scenario, 1990 First Assessment Report by IPCC).

It is difficult to predict how climate will change within specific regions of the world (the 1990 FAR by IPCC gives some best regional estimates). The climate models are still unable to simulate reliably the regional details of climate. But there are reasons to expect that some parts of the world will experience climate chanegs significantly greater than the global average. The greatest warming is likely to occur in winter at high latitude (60-90°), especially in the northern hemisphere, and the least warming in the tropical latitudes in summer. Climate models agree that global average precipitation should increase, with higher precipitation tending to occur in high latitudes in winter (see Table 2.4).

Region	Temperature change (as a multiple of global average increase)		Precipation change
	Summer	Winter	
High latitudes (60-90 deg)	0.5 x to 0.7x	2.0x to 2.4x	Enhanced in winter
Mid latitudes (30-60 deg)	0.8x to 1.0x	1.2x to 1.4x	Possible reduced in summer
Low latitudes (0-30 deg)	0.9x to 0.7x	0.9x to 0.7x	Enhanced in places with heavy rainfall today

Table 2.4: Regional scenarios for climate change (Commonwealth, 1989)

It is also likely that changes in climate will change the frequency of extreme events such as severe tropical storms, floods, droughts or extremes of heat. Some changes in climatic extremes could be relatively large. In the changed frequencies of short-term, climatic extremes the social impacts of long-term climate change are most likely to be manifested.
### **Sea Level Changes**

There is a general consensus that global warming will be accompanied by a rise in global mean sea level. The best estimate for the policy scenario "Business As Usual" is that this will lie in between about 8 and 29 cm by 2030, and between 31 and 110 cm by 2100, with 66 cm as the best estimate of the average rise by 2100 (Fig 2.13). The main factors changing the ocean water volume are likely to be the melting of mountain glaciers and the expansion of the warming seas (see Table 2.4). Because of the slow process of heat transfer from atmosphere to ocean, and of the very long response times of polar ice, even if global warming stopped abruptly in 2030, global sea-level would continue to rise for many decades, and possibly hundreds or even many hundreds of years (Warrick et al, 1988).



Figure 2.13: Global sea-level rize for Policy Scenario Business-as -Usual

	Thermal Expansion	Mountain Glaciers	Greenland	Antartica	Total
High	14.9	10.3	3.7	0.0	28.9
Best Estimate	10.1	7.0	1.8	0.6	18.3
Low	6.8	2.3	0.5	-0.8	8.7

Table 2.4: Factor contributing to sea-level rice in (cm) 1985-2030

# 3. Phenomenology of coastal inlets and tidal basins

# 3.1 Delineation of the coastal system type

In nature a large variety of coastal systems exists. As described in section 2.2 a first-order classification (related to space scales of 1000's of km's) is based on global plate tectonics. The second-order features (related to space scales of 100's of km's) are numerous. One way to distinguish second-order features consists of looking at the hydraulic boundary forcings to the coastal system. Figure 3.1 indicates which coasts are subject to either ocean storms and swells (*storm-dominated*), to ocean currents (*ocean-dominated*), to locally generated wind waves (*wave-dominated*), or to tides (*tide-dominated*). Note that this classification does not take into account the influence of rivers (sediment supply, shorenormal circulation), of sediment type (silt, sand, shingle, rock), or of individual plate tectonics (out of equilibrium morphology).



Figure 3.1: World distribution of wave types

The diversity is the result of the simultaneous occurrence of all coastal system determining factors in a nearly infinite number of combinations. That is why there exists no truly unique classification system of coastal ocean systems; and that is why one encounters a large variety of classification systems in literature. A classification which comes close to the one that is central in our lectures and that is relevant to the typical Netherlands situation (being an example of a marginal sea system; other ones being the Mexican gulf and the Chinese coast) and to trailing edge coasts (such as the USA east coast) is one that distinguishes river-dominated, tide-dominated and wave-dominated coastal systems (see Figure 3.2 for example sketches). In the hydraulic engineering education we focus on:

- tide-dominated systems (tidal bays and tidal lagoons);
- tide- and river-dominated systems (tidal rivers and estuaries);
- wave-dominated systems (uninterrupted coastal stretches);
- tide- and wave-dominated systems (coastal seas, coastal inlets, shorefaces).

In these lectures we make a further restriction by focusing on relatively stable coastal inlets and tidal basins, where we define the latter to consist of tidal lagoons with approximately equal influx and outflux of water (e.g. Wadden Sea basins) and of estuaries with negligible influence of river run-off (western section of the Westerschelde, Oosterschelde, Haringvliet). The reason being we wish to refrain from giving due attention the fresh-salt water mixing problem, a topic which would require a rather extensive treatment. Before describing tidal basin characteristics in Section 3.3 and the coastal inlet characteristics in Section 3.4, we quote a more general description of "Tidal environments" given by Carter (1988).



Figure 3.2: Example sketches different coastal systems

# 3.2 Tidal environments: bays, lagoons and estuaries

(source: R.W.G. Carter; Coastal Environments, 1988)

Tidal conditions dominate where wave energy is relatively low. This may occur due to restricted fetch or where offshore conditions trap or deflect incident wave energy. Such environments include tidal bays, lagoons and estuaries. While there are affinities and overlaps between all three, each possesses a certain distinctiveness (see table 3.1), particularly relating to the mode of wave energy dissipation or exclusion and the presence of fresh and salt water, causing marine influences to be tempered by terrestrial ones.

Environment	Distinctive attributes
Tidal bays	
Baie de St. Michel	High levels of wave energy dissipation;
Normandy / Brittany	sediment fines onshore;
West coast of South Korea	little freshwaer run-off
<i>Tidal lagoons</i> Wadden Sea, Netherlands/Germany/Denmark Laguna Madre, Texas	Waves exluded by barriers; Tidal flows via passes; infilling wetlands; little freshwater run-off
<i>Estuaries</i> Bay of Fundy, Nova Scotia Bristol Channel	Waves exluded by barriers or sand shoals; high freshwater run-off

Table 3.1: Distinctive attributes of tidal environments

#### 3.2.1 Lagoons

Tidal lagoons are water bodies entrapped behind coastal barriers. They have a surface or sub-surface connection with the sea allowing water levels to be modulated by the tide, although considerable distortion and attenuation of the tidal wave may occur within the confines of the lagoon. There are many types of tidal lagoons from almost semi-enclosed bays, like the Wadden Sea, to near-estuaries with significant freshwater throughputs. Despite the variety, the lagoon must not be viewed as a link between bays and estuaries. Almost 12% of the world's coastline is made up of barriers, many of them enclosing lagoons. There are many studies investigating barrier-lagoon systems, and it is clear that no one process prevails in their information and maintenance, rather a set of co-dominate and sub-ordinate processes. It is divided into three parts: barriers, lagoons and tidal passes, but you should bear in mind that such a tripartite is somewhat false, as all three are highly intertwined.

There are three basic categories of lagoons, based on the water volume and the mode of water exchange between the sea and the lagoon (see table 3.2). In most cases the tidal fluxes in and out of the lagoon are balanced over a tidal period. Water flows into the lagoon with the flood and out during the ebb. Propagation of the tidal wave is affected by the geometry and nature of the barrier as well as the traces of the previous incursion. The volume of exchanged water is known as the tidal prism. This may be estimated crudely by measuring the lagoon area up to the tidal limit and multiplying by the tidal range. In some cases the seawater input exceeds due to evaporation. These lagoons are called sabkhas and are common on low-latitude arid coasts.

Hydraulic balance	Barrier form
Inflow = outflow	Pass - back barrier
(tidal)	Seepage - throughflow lagoon
(wave set-up)	Reef - atoll
Inflow > outflow	Sabkha
Inflow < outflow	Pass - estuary
	Seepage - liman

Table 3.2: Types of lagoons

In Figure 3.3, Carter (19..)gives an overview of the Wadden Sea. Many of the Wadden Sea passes have distinct catchment areas separated by a low watershed or wantide zone, conducive to fine-grain deposition. Large tidal eddies are common along the back barrier flanks, sometimes superimposed on a unidirectional stream creating a spiralling current.



Figure 3.3. : The Wadden Sea tidal dynamics model, displaying the types of tidal passes and the wantide zones. The movement of tide is from left to right.

### 3.2.2 Estuaries

Discharge of terrestrial runoff and its associated transport products to the sea results in a mingling of processes and a mixing of mass. Mingling may take place within a discrete basin or estuary, often semi-enclosed by a seaward barrier. An estuary is a somewhat 'hybrid' environment incorporating terrestrial and marine, lake and ocean dynamics.

Estuaries are semi-enclosed water bodies, connected to the sea, within which seawater is measurably diluted by freshwater. Interaction of two chemically and physically-different water masses gives rise to complex sedimentological and biological process and patterns. Estuaries, unlike river mouths, tend to be tide-dominated.

Almost without exception estuaries are basins of non-marine origin that have been invaded by the sea. Many estuaries are the result of marine flooding due to the global rise in the postglacial sea-level. Some of them are further constricted by the development of spits, shoals or barriers across their mouths. Processes of basin formation include tectonic subsidence, fluvial erosion and glacial action.

The most fundamental characteristic of an estuary is the interaction between salt- and freshwater. This leads to a common classification by salinity structure relating to the degree of separation or mixing of the two water masses. Three different mixing regimes are identified: stratified, partially mixed and mixed or homogeneous.

In stratified estuaries the fresh- and saltwater masses remain distinct. As the less dense fresh water meets the salt water, so it overrides it. The landward-facing gradient of the salt wedge is proportional to the hydraulic force of the freshwater internal waves, which may break causing entrainment of saltwater particles within the seaward-flowing fresh water. Eventually this process mixes the two water masses. To maintain continuity, salt water moves upstream, creating opposing currents at the wedge front. Near the bottom this counteraction favours deposition.

Tidal waves propagating into estuaries may be either progressive or standing waves, or a mixture of the two. In long narrow estuaries a multi-nodal standing wave and tide may develop.

# 3.3 Tidal basins

Tide-dominated coasts, with relatively stable inlets and basins, are characterised by a large number of inlets having ebb- and flood-tidal deltas, giving access to tidal basins. These were created by breakthroughs and flooding of coastal plains as sea level rose or by abandonce of river branches forming deltas. Bottom subsidence of the coastal plains due to human interferences (peat-harvesting, empoldering, water, oil and gas extraction) is sometimes an additional cause for the creation of tidal basins.

The characteristic morphology of tidal basins is that of a meandering, braided and/or branched channelsystem, ebb- and flood chutes, intertidal sand and mud flats, and marshes. Along barrier island coasts the basins are often rectangular or near square, and the channel structure is often more branched than braided. The branching structure is found to show fractal characteristics. Not seldomly, rivers discharge in tidal basins. When this concerns larger rivers the tidal basin is often funnel shaped, and the channel structure is more braided than branched.

If in the latter case the river discharge is large, there exists a transitional region in the basin between salt and fresh water. As described in section 3.2, these basins are called estuaries. In the case of small and medium discharges the salt-fresh water transition region is located near the river mouth, which is then called the estuary. In the more seaward region the tidal water motion dominates and the water is mainly saline. In the more landward region the river flow dominates and the water is mainly fresh. The pressure gradient associated with the density difference between the saline (of sea origin mainly) water and the upstream fresh river water drives a vertical flow circulation, the so-called estuarine circulation. Along the bottom the flow is net (i.e tidally averaged) landward, while it is seaward along the surface. This phenomenon contributes to the formation of the turbidity maximum, which is the concentration of fine sediments at the end of the salt water wedge.

In short tidal basins the tidal wave is reflected and it has a standing character. When these basins are long enough the tidal wave resonates. When they are even longer the tidal wave is dampened by friction and the reflected wave is weak, so that the tidal wave has a propagating character. However, the tidal propagation is of such nonlinear nature, that we may not really speak of a wave-like motion. In most tidal basins the tidal discharges are largest in the inlet gorge, while decreasing in landward direction. This is associated with a decrease of the discharging cross-sectional area, which leads to the typical funnel-shaped river mouth. Tidal-asymmetry and estuarine circulation also play a role in this shaping.

Caused by the large tidal discharges in the inlet gorge there is a strong sediment exchange between the basin and the adjacent coast. In the divergent ebb-tidal discharging flow, the flow velocities decrease and sediment is deposited, thus creating the ebb-tidal or outer delta which displays a variety of dynamic phenomena (see section 3.4).

Also inside the tidal basin there exists a morphologic activity, primarily driven by the interaction between bottom morphology and tidal motion. This interaction is the cause of a complex three-dimensional structure of residual circulations, which are both cause and result of the morphologic structures of the basins. In a meandering channel residual circulations are the cause of a spiralling flow structure. Sedimentation, erosion and sand- and marshflats are connected with these flow-structures.

Within the basin the tide is deformed by bottom friction and other non-linear effects associated with the basin geometry. This tidal distortion causes both an asymmetry between high- and low-water flow change. These asymmetries greatly impact on the net movement of sediment, both sand and silt. Also, the turbidity maximum is determined by the tidal asymmetry. In tidal basins with an extremely strong tide (amplitudes of the order of 10 m) the tidal wave is deformed so strongly that a tidal bore may develop. The tidal flood wave is than of such strength that a "wall of water" enters the basin. Examples of such basins are the Amazone, the Tsientang and the Severn.

Examples of tidally dominated coastal systems of the estuarine type are the Schelde, the Thames, the Elbe (all three in the winter season), the Seine, the Yangtse, the Mekong, the Ganges-Brahmaputra, the Fly River, Chesapeake Bay and the Rio de la Plata. Examples of tidal basins without strong river influence are the Irish Sea, the Channel, the (Eastern) Wadden Sea, the Oosterschelde, the Gulf of St.Lawrence, Colorado River, San Francisco Bay (the latter two both in summer).

## 3.4 The ebb tidal delta or outer delta

(sources: Steyn & Sha / vd Berg)

#### 3.4.1 Introduction

A typical ebb tidal delta includes a main ebb channel, channel margin linear bars, a terminal lobe, swash platforms and bars, and marginal flood channels (Figure 3.4). It is noted that the flood channels as these occur, for instance, at the inlets at the SW of the Netherlands are not marginal at all, but guite distinct.

The channel margin linear bars flank the ebb tidal channel and are deposits built up by the interaction between flood- and ebb tidal currents with wave-generated currents. In the Dutch inlets these distinct bars are not recognisable; in stead, there are wide flats. The terminal lobe is a rather steep seaward-sloping body of sand, which forms the outer end of the ebb tidal delta. The main ebb channel is flanked by swash platforms, which are broad sheets of sand. On these swash platforms, isolated swash bars can be recognised, built up by swash action of waves. Marginal flood channels usually occur between the barrier islands coast and the swash platforms. The boundaries of the ebb tidal delta can be found via the no-inlet bathymetry: where the differences in bottom height is nil between the actual and no-inlet bathymetry.



The occurrence of ebb- and flood dominated tidal channels is of prime importance to the morphology of the ebb tidal delta. Postma (1967) explains this phenomenon by the time-velocity asymmetry of tidal currents: the maximum ebb- and flood currents do not occur at mid tide. Often the maximum ebb currents occur near low water. The flood currents therefore, choose the way of least resistance, i.e. around the margin of the delta. The overall morphology of the ebb tidal delta depends on the interaction of tidal currents and waves. Generally speaking, the ebb tidal delta morphology is determined by the (dynamic) balance between a net offshore directed sediment flux induced by the inlet currents (ebb dominance) and a net onshore directed sediment flux induced by offshore waves. Although, these principles can not solely explain the ebb tidal delta morphology, it gives a basic explanation for the apparent mechanism that "holds the ebb tidal delta together".

### 3.4.2 Sediment

The great importance of the ebb tidal delta can be indicated by the volume of sand that is accumulated in this delta. This volume will be larger in case of low onshore directed wave energy and large tidal forces. Under these circumstances the ebb tidal delta can extend far seawards without a distinct terminal lobe. Quite often the volume of sand deposited in an ebb tidal delta is much larger then the total volume of the adjacent barrier beaches. Although the governing physical processes are not yet fully understood, sediment exchange takes place between the adjacent barrier beaches and the ebb tidal delta. The volume of sand stored in the ebb tidal delta is related to the tidal prism of the backbarrier system and the wave climate, see Figure 3.5 Even small changes herein (by nature or human interference's) may result in changes of the volumetric contents of the ebb tidal delta. For example, suppose the tidal prism of an inlet is enlarged after flooding of part of the hinterland. According to the empirical relationships (Figure 3.5), this will result in an enlargement of the volumetric contents of the ebb tidal delta.



Figure 3.5: Emperical relationship between tidal prism and the volume of sand in the outer delta

The sediment required for this enlargement may originate from the adjacent barrier coast, the backbarrier system or from offshore. Most probably it will be a combination of these three sources, the distribution of which is very hard to determine without thorough knowledge of the underlying physics. Probably, erosion of tidal gullies will contribute relatively most to the 'sand demand' of the (ebb) tidal delta, because of the relatively small morphological time scales of such adjustments. Due to increased flow velocities the tidal channels will (immediately) start to erode. The eroded material will mainly be deposited at the ebb tidal delta where the flow decelerates. On the other hand, the contributions of the adjacent barrier coasts are much more vulnerable as small changes in the sediment balance may cause severe coastline retreats.

#### 3.4.3 Hydrodynamic classification

The tidal range outside an inlet depends primarily on the ocean tides and their interaction with the continental shelf. The wave conditions are generated seaward (and thus independently) of the inlet. Since both parameters are independent of the inlet system configuration they are very suitable to be used for classification.

An inverse relation between tidal range and the length of a barrier island seems present: large tidal ranges apparently result in shorter barrier islands. Hayes e.a. (1979) used the following classification:

- microtidal, tidal range < 1.0 m
- low mesotidal, tidal range 1.0 2.0 m
- high mesotidal, tidal range 2.0 3.5 m
- low macrotidal, tidal range 3.5 5.5 m
- high macrotidal, tidal range > 5.5 m

Wave action is generally considered to act as a bulldozer on the tidal inlet morphology (Hageman, 1969); it moves sediment onshore and limits the area over which the ebb tidal delta can spread out. The wave climate is generally characterised by the mean significant wave height  $H_s$  on a yearly average basis:

0	low wave energy,	H_< 0.6 m
0	medium wave energy,	0.6 m < H <sub>s</sub> < 1.5 m
0	high wave energy,	H <sub>s</sub> > 1.5 m

The actual classification of tidal inlets is based on a combination of the tidal range and wave energy classification as described above, because the relative effect of waves and tides is really important. Hayes distinguishes the following five classes of tidal inlets with respect to tidal/wave dominance:

- 1. wave dominated inlets
- 2. mixed energy wave dominant
- 3. mixed energy tide dominant
- 4. tide dominated low
- 5. tide dominated high

This hydrodynamical classification is shown in Figure 3.6. Each class develops its own specific morphologic features, viz.:

- Ad 1 Wave dominant inlets have long continuous barriers, with only few tidal inlets and a lot of washovers.
- Ad 2 Mixed energy inlets (wave dominant) have a larger number of inlets and a smaller number of washovers. The size of the ebb-tidal delta will become somewhat larger.
- Ad 3 Mixed energy coasts (tide dominant) have abundant tidal inlets, larger ebb tidal deltas and usually drumstick barriers.
- Ad 4 Tide dominated inlets (low dominance) occasionally show wave built bars. Transitional forms can be recognized.

Ad 5 Tide dominated (high dominance) are characterised by predominant tidal current ridges, extensive salt marshes and tidal flats. Inlets of this type often have large ebb tidal deltas and very deep inlet gorges.



\* "Het Friesche Zeegat"

Figure 3.6: Hydrodynamical classification of tidal inlets.

For classification of tidal inlets not only the above relative dominance of waves or tides is important, but also the surface area of the inlet basin.

#### 3.4.4 Geometry

The geometry of ebb tidal deltas largely depends on offshore wave climate and characteristics of the tidal regime. This is described by van den Berg and Sha (1993), using the tidal inlets in the south-western part of the Netherlands and those at the western Wadden Sea as study material.

The tidal wave propagates along the North Sea coast in north-east direction, while the amplitude of the tidal wave rapidly diminishes from about 4 m near Vlissingen to 1.7 m at Hoek van Holland. A minimum value of less than 1.4m is reached south of Den Helder. The tidal range then gradually increases north-eastward along the West and East Frisian Islands coast to about 3 m.

Along the West and East Frisian Islands the longshore tidal currents reach their maximum at about mid-tide water level. A rather contrasting situation is found along the SW-part of the Netherlands. Here, the relatively large difference in tidal range results in significant longshore water surface gradients at low and high water levels. For this reason and because of the hysteresis effects the curve of the horizontal tide lags about three hours behind the curve of the vertical tide, resulting in a coincidence of maximum longshore directed flood- and ebb-tidal currents with the high and low water levels respectively.

In Figure 3.7 possible inlet flow patterns over the tidal cycle are shown for these two basic cases. It is noted that a flat bottom topography and a tidal wave propagation from left to right is assumed.

Case 1: no phase differences are present (Wadden Sea inlets). Two basic flow patterns (marked as A and B) can be recognised. The flow concentrates on the left side of the inlet. Case 2: the phase difference equals a quarter of the tidal period (SW-part of the Netherlands). Now, eight different flow patterns (marked as A to H) can be recognised.

The effect of waves on this concept is shown schematically in the lower drawing of Figure 3.7. Four situations are considered here: A and B correspond with the Wadden Sea inlets, i.e. no phase differences, and C and D correspond with the SW-coast, i.e. with a phase difference of a quarter of the tidal cycle. The direction of the energy flux is from the right (so opposing to the direction of the tidal wave propagation) for situation A and C, and from the left (so in line with the tidal wave propagation) for situations B and D. The table in figure.3.7 gives classification of several Dutch inlets according to this scheme. This classification has been compared by van den Berg (1987) with the appearance of these inlets in 1850, well before human interventions in the systems. The agreement is good. It shows that:

- the inlets in the south western part of the Netherlands were typically type C inlets (with a slight southward located ebb channel due to wave flux coming from the north), and that
- the Wadden Sea inlets were typically type B inlets (with more distinct westward located main ebb channel due to the zero-phase lag of the tidal currents at open sea and in the inlet gorge; the predominant wave energy flux reduces this effect).

It should be noted however, that the difference in morphology between type B and C is not very large. The only noticeable difference is the orientation of the main ebb-channel, which is subject to periodic changes as well.



Figure 3.7: Effect of interaction of tidal currents on inlet morphology

#### 3.4.5 Processes at the outer delta

The main processes acting in the outer delta are the waves and the tides. The morphology of the outer delta depends on the interaction of these tidal currents and waves and is determined by the balance between the offshore transport, caused by the ebb tidal currents, and the wave action. The sediment circulation at the outer delta can be thought of as an intermittent process, according to de Vriend and Bakker (19..), in which the sediment is transported, deposited and picked up, depending on the prevailing conditions. For transport near the gorge the representative conditions are moderate waves and a tide slightly above the mean (Steijn and Hartsuiker 1992) whereas the transport along the delta margin requires higher representative waves. If the inlet is in equilibrium then a balance exists, the sediment fluxes due to tidal and wave effects form a closed circulation system.

### By-passing of tidal inlets

If the waves approach the coast at an angle, there will be a longshore drift along the coast. The outer delta forms a barrier for the longshore transport of sediment which may enter the inlet via the flood channel or may be transported from the updrift island to the downdrift island either around or over the delta. This result in the long term sediment transport paths that are not closed but spiral or zigzag along the delta as indicated in Figure 3.8.



Figure 3.8: Residual drift via the shoals on the terminal lobe of the outer delta

Sediment by-passing is defined as the process which allows material, after a short interruption caused by an inlet, pass, channel, jetty, or other type of littoral barrier, to become part again of the normal littoral drift zone a short distance downdrift from the littoral barrier. Bruun & Gerritsen (1959) describe the two main principles of by-passing by natural action:

- 1. by-passing via offshore bars, and
- 2. by-passing by tidal flow action.

### 1. Bar-by-passing

The upper sketch of Figure 3.9 shows an inlet with a submerged bar. Littoral drift "simply" continues over the submerged bar to the down-drift barrier island. The integrated longshore drift capacity is kept at the same level. If the littoral drift (and its distribution cross-shore) is known for profile AA, i.e. the normal beach, it is possible to estimate the required dimensions (depth and width) of the submerged bar. In the lower sketch of Figure 3.9, this principle is shown. By increasing amounts of littoral drift the depth of the bar will decrease and its width will increase. The depth over the bar is usually limited to the breaker depth for storm waves.



Figure 3.9 : By-passing principle on a submerged bar

This type of by-passing can only occur in situations with considerable wave action. Bar bypassing is rather hazardous for navigation, as any (dredged) channel will be subject to sand deposition. For navigation purposes these type of tidal inlets are often "improved" by jetties. However, mostly these improvements only are temporarily effective and require maintenance dredging after some time.

## 2. Tidal flow by-passing

When tidal flow plays an important part in the by-passing of material, sand transfer can take place in two different ways, viz.:

migration of channels and bars. Tidal channels can migrate over the inlet throat. Figure 3.10 shows two phases of an inlet/channel system. A new channel has developed and another has disappeared. Generally, tidal channels migrate in downdrift direction. One of the reasons for this may be that littoral material is deposited on the updrift banks of the channels forcing the shift down-drift. Another process forcing curved channels to migrate is the meandering capacity, as also occurs in rivers. The bars in between the channels may follow this migration and occasionally join the downdrift barrier coast.

• transport by tidal flow in the channels. The vectors in Figure 3.10 indicate the resultant direction of sediment transport in the flood- and ebb channels.



VECTORS INDICATE Resultant direction of Sand Transport in Flood and Erb Channels

Figure 3.10: Migration of tidal inlet channels

$$r = \frac{P}{M_{\text{max}}}$$

in which: P is the tidal prism (cubic meters per tidal cycle) M<sub>tot</sub> the total littoral drift (cubic meters per year).

The stability of the inlet (in terms of by-pass capacity) expressed in this latter expression for is as follows:

r < 20:	inlets become unstable non-permanent overflow channels
20 < r < 50:	the inlets are typical bar-by-passers
50 < r < 150:	the entrance bars are still pronounced (combination of bar-by-passing and flow-by-passing);
r > 150	the inlets are predominant tidal flow by-passers (little bar and good flushing).

Equation 2.2 is a commonly used and accepted criterion. Bruun (1991) mentions that he obtains more and more examples that fit in the description. Backgrounds of the expression are given in Bruun, e.a. (1974).

Bruun & Gerritsen (1959) propose a parameter r to indicate the type of by-passing:

$$r = \frac{M_{mean}}{Q_{max}}$$
(3.1)

in which :

 $\ensuremath{\mathsf{M}_{\text{mean}}}$  is the mean value for the longshore sediment transport rate to the inlet

 ${\rm Q}_{\rm max}$  is the maximum discharge through the inlet during spring tide.

In case r < 10 - 20, predominantly tidal flow by-passing will occur. In case r > 200 - 300, predominantly bar bypassing will occur. For intermediate values of r an intermediate type of bypassing will occur. Later Bruun and Gerritsen (1960) and Bruun (1978) converted the expression for r to:

(2.2)



3-14 3-14 In Table 3.3 some values for r are given (from: Bruun, 1978). The tidal inlets along the West-Frisian Barrier islands all show a large value for r, which indicates that tidal flow by-passing is dominant. In general, increased values of r produce greater seaward displacement of inlet bars (Oertel, 1988). The first eight inlets in Table 3.3 are Wadden Sea inlets. The given values for r show that r is decreasing in eastward direction. This seems to correspond with the occurrence of sand ridges at the more easternly located inlets.

Inlet	Tidal Prism	Total Littotal	r	Stability	Comments
	Р	Drift M <sub>tot</sub>			
	[m³/half cycle]	[m³/y] *10 <sup>6</sup>			
	*10*				
Inlet of Texel, Holland	1.000	~1	~ 1.000	good	1
Eyerlandse Gat, Holland	200	~ 1	~ 200	good	Wadden
Inlet of the Vlie, Holland	900	~1	~ 900	good	Sea Inlets
Borndiep, Holland	500	~1	~ 500	good	going from
Friesche zeegat, Holland	300	< 1 (est.)	~ 400	good	West to
Nordeneyer Seegat, Germany	160	< 1 (est.)	~ 250	good	East
Wichter Ee, Germany	42	< 1 (est.)	~ 60	poor/fair	
Otzumer Balje, Germany	110	< 1 (est.)	~150	fair	↓
John's Pass, Florida	14	0.1	140	fair	
Longboat Pass, Florida	20	0.1	200	good	
Aveiro, Portugal	60	1	60	poor/fair	
Big Pass, Florida	10	0.01	100	fair	
Masonboro Inlet, Florida	20	0.3	70	poor/fair	
Saratosa Pass, Florida	3	0.1	30	poor	
Penang Harbour, Malaysia	700	0.6	~ 1.000	good	
Krishnapatam, India	10	0.6	17	poor	
ThyborØn, Denmark	100	0.8	120	fair	jettied inlet
East Pass, Florida	40	0.1	400	good	weir jetty
Oregon Inlet, NCar	60	1	60	poor/fair	
Tan My, Vietnam	47	1.6	30	poor	

Table 3.3: Stability figures of some tidal inlets.

Following the early work of Bruun and Gerritsen (1959), FitzGerald (1988) proposed three explanatory models for inlet sediment by-passing along mixed energy coasts (Figure 3.11).

### Model 1: inlet migration and spit breaching

Littoral material transported to a tidal inlet can cause constriction of the inlet throat. The decrease in cross-sectional area will result in greater scouring capacities, which will cause erosion of the downdrift barrier beaches. The rate of migration is dependent on sediment supply, wave energy, tidal currents and the composition of the channel bank. The inlet channel in this way is elongated, which results in a lower discharge efficiency. When after a storm a short-cut has been created, this new channel will take over from the old one. The new location of such an inlet is rather difficult to predict, but it is quite often located just in front of the ebb channel. Once the old inlet has attached to the downdrift coast, a large quantity of sand has by-passed the inlet. Most inlets migrate downdrift, however, some examples exist of updrift migration.

### Model 2: stable inlet processes

Stable inlets have a stable throat position and a non-migrating main ebb channel. Sand bypassing at these inlets occur through the formation of bars which migrate and attach to the downdrift coast. These large bars have been formed by the stacking and coalescing of wave built accumulations of sand (swash bars) on the ebb tidal delta platform. Due to flood dominance at these parts of the ebb tidal delta in combination with breaking and shoaling waves, the swash bars move onshore. New bars will constantly develop due to a more or less continuous sand delivery by the ebb tidal channel. Thus, sediment by-passing in this model depends on sand deliveries through the channel system. The rate of migration of the swash bars decreases when they reach the coast. FitzGerald (1988) explains this as the swash bars gain a greater intertidal exposure when they migrate up the shoreface. Thus, wave swash operates over an increasingly shorter period of the tidal cycle. The deceleration of the migration of swash bars results in the stacking and coalescing of individual bars to large bar complexes. These bars can be 1 - 2 km long and 100 - 250 m wide. The size of the bar complexes and the time required for the bars to migrate onshore decreases as the size of the inlet decreases.

#### Model 3: Ebb-tidal delta breaching

Tidal inlets that by-pass sediments by ebb-tidal delta breaching have a stable inlet throat position, but a variable position of the main ebb tidal channel. The main ebb channel will be pushed towards the downdrift coast by the predominant littoral drift. In a similar way as described for model 1, the flow efficiency of this elongated channel decreases and hydraulic gradients over the bars steadily increase. The inlet will divert its flow to a more seaward direction via so-called spillover lobe channels. Consequently, the discharge through the old ebb channel decreases. The result is that a large portion of the ebb tidal delta sand bodies is by-passed.

FitzGerald (1988) mentions that in real tidal inlets all three by-pass processes can occur simultaneously or may dominate alternately. The movement of sand along the terminal lobe by wave action is almost always present at mixed-energy coasts. An interesting feature of bar complexes that approach a downdrift coast is the formation of tombolos on this coast. An example of this is shown for the north-eastern end of Kiawah Island, South Carolina (FitzGerald, 1988). Basically, the bar complex acts as a sort of offshore breakwater, reducing the littoral drift capacity between the bar and the coast (sheltering effect).

#### 3.4.6 Human interference

#### Removal of sand from the outer delta

The sand in the outer delta may be used as a sand source for, say, beach nourishment or it may be necessary to dredge navigation channels through the outer delta. In both of these cases the sand is removed from the inlet system and the inlet will have been disturbed from the "equilibrium state". Since the tidal prism and the protrusion ratio of the delta remain constant, the supply of sand into the inlet via the flood channels does not change. The delta will retreat rapidly, while the barrier islands show no erosion in the first few years. This trend reverses however, as the outer delta will demand sand in order to return to its equilibrium state. Assuming that the inlet is a "closed system" then the demanded sand will be supplied by the barrier island coastlines.

The erosion of the barrier islands will begin on the foreshore, the beach profile steepens and gradually the erosion of the beach will become evident. Since the erosion of the barrier islands occurs tens of years after the dredging, the cause of the erosion may not be linked to the dredging operation.

In order to minimise this additional erosion it may be wise to consider the option of dredging sand from the inner delta instead of from the outer delta. The final result, after 100 years, is the same as for the previous case but that the delta retreats more slowly and to a lesser extent. This will reduce the additional erosion on the downdrift island.

If the amount of sand removed from the inlet system per year is less than or equal to the total volume of sand transported into the inlet per year via the flood channels, then the effects of dredging the inner and outer delta will be the same.

### Effects of the construction of a groin on the updrift island

The construction of a groin on the updrift island will trap sediment that previously entered the inlet via the updrift flood channels. Not only can one expect lee-side erosion but the outer delta and the downdrift island will also erode.

The outer delta act as a buffer for the other elements of the inlet system. When sand is removed from one portion of the inlet the outer delta will respond by sharing sand with the other elements but this supply is only a "short term loan" which will be repaid by the adjacent coastlines in the longer term.

So erosion of the outer delta can be expected to occur rapidly, while on the downdrift island this will take place after tens of years.

In the case of a longer groin, more sand will be trapped behind the groin and it will also take more time before by-passing begins and consequently the erosion of the outer delta and the downdrift island will be more severe.

# 3.5 Empirical relations for tidal inlets and tidal basins

Morphodynamics of tidal inlets can be defined as the dynamic behaviour of all morphological units of an inlet, such as ebb tidal delta, tidal channels, flood tidal delta and barrier islands. This behaviour is a very complex one, dependent of a lot of factors.

The stability of the morphological units can be described with empirical relations. The geometric properties are related to hydraulic boundary conditions.

O'Brien derived a relationship for inlets along the Pacific coast, the cross-sectional area in the throat, A<sub>c</sub>, is related to the tidal prism, P, at spring tide. From data of a tidal channel in the Wadden Sea it appeared that also along a channel the flow area was related to the tidal volume passing the local cross section. Later on it is was found that for various tidal channels in the Wadden Sea and in the estuaries in the Dutch Delta in the South this relationship was valid.

This relationship is presented by:

$$A_{MSL} = c_A P$$

in which:

A relationship for channel volume of a tidal basin or estuary and the tidal prism is presented by:

$$V_{\rm c} = c_{\rm c} P^{\frac{3}{2}}$$
 (3.3)

in which:

V<sub>c</sub> = channel volume below MSL c<sub>c</sub> = empirical coefficient

For example, the empirical coefficient  $c_c$  is 65 10<sup>-6</sup> for the Wadden Sea and 73 - 80 10<sup>-6</sup> for the Eastern Scheldt and the Grevelingen, see Figure 3.12.

(3.2)



Figure 3.12: Channel volume - mean tidal volume relation Eastern Scheldt and Grevelingen

Another important relationship is the one for the sand volume stored in the outer deltas in front of the tidal inlet. This relationship is derived for outer deltas in the USA and reads:

$$V_0 = c_0 P^{1.23}$$
(3.4)

in which: Vo

Co

sand volume stored in the outer delta
empirical coefficient

Finally the relationship between the relative tidal flat area and the total basin area is given. The cross sectional area of the channels is proportional to the tidal prism and the length of the channels is proportional to the square root of the basin area:

$$\begin{array}{c} A_{c} \therefore P \\ L_{c} \therefore \sqrt{A_{b}} \therefore \sqrt{P} \end{array} \rightarrow V_{c} = c_{c} P^{\frac{3}{2}} \end{array} \tag{3.5}$$

The flats area is equal to the total basin area minus the channels area:

$$A_{f} = A_{b} - A_{c} \approx A_{b} - \alpha \frac{P\sqrt{A_{b}}}{D_{c}} \approx A_{b} - \beta \frac{H_{m}}{D_{c}} A_{b}^{\frac{3}{2}}$$
(3.6)

in wich

A <sub>f</sub>	= the flats 'area 1.e. the area above MSL [m <sup>2</sup> ]
A <sub>b</sub>	= the gross basin area (flats and channels) [m <sup>2</sup> ]
A <sub>ch</sub>	= the horizontal area covered by all channels [m <sup>2</sup> ]
α, β	= constants of proportionality [m <sup>-1</sup> ]
D <sub>ch</sub>	= characteristic channel depth [m]
H <sub>m</sub>	= mean tidal range [m]

Renger and Partenscky found for the German Bight:

$$A_{f} = A_{b} - 0.025 A_{b}^{\frac{3}{2}}$$
(3.7)

Such a relationship was also found for the estuaries in the south of the Netherlands, however, this one differs from the one in the Wadden Sea, see Figure 3.13. An explanation for the trend of the relation could be the increasing activity of local wind waves in larger basins.



Figure 3.13.: Morphologic relationships

#### 3.5.1 Application of relationships

The relationships are very helpful to determine the response of the inlet to changes in the dynamic equilibrium.

The relations of the relative tidal flat area and the height of the tidal flats may be helpful in the interpretation in what way the development will occur in combination with knowledge on general sediment transport theories and hydraulics. Two examples will be given.

#### 1. Closure of a part of the tidal basin.

The closure of a part of the tidal basin will result in a reduction of the channel volume, V<sub>c</sub>, and the tidal prism, V. The inlet will adapt itself to the new situation, see Figure 3.14. The two lines represent the relationship of the channel volume and the tidal prism and the relationship of the sand volume of the outer delta and the tidal prism. Assume an equilibrium situation with tidal prism V and corresponding channel volume V<sub>msl</sub> and sand volume of the outer delta V<sub>od</sub>. At a certain time, a part of the tidal prism  $\Delta V$ . A new equilibrium will arise at the equilibrium lines for tidal prism V –  $\Delta V$ . This means that the channel volume is a m<sup>3</sup> too big and the sand volume of the tidal basin is b m<sup>3</sup> too big. The sand of the outer delta is available for the adaptation of the channels. The rest, in case a - b > 0 has to be supplied from outside.



Figure 3.14: Morphologic relationships

## 2. Accretion of new land.

Natural accretion along the borders of the basin, due to the deposition of silt is a very slow process. This process causes a gradual reduction of the tidal prism, see Figure 3.15. Due to this reduction the channel volume is a m<sup>3</sup> too large and the sand volume of the ebb tidal delta is b m<sup>3</sup> too big. This results in a sand demand from outside.



Figure 3.15. :Morphologic relations

With respect to the effects of increased sea level rise the answers are more difficult to assess. From storage curves of the different basins it follows that, without adaptation of the level of the tidal flats, the tidal volume will increase and the area of intertidal zones will decrease. The latter effect will be most serious in the Wadden Sea, especially in the western part where the levels of the tidal flats are low. If this scenario is realistic, the channels in the basin will widen and sand will be transported partly to the outer delta, which will extend, and partly will become available for accretion of the North Sea Coast adjacent to the tidal inlet.

However, an increase of sea level rise also will effect the level of the tidal flats. In a relative sense the disturbance of the characteristic water depth at the flats will be much greater than in the channels. Hence, the sediment transports in the channel will be far less effected than those on the flats. Consequently, it seems realistic to assume that the response of Nature will be the strongest on the tidal flats. If it is assumed that the levels of the tidal flats can follow the extra sea level rise, this implies that the tidal volume of the basin remains unchanged, whereas the volume of the channels increases. Thus, this scenario results in a demand of sand from outside.

The above mentioned relationships only give an indication of the new equilibrium between morphology and hydrodynamic conditions in case of changes. They don't give any information how the adaptation will take place and what time this will take. In general, adaptation processes show a logarithmic character, see Figure 3.16. The expression that shows the adaptation history is given by:

$$\frac{X}{X_0} = e^{-\frac{t}{\tau}}$$
(3.9)

In which:

 $\begin{array}{ll} t & = \mbox{ time since disturbance of existing equilibrium} \\ X & = \mbox{ quantity representing difference from new equilibrium} \\ X_0 & = \mbox{ initial difference from new equilibrium} \\ \tau & = \mbox{ characteristic time for adaptation} \\ \tau & = \frac{X_0}{\Delta X_0} \\ \mbox{ In which:} \\ \Delta X_0 & = \mbox{ initial rate of adaptation.} \end{array}$ 

This approach is not valid in case of change in sea level rise as this is a process of continuous growing disturbance of the existing equilibrium. In this case it is not realistic to assume that this will result in an immediate response of the sea bed in a tidal basin.



Figure 3.16: Accretion of the Dollard

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## 4. FYSICA VAN GETIJDEBEKKENS

### 4.1 Karakteristiek van getijdebekkens

Kustlaagvlaktes zijn wereldwijd de gebieden met de grootste bevolkingsaanwas. Vooral rond zeegaten en getijdebekkens is er een zeer actieve wisselwerking tussen natuurlijke en maatschappelijke dynamiek. Door inpoldering en ontginning en door andere vormen van ontwikkeling neemt de kwetsbaarheid van kustlaagvlaktes vaak sterk toe. Kennis van de fysische en ecologische processen in deze gebieden is dan ook noodzakelijk om een efficiënte en duurzame ontwikkeling mogelijk te maken. Deze kennis moet dan wel vertaald kunnen worden naar kwantitatieve voorspellingen. In het college worden hiervoor de belangrijkste fysische bouwstenen behandeld. De nadruk ligt op de ontwikkeling van fysisch inzicht, meer dan op numerieke voorspellingsmethoden. Dit fysisch inzicht is vooral van belang om uitspraken te kunnen doen over de veranderingen die zich afspelen op de schaal van het getijdebekken zelf. Het gaat daarbij niet alleen om veranderingen in de waterbeweging, maar meer nog om veranderingen in de fysieke structuur van het bekken, d.w.z. van de morfologie. Vragen die daarbij in de praktijk gesteld worden zijn b.v. :

- hoe reageert een getijdebekken op zeespiegelstijging (volgt het bekken de zeespiegel door aan te zanden, of niet ?)
- wat zijn de gevolgen van baggerwerken t.b.v. scheepvaart of zandwinning (is er versterkte aanzanding, neemt de zoutindringing toe, en in welke mate ?)
- wat gebeurt er na een dijkdoorbraak (wordt er een nieuw getijdebekken gevormd ?)
- wat zijn de consequenties van inpoldering (sluit het zeegat zich af, vindt er aanzanding of aanslibbing plaats, verandert de verversingstijd ?)

Getijdebekkens zijn bijna altijd ontstaan als door de zee overstroomde kustvlaktes. Ze komen voor bij laaglandkusten waar een sterk getij heerst en waar geen grote sedimenttoevoer vanuit rivieren plaatsvindt. Voorbeelden zijn: Chesapeake Bay, San Francisco Bay, Waddenzee, Baie d'Arcachon. Soms vallen getijdebekkens samen met rivierdelta's, zoals bij de Ooster- en Westerschelde, maar vaak is dat ook niet het geval, zoals bij de Waddenzee.

De waterbeweging in getijdebekkens wordt in veel sterkere mate bepaald door het getij dan door de rivierafvoer. Ook de sedimentatie hangt voornamelijk af van aanvoer vanuit zee. Het grovere sediment (zand) wordt vooral in het zeewaartse deel afgezet, terwijl het fijnere sediment (slib) afgezet wordt in de meer beschutte landwaartse gebieden.

De lengte van getijdebekkens is meestal veel kleiner dan de golflengte van het getij. Het getijverschil in het bekken is dan van dezelfde orde als op zee. Bij langere bekkens kan resonantie optreden wanneer de bekkenlengte ongeveer een kwart getij-golflengte bedraagt; de getijamplitude neemt dan toe. Door wrijvingsverliezen wordt het getij gedempt en heeft de getijgolf in de monding deels het karakter van een lopende golf (vooral bij lange bekkens) en deels van een staande golf (vooral bij korte bekkens).

Het zeewater stroomt het bekken in en uit via geulen die (grote) plaatgebieden doorsnijden. Deze platen vervullen de rol van komberging: bij laagwater vallen de platen deels droog, terwijl bij hoogwater alleen de meest landwaartse, hoger gelegen begroeide delen boven water uitsteken, de z.g. kwelders of schorren. De geulen vormen meanders, vooral in de meer zandige delen van het bekken. De maximale eb- en vloedstromen volgen niet precies dezelfde weg, vooral als gevolg van verschillen in centrifugaalkrachten, maar ook door aardrotatie. Hierdoor ontstaan eb- en vloedscharen, geuldrempels en geulvertakkingen. Gedurende een getijperiode zijn watermassa's onderhevig aan circulaties door het plaat-geul systeem die van groot belang zijn voor de verversing van het bekken met zeewater. Ook het tijdelijk achterblijven van watermassa's in z.g. dode zones (platen, schorren, geulvertakkingen) draagt bij tot menging. Verticale zout-zoet gelaagdheid treedt in de meeste getijdebekkens nauwelijks op door de dominantie van getij-instroom over rivierinstroom en door de sterke horizontale uitwisselingsprocessen van watermassa's. De topografische karakteristieken van getijdebekkens (b.v. verhouding plaat/geul-oppervlak) beïnvloeden de getijvoortplanting. Hierdoor kan de grootte van de maximale vloedstroom t.o.v. de ebstroom versterkt worden, of omgekeerd. Dit is van belang voor het resulterende sedimenttransport van of naar zee en daarmee voor de morfologische ontwikkeling van het getijdebekken in de tijd. Door deze terugkoppeling bepaalt het bekken in wezen zelf de ont-wikkeling die gaat plaatsvinden; van buitenaf opgelegde omstandigheden zoals sedimentbeschikbaarheid en stormgetijden spelen natuurlijk ook een rol. Veel getijdebekkens zijn depositiegebieden voor fijn sediment. Dit komt door de beschutte ligging, maar de toevoer van fijn sediment hangt vooral af van de getijasymmetrie. Van belang is in het bijzonder de verhouding tussen de kenteringsduren voorafgaande aan vloed en eb: deze bepalen ten dele hoeveel sediment in het begin van de vloed en de eb vervoerd wordt. Golfwerking en stormvloeden kunnen een sterke erosie en export van sediment veroorzaken.

Sommige getijdebekkens zijn instabiel en verzanden of verslibben in korte tijd, zoals destijds het Zwin. In andere bekkens is er een bijna evenwicht tussen import en export van sediment. Dit is b.v. het geval voor de Oostelijke Waddenzee die al duizenden jaren ongeveer de huidige vorm heeft. Deze quasi-stabiliteit heeft een dynamisch karakter. De ligging van platen en geulen fluctueert over tijdschalen van tientallen tot honderden jaren.

## 4.2 Inhoud van het college

De fysica van getijdebekkens wordt in vier stappen behandeld.

Als eerste stap wordt inzicht gegeven in de complexe geometrische structuur van getijdebekkens, die wordt gevormd door meanderende en vertakkende geulen. Deze doorsnijden plaatgebieden die deels zijn opgebouwd uit zandig sediment en deels uit slibrijk materiaal, met plantengroei op de hogere landwaarts gelegen delen. Uitgelegd wordt hoe een plaat-geul patroon zich ontwikkelt en hoe zich een voortdurend fluctuerend, dynamisch evenwicht instelt. De stromingsstructuur in geulbochten wordt behandeld; aangegeven wordt hoe dit leidt tot de vorming van eb- en vloedscharen. Tevens blijkt dat hierdoor gemiddeld over het getij grootschalige circulaties ontstaan die aan de bekken-topografie zijn gerelateerd.

Als tweede stap wordt de voortplanting van het getij in het bekken beschreven. Een analytische beschrijving is alleen mogelijk voor gesimplificeerde geometrieën. In het college wordt de één-dimensionale getijvoortplantingsvergelijking afgeleid voor het geval dat de geul-plaat geometrie vrijwel niet in langsrichting verandert. De nadruk ligt op de vervorming van de getijgolf bij voortplanting en terugkaatsing in het bekken, waarbij bodemwrijving een dominante rol speelt. Door deze getijvervorming ontstaat een asymmetrie tussen de eb- en de vloedstroom. Deze asymmetrie is verantwoordelijk voor een netto getij-gedreven transport van sediment in het bekken.

Als derde stap wordt de sedimentatie in getijdebekkens behandeld. Inzicht wordt gegeven in de faktoren die bepalend zijn voor de import of export van sediment. Deze kunnen verschillend zijn voor grof en fijn bodemmateriaal, vooral door de verschillen in responsiesnelheid van sedimentatie en erosie op fluctuaties in de stroomsnelheid. In een intermezzo wordt kort ingegaan op enkele belangrijke eigenschappen van sedimentdeeltjes - in de range van slib tot zand. Import of export hangt verder ook af van de bekken-topografie; verschillende typen topografie worden behandeld. Aangegeven wordt op welke wijze door getij-asymmetrie een resulterend transport van grof en fijn materiaal plaatsvindt. De ééndimensionale getijvoortplantingsvergelijking legt een verband tussen de getijasymmetrie en bepaalde geometrische karakteristieken van het bekken. Hieruit volgt een relatie tussen enerzijds de geometrie van het bekken en anderzijds het resulterend transport van sediment. Deze relatie maakt het mogelijk uitspraken te doen over de morfologische stabiliteit van getijdebekkens.

De vierde en laatste stap betreft de verversing van getijdebekkens. Inzicht wordt gegeven in de processen die een rol spelen bij de menging en verversing. Het begrip dispersie wordt geïntroduceerd aan de hand van een "tidal random walk" beschrijving van watermassa's. Dit tidal random model wordt toegepast om de grootte van de getijgemiddelde menging af te

schatten. Enerzijds betreft dit dispersie door de combinatie van getijtransport en topografiegebonden circulaties. Anderzijds gaat het om dispersie door de uitwisseling van met het getij getransporteerde watermassa's met achterblijvende watermassa's op buiten de hoofdgeul gelegen gebieden.

### Voorkennis

Theorie van lineaire golfvoortplanting, basisvergelijkingen vloeistofmechanica, sedimenteigenschappen.

Ontstaan	overstroomde kustvlakte, ingeklonken laaglanden
getijdebekken	fungeren als "vloedkuil"
Getijgolf	landwaarts vooral staande golf, zeewaarts vooral
	lopende golf; resonantie onderdrukt door demping
	gereflecteerde golf
Getijvoortplanting	volgt de hoofdgeul, eendimensionaal van karakter
Getij-asymmetrie	vervorming getij door niet-lineair variërende
	bodemwrijving en door wisselwerking met
	geometrie
Morfologie	meanderend, vertakt geulsysteem, drempels,
-	platen, slikken, schorren/kwelders
Getij-geinduceerde	spiraliserend stromingspatroon in geulmeanders,
stroming	rest-circulaties in eb/vloedschaar cellen zorgen
-	voor plaatopbouw en drempelvorming
Rivier-invloed	ondergeschikt aan getij-invloed, geen sterke
	gelaagdheid, dichtheid-gedreven circulatie heeft 3-
	dimensionale structuur, is van ondergeschikt
	belang voor zoutindringing
Morfo-dynamiek	op grote schaal door niet-lineaire wisselwerking
	van getij-voortplanting met morfologie, lokaal door
	niet-lineaire wisselwerking van secundaire
	stromings-structuur met plaat-geul geometrie.
	Leidt tot cyclische variaties in eb/vloedschaar-
	patroon.
Sedimentatie	zanddepositie meer zeewaarts, slibdepositie meer
	landwaarts (aan einde transport-weg); in
	afgesneden geultakken, op hogere, begroeide of
	beschut liggende platen, bij wantijen
Menging en verversing	door topografisch geïnduceerde schering in het
	getijsnelheidsveld, door secundaire circulaties in
	eb/vloedschaarcellen en geulmeanders, door
	uitwisseling met dode zones, door chaotische
	transportpaden

Tabel 4.1: Karakteristiek getijdebekkens

Een karakteristiek getijdebekken is de Westerschelde, zie figuur 4.1.



Figuur 4.1: Plaat-geul systeem van de Westerschelde

## 4.3 De complexe geometrie van getijdebekkens

#### 4.3.1 Bochtstroming en plaatopbouw

De morfodynamiek van getijdebekkens is zeer goed beschreven in het artikel van van Veen uit 1950, "Eb- en vloedschaarsystemen in de Nederlandse getijdewateren". De karakteristieke topografie met meanderende, vertakte geulen, eb- en vloedscharen en platen is voor een belangrijk deel toe te schrijven aan horizontale en transversale circulatiestromen die worden opgewekt bij geulbochten. Deze circulatiestromen spelen een actieve rol bij de morfodynamische ontwikkeling van geulmeanders en van eb- en vloedschaarsystemen. In figuur 4.2 is een geulbocht schematisch weergegeven, zowel in dwarsdoorsnede als in langsdoorsnede.



Figuur 4.2: restcirculaties in een geulbocht

Een eenvoudige beschrijving van transverale circulatie in een geulbocht wordt verkregen als verondersteld wordt dat de stroming (eb of vloed) zich onder invloed van bodemwrijving en oppervlakteverhang heeft aangepast aan de geulbocht en dat zowel de geul- als de stromingsstructuur hoekonafhankelijk zijn in een cylindrisch coördinatenstelsel. De krachtenbalans dwars op de stroming is dan als volgt:

$$-\frac{u^2}{R} + fu + \frac{l}{\rho} p_y - (Nv_z)_z = 0.$$
 (2.1)  
centrifugaalkracht Coriolis drukgradiënt turbulente viscositeit

Hier is:

*R* de straal van de geulbocht (deze is veel groter verondersteld dan de geulbreedte),

*u* de snelheid langs de geulas en  $x = R \theta$  de langscoördinaat,

- z de verticale coördinaat (bodem: z = -h, oppervlak  $z = \eta$ ),
- v de dwarssnelheid en y de dwarscoördinaat.

De drukgradiënt in dwarsrichting is het resultaat van dwarshelling van wateroppervlak en van isopycnen. De vergelijking beschrijft het ontstaan van een dwarscirculatie v(y,z) onder invloed van verticale schering in de langssnelheid u(y,z). Zowel de centrifugaalkracht als de Coriolis-versnelling zijn van belang. De Coriolis-bijdrage is bij eb en vloed omgekeerd van teken (getijgemiddeld neutraal); de centrifugaalkracht werkt bij eb vloed echter in dezelfde richting. Dichtheidsverschillen kunnen in de krachtenbalans een belangrijke rol spelen; ze beïnvloeden de verticale structuur van de langssnelheid (dichtheidsstroming), de grootte en verticale structuur van de eddy-viscosity N en de grootte en verticale structuur van de transversale druk  $p_y$ . Bij sterke gelaagdheid kan de schering in de centrifugaalkracht gecompenseerd worden door een dwarsverhang van het grensvlak; er wordt dan geen grote dwarscirculatie opgewekt. Bij kleine gelaagdheid wordt de dwarscirculatie juist versterkt als gevolg van de demping van de eddy-viscosity. Bij de volgende aannamen:

gelaagdheid wordt buiten beschouwing gelaten (d.w.z.  $p_y = g \rho \eta_y$  ),

waterstandsvariaties worden verwaarloosd (  $\eta << h$  ),

N wordt constant verondersteld,

de schering in de langssnelheid wordt constant verondersteld (u(z) = u(0)(1 + z / h)), en bij oppervlak en bodem worden resp. slip- en no-slip randvoorwaarden opgelegd ( $v(z = -h) = 0, v_z(z = 0) = 0$ ),

dan volgt uit tweemaal integreren van de krachtenbalans over de diepte de volgende uitdrukking voor de dwarscirculatie:

$$v(z) = \frac{h^2 u(0)}{24N} \left[ -\frac{u(0)}{5R} (10 z^{*4} + 40 z^{*3} + 33 z^{*2} - 3) + \frac{f}{2} (8 z^{*3} + 9 z^{*2} - 1) \right], \qquad z^* = \frac{z}{h}$$
(2.2)

Deze circulatie is bij de bodem gericht naar de binnenbocht en genereert hierdoor een sedimenttransport van de buitenbocht, die erodeert, naar de binnenbocht, die aanzandt. Door deze transversale circulatie heeft de getijstroming in een geulbocht een spiraliserend karakter. De Coriolis-versnelling versterkt de dwarscirculatie wanneer vanuit de hoofdstroom gezien de geulbocht tegen een rechteroever aanligt (bij eb en vloed dus tegengestelde geulbochten). Door aardrotatie zal op het Noordelijk halfrond de vloedstroom in een breed getijdebekken meer geconcentreerd zijn langs de rechteroever, vanuit de vloed gezien, en de ebstroom meer geconcentreerd zijn aan de rechteroever, vanuit de eb gezien (Kelvin-golf type getijvoortplanting).

Door bochtstroming ontstaat ook een dwarsverhang  $\eta_v =$ 

$$\frac{9}{20}\frac{u^2(0)}{gR} - \frac{5}{8}\frac{fu(0)}{g} \quad . \quad (2.3)$$

#### 4.3.2 Eb/vloed-scharen en horizontale circulaties

De waterstand in de buitenbocht is iets hoger dan aan de binnenbocht. Dit dwarsverhang veroorzaakt dus tevens een langsverhang. Langs de buitenbochtse oever neemt de waterstand toe in langsrichting gaande naar het centrum van de bocht; langs de binnenbochtse oever neemt de waterstand daarentegen af in langsrichting gaande naar het centrum van de bocht. Deze langsverhangen drijven een horizontale circulatie aan, die langs de buitenbochtse oever van de bocht af is gericht en langs de binnenbochtse oever naar de bocht toe. Deze horizontale circulatie wordt ook wel "headland eddy" genoemd. Deze eddy is meestal gesuperponeerd op een sterke getijstroom; het instantane beeld is dat van een getijstroom die de neiging heeft om "uit de bocht te schieten", zowel bij eb als bij vloed. De horizontale circulatie heeft op het sedimenttransport hetzelfde effect als de transversale circulatie: er wordt netto sediment getransporteerd van de buitenbocht naar de binnenbocht. Beide circulaties veroorzaken uitbochting van de geul en plaatopbouw aan de binnenzijde. Geuluitbochting is dus een zelfverstekend fenomeen. Rechte geulen zijn niet stabiel; een uitwijking in de geulligging zal de neiging hebben vanzelf te groeien. Door hetzelfde mechanisme zal een brede geul die een S-bocht maakt de neiging hebben zich te splitsen in een eb- en een vloedgedomineerde geul.

Door het doorschieten van de stroming aan het einde van de geulbocht ontstaan eb- en vloedscharen in de platengebieden die aan de binnenzijde liggen van aangrenzende geulbochten. Vloedscharen zijn o.h.a. sterker ontwikkeld dan ebscharen. Dit komt doordat de gemiddelde waterstand gedurende de vloed (en vooral aan het einde van de vloed) hoger is dan gedurende de eb. Het water stroomt bij vloed dus meer over de platen, terwijl de ebstroom hoofdzakelijk de hoofdgeul volgt. Figuur 4.3 toont een karakteristieke structuur van eb- en vloedscharen in een meanderend geulensysteem, met daarin aangegeven het diepte-gemiddelde reststroompatroon.

De platen worden vooral via eb- en vloedscharen met zand gevoed. Door deze zandvoeding ontstaan drempels tussen de eb- en vloedscharen. Wanneer deze drempel doorbreekt wordt de geulbocht afgesneden. Er vormt zich een nieuwe, rechtere hoofdgeul, terwijl de oude hoofdgeul verzandt. Maar in de loop van de tijd zal de nieuwe geul weer gaan uitbochten. De morfologie van getijdebekkens is niet statisch maar dynamisch; cycli van geuluitbochting en geulafsnijding zijn een belangrijk onderdeel van deze interne dynamiek.



Figuur 4.3: Meanderende getijgeul met eb- en vloedscharen

# 4.4 Voortplanting en terugkaatsing van de getijgolf

4.4.1 De ééndimensionale getij-voortplantingsvergelijking

In getijdebekkens is er over het algemeen sprake van een hoofdgeul, waarlangs het meeste water het bekken bij vloed en eb in- en uitstroomt. Het getij plant zich in het bekken voornamelijk langs deze geul voort. Als we de lengte-as (x-as) van het bekken langs de geul-as leggen, dan kan de getijvoortplanting bij benadering worden beschreven door een stelsel één-dimensionale vergelijkingen. Dit zijn de balans-vergelijkingen voor massa en voor impuls in de x-richting.

Voor de afleiding van deze vergelijkingen wordt de dwarsdoorsnede van het bekken A(x,t) geschematiseerd weergegeven, waarbij onderscheid wordt gemaakt tussen twee delen. Dit is enerzijds het stroomvoerende deel van de dwarsdoorsnede, die al het water vervoert, en anderzijds het kombergende deel, waar het water niet in langsrichting stroomt, zie figuur 4.4. De stroomvoerende dwarsdoorsnede heeft een oppervlak  $A_S(x,t)$ , die het produkt is van een stroomvoerende breedte  $b_S(x,t)$  en een representatieve instantane diepte H(x,t). De representatieve instantane diepte is de som van de representatieve stroomvoerende waterdiepte h(x,t) t.o.v. een horizontaal middenstandsvlak en de de getijstand  $\eta(x,t)$ . De gemiddelde stroomsnelheid in de stroomvoerende dwarsdoorsnede wordt aangeduid als  $u_S(x,t)$ . In formulevorm:

Stroomvoerende doorsnede  $A_S = b_S H = b_S (h + \eta)$ , Debiet  $Q = Au = A_S u_S$ .

De dubbele overstreping van de stroomsnelheid u staat voor middeling over de totale dwarsdoorsnede.

De één-dimensionale balansvergelijkingen worden afgeleid voor een infinitesimaal korte sectie in het bekken.

De massabalans (continuïteits-vergelijking) luidt:

 $(A_s u_s)_x + b_K \eta_t = 0$ volumeverandering volumeverandering door in- en uittredend transport door waterstandsvariatie

De massabalansvergelijking kan ook geschreven worden als

$$\frac{1}{b_x}(b_sHu_s)_x + \eta_t = 0$$

De impulsbalans luidt

$$(A_{S}u_{S})_{t} + (A_{S}u_{S}^{2})_{x} + (b_{K}-b_{S})\eta_{t}u_{S} + gA_{S}\eta_{x} + \frac{b_{S}}{\rho}\tau_{b} = 0$$

impulsverandering u

uit-intredend impulstransport in langs- en dwars-richting drukgradiënt bodemwrijving

ha

De laatste vergelijking kan worden vereenvoudigd tot

$$u_t + uu_x + g\eta_x + c_D \frac{|u|u}{H} = 0$$

Voortaan wordt de index S bij de stroomsnelheid weggelaten.

Ook voor de getij-energie kan een balansvergelijking worden opgesteld.

De getij-energie per eenheid van lengte is gegeven door

$$E = \langle b_K | \frac{l}{2} g \rho \eta^2 + A_S | \frac{l}{2} \rho u^2 \rangle$$

waar <> staat voor middeling over de getijperiode. We gaan uit van een cyclisch getij, zodat  $E_t = 0$ .

De energiebalans luidt

$$\frac{\partial}{\partial x} < A_s u \left(\frac{1}{2} \rho u^2 + \rho g \eta\right) > + < (b_K - b_S) \eta_t \frac{1}{2} \rho u^2 > = - < b_S \tau_b u >$$
  
in-uittredend energietransport in langs- en dwarsrichting energiedissipatie  
door bodemwrijving

Deze vergelijking laat zich niet schrijven in de gebruikelijke vorm van een balanstransportvergelijking van energiedichtheid. Dit hangt samen met het feit dat getij-energie niet alleen door advectief transport wordt overgedragen, maar ook door druk. Voor een symmetrische (sinusoïdale) getijgolf geldt:

$$< A_{s}u \frac{1}{2}\rho u^{2} > \approx < (b_{K}-b_{S}) \eta_{t} \frac{1}{2}\rho u^{2} > \approx 0$$
.

Als verder verondersteld wordt dat het getij zich voortplant als een lopende golf, en als de bovenafvoer wordt verwaarloosd, dan is het linkerlid van de energiebalans ongeveer gelijk aan

$$< A_S \rho g u \eta > \approx b_S c E, \quad E = \frac{1}{2} g \rho a^2$$

Hier blijkt dat in een symmetrische, lopende getijgolf de getijenergie wordt getransporteerd met de golfsnelheid  $c = \sqrt{gh}$ .



Figuur 4.4: karakteristiek getijrivieren

#### 4.4.2 Schalings-analyse

Om vereenvoudigingen in de getijvoortplantingsvergelijkingen te kunnen doorvoeren wordt een schalingsanalyse uitgevoerd, waaruit blijkt welke termen onder welke omstandigheden van vergelijkbare grootte-orde zijn, en welke termen in eerste instantie verwaarloosd kunnen worden.

De volgende schalingen worden ingevoerd:

$$\stackrel{=}{u} = \{u\} \ u^*, \ \eta = \{\eta\} \ \eta^*, \ H = \{H\} \ H^*, \ t = \{t\} \ t^*.$$

Alle variabelen aangeduid met een \* zijn van grootte-orde 1. De accolades staan voor de grootte-orde van de niet-geschaalde variabelen; dit zijn constante waarden. De geometrie van de getijrivier wordt bij benadering uniform verondersteld. Er resteren dan 2

lengteschalen: de getijgolflengte  $L = \frac{2\pi}{\omega} \sqrt{gh}$ , en de lengteschaal van getij-indringing.

We kiezen voor  $\{x\}$  deze laatste lengteschaal, omdat die het meest bepalend is voor de variatie van de getijparameters.

Verder is  $\{\eta\} = a, \{t\} = 1/\omega, \{H\} = h$ .

Uit substitutie in de continuïteitsvergelijking volgt  $\{u\} \approx \frac{b_K}{b_S} \frac{a}{h} \otimes \{x\}$ .

Noem  $P = \frac{b_K}{b_S} \left( 2\pi \left\{ x \right\} / L \right)^2$ , dan wordt de impulsbalans

$$P u_{t^{*}}^{*} + P \frac{b_{K}}{b_{S}} \frac{a}{h} u^{*} u_{x^{*}}^{*} + \eta_{x^{*}}^{*} + P \frac{b_{K}}{b_{S}} \frac{a}{h} \frac{\{x\}}{h} c_{D} \frac{|u^{*}| u^{*}}{H^{*}} = 0$$

1 + 1 -

Voor karakteristieke orde-groottes

$$a / h = O[10^{-1}], b_K / b_S = O[1], h \approx O[10m]$$

geldt dat de tweede term in het linkerlid kan dan worden verwaarloosd. Als ook de wrijving verwaarloosbaar is, dan volgt  $P \approx 1$  en  $\{x\} \approx L/2\pi$ . Bij sterke wrijving geldt dat de wrijvingsterm van dezelfde orde-grootte moet zijn als de verhangterm:

$$P \frac{b_K}{b_S} \frac{a}{h} \frac{\{x\}}{h} c_D \approx 1 , \quad \{x\} \approx \left(\frac{2\pi}{T\sqrt{g}} \frac{\sqrt{h}}{\frac{a}{h} c_D}\right)^{\frac{1}{3}} \approx 0.8 \frac{L}{2\pi}$$

(gesteld dat  $a/h \approx 0.1$ ,  $c_D = 0.003$ ).

Bij sterke wrijving is  $\{x\}$  kleiner is dan bij geringe wrijving; wrijvingsinvloeden verminderen de getij-indringing dus in aanzienlijke mate.

Er is sprake van wrijvingsgedomineerde stroming bij:  $c_D$  groot, *h* klein, *a/h* groot. De eerste twee termen in de impulsbalansvergelijking kunnen dan in eerste benadering worden verwaarloosd. In veel getijdebekkens is dit het geval.

In de meeste getijdebekkens kan de geometrie niet constant worden verondersteld over de lengte van getij-indringing. Dit betekent dat in de schalingsanalyse ook een geometrische lengteschaal ingevoerd moet worden. Als deze geometrische lengteschaal veel kleiner is dan de getij-indringingslengte dan spelen in de massa-balansvergelijking andere termen een belangrijke rol dan in bovenstaande analyse. Lopende getijgolf bij wrijvingsgedomineerde stroming

Bij wrijvings-gedomineerde stroming en uniforme geometrie kunnen de getijvergelijkingen worden vereenvoudigd tot:

$$g \eta_x + c_D \frac{|\mathbf{u}|u}{H} = 0$$
,  $b_S H u_x + b_K \eta_t = 0$ 

Door eliminatie van η uit deze vergelijkingen volgt:

$$u_t = \Xi \ u_{xx}$$
,  $\Xi = \frac{g \ b_S \ H^2}{2 \ b_K \ c_D \ |u|}$  = diffusiecoëfficient.

Dit is een diffusievergelijking. Het getij plant zich in eerste benadering dus niet voort als een golf, maar wordt in het bekken gediffundeerd. Perioden rond stroomkentering worden echter niet goed beschreven door de getij-diffusie vergelijking; wrijving is in deze perioden niet dominant, het getij plant zich dan bij benadering voort als een golf. Vooral door de afhankelijkheid van de stroomsterkte varieert de getij-diffusiecoëfficient sterk gedurende de getij-periode. De getij-diffusievergelijking is dus niet lineair, en kan daarom niet met elementaire methoden opgelost worden. Het niet-lineare karakter van de vergelijking betekent dat de getijgolf tijdens de diffusie door het bekken vervormd wordt en zijn sinusoïdale karakter verliest. Dit brengt ook met zich mee dat de symmetrie tussen de eb en de vloed verbroken wordt. Het gedrag van de model-golf die door de diffusievergelijking beschreven wordt benadert de echte getijgolf alleen in perioden waar de stroomsnelheid voldoende groot is, en de wrijvingsterm belangrijker is dan de versnellingstermen. We veronderstellen dat dit gedurende het grootste deel van het getij het geval is, zodat de model-golf een redelijke algemene indruk geeft van de echte getijgolf.

In het vervolg zoeken we naar een benaderde oplossing van de niet-lineaire getijdiffusievergelijking. Uit oplossing van de diffusievergelijking zal blijken dat de maximale vloedsnelheid optreedt kort vóór HW (een achtste getijperiode); tussen de tijdstippen van maximale ebsnelheid en LW bestaat een zelfde tijdsinterval. Rond maximale vloedsnelheid en ebsnelheid is de variatie van de stroomsnelheid gering. Rond HW en LW is de variatie van de waterstand gering. Dat betekent dat er zowel tussen maximale vloed en HW als tussen maximale eb en LW een interval bestaat waarin H en u weinig variëren. Voor deze beide perioden beschouwen we H en  $u_{\tau}$  als constanten, respectievelijk

 $H \approx H^{\pm} = h + \eta^{\pm}$ ,  $u_T \approx u_T^{\pm}$  als constanten voor de respectieve perioden. De getijdiffusiecoëfficiënt mag hier dan bij benadering constant verondersteld worden. We veronderstellen dat de stroming in die beide intervallen wrijvingsgedomineerd is.

We beschouwen nu een getijdebekken dat zo lang is dat de getijgolf aan het bekkeneinde bijna is uitgedempt. Dit was bijvoorbeeld het geval in de voormalige Zuiderzee, waar de via het Marsdiep inkomende getijgolf achterin (bij het IJ en bij de Veluwerand) grotendeels was uitgedempt. Ook in getijrivieren is sprake van een geleidelijk uitdempende lopende getijgolf, voorbeelden zijn de Lek en Waal in Nederland en de St. Lawrence rivier in Canada. In deze getijrivieren is de teruggekaatste getijgolf verwaarloosbaar klein, zodat het getij het karakter heeft van een naar binnen diffunderende golf. In de perioden dat H en u weinig variëren kan deze golf geparametreerd worden als

 $u \propto e^{i\kappa x - i\omega t}$  (in complexe notatie; alleen het reële deel telt).

Uit substitutie in de getijvergelijkingen volgt voor het HW-interval (resp. het LW-interval) bij benadering:

$$\eta = a \ e^{-kx + ik(x - c(t - \frac{T}{s}))}, \quad u = -u_R + u_T, \quad u_T = \frac{1}{\sqrt{2}} \ \frac{a}{h} \ c \ e^{-kx + ik(x - ct)}$$

Bovenstaande uitdrukkingen gelden alleen voor respectievelijk de HW en de LW perioden; het golfgetal k en de voortplantingssnelheid c zijn voor deze perioden gegeven door:

$$k^{\pm} = \sqrt{\frac{c_D \ b_K^{\pm} \omega \left| -u_R + u_T^{\pm} \right|}{g b_S^{\pm} H^{\pm 2}}}, \qquad c^{\pm} = \frac{\omega}{k^{\pm}} ,$$
$$H^{\pm} = h + \eta^{\pm}, \quad \eta^{\pm} \approx \pm a (1 - e^{-1}), \quad u_T^{\pm} \approx \pm \frac{1}{2} (1 - e^{-1}) \frac{a \omega}{h k^{\pm}}$$

De laatste uitdrukkingen zijn verkregen door te veronderstellen dat de getijamplitude over het traject van getij-indringing ongeveer exponentieel wordt gedempt.

De stroomsnelheid heeft twee componenten:  $u_R$ , de stroomsnelheid veroorzaakt door rivierafvoer alleen, en  $u_T$ , de stroomsnelheid ten gevolge van het getij. De eerste is in getijdebekkens over het algemeen klein t.o.v. de laatste. In het vervolg veronderstellen we meestal  $u \approx u_T$ .

In getijrivieren, waar de rivierstroomsnelheid en de getijstroomsnelheid van dezelfde orde zijn, geldt deze veronderstelling niet. In getijrivieren is het golfgetal (tevens de omgekeerde dempingslengte) aanzienlijk kleiner voor de HW-periode dan voor de LW-periode:

 $k^+ \prec k^-$ . Dit wordt vooral veroorzaakt door de veel grotere stroomsnelheid bij maximale eb dan bij maximale vloed. Hierdoor is de demping in de eb-periode rond LW sterker dan in de vloed-periode rond HW. Tevens is de voortplantingssnelheid  $c^* = \omega / k^*$  van het HW aanzienlijk groter dan de voortplantingssnelheid  $c^- = \omega / k$  van het LW. Door dit verschil in voortplantingssnelheid vervormt de getijkromme: de periode van waterstands-stijging is korter dan de periode van waterstandsdaling. Dit impliceert dat er in stroomopwaartse richting altijd sprake is van een relatieve toename van vloedsnelheden en een relatieve afname van ebsnelheden, zie figuren 5 en 6.

In lange getijdebekkens waar de rivierafvoer verwaarloosbaar is t.o.v. het getijdebiet zal o.h.a. ook getijasymmetrie ontstaan, maar deze wordt door andere faktoren beïnvloed. Enerzijds is dit de verhouding tussen kombergende en stroomvoerende breedte, die bij HW en LW sterk verschillend kan zijn. Dit speelt ook voor kortere getijdebekkens, waar de getijgolf een staand karakter heeft. Hier wordt in het vervolg meer in detail op ingegaan. Anderzijds is in lange getijdebekkens het faseverschil tussen waterstand en stroomsnelheid van invloed op de getijasymmetrie. In een lopende golf is dit faseverschil kleiner dan  $\pi/2$ , zodat de waterstand gemiddeld over de vloed hoger is dan gedurende de eb. Omdat bij eb eenzelfde volume water het bekken verlaat als er tijdens vloed is ingestroomd, maar bij een kleinere stroomvoerende doorsnede, moeten de ebsnelheden gemiddeld groter zijn dan de vloedsnelheden. Dit zou impliceren dat bij lange getijdebekkens het sedimenttransport bij eb altijd groter is dan bij vloed. In het vervolg zullen we zien dat andere geometrische effekten deze ebdominantie teniet kunnen doen. Niettemin is het ebdominante gedrag van een gedempte getijgolf een belangrijke oorzaak voor de vorming van een getijdebekken in laagvlaktes na dijkdoorbraak.



Vervorming getij bij snellere voortplanting HW tov LW; periode van getijstijging wordt korter en van getijdaling langer.





Bij snellere voortplanting HW tov LW wordt het HW minder snel gedempt dan het LW.





getij stroomopwaarts

resultaat van getijvorming en fase-verschuiving u-η Figuur 4.5: Vervorming getijgolf in getijrivier


Figuur 4.6: Getijvervorming op de rivier de Lek

#### 4.4.3 Staande getijgolf, resonantie

Als de getijgolf niet sterk gedempt wordt in het bekken vindt terugkaatsing plaats aan het landwaartse bekkeneinde. We beschouwen nog steeds een uniform getijdebekken waarin de getijgolf vooral de hoofdgeul volgt, zodat de getijvoortplanting bij benadering ééndimensionaal beschreven kan worden. De stroming wordt geschematiseerd volgens een bij benadering uniform profiel in het stroomvoerende deel van de doorsnede en daarbuiten géén netto stroming in x-richting. De 1-dimensionale waterbewegingsvergelijkingen kunnen dan worden geschreven als:

massabalans  $h_K u_x + \eta_t = 0$ , impulsbalans  $u_t + g\eta_x + c_D \frac{|u|u}{H} = 0$ .

De impuls-transportterm  $uu_x$  is hier verwaarloosd, maar de versnellingsterm niet. Hier is u niet de dwarsdoorsnede-gemiddelde stroomsnelheid, maar het quotiënt van debiet en stroomvoerende doorsnede:  $u = Q / A_S$ . Verder is  $h_K$  een equivalente diepte die het verschil tussen stroomvoerende en kombergende diepte verdisconteert,  $h_K = A_S / b_K = Hb_S / b_K$ . Bij zeer klein getij (a/h << 1) kan  $h_K$  als een constante

worden opgevat.

Eerst beschouwen we het (meestal niet realistische) geval van verwaarloosbare wrijving,  $c_D = 0$ . In dat geval wordt de getijgolf onvervormd gereflecteerd bij het bekkeneinde x=l; het resultaat is een staande getijgolf:

$$\eta(x,t) = \frac{1}{2}a_l \left[\cos(k(x-l) - \omega t) + \cos(k(x-l) + \omega t)\right] = a \frac{\cos(k(x-l))}{\cos kl} \cos \omega t$$
$$u(x,t) = \frac{1}{2}\frac{a c}{h_K \cos kl} \left[\cos(k(x-l) - \omega t) - \cos(k(x-l) + \omega t)\right] = U \frac{\sin(k(x-l))}{\sin kl} \sin \omega t$$

Hier is 
$$a_l = \frac{a}{coskl}$$
,  $U = u(0,0) = \frac{ac}{h_K} \tan kl$ 

De snelheid aan de landwaartse rand x=l is nul. Aan de landwaartse rand is de getijamplitude groter dan aan de zeerand. Waterstand en snelheid variëren 90° uit fase; de gemiddelde waterstand bij eb en vloed is dus dezelfde. Verder betekent dit dat er geen influx of outflux van getijenergie  $E = \frac{1}{2}\rho u^2 + \rho g\eta$  optreedt, want < uE >= 0.

Als de bekkenlengte gelijk is aan een kwart van de getijgolflengte treedt resonantie op:

$$l = \frac{1}{4}L = \frac{1}{4}T\sqrt{gh_K} = \frac{\pi}{2k} \implies coskl = 0 \implies U, a_l \rightarrow \infty$$

In dit geval mag bodemwrijving zeker niet verwaarloosd worden.

#### 4.4.4 Gedempte teruggekaatste getijgolf

In de meeste getijdebekkens kan het getij beschreven worden als superpositie van een ingaande en teruggekaatste gedempte getijgolf. Deze wordt uit de vergelijkingen afgeleid voor een klein getij ( $a/h \ll 1$ ) en voor een gelineariseerde beschrijving van de wrijvingsterm:

$$c_D \frac{|u|u}{H} \rightarrow r \frac{u}{h}$$
,  $r \approx \frac{8}{3\pi} c_D |u|$ 

Deze laatste grootheid (de gelineariseerde wrijvingscoëfficiënt) wordt constant verondersteld.

Substitutie in de getijvergelijkingen geeft de volgende oplossing:

$$\eta = \frac{1}{2} a_{l} \left[ e^{-\mu(x-l)} \cos(k(x-l) - \omega t) + e^{\mu(x-l)} \cos(k(x-l) + \omega t) \right] \quad \text{en}$$

$$u = \frac{1}{2} \frac{a_{l}}{h_{K}} \frac{\omega}{\sqrt{k^{2} + \mu^{2}}} \left[ e^{-\mu(x-l)} \cos(k(x-l) - \omega t - \phi) - e^{\mu(x-l)} \cos(k(x-l) + \omega t + \phi) \right]$$

Hier is 
$$a_l = \frac{a}{\sqrt{\cos^2 kl \cosh^2 \mu l + \sin^2 kl \sinh^2 \mu l}}$$
, en

$$k = \frac{\omega}{\sqrt{2g h_K}} \sqrt{1 + \sqrt{1 + \beta^2}}, \quad \mu = \frac{\omega}{\sqrt{2g h_K}} \sqrt{-1 + \sqrt{1 + \beta^2}}, \quad \cos\phi = \frac{k}{\sqrt{k^2 + \mu^2}}$$
  
met  $\beta = \frac{r}{h\omega}$ .

In het geval dat de wrijvingsterm dominant is over de versnellingsterm (het meest voorkomende geval) is  $\beta > 1$ . Soms geldt zelfs  $\beta >> 1$ , zodat  $k \approx \mu \approx \sqrt{\frac{r\omega}{2ghh_K}}$ In dat laatste geval is u ongeveer 45° uit fase met  $\eta$ . Dit komt overeen met het beeld van een getijgolf die zich voortplant door diffusie. Ook als  $kl = \frac{\pi}{2}$  blijven  $a_l$  en U eindig. Het feit dat energie in het bekken wordt gedissipeerd komt tot uiting in de netto influx van getijenergie:  $\langle uE \rangle > 0$ 

#### 4.4.5 stralingsdemping

Behalve door bodemwrijving vindt er ook energie-verlies plaats door uitstraling in de monding. Het bekken fungeert als een impuls-bron  $u = -U \sin \omega t$  voor de open zee. Deze impuls-uitstraling veroorzaakt een aanpassing van het getij bij de bekken-ingang

$$\eta \rightarrow \eta + \eta_r$$
,  $\eta_r = a_r \cos \omega t$ 

Uit de voortplantingsvergelijking van de impuls-bron in de open zee volgt

$$a_r = F_r \frac{hU}{c}$$

Hier is  $u_0$  de door bekken-resonantie opgewekte stroomsnelheid in de monding,

$$u_0 = -\frac{a+a_r}{h}c tankl$$
, en  $F_r$  is een faktor gegeven door

$$F_r = -\frac{2}{\pi} \frac{1}{k} \frac{h}{h_z} \frac{d}{dr} H_0^{(2)} \left( \frac{\omega r}{\sqrt{gh_z}} \right) \bigg|_{r=\frac{1}{2}b}, \qquad H_0^{(2)} = Hankelfunctie$$

De Hankelfunctie  $H_0^{(2)}\left(\frac{\omega r}{\sqrt{gh_z}}\right)$  beschrijft het radiale verloop van een getijgolf die

uitstraalt naar een oneindig grote zee met uniforme diepte  $h_z$ . Deze golf onttrekt getijenergie uit het bekken:  $< uE > |_{r=\frac{1}{2}b} < 0$ .

Uit de voorgaande vergelijkingen volgt  $a_r = -a \frac{F_r \tan kl}{1 + F_r tankl}$ .

De opslingering in het bekken blijft dus beperkt tot

$$\left|\frac{a_l}{a+a_r}\right| = \frac{1}{\left|\cos kl + F_r \sin kl\right|} < \infty.$$

## 4.5 Getij-vervorming; eb-vloed asymmetrie

In de meeste getijdebekkens wijken de getijkrommen voor waterstand en stroomsnelheid (of debiet) sterk af van een sinusoïdaal verloop. Dit is te zien in figuur 4.7, voor een aantal Nederlandse getijdebekkens. De achterliggende processen worden in deze paragraaf besproken. We zullen zien dat de oorzaak van de getijvervorming ten dele in de morfologische karakteristieken van de bekkens zelf gelegen is.

#### 4.5.1 Invloed bodemwrijving

Bodemwrijving is vooral van belang bij lange getijdebekkens (lengte van tenminste enkele tientallen kilometers) met maximale getijsnelheden van tenminste 1 m/s. De bodemwrijving introduceert eb-vloed asymmetrie in het dat geval de getijamplitude niet t.o.v. de bekkendiepte verwaarloosd wordt, d.w.z. als de dieptes H en  $H_K = Hb_S / b_K$  over het getij variëren en dus niet als constantes worden beschouwd. In dat geval zijn de getijvergelijkingen niet meer lineair, en laten ze zich niet meer eenvoudig oplossen. Om toch een beeld te krijgen van de getijvervorming en de getijasymmetrie gebruiken we een soortgelijke veronderstelling als eerder bij de lopende getijgolf. In een periode gelegen tussen hoog water (HW) en hoogwater kentering (HWK) beschouwen we H en  $H_K$  als

constantes met waarden  $H^+$  en  $H_K^+$ . Andere afgeleide grootheden in deze periode die constant worden verondersteld zijn aangeduid met de superscript + . Ook in een periode gelegen tussen laag water (LW) en laagwater kentering (LWK) beschouwen we H en  $H_K$ 

als constantes, nu met waarden  $H^-$  en  $H_K^-$ . Andere afgeleide grootheden in deze periode die constant worden verondersteld zijn aangeduid met de superscript - .

We gaan uit van de eerder afgeleide uitdrukkingen voor u en  $\eta$  :

$$\eta = \frac{1}{2} a_{l} \left[ e^{-\mu(x-l)} \cos(k(x-l) - \omega t) + e^{\mu(x-l)} \cos(k(x-l) + \omega t) \right] \quad \text{en}$$

$$u = \frac{1}{2} \frac{a_{l}}{H_{K}} \frac{\omega}{\sqrt{k^{2} + \mu^{2}}} \left[ e^{-\mu(x-l)} \cos(k(x-l) - \omega t - \phi) - e^{\mu(x-l)} \cos(k(x-l) + \omega t + \phi) \right]$$

Deze uitdrukkingen blijven geldig voor bovengenoemde twee perioden, maar met voor elke periode een verschillend golfgetal, voortplantingssnelheid en fase (resp.  $k^{\pm}, c^{\pm}, \phi^{\pm}$ ).



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In deze uitdrukkingen is de tijd-as zodanig gekozen dat t=0 samenvalt met HW aan het bekkeneinde x=l en dat t = T/2 samenvalt met LW aan het bekkeneinde. Ten opzichte van deze tijdstippen kunnen de tijdstippen van kentering bepaald worden, resp. HWK en LWK, door oplossing van de vergelijking u(x,t) = 0:

$$e^{-\mu(x-l)}\cos(k(x-l) - \omega t - \phi) = e^{\mu(x-l)}\cos(k(x-l) + \omega t + \phi)], \text{ of } \cos(\omega t + \phi)\cos(k(x-l)[e^{-\mu(x-l)} - e^{\mu(x-l)}] = -\sin(\omega t + \phi)\sin(k(l-x)[e^{-\mu(x-l)} + e^{\mu(x-l)}])$$

In de relatief korte bekkens die we hier beschouwen geldt altijd dat  $(\,\mu l)^2\,\,$  » 1 ,  $\,(kl)^2\,$  » 1 .

In dat geval geldt: 
$$\tan(\omega t + \phi) = \frac{\tanh(\mu(x-l))}{\tan(k(x-l))} \approx \frac{\mu}{k}$$

Hieruit blijkt dat kentering overal in het bekken ongeveer gelijktijdig optreedt. Omdat, als we het bekkeneinde naderen ( $x \rightarrow l$ ), hoogwater en hoogwaterkentering samenvallen, hebben we:

$$t_{HWK}(x) - t_{HWK}(l) \approx t_{HWK}(x) - t_{HW}(l) \approx 0,$$
  

$$t_{LWK}(x) - t_{LWK}(l) \approx t_{LWK}(x) - t_{LW}(l) \approx 0$$

Door de vergelijking  $\eta_t = 0$  op te lossen kunnen we hoogwater en laagwater tijdstippen in het bekken bepalen op verschillende locaties ten opzichte van elkaar. Dit gaat op dezelfde wijze als voor u=0. Het resultaat is

$$\tan(\omega t) = -\tan(k(x-l))\tanh(\mu(x-l)) \approx -k\mu(x-l)^2 \approx \omega t$$

Dat betekent dat:

$$t_{HW}(x) - t_{HW}(l) \approx t_{HW}(x) - t_{HWK}(l) \approx \frac{k^{+}\mu^{+}}{\omega} [l^{2} - (x - l)^{2}]$$
  
$$t_{LW}(x) - t_{LW}(l) \approx t_{LW}(x) - t_{LWK}(l) \approx \frac{k^{-}\mu^{-}}{\omega} [l^{2} - (x - l)^{2}]$$

Laten we nu veronderstellen dat het getij aan de zeerand (x=0) symmetrisch is, d.w.z.

$$t_{HW}(0) - t_{LW}(0) = \frac{1}{2}T \quad .$$

Hieruit volgt dat het verschil in tijdsduur tussen eb en vloed in het zeegat bij benadering is gegeven door

$$\Delta t_{flood} - \Delta t_{eb} \approx \frac{2\pi}{\omega} l^2 \left[ k^+ \mu^+ - k^- \mu^- \right] .$$

Door substitutie van de uitdrukkingen van k en  $\mu$  vinden we tenslotte

$$\Delta t_{flood} - \Delta t_{eb} \approx \pi \frac{rl^2}{g} \left[ \frac{1}{H^+ H_K^+} - \frac{1}{H^- H_K^-} \right]$$
, met andere woorden:

 $\Delta t_{flood} - \Delta t_{eb} \propto -H^+ H_K^+ + H^- H_K^-$ 

Een verschil tussen vloedduur en ebduur betekent dat de gemiddelde stroomsnelheid bij vloed verschilt van de gemiddelde stroomsnelheid bij eb, en dat het gemiddelde sedimenttransport bij vloed verschilt van het gemiddelde sedimenttransport bij eb. Bij getijdebekkens is de kombergende breedte bij HW vaak veel groter dan de kombergende breedte bij LW, vooral als er grote platengebieden zijn. In dat geval is

 $(b_K / b_S)^+ >> (b_K / b_S)^-$ . Als bovendien de geulen diep zijn (*a/n* << 1), dan volgt dat  $H^+ \approx H^-$ , d.w.z.  $(HH_K)^+ << (HH_K)^-$ . Dit betekent dat de vloedperiode langer is dan de ebperiode. Bijgevolg is de maximale ebsnelheid groter dan de maximale vloedsnelheid. Voor getijdebekkens met weinig kombergend plaatoppervlak en ondiepe geulen geldt het omgekeerde

 $((b_K / b_S)^+ \approx (b_K / b_S)^-, H^+ > H^-, (HH_K)^+ > (HH_K)^-)$ ; de maximale vloedsnelheid is groter dan de maximale ebsnelheid. In getijdebekkens is er dus sprake van getijasymmetrie en resulterend sedimenttransport die zijn gerelateerd aan de bekkenmorfologie.

#### 4.5.2 Korte getijdebekkens; invloed bekkengeometrie

De bekkengeometrie beïnvloed ook in kortere getijdebekkens de getij-asymmetrie. Het gaat om en ander soort asymmetrie dan voor de lange getijdebekens, namelijk een verschil tussen de stroomsnelheidsvariatie rond HW- en LW-kentering. Zoals later zal blijken is deze asymmetrie van groot belang voor het netto transport van fijn sediment in het getijdebekken. In het geval een kort getijdebekken (lengte veel kleiner dan een kwart van de getijresonantielengte, amplitudes van inkomende en gereflecteerde getijgolf van dezelfde orde) is het getijfaseverschil tussen zeerand en bekkeneinde klein en kunnen waterstandsverschillen over het bekken verwaarloosd worden. De continuïteits-vergelijking kan dan worden geïntegreerd over de lengte van het bekken:

$$u(x,t) = \frac{\eta_t(t)}{A_s(x,t)} \int_x^l b_K(x,t) dx \approx (l-x) \frac{\eta_t(t)}{H_K(t)}$$

Hier is voor de laatste uitwerking de dwarsdoorsnede bij benadering uniform verondersteld. In een kort bekken zijn  $\eta$  en u uit fase; bij HWK (hoogwater-kentering) en LWK geldt dat  $\eta_t \approx 0$ . Hieruit volgt voor het snelheidsverloop rond HWK t.o.v. het snelheidsverloop rond LWK:

$$\frac{|u_t|_{HWK}}{|u_t|_{LWK}} \approx \frac{H_K^-}{H_K^+} \frac{|\eta_t|_{HWK}}{|\eta_t|_{LWK}}$$

Stel dat het getij op zee rond HW symmetrisch is t.o.v. LW, dan is de laatste factor ongeveer gelijk aan 1. Bovenstaande vergelijking zegt ons dan dat in een bekken met uitgestrekte

kombergende platen en diepe geulen ( ${H_K}^+ < {H_K}^-$ ) er een snel verloop van de getijsnelheid is rond HWK t.o.v. LWK: de HW-kenteringsduur is korter dan de LW-kenteringsduur. In een bekken met ondiepe geulen en weinig kombergende platen

 $({H_K}^+ > {H_K}^-)$  is er een relatief langzaam verloop van de getijsnelheid rond HWK t.o.v. LWK: de HW-kenteringsduur is langer dan de LW-kenteringsduur. In het eerste geval zal er een sterkere sedimentatie plaats vinden bij LWK dan bij HWK, in het tweede geval zal er een sterkere sedimentatie optreden bij HWK.

#### 4.5.3 Invloed zee-getij

De asymmetrie van het snelheidsverloop in een getijdebekken hangt sterk af van het verloop van de waterstand op zee. Dit geldt vooral voor korte bekkens, maar in iets mindere mate ook voor langere getijdebekkens.

Een snellere getijrijzing dan getijdaling op zee veroorzaakt hogere vloed- dan ebsnelheden in het bekken. Een langzamere rijzing heeft het omgekeerde effect.

Een langdurig HW op zee veroorzaakt een lange kenteringsduur en een kortdurend HW een korte kenteringsduur. Hetzelfde geldt voor het LW. Vervorming van de getijgolf langs de kust beïnvloedt de eb-vloed asymmetrie in de aanleggende bekkens en zodoende het netto in- of uitgaande sedimenttransport en de morfologische ontwikkeling. Figuur 4.8 laat het

getijverloop zien voor de Nederlandse kust en in de aangrenzende getijdebekkens; hierin is te zien dat de getij-asymmetrie in het bekken doorwerkt.



Figuur 4.8: Getijverloop van de waterstand langs de Nederlandse kust en in de aangrenzende bekkens

## 4.6 Intermezzo: sedimenteigenschappen

Het bodemmateriaal in kustgebieden wordt gevormd door verschillende typen sedimenten. Een belangrijke karakteristiek waardoor sedimentdeeltjes zich onderscheiden is de diameter. Deze kan sterk varieren, van slib (diameter ca. 10<sup>-6</sup> m) tot zand (diameter ca. 10<sup>-4</sup> m) en grind (diameter ca. 10<sup>-3</sup> m). Het soortelijk gewicht (s.g.) van de sedimentdeeltjes bedraagt gemiddeld 2,5 tot 3 keer het s.g. van water; maar zware mineralen met een veel groter s.g. kunnen ook voorkomen.

Doordat het s.g. van sedimentdeeltjes groter is dan van water, zullen de deeltjes bezinken, bij afwezigheid van turbulente of opstijgende waterbewegingen. Voor kleine sedimentdeeltjes kan de gemiddelde valsnelheid in water bij benadering worden afgeleid van de wet van Stokes



Deze geldt voor Reynoldsgetallen kleiner dan 1, d.w.z. voor kleine sedimentdeeltjes. Zulke kleine sedimentdeeltjes hebben echter de neiging te klonteren, zodat vlokken met een veel grotere diameter gevormd worden (figuur 4.9). Vlokken hebben daardoor een veel grotere valsnelheid dan de samenstellende deeltjes, ook al is de vlok-dichtheid geringer dan de dichtheid van de samenstellende deeltjes.

Figuur 4.9: Vlokvormin van sedimentdeeltjes

Bij vlokvorming spelen meerdere faktoren een rol:

- de deeltjes-concentratie. De vlokgroei is sterker en de omvang van de vlokken neemt toe naarmate deeltjes vaker met elkaar in contact komen. Uit metingen blijkt dat hierdoor de valsnelheid van vlokken gevormd bij hogere concentratie groter is dan bij lagere concentratie ( ongeveer lineair of kwadratisch toenemend met de concentratie). Bij zeer hoge concentratie (b.v. dicht bij de bodem) hinderen deeltjes elkaar bij het uitzakken en neemt de valsnelheid af met toenemende concentratie. Men spreekt van "hindered settling".
- de turbulentiegraad. Turbulentie draagt bij aan de frequentie waarmee deeltjes elkaar ontmoeten en bevordert daarmee vlokgroei. Bij zeer hoge turbulentie en dito schuifspanning (b.v. bij de bodem) worden grote vlokken echter stukgetrokken, waardoor de valsnelheid afneemt en sedimentdeeltjes weer worden opgewerveld.
- het zoutgehalte. Kleideeltjes bezitten oppervlakte-ladingen waardoor deeltjes onderling kunnen hechten. De aanwezigheid van Cl<sup>-</sup> ionen bevordert dit aanhechtingsproces.
- de aanwezigheid van organische stoffen en bacteriën en andere organismen. Vlokken worden onderling vaak verbonden door grote moleculen (suikers, eiwitten) en draden die door organismen worden afgescheiden in het water.

Vlokken zijn sterk variabel van samenstelling. De diameter kan zo'n 1000 keer groter zijn dan die van de samenstellende deeltjes. Zulke grote vlokken zijn echter heel fragiel. Voor de valsnelheid is niet alleen de diameter van belang, maar ook de dichtheid (watergehalte) en de "drag" (hydrodynamische vorm). In de praktijk kunnen valsnelheden van (gevlokte) sedimentdeeltjes alleen proefondervindelijk bepaald worden (figuren 10 en 11).



Figuur 4.10: Valsnelheid van sedimentdeeltjes als functie van de concentratie (fijn cohesief sediment)



Figuur 411: Verscheidenheid aan valsnelheden van fijn sediment in verschillende kustsystemen

Door het uitzakken van sedimentdeeltjes kan bij de bodem een suspensie ontstaan met een significant hogere dichtheid dan het bovenliggende water (bij voldoende turbulente menging in de bodemlaag). Als turbulente uitwisseling tussen onder- en bovenlaag door het dichtheidsverschil voldoende wordt onderdrukt, ontstaat een scherp grensvlak. Versneld uitzakken door vlokvorming in de bovenlaag en "hindered settling" (botsingen) in de onderlaag dragen bij aan de vorming van een troebele bodemlaag ("lutocline"). Deze laag wordt gestabiliseerd door de aanwezigheid van een dichtheidsverschil met het bovenliggende water.

De mate waarin gesedimenteerd materiaal later weer door de stroming opgepakt wordt is ook van een groot aantal faktoren afhankelijk. Dit is, onder andere, de consolidatie- of pakkingsgraad van het sediment, die toeneemt naarmate het sediment ouder is, onder het eigen gewicht wordt samengedrukt en ontwatert. Bij een losgepakte, waterverzadigde bodem maakt de waterspanning dat deeltjes van elkaar los kunnen komen, zodat ze gemakkelijk opgewerveld en meegenomen worden door de stroming. Een bodem waarbij de poriën tussen grovere sedimentdeeltjes zijn opgevuld met fijne sedimentdeeltjes laat zich minder gemakkelijk eroderen. De bodem is dan namelijk van de waterkolom afgesloten, zodat drukfluctuaties worden opgevangen door de fijne sedimentdeeltjes en nauwelijks doorgegeven worden naar de waterspanning in de bodem. Ook spelen cohesieve eigenschappen een rol. Deze worden vooral veroorzaakt door organismen die kleefstoffen afscheiden; op de bodem kunnen b.v. algen-matten worden gevormd. Daarnaast zijn er allerhande organismen die het sediment omwoelen (men spreekt van bioturbatie). Ook spelen door bacteriën veroorzaakte chemische omzettingen in de bodem een rol (o.a. gasvorming). Het nettoeffekt van biota op erodeerbaarheid van bodems kan zowel positief als negatief zijn.

De schuifspanning veroorzaakt door stroming langs de bodem is primair verantwoordelijk voor erosie van sedimentdeeltjes. Turbulente waterbewegingen ("bursts" en "sweeps") spelen hierbij een belangrijke rol. Deze zorgen er ook voor dat sedimentdeeltjes in suspensie komen. Beneden een bepaalde schuifspanning, die afhangt van de cohesie en pakkingsgraad, en van de grootte van de sedimentdeeltjes, wordt er geen sediment van de bodem geërodeerd. Figuur 4.12 geeft een indruk van de minimale ("critische") stroomsnelheid die nodig is om bodemerosie te bewerkstelligen, voor verschillende typen sediment en verschillende sedimentdiameters. De figuur 4.laat echter niet zien dat ook golfwerking een grote invloed heeft op de erodeerbaarheid van de bodem; in aanwezigheid van golven treedt erosie op bij een veel geringere stroomsnelheid dan in afwezigheid van golven. Dit heeft vooral te maken met door golven veroorzaakte fluctuaties in de waterspanning in de bodem.



Erosion, transport, and deposition regimes for mean current velocities versus grain size. Solid lines represent critical velocities required for erosion, transport, and deposition of bed sediment while dashed lines represent extrapolated trends. Lines A-C and critical velocities for various stages of consolidation represented by decreasing water content. Adapted from Hjulstrom (1939), Postma (1967), and McDowell and O'Connor (1977).

Figuur 4.12: "critische" stroomsnelheid waarbij sedimentdeeltjes worden opgewerveld of worden afgezet, als functie van de sedimentdiameter

Wanneer de stroomsnelheid de critische waarde voor bodemerosie overschrijdt, dan wordt bodemmateriaal door de stroming meegenomen. Een deel van het transport vindt plaats langs de bodem (rollend), en een deel van het sediment wordt door turbulentie in suspensie gebracht. De verhouding van suspensietransport en bodemtransport hangt af van de grootte van de stroomsnelheid en de grootte van de sedimentdiameter, zie figuur 4.13. Het bodemtransport staat in nauw verband met kleinschalige structuren in de bodemligging. Afhankelijk van sedimenttype, sedimentdiameter en stroomsnelheid worden verschillende bodemstructuren gevormd, zie tabel 4. 1. Deze bodemstructuren hebben grote invloed op de bodemruwheid die de stroming ondervindt. Uit waarnemingen blijkt dat bij permanente stroming de schuifspanning  $\tau_b$  tussen water en bodem goed kan worden beschreven door de formule

 $\tau_b \approx \rho c_D |\mathbf{u}| \mathbf{u}$ .

De waarde van de schuifspanningscoëfficient  $c_D$  ("bodemruwheid") ligt o.h.a. tussen 0.002 en 0.004, afhankelijk van sedimentdiameter en kleinschalige bodemstructuren. Een fysische beschrijving van de wisselwerking tussen stromingseigenschappen, bodemruwheid, bodemtransport en de vorming van bodemstructuren is zeer complex en grotendeels nog onbekend. In de praktijk wordt gebruik gemaakt van empirische verbanden, die voor veel voorkomende stromingssituaties en sedimentaire omgevingen proefondervindelijk zijn vastgesteld.

Voor bodemtransport geldt bij benadering de empirische formule (geldig voor kleine bodemhellingen)

**S** 
$$\approx \rho_{sed} \beta |\mathbf{u}|^{b-1} \mathbf{u}$$
,  $b = 3 - 5$ ,  $\beta = 10^{-7} - 10^{-4} [m^{2-b} s^{b-1}]$ 

In het college gebruiken we meestal b = 3.

Voor suspensietransport van middel/grof, niet cohesief bodemmateriaal (zand) wordt vaak een soortgelijke formule gebruikt, zie het college zandkusten. Voor suspensietransport van fijn (cohesief) sediment moeten andere formules worden gebruikt, die rekening houden met de tijdsduur van opwerveling en bezinking. Zie hiervoor de volgende sectie



Relationship between flow velocity, grain size, and state of movement for uniform sediment of 2.65 density. The curves representing the relative concentration of suspended sediment indicate the ratio between the concentration at onehalf the water depth and a reference level near the bed. Dotted lines represent diffuse boundaries. Velocity is measured at 100 cm above the bed; after Reineck and Singh (1980) and Sundborg (1967).

Figuur 4.13 : Type sedimenttransport (bodemtransport of suspensietransport) als funtie van stroomsnelheid en sedimentdiameter

## 4.7 Transport van fijn sediment

De netto hoeveelheid sediment die per eenheid van breedte in 1 getijperiode door het vlak  $x_0$ wordt getransporteerd is gegeven door:

$$M(x_0) = \int_0^T h(x_0, t) u(x_0, t) c(x_0, t) dt$$

Hier is c = dieptegemiddelde concentratie van fijn sediment in suspensie u = dieptegemiddelde snelheid.

Deze uitdrukking is alleen geldig als de stroomsnelheid vrijwel uniform over de dwarsdoorsnede wordt verondersteld. Bij dichtheidsgedreven circulatie is dat zeker niet het geval en moeten in de integraal correlatietermen tussen verticale en laterale variaties in stroomsnelheid en sedimentconcentratie worden meegenomen.

In een meebewegend assenstelsel ( $\frac{D}{Dt} = \frac{\partial}{\partial t} + u \frac{\partial}{\partial x}$ ) geldt:

 $\frac{Dc}{Dt} = \frac{l}{h} [Er - Se] .$ erosie - sedimentatie verandering sedimentvracht

Voor fijn, cohesief sediment zijn uit waarnemingen de volgende relaties afgeleid:

sedimentatie 
$$Se = W_s c(1 - \frac{\tau}{\tau_{Se}}) \Theta(\tau_{Se} - \tau)$$
,  
erosie  $Er = \mu_{Er} (\frac{\tau}{\tau_{Er}} - 1) \Theta(\tau - \tau_{Er})$ .

erosie

Hier is  $\theta$  de stap-functie,  $\theta(x)=0$  als x<0,  $\theta(x)=1$  als  $x\geq 0$ .

De coëfficiënten  $W_{se}$  [m/s] en  $\mu_{Er}$  [kg/m<sup>2</sup>s] en de critische schuifspanningen  $\tau_{se}$  en  $\tau_{Er}$ zijn afhankelijk van het type en concentratie sediment, de bodemtoestand (m.n. consolidatiegraad) en andere invloeden (o.a. chemisch-biotisch).

Nu zijn we hier niet zozeer geïnteresseerd in de meest accurate beschrijving van het sedimentatie-erosie proces, maar alleen in de tijdschaal van erosie en sedimentatie. Dit is een karakteristiek waardoor de getijdynamiek van fijn sediment essentieel verschilt van die van grover sediment, zoals zand. Voor zand is deze tijdschaal een orde kleiner dan de getijperiode, terwijl voor fijn sediment de tijdschalen in dezelfde orde liggen. We noemen voor fijn sediment deze tijdschaal T<sub>sed</sub>. Als we veronderstellen dat zich bij constante stroomsnelheid na verloop van tijd een evenwichtsconcentratie  $c_{eq}$  instelt, dan is  $T_{sed}$  de tijdschaal waarop de concentratie in suspensie de evenwichtsconcentratie benadert. In formulevorm geeft dit:

$$\frac{Dc}{Dt} = \frac{l}{T_{sed}} (c_{eq} - c) .$$

De evenwichtsconcentratie ceq kan uit waarnemingen worden bepaald; het blijkt dat vaak geldt  $c_{eq} \approx \beta |u|^{b-l}$ , met  $b \approx 3$  (afhankelijk van het type sediment).

Er moet in principe onderscheid gemaakt worden tussen de relaxatietijdschalen  $T_{sed}$  voor erosie en voor sedimentatie, resp.  $T_{Er}$  en  $T_{Se}$ . De laatste is o.a. afhankelijk van de waterdiepte,  $T_{Se} \approx h / w_s$ .



TRANSPORT (C) KARAKTERISTIEKE VERTIKALE VERDELINGEN VAN HET TRANSPORT VAN GESUSPENDEERDE DEELINES DP VERSCHILLENDE TUDSTIPPEN IN DE GETYPERIODE

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Dat betekent dat de sedimentatie bij HW in omvang sterk kan verschillen van de sedimentatie bij LW. Dit is met name het geval als het natte oppervlak van het bekken veel groter is bij HW dan bij LW, d.w.z. dat  $H_K^+ \prec H_K^-$ . Het effekt van deze grotere sedimentatie bij HW dan bij LW kan worden versterkt door het effekt van een langere kenteringsduur bij HW dan bij LW (of worden tegengewerkt als de kenteringsduur bij HW korter is dan bij LW).

De relaxatie-vergelijking voor de gesuspendeerde concentratie laat zien dat een asymmetrie in de kenteringsduren rond HW en LW een resulterend transport van fijn sediment kan veroorzaken. Veronderstel b.v. dat de kenteringsduur rond HW veel langer is dan rond LW. We hebben eerder gezien dat dit een karakteristiek is van een getijdebekken met ondiepe geulen en weinig kombergende platen. In dat geval is er rond HWK een sterke sedimentatie; in het begin van de eb is er weinig materiaal gesuspendeerd in de waterkolom en de concentratie neemt slechts geleidelijk toe. Pas later in de ebperiode is er veel fijn sediment opgewerveld dat wordt getransporteerd. Rond LW kentering sedimenteert weer een deel van dit sediment, maar niet zoveel vanwege de korte kenteringsduur. Wanneer de vloedstroom aantrekt transporteert deze dus gelijk een aanzienlijke hoeveelheid gesuspendeerd materiaal. De gesuspendeerde concentratie neemt vervolgens door opwerveling nog verder toe. Gemiddeld over de vloed is de concentratie gesuspendeerd sediment dus hoger dan gedurende de eb. Over het hele getij genomen is er sprake van een netto landwaarts gericht transport. Een voorbeeld van het omgekeerde, een kort durende HW-kentering en dominerend zeewaarts gericht slib-transport is te zien in figuur 4.14, voor het Thames-estuarium.

In het geval dat  $T_{Er} = T_{Se} = T_{sed} = constant$  kan de erosie-sedimentatie vergelijking worden geïntegreerd,

$$c(t) = c(t_0) + \frac{e^{-\frac{t}{T_{sed}}}}{T_{sed}} \int_{t_0}^t e^{-\frac{t'}{T_{sed}}} (c_{eq}(t') - c(t_0))dt'$$

Het gaat hier om de gesuspendeerde concentratie in een met het getij meebewegend assenstelsel. Voor de berekening van het getij-geïntegreerde transport kan dit resultaat niet zonder meer worden gebruikt omdat in de transport-integraal de vaste doorsnede  $x_o$  wordt beschouwd.

In plaats van een transport-integraal door een vaste doorsnede beschouwen we daarom transport door een vlak dat meebeweegt met de gemiddelde getijsnelheid u(x,t). De positie van het vlak wordt aangeduid met X(t), en we hebben:

#### dX / dt = u(X,t).

Als de getijbeweging cyclisch wordt verondersteld, en als de rivierafvoer wordt verwaarloosd, dan is  $X(t=0) = X(t=T) = x_0$ . Het bewegende vlak keert dus na een getijbeweging op zijn uitgangspositie terug. De hoeveelheid materiaal die door het bewegende vlak wordt getransporteerd is gemiddeld over de getijperiode daarom identiek aan het netto transport door het vaste vlak:

$$M(x_0,t) = \int_0^T h(x_0,t)u(x_0,t)c(x_0,t)dt = \int_0^T h(X(t),t)u(X(t),t)c(X(t),t)dt$$

De laatste integraal kan worden herschreven door te redeneren dat door het bewegende vlak alleen materiaal wordt getransporteerd dat op de bodem ligt. Dus materiaal dat vóór het bewegende vlak wordt geërodeerd gaat niet door het vlak en moet in mindering worden gebracht; daarentegen, materiaal dat vóór het vlak sedimenteert gaat vervolgens door het bewegende vlak en moet dus bij het transport worden opgeteld. Als wordt verondersteld dat er geen netto sedimentatie of erosie plaatsvindt over het traject dat het vlak in een getij doorloopt, dan laat de transportintegraal zich schrijven als:

$$M(x_0,t) = \int_0^T dt \int_{x_0}^{x(t)} (Se - Er) dx .$$

Uit deze uitdrukking blijkt dat sedimentatie bij HWK een rol speelt in het landwaartse deel van het bekken (tussen  $x_o$  en X(HWK)) en sedimentatie bij LWK in het zeewaartse deel (tussen X(LWK) en  $x_o$ ). Iets dergelijks geldt ook voor erosie. Met name voor fijn sediment geldt dat als in het landwaartse deel van het bekken bij HWK een sterkere sedimentatie en langzamere erosie plaatsvindt dan in het zeewaartse deel bij LWK, dan veroorzaakt dit een resulterend landwaarts sedimenttransport. Voor fijn sediment is dus een asymmetrie in kenteringsduur in de HWK en LWK posities van belang.

Bovenstaande wordt in de volgende formules samengevat, waarbij ook het effekt van het nat oppervlak (dus gemiddelde bekkendiepte) op de omvang van de sedimentatie is meegenomen. Er is netto vloed-transport als:

$$\left|\frac{du}{dt}(LWK, X_{LWK})\right| > \left|\frac{du}{dt}(HWK, X_{HWK})\right|, \quad H_K(LWK, X_{LWK}) > H_K(HWK, X_{HWK})$$

en netto eb-transport als:

$$\left|\frac{du}{dt}(LWK, X_{LWK})\right| < \left|\frac{du}{dt}(HWK, X_{HWK})\right|, \quad H_K(LWK, X_{LWK}) < H_K(HWK, X_{HWK})$$

(HWK = kentering voorafgaand aan de eb, LWK = kentering voorafgaand aan de vloed). Een groot kombergend plaatoppervlak heeft twee tegengestelde invloeden op het netto transport van fijn sediment: aan de ene kant veroorzaakt de grote komberging een korte kenteringsduur in de stroomgeul bij HWK; aan de andere kant kan er in die korte periode toch een sterke bezinking plaatsvinden door de geringe waterdiepte. Dit laatste effekt heeft vaak de overhand.

Voor grover sediment (zand) speelt de kenteringsduur niet zo'n grote rol omdat de sedimentatie-tijdschaal veel korter is. Hier is het verschil tussen concentraties zwevend of rollend sediment bij eb en vloed van belang voor resulterend transport. Er is o.h.a. netto vloed-transport als  $u_{max}(vloed) > u_{max}(eb)$ , en netto eb-transport in het omgekeerde geval.

#### 4.7.1 Samenvatting relatie sedimenttransport-morfologie

Uit de voorgaande paragrafen kunnen kwalitatieve relaties worden afgeleid tussen netto sedimenttransport en morfologische karakteristieken van getijdebekkens. Deze zijn in bijgaande tabel 4. samengevat, waarbij ook de rol van externe condities (getijden, golven) is aangegeven.

NETTO VLOED- TRANSPORT	NETTO EB- TRANSPORT	GETIJASYMMETRIE
weinig kombergend plaatoppervlak	groot plaatoppervlak	u <sub>max</sub> vloed/eb kenteringsduur HW/LW
groot kombergend oppervlak	klein plaatoppervlak	sedimentatie HW/LW (dit effekt domineert vaak het voorgaande)
sterke landwaartse diepte-afname	landwaarts diep	sedimentatie HW/LW
ondiepe geulen	diepe geulen	u <sub>max</sub> vloed/eb kenteringsduur HW/LW
landwaarts afnemende stroming	landwaarts sterke stroming	kenteringsœuur HW/LW
lange HW-periode op zee	korte HW-periode op zee	kenteringsduur HW/LW
beschutte ligging, weinig golven	open, veel golven	sedimentatie HW/LW

#### Condities voor netto transport van zand

NETTO VLOED- TRANSPORT	NETTO EB- TRANSPORT	GETIJ-ASYMMETRIE
weinig kombergend plaatoppervlak	groot plaatoppervlak	u <sub>max</sub> vloed/eb
lange ondiepe geulen	lange diepe geulen	u <sub>max</sub> vloed/eb
snelle getijrijzing op zee	snelle getijdaling op zee	u <sub>max</sub> vloed/eb
landwaarts afnemende stroming	landwaarts sterke stroming	gradiënt sedimenttransport

Tabel 4.2: Condities voor netto transport van fijn sediment (slib)

In figuur 4.15 zijn verschillende bekkentypen geschetst.



O(Z)= bekkenoppervlak [m<sup>2</sup>] als functie van de afstand tot het wateroppervlak z

Figuur 4.15: Bekken Typen

.

## 4.8 Grootschalige morfodynamiek

#### 4.8.1 Morfologisch evenwicht

Er is morfologisch evenwicht als de eb- en vloedtransporten van sediment gelijk zijn. Dit hangt o.a. af van de externe condities, maar als deze neutraal worden verondersteld dan kunnen verschillende typen bekkens aan de evenwichtsvoorwaarde voldoen:

- ondiepe geulen en groot kombergend plaatoppervlak (b.v. Waddenzee)
- diepe geulen en klein kombergend plaatoppervlak (b.v. Oosterschelde)
- tussenliggende geometrieën

De geometrische karakteristieken van het bekken zijn dan dusdanig dat de vloed- en ebperioden ongeveer gelijk zijn. Als de getijasymmetrie op zee niet al te sterk is dan impliceert dit

$$\left(HH_{K}\right)^{+}\approx\left(HH_{K}\right)^{-}$$

Omdat in veel getijdebekkens de stroomvoerende breedte bij HW (+) niet sterk afwijkt van de stroomvoerende breedte bij LW (-) is de voorwaarde voor getijsymmetrie bij benadering

$$\frac{H^+}{H^-} = \alpha \sqrt{\frac{b_K^+}{b_K^-}}$$

Om de vloeddominantie door snellere getijrijzing op zee te compenseren zal in de praktijk de faktor  $\alpha$  iets groter zijn dan 1.

4.8.2 Morfologische stabiliteit

Bij gegeven getijrandvoorwaarde kan in principe het getijgemiddelde sedimenttransport *<S>* uitgedrukt worden als functie van de bekkentopografie *< S >= f (b<sub>S</sub>, h<sub>S</sub>, b<sub>P</sub>, h<sub>P</sub>)*. Hier zijn  $b_S, h_S, b_P, h_P$  respectievelijk de stroomvoerende breedte (geulbreedte), stroomvoerende diepte (geuldiepte), de plaatbreedte ( $b_P = b_K^+ - b_S^+$ ) en de plaatdiepte. We veronderstellen dat het bekken dusdanig uniform is in langsrichting dat de bekkentopografie overal beschreven kan worden door deze karakteristieken in een enkele dwarsdoorsnede *A*. We

gebruiken de kortere notatie  $\langle S \rangle = \langle S(A) \rangle$ . Bij morfologisch evenwicht geldt  $\langle S(A^{eq}) \rangle = 0$ .

Als we dit evenwicht licht verstoren, zodat  $A = A^{eq} + A'$ , met  $A' << A^{eq}$ , dan is

$$< S(A) > \approx < S(A^{eq}) > +A' \frac{\partial < S >}{\partial A}\Big|_{eq} = A' \frac{\partial < S >}{\partial A}\Big|_{eq}$$

Deze laatste term is een korte schrijfwijze voor

$$A'\frac{\partial < S >}{\partial A} = b'_S \frac{\partial < S >}{\partial b_S} + h'_S \frac{\partial < S >}{\partial h_S} + b'_P \frac{\partial < S >}{\partial b_P} + h'_P \frac{\partial < S >}{\partial h_P}$$

De accenten staan hier voor een kleine verstoring t.o.v. de evenwichtswaarden. De sedimentbalansvergelijking luidt

$$\frac{\partial A}{\partial t} = \frac{\partial < S >}{\partial x}$$

Uit integratie van x tot het bekkeneinde / volgt:

$$\frac{\partial A'}{\partial t} \approx - \frac{1}{l-x} < S(A) > \approx - \frac{A'}{l-x} \frac{\partial < S >}{\partial A}\Big|_{eq},$$

waarbij 
$$\frac{\partial A'}{\partial t} = b_S \frac{\partial h'_S}{\partial t} + h_S \frac{\partial b'_S}{\partial t} + b_P \frac{\partial h'_P}{\partial t} + h_P \frac{\partial b'_P}{\partial t}$$

Als de verschillende morfologische karakteristieken dynamisch onafhankelijk van elkaar zijn dan volgt uit bovenstaande vergelijking dat er alleen sprake is van stabiel morfologisch evenwicht onder voorwaarde dat alle afgeleiden positief zijn:

$$\frac{\partial \langle S \rangle}{\partial b_S}\Big|_{eq}, \frac{\partial \langle S \rangle}{\partial h_S}\Big|_{eq}, \frac{\partial \langle S \rangle}{\partial b_P}\Big|_{eq}, \frac{\partial \langle S \rangle}{\partial h_P}\Big|_{eq} > 0.$$

Als één van de afgeleiden negatief is dan zeggen bovenstaande vergelijkingen ons dat de verstoring initieel een exponentiële groei ondervindt.

In een bekken met groot kombergend oppervlak en diepe geulen is het sedimenttransport netto in zeewaartse richting; morfologisch evenwicht vereist ondiepe geulen. Een positieve verstoring (toename) van de evenwichtsdiepte  $h_s$  van de geul zal dus leiden tot netto sedimentexport en veroorzaakt verdere verdieping; dit is een geval van instabiel morfologisch evenwicht.

Dit resultaat volgt ook uit de eerdere redenering dat voor lange, wrijvingsgedomineerde bekkens het getijgemiddelde sedimenttransport <S> toeneemt of afneemt al naar gelang het verschil in tijdsduur van eb en vloed toeneemt of afneemt,

$$\delta < S > \infty \ \delta \left[ H^{+} H_{K}^{+} - H^{-} H_{K}^{-} \right] \propto \ \delta \left[ H^{+2} \frac{b_{S}^{+}}{b_{K}^{+}} - H^{-2} \frac{b_{S}^{-}}{b_{K}^{-}} \right]$$

Als de geuldiepte toeneemt, dan neemt vooral H toe, dus  $\frac{\partial < S >}{\partial h_S} < 0$ .

Een positieve verstoring van de geulbreedte b<sub>s</sub> is equivalent met een relatieve toename

van het kombergend oppervlak op laag water,  $b_{\kappa}$ , dus hier geldt dat  $\frac{\partial < S >}{\partial b_S} > 0$ . In

dit geval ontstaat een netto sedimentimport die de oorspronkelijke verstoring teniet kan doen. Het feit dat getijdebekkens bestaan betekent dat ze een grote mate van stabiliteit kunnen bezitten. Deze stabiliteit vereist dat geulbreedte, geuldiepte, plaatbreedte en plaatdiepte dynamisch gekoppeld zijn. Het zijn vooral de transversale transporten in het getijdebekken die bij deze dynamische koppeling een rol spelen. Deze interne dynamiek kan de morfologische stabiliteit van het bekken herstellen als b.v. de geuldiepte in positieve of negatieve zijn wordt verstoord.

In figuur 4.16 is te zien dat voor de Nederlandse getijdebekkens de verhouding tussen gemiddelde geuldiepte en kombergend oppervlak dusdanig is dat eb- en vloedtransport van sediment ongeveer in evenwicht zijn.

#### 4.8.3 Lokaal morfologisch evenwicht

Bovenstaande stabiliteitsanalyse heeft betrekking op verstoringen op bekken-schaal. Vaak hebben verstoringen van diepte of breedte echter een lokaal karakter. Dergelijke verstoringen kunnen door vloed- of ebtransport worden vereffend zonder dat er sprake is van netto getijgemiddeld zandtransport. Daar waar een stroomverlamming optreedt wordt vaak een drempel gevormd en een lokale vernauwing wordt meestal uitgeschuurd. Hieruit volgt dat nog een andere evenwichtsvoorwaarde in acht genomen moet worden, n.l. dat er geen sterke gradiënten in eb- of vloedtransport mogen optreden. We gaan dit nu nader onderzoeken. Daarvoor stellen we dat bij een bepaalde dwarsdoorsnede het bekken in morfologisch evenwicht is:

$$A(x) = A^{eq}(x)$$
,  $< S(x) >_{vloed} = - < S(x) >_{eb} \equiv S(A^{eq})$ ,

Nu wordt vervolgens een lokale verstoring  $A = A^{eq} + A'$  met lengteschaal *l'* aangebracht. We veronderstellen dat deze lengteschaal veel kleiner is dan de lengteschaal  $I_m$  waarover het sedimenttransport evenwicht bereikt ( $I_m$  is de door de stroming afgelegde weg waarbinnen een surplus aan suspensie wordt afgezet of waarbinnen een tekort aan suspensie van de bodem wordt opgenomen).

We beschouwen stroming in vloedrichting. Een bodemverstoring tussen x en x+l' neemt dan toe of af volgens:

$$\frac{\partial A'}{\partial t} = \frac{1}{l'} \left[ S(x+l') - S(x) \right] \approx \left. \frac{A'}{l_m} \frac{\partial S}{\partial A} \right|_{e_a}$$

Hierbij is gebruik gemaakt van  $S(x + l') \approx S(x) + \frac{l'}{l_m}(S(A) - S(A_{eq}))$ .

Voor zandtransport geldt, in geval van een ongeveer sinusoïdaal getij,  $S \propto \tau_{max}^{\frac{3}{2}}$ , waar  $\tau_{max}$  staat voor de maximale bodemschuifspanning. Voor een kleine verstoring geldt dan de stabiliteitsvoorwaarde

$$\frac{1}{A'} \frac{\partial A'}{\partial t} \propto \left. \frac{1}{l_m} \frac{\partial S}{\partial A} \right|_{eq} \propto \frac{\partial \tau_{\max}}{\partial A} \right|_{eq} < 0 .$$

Op het eerste oog lijkt dat aan deze voorwaarde vanzelfsprekend voldaan wordt: de stroomsnelheid neemt toe als de dwarsdoorsnede wordt verkleind. Toch is dat niet altijd het geval, bijvoorbeeld als er in het getijdebekken meerdere parallelle geulen lopen. In de paragraaf over zeegaten wordt aangetoond dat ook in het geval van een enkele geul instabiliteit kan optreden.

Vereffening van een lokale verstoring van de dwarsdoorsnede brengt automatisch met zich mee dat er zand uit de nabije omgeving onttrokken wordt of dat er zand wordt toegevoegd. In feite verplaatst de verstoring zich hierdoor. Als de eb- en vloedsnelheden van dezelfde orde zijn dan is er echter geen netto verplaatsing. Wel vindt diffusie plaats, omdat zowel bij eb als bij vloed het effekt van de verstoring over een grotere afstand in de stroomrichting verplaatst wordt. Het is in deze zin dat een verstoring in een morfologisch stabiele geul vereffend wordt.

In het bovenstaande zijn we ervan uitgegaan dat er geen secundaire circulatiestromingen door de verstoring worden opgewekt. Deze veronderstelling is niet vanzelf gerechtvaardigd: in getijdebekkens zijn door de topografie opgewekte circulatiestromen talrijk aanwezig. We zijn er echter van uitgegaan dat deze circulatiestromen niet door de aangebrachte verstoring wezenlijk aangetast worden. In werkelijkheid kan dat wel het geval zijn, en ontstaan door positieve terugkoppeling tussen topografie en restcirculaties allerlei bodemstructuren, zoals bodemribbels, zandgolven en geulmeanders.

Wanneer er sprake is van n parallelle geulen  $A_1, A_2, ...$  met schuifspanningen  $\tau_1(A_1, A_2, ...), \tau_2(A_1, A_2, ...), ...$  dan kan stabiliteit worden afgeleid uit de vergelijkingen:

$$\frac{dA'_i}{dt} = \gamma \sum_{j=1}^n A'_j \left. \frac{\partial \tau_i}{\partial A_j} \right|_{eq}, \quad i = 1, \dots n.$$

Stabiliteit vereist dat alle eigenwaarden van dit stelsel negatief zijn. In het geval dat eenzelfde kombergend gebied (vloedkuil) door meerdere geulen is verbonden met de zee dan kan

worden aangetoond dat tenminste één van de eigenwaarden positief is. Het systeem is dus instabiel: één van de geulen gaat zich ten koste van de andere verdiepen, de andere zeegaten zullen zich sluiten. Onder gewijzigde omstandigheden, b.v. bij storm, kan echter opnieuw een systeem met meerdere zeegaten worden gevormd.



Figuur 4.16: Relatie tussen bekken diepte en oppervlakte voor Nederlandse getijdebekkens (voor afkortingen zie fig. 4.3)



Figuur 4..17: Relatie tussen doorsnede zeegat en getijprisma voor Nederlandse getijdebekkens

#### 4.8.4 Stabiliteit van het zeegat

Uit waarnemingen blijkt dat de evenwichts-bodemschuifspanning  $\tau_{max}(eq)$  die door het getij wordt uitgeoefend op de bodem in een zeegat overeenkomt met een maximale stroomsnelheid  $u_{max}$  van ongeveer 1 m/s. Als de getijsnelheid in het zeegat onvoldoende groot is wordt het zeegat door golfgedreven zandtransport afgesloten. Wanneer de getijschuifspanning groter is dan de evenwichtsschuifspanning dan zal de geul zich verdiepen, totdat de getij-schuifspanning zover is afgenomen dat de evenwichtswaarde is bereikt. Hierbij wordt meestal zand zeewaarts afgevoerd en afgezet in de buitendelta. Even buiten het zeegat domineert de ebstroom vaak over de vloedstroom, ook al zijn de eb- en vloeddebieten even groot. De reden is dat de ebstroom meer in de hoofdgeul is geconcentreerd en het getijdebekken als een "jet" verlaat, terwijl de vloedstroom vanuit een wijde omgeving van het zeegat wordt gevoed.

Er kunnen zich twee situaties voordoen waarbij de getij-schuifspanning aanvankelijk kleiner is dan de evenwichtswaarde:

- 1. de geuldoorsnede is groter dan de evenwichtsdoorsnede; in dat geval vindt er sedimentatie plaats in het zeegat totdat de evenwichtsdoorsnede bereikt is.
- 2. de geuldoorsnede is kleiner dan de evenwichtsdoorsnede; ook in dat geval vindt er sedimentatie plaats, net zolang totdat het zeegat volledig afgesloten is.

Dit wordt in het vervolg gedemonstreerd.

Eerst berekenen we de evenwichtsdoorsnede door aan te nemen dat het achterliggende getijdebekken wordt gevuld en geleegd via een stroomgeul waarin het waterstandsverhang gerelateerd is aan de bodemwrijving:

$$bhu(t) = O\eta_t(bekken),$$
  

$$gh[\eta(zee) - \eta(bekken)] \approx \frac{8}{3\pi} c_D l u_{\max} u(t).$$

Stel  $\eta(zee) = a \sin \omega t$  dan levert oplossing van bovenstaande vergelijkingen de volgende relatie tussen maximale getijsnelheid in het zeegat  $u_{max}$  en geuldoorsnede A:

$$\tau \propto u_{\text{max}}^{2} = \frac{1}{2} \left( \frac{3\pi g}{8 c_{D} \omega} \frac{A^{2}}{b l O} \right)^{2} \left[ -1 + \sqrt{1 + \left( \frac{3\pi c_{D} \omega^{2}}{4g} \frac{a b l O^{2}}{A^{3}} \right)^{2}} \right]$$

Hier is O het bekkenoppervlak en I de lengte van de instroomgeul. Als voldaan is aan de voorwaarde

$$\left(\frac{4g\omega}{3\pi c_D}\frac{Oa^2}{bl}\right)^{\frac{1}{3}} > u_{eq} \approx 1 m/s$$

dan zijn er twee evenwichts-doorsneden  $A_1^{eq}$  en  $A_2^{eq}$  , waarbij

$$A_1^{eq} > A_c$$
,  $A_2^{eq} < A_c$  en  $A_c = \frac{l}{\sqrt{2}} \left( \frac{3\pi c_D \omega^2}{4g} abl O^2 \right)^{\frac{l}{3}}$ .

Veronderstel dat de geuldoorsnede A met één van de evenwichtsprofielen overeenkomt, en

vervolgens een kleine verstoring A' ondergaat. Een positieve verstoring zal gaan groeien als door de verstoring de schuifspanning toeneemt, een negatieve verstoring zal gaan groeien in het omgekeerde geval.

Voor  $A = A_2^{eq}$  is  $d\tau / dA < 0$ , het zeegat is stabiel. Voor  $A = A_1^{eq}$  is  $d\tau / dA > 0$ , het zeegat is instabiel. Als A' < 0 wordt het zeegat afgesloten; als A' > 0 verruimt het zeegat tot de stabiele doorsnede  $A_2^{eq}$ . (Wanneer er sprake is van meerdere parallelle geulen dan blijkt uit een soortgelijke stabiliteitsanalyse, waarbij dwarscirculaties worden verwaarloosd, dat er geen stabiel evenwicht mogelijk is).

Als de instroming niet sterk door bodemwrijving in het zeegat wordt beïnvloed dan is de maximale getijsnelheid  $u_{max}$  evenredig met het quotiënt van getijprisma en geuldoorsnede. Uit de empirische relatie  $u_{max} = O(1 \text{ m/s})$  volgt dat voor veel getijdebekkens dit quotiënt een vaste karakteristieke waarde heeft. Figuur 4.17 laat zien dat deze relatie voor de Nederlandse getijdebekkens redelijk opgaat. Omdat sterke longitudinale variatie van vloed-en ebtransport niet is toegestaan (zoals aangegeven in de vorige paragraaf) geldt ook verder landwaarts in het bekken dat  $u_{max} = O(1 \text{ m/s})$ .

## 4.9 Mengprocessen

#### 4.9.1 Structuur van het stroomsnelheidsveld

Door plaat/geul- en eb/vloedschaarsystemen bezitten getijdebekkens meestal een complexe topografie. Die complexiteit is weerspiegeld in de stromingsstructuur; dit geldt vooral voor het horizontale (dieptegemiddelde) stromingsveld. Vloeistofpakketjes doorlopen gedurende de getijperiode vaak een onregelmatig pad, dat wordt beïnvloed door horizontale en transversale circulaties, en door tijdelijk verblijf in buiten de hoofdstroom gelegen gebieden, zoals platen, kreken e.d. Vloeistofpakketjes zullen dus zelden na een getijperiode op dezelfde plaats terugkeren, ook al is het getij cyclisch en de bovenafvoer verwaarloosbaar klein. Ook nabijgelegen vloeistofpakketjes kunnen tijdens de getijperiode sterk verschillende paden afleggen, en sterk van elkaar verwijderd raken. Dit proces is synoniem met menging, die zich b.v. manifesteert door een korte verversingstijd van het bekken met zeewater, of door een snelle verdunning van geloosde stoffen.

Een exacte beschrijving van die menging vereist het nauwgezet volgen van alle verplaatsingen van vloeistofpakketjes in het bekken. In de praktijk is dat niet goed mogelijk, zelfs grote computermodellen hebben er moeite mee. Om de de globale karakteristieken van menging te bepalen, is zo een gedetailleerde analyse niet noodzakelijk. In plaats daarvan zullen we proberen statistische uitspraken te doen over de gemiddelde netto verplaatsing van vloeistofpakketjes in een getijperiode. We interesseren ons daarbij vooral in de netto verplaatsing in longitudinale richting, d.w.z. langs de geulas. We zullen veronderstellen dat de tijdschaal waarover vloeistofpakketjes het getijdebekken in longitudinale zin doorlopen veel langer is dan de tijdschaal waarop verplaatsing over de breedte of diepte plaatsvindt. Dit is o.a. het geval als het verschil in zoutgehalte tussen zeegat en landwaarts bekkeneinde veel groter is dan het verschil in zoutgehalte tussen de oevers of tussen bodem en oppervlak, wanneer landwaarts zoetwater op het bekken wordt toegevoerd. We beschouwen een getijdebekken met een cyclisch getij, waar aan het landwaartse einde zoetwater wordt toegevoerd met een constant debiet Q<sub>R</sub> dat veel kleiner is dan de getijdebieten.

Als gevolg van de netto uitstroom is er gemiddeld gedurende een getijperiode een zeewaartse zoutflux door elke willekeurige dwarsdoorsnede  $x_0$  gelijk aan

$$\Phi_R(x_0) = Q_R \overline{\overline{S}}(x_0), \quad \overline{\overline{S}} = \frac{1}{\langle A \rangle} \frac{1}{T} \int_{t}^{t+T} dt' \iint_{A(x_0,t')} S(x_0, y, z, t') dy dz.$$

machanismen	beschrijving van netto vernlaatsing	wanneer van
Mechanismen	von waternakketies	beland
voor menging en	vali waterparkeyes	beiding
	ver ereine deer rivieref oor die	actiirivioren
rivieratvoer	verversing door rivierarvoer, die	geujiwieren
	zeewaterindringing tegengaat; diffusie	
	speelt een ondergeschikte rol	
turbulentie	menging door wervels op	in alle
	verschillende tijd- en ruimteschalen,	getijsystemen op
	met een 3D-structuur of met verticale	kleine ruimte- en
	draai-as. De diffusiecoëfficiënt ligt in	tijdschalen; in het
	de orde van $D \sim h_{\sqrt{\tau/\rho}} \sim O[1 m^2/s]$	bijzonder in
	<b>v</b> , <b>i</b>	combinatie met
		schering in het
		snelheidsveld
restcirculaties op	waterpakketjes worden met de	Grootschalige
bekken-schaal	restcirculatie door het bekken	wind-gedreven
	getransporteerd. Een diffusie-	circulatie in
	beschrijving is vaak niet van	kustzeeën;
	toepassing	zoutwig-estuaria
		met sterke
		entrainment
residuele	het diffunderende effekt van turbulente	getijdebekkens met
schoringsdiffusie	wervels wordt versterkt door het	complexe
rosteiroulatiocellen	verspreiden van waternakketies over	topografie
in combinatio met	verschillende restcirculatiecellen	opograno
turbulanta diffusio	Dispersiecoëfficiënt in de orde van D	
	$\sim O[100 m^2/s]$	
	~ O[10011175] hot diffunderende offekt van turbulente	getiidebekkens met
geuj-	diffusio wordt vorstorkt door	complexe
scheningsolliusie.		topografie en
schering in het	verspreiding van waterpakkeijes over	storke getijstroming
getij-sneineldsveid	gradienten in het getijsheineldsveid,	Sterke genjanorning
in combinatie met	waardoor dillerentiele advectie	
turbulente diffusie	geaurenae net getij optreedt.	
	Dispersiecoefficient in de orde van $D$	
	~ O[100 m²/s]	actiidabalkkana mat
aeterministische	waterpakketjes volgen verschillende	
menging		complexe
	in getijfase bestaan, b.v. stroombanen	geometrie, i.n.b.
	door geulen, over platen, van- en nagr	
	Kompergingsgebieden, etc.	geulenstelsel
	Dispersiecoefficient in de orde van $D$ ~ $O[100 m^2/s]$	
chaotische	waterpakketjes volgen elke	getijdebekkens met
transportpaden	getijperiode verschillende	een grote getijweg
	stroombanen en ondergaan daardoor	t.o.v. de
	guasi-stochastische netto	topografische
	verplaatsingen. Dispersiecoëfficiënt D	structuren
	~ O[100 m <sup>2</sup> /s]	

Tabel 4.3: Beschrijving mengprocessen

De totale flux  $\Phi(x_0)$  moet echter gelijk zijn aan nul. Het compenserende landwaartse transport wordt veroorzaakt door mengprocessen, we spreken van dispersief transport  $\Phi_D$ . De totale flux is gegeven door

$$\Phi = \Phi_R + \Phi_D = \frac{1}{T} \int_{t}^{t+T} dt' \iint_{A(x_0,t')} u(x_0, y, z, t') S(x_0, y, z, t') dy dz.$$

## 4.10 Stochastische beschrijving van dispersie

We herschrijven de integraal op basis van een statistische schatting van de netto verplaatsing van een vloeistofpakketje in n getijperioden. De waarschijnlijkheid dat een vloeistofpakketje dat zich op *t*=0 bevindt in (*x*,*y*,*z*) zich na n getijperioden landwaarts bevindt van het vlak  $x_0$  noemen we  $p_n(x,y,z,t:x_0-x)$ . De totale flux kan dan geschreven worden als

$$\Phi = \frac{1}{nT} \left[ \int_{-\infty}^{x_0} dx \iint_A p_n S \, dA - \int_{x_0}^{\infty} dx \iint_A (1 - p_n) S \, dA \right]$$
  
=  $\frac{1}{nT} \left[ \int_{-\infty}^{x_0} A \, \overline{p_n} \, \overline{\overline{S}} \, dA - \int_{x_0}^{\infty} A \, (1 - \overline{p_n}) \, \overline{\overline{S}} \, dA \right] + \frac{1}{nT} \int_{-\infty}^{\infty} dx \iint_A p_n S' \, dA$ 

Hier is  $S(x, y, z, t) = \overline{\overline{S}}(x, t) + S'(x, y, z, t)$ .

Noem  $\tau_A$  de tijdschaal voor menging over de dwarsdoorsnede en  $\tau_I$  de tijdschaal voor menging over het hele bekken (de verversingstijdschaal) en veronderstel  $\tau_A < nT < \tau_I$ , dan kan de laatste integraal verwaarloosd worden. In het vervolg laten we de dubbele overstreping weg, *S* staat nu voor het dwarsdoorsnede-gemiddelde zoutgehalte. We veronderstellen dat het zoutgehalte geleidelijk varieert over de lengte van zoutindringing. Uit waarnemingen blijkt ook dat in veel getijdebekkens het zoutgehalte over een grote afstand bijna lineair varieert langs de geulas. Daarom substitueren we de Taylor-reeks

$$S(x,t) = S(x_0,t) + (x - x_0) S_x + \dots$$

en vinden we

$$\Phi_{R} = Q_{R}S(x_{0},t), \qquad Q_{R} = \frac{1}{nT} \left[ \int_{-\infty}^{x_{0}} A \,\overline{\overline{p_{n}}} \, dx - \int_{x_{0}}^{\infty} A \,(1 - \overline{\overline{p_{n}}}) \, dx \right]$$
  
$$\Phi_{D} = -DAS_{x} + \dots, \qquad D = \frac{1}{2nT} \int_{-\infty}^{\infty} x^{2} \frac{\partial \overline{\overline{p_{n}}(x_{0} - x, t; x)}}{\partial x} \, dx$$

Het dispersief transport  $\Phi_D$  is bij benadering evenredig met de longitudinale gradiënt van het zoutgehalte, gemiddeld over de dwarsdoorsnede. De evenredigheidsconstante, de dispersiecoëfficiënt D, staat voor de gemiddelde kwadratische verplaatsing van vloeistofpakketjes in een getijperiode,

$$D = \frac{1}{2nT} E[(\widetilde{X}_n - x_0)^2].$$

Hier is  $E[(\tilde{X}_n - x_0)^2]$  de gemiddelde waarde voor de netto kwadratische

longitudinale verplaatsing van vloeistofpakketjes die volgt uit een groot aantal steekproeven waarbij de pakketjes vanuit een willekeurige initiële positie in de dwarsdoorsnede vertrekken, en

$$\widetilde{X}_n = x_0 + \int_0^{nT} u(\widetilde{X}, \widetilde{Y}, \widetilde{Z}, t) dt$$

is een stochastische variabele die staat voor de longitudinale positie van een vloeistofpakketje na *n* getijperioden,  $(\widetilde{X}, \widetilde{Y}, \widetilde{Z}, t)$  is de positie van het vloeistofpakketje op tijdstip *t*.

In feite beschrijven deze formules een random walk over n getijperioden.

## 4.10.1 Dispersie door horizontale circulaties en door schering in de getijsnelheid

We gaan bovenstaande uitdrukkingen vereenvoudigen en doen daarvoor de volgende aannamen, die voor de meeste getijdebekkens goed opgaan:

- de tijdschaal voor menging over de diepte T<sub>z</sub> is veel korter dan de tijdschaal voor menging over de breedte T<sub>y</sub>; menging over de diepte draagt niet direct bij aan longitudinale dispersie. Er is wel indirect een bijdrage van dieptemenging aan breedtemenging via transversale circulatie. We beschouwen daarom alleen dieptegemiddelde snelheden en veronderstellen dat breedtemenging een diffusief proces is waar o.a. transversale circulaties een belangrijke bijdrage aan leveren.
- longitudinale gradiënten in het snelheidsveld zijn klein in vergelijking met transversale gradiënten. Daarom wordt het snelheidsveld uniform in x-richting verondersteld. Later komen we op deze aanname terug. Verder veronderstellen we dat de dwarsdoorsnede gemiddelde getijsnelheid geïntegreerd over een getijperiode niet tot een netto verplaatsing leidt.

Noemen we  $\widetilde{u} = u - \overline{u}$ , dan is

$$E[(\widetilde{X}_n - x_0)^2] = \int_0^{nT} \int_0^{nT} E[\widetilde{u}(\widetilde{Y}, t)dt \, \widetilde{u}(\widetilde{Y}', t')dt'] .$$

Als we dispersie beschouwen als een stochastisch proces dan is de correlatie tussen de verplaatsingen  $\widetilde{dX} = \widetilde{u}(\widetilde{Y},t)dt$  en  $\widetilde{dX'} = \widetilde{u}(\widetilde{Y'},t')dt'$  gelijk aan de waarschijnlijkheid  $\rho(\widetilde{Y},t;\widetilde{Y'},t')dy'$  dat een vloeistofpakketje, dat zich op tijdstip t bevindt in  $\widetilde{Y}$ , zich op tijdstip t' bevindt in  $[\widetilde{Y'},\widetilde{Y'}+dy']$ . Hieruit volgt:

$$E[(\widetilde{X}_n - x_0)^2] = \int_0^{nT} dt \int_0^{nT} dt' \frac{1}{b} \int_0^b dy \int_0^b dy' \ \rho(y,t; y',t') \ \widetilde{u}(y,t) \ \widetilde{u}(y',t')$$

Als de transversale menging beschreven kan worden door een diffusieproces met transversale diffusiecoëfficiënt  $K_y = \frac{b^2}{2T_y}$ , d.w.z.

$$\frac{\partial \rho}{\partial t} = K_y \frac{\partial^2}{\partial y^2} , \quad K_y \left. \frac{\partial \rho}{\partial y} \right|_{y=0,b} = 0 , \quad \rho(y,t;y',t) = \delta(y-y') ,$$

dan volgt:

$$\rho(y,t;y',t') = \frac{1}{b} + \frac{2}{b} \sum_{k=1}^{\infty} \cos \frac{k\pi y}{b} \cos \frac{k\pi y'}{b} e^{-\frac{k^2\pi^2|t-t'|}{2}T_y}.$$

Veronderstel dat er in het snelheidsveld  $\tilde{u}(y,t)$  zowel een horizontale circulatie  $u_c(y)$  als een getijcomponent met laterale schering  $u_{\tau}(y,t)$  aanwezig is. Als de getijsnelheid een sinusoïdaal verloop heeft dan kan het snelheidsveld worden weergegeven door :

$$\widetilde{u}(y,t) = u_C(y) + u_T(y) \sin \frac{2\pi t}{T}$$

Fourier ontwikkeling hiervan geeft:

$$\widetilde{u}(y,t) = \sum_{k=1}^{\infty} u_{C,k} \cos \frac{k\pi y}{b} + \sin \frac{2\pi t}{T} \sum_{k=1}^{\infty} u_{T,k} \cos \frac{k\pi y}{b}$$

Substitutie en integratie leidt tot een uitdrukking voor de dispersiecoëfficiënt *D* waarin de bijdragen van de horizontale circulaties en de getijschering bij elkaar worden opgeteld:

$$D = D_C + D_T$$
, waar:

$$D_{C} = \frac{T_{y}}{\pi^{2}} \sum_{k=l}^{\infty} \frac{1}{k^{2}} u^{2}_{C,k}$$
$$D_{T} = \frac{T}{8\pi} \sum_{k=l}^{\infty} \frac{\alpha_{k}}{l + \alpha_{k}^{2}} u^{2}_{T,k} , \qquad \alpha_{k} = \frac{\pi k^{2}}{4} \frac{T}{T_{y}} .$$

Uit deze uitdrukkingen blijkt het volgende:

- de dispersie door horizontale circulaties is groter naarmate de tijdschaal voor menging over de dwarsdoorsnede van het bekken langer is. Circulaties die de hele breedte van het bekken beslaan (Fourier-component k=1) dragen meer bij dan kleinschalige circulaties.
- aan de dispersie door schering in de getijsnelheid wordt vooral bijgedragen door Fourier componenten  $k \approx \sqrt{T_y/T}$ . Omdat de tijdschaal voor menging over de dwarsdoorsnede vaak meerdere getijperioden bedraagt zijn het vooral Fourier componenten k=2 of k=3 die van belang zijn. Deze komen vaak overeen met het achterblijven van watermassa's op platen of andere topografische structuren langs de geuloever. Hierbij is van belang dat langs de geuloever transversale circulatie zorgt voor een effectieve menging van achterblijvende watermassa's met de hoofdstroom. In feite is menging over de hele breedte van het bekken minder van belang dan menging over de laterale grenslaag tussen de hoofdstroom en langs de oever achterblijvende watermassa's.

De variatie in longitudinale richting van het snelheidsveld is in de analyse verwaarloosd. Dit is o.h.a. niet gerechtvaardigd: gedurende het dispersieproces doorlopen vloeistofpakketjes grotere afstanden dan de lengteschaal waarop het snelheidsveld uniform beschouwd mag worden. Voor de dispersie door horizontale circulaties betekent dit in feite dat de effectieve circulatie-sterkte gemiddeld over een grotere lengte dan de lengteschaal van de circulatie in rekening gebracht moet worden. Op een soortgelijke wijze moet ook het effect van geuloever-schering over een grotere lengte dan die van de afzonderlijke structuren worden gemiddeld.

In figuur 4.18 zijn experimenteel bepaalde waarden voor de dispersie-coëfficiënt in de getijdebekkens van de Eems-Dollard en de Oosterschelde (voor gedeeltelijke afsluiting) weergegeven. De grootte-orde van deze coëfficiënten komt overeen met waarden die verkregen worden uit de voorgaande theoretische uitdrukkingen.



Figuur 4.18: Waarden van de longitudinale dispersiecefficient [m<sup>2</sup>/s] in de Oosterschelde en de Eems-Dollard

#### 4.10.2 Chaotische transportpaden

De stochastiek van het dispersieproces berust ten dele op turbulente fluctuaties in het snelheidsveld. Deze turbulente fluctuaties zijn kleinschalig van aard en veroorzaken op zich geen sterke dispersie. De dispersie wordt gegenereerd door wisselwerking tussen de kleinschalige turbulentie en grootschalige schering in het snelheidsveld gerelateerd aan horizontale circulaties en laterale gradiënten in de bodemschuifspanning. Maar ook zonder turbulente fluctuaties kan er sprake zijn van grote dispersie. De menging over de dwarsdoorsnede wordt meer beïnvloed door horizontale en transversale circulaties dan door kleinschalige turbulentie. Door de topografische complexiteit van getijdebekkens en de interactie van hoofdstroming en secundaire circulaties kunnen de resulterende verplaatsingen gekarakteriseerd worden als deterministische menging of "chaos". De bovenbeschreven stochastische benadering van het dispersieproces blijft ook geldig als er sprake van chaos. Hierbij moet de tijdschaal van het dispersieproces wel veel groter gekozen worden dan de tijdschalen die gemoeid zijn met deterministische menging. Een voorbeeld van deterministische menging is de relatieve verplaatsing van watermassa's die wordt veroorzaakt door fase-verschillen tussen verschillende waterbewegingen in het bekken. Zo bestaan er vaak fase-verschillen tussen de waterbewegingen in verschillende geultakken, en bestaat er een fase-verschil tussen de stroming in de hoofdgeul en de stroming van en naar aanliggende platen en kombergingsgebieden. In één van de oefeningen wordt hierop ingegaan.

#### 4.10.3 Tijdschalen voor menging en verversing

Door de aanwezigheid van menging en transportprocessen en doordat getijsystemen in open verbinding staan met de oceaan is de verblijftijd van waterpakketjes in een getijsysteem van beperkte duur. Getijsystemen worden geregeld ververst. Door het stochastische karakter van mengprocessen is de verblijftijd van waterpakketjes verschillend, ook al bevinden ze zich op een bepaald tijdstip op ongeveer dezelfde plaats. De verblijftijd van individuele waterpakketjes kan dus worden opgevat als een stochastische variabele, waaraan een bepaalde kansverdeling is geassocieerd. Deze kansverdeling is afhankelijk van de plaats en het tijdstip waarop het waterpakketje het getijsysteem binnenkomt. We kunnen daarbij onderscheid maken tussen twee tijdschalen, de leeftijd en de verblijftijd.

De leeftijd van een waterpakketje is de tijd die het in het systeem heeft doorgebracht voordat het op een bepaalde positie is aangekomen. De leeftijdsverdeling  $p_a$  hangt dus af van de actuele positie van het waterpakketje r en van de plaats van binnenkomen in het systeem  $r_o$  en de tijd t die sindsdien is verstreken,  $p_a = p_a(r_o; r, t)$ . Omdat de waterbeweging in een getijsysteem niet stationair is speelt ook het tijdstip van binnenkomen een rol; als we veronderstellen dat de rivierafvoer constant is en de getijbeweging cyclisch, dan gaat het alleen om de fase  $\phi$  van het getij. Ook als we veronderstellen dat de gemiddelde leeftijd veel groter is dan de getijperiode dan mag deze getijfase vaak toch niet verwaarloosd worden. De leeftijdsverdeling kan worden afgeleid uit een experiment waarbij op tijdstip t=0een hoeveelheid stof met een massa 1 in het systeem bij  $r_o$  wordt geloosd. De leeftijdsverdeling is dan identiek aan de concentratieverdeling,

 $p_a(\mathbf{r}_0; \mathbf{r}, t) = c(\mathbf{r}_0, \phi; \mathbf{r}, t)$ . Hieruit kan de gemiddelde leeftijd bepaald worden; deze is gegeven door

$$\tau_a(\mathbf{r}_0,\mathbf{r}) = \int_0^\infty t \, c(\mathbf{r}_0,\phi;\mathbf{r},t) \, dt \quad .$$

De verblijftijd is de tijd die verstrijkt voordat een waterpakketje dat zich op t=0 bevindt in  $r_o$  het systeem verlaat. De verblijftijdsverdeling hangt dus af van de positie  $r_o$  en van de getijfase  $\phi$  bij t=0. De verblijftijdsverdeling  $p_r$  kan worden afgeleid uit hetzelfde experiment als tevoren, een lozing van massa 1 op t=0 bij  $r_o$ . De verdeling is dan gegeven door de hoeveelheid stof die op tijdstip t het systeem per tijdseenheid verlaat.

$$p_r(\mathbf{r}_0, t) = -\frac{\partial}{\partial t} \int_V d^3 \mathbf{r} c(\mathbf{r}_0, \phi; \mathbf{r}, t) ,$$

waar *V* het volume is van het getijsysteem. De gemiddelde verblijftijd is gegeven door

$$\tau_r(\mathbf{r}_0) = \int_0^\infty t p_r(\mathbf{r}_0, t) dt$$

Deze uitdrukking kan worden vereenvoudigd door niet uit te gaan van een instantane lozing in  $r_o$  maar van een continue lozing met een klein debiet  $Q_o$  en concentratie  $c_o$ . We noemen de resulterende getijgemiddelde concentratieverdeling  $c_{Q_{aco}}(\mathbf{r}_0;\mathbf{r})$ . Dan is

$$\tau_r(\mathbf{r}_0) = \lim_{Q_0 \to 0} \frac{1}{Q_0 c_0} \int_V d^3 \mathbf{r} \ c_{Q_0 c_0}(\mathbf{r}_0; \mathbf{r}) \ .$$

Uit deze uitdrukking blijkt dat de gemiddelde verblijftijd in het bekken van een waterpakketje,

dat bij het bekkeneinde is ingestroomd, experimenteel kan worden bepaald uit de stationaire getijgemiddelde zoutverdeling bij kleine rivierafvoer, want

$$\tau_r(l) \approx \frac{V_{zoet}}{Q_R}$$
, waar  $V_{zoet}$  staat voor de volumefractie zoet water in het bekken

Voor lange getijdebekkens, waarbij de tijdschaal voor menging over de lengte veel groter is dan voor menging over de dwarsdoorsnede, kunnen de 1D-benaderingen van bovengenoemde vergelijkingen worden gebruikt:

$$\tau_{a}(x_{0}, x) = \int_{0}^{\infty} t C(x_{0}, \phi; x, t) dt \text{ en}$$
  
$$\tau_{r}(x_{0}) = \lim_{Q_{0} \to 0} \frac{1}{Q_{0}c_{0}} \int_{0}^{t} dx A(x) C_{Q_{0}c_{0}}(x_{0}; x)$$

Hier is C de over de dwarsdoorsnede gemiddelde concentratie, en A de getijgemiddelde dwarsdoorsnede.

Door oplossing van de 1D dispersie-vergelijking,

$$AC_t + (Q_RC)_x = (DAC_x)_x$$
,  $C = 0$  bij de zeerand  $x=0$ ,

kunnen de gemiddelde leeftijd en verblijftijd gerelateerd worden aan de dispersiecoëfficiënt D. Voor een uniform getijdebekken (A en D onafhankelijk van x) zijn deze relaties:

$$\tau_a(l,x_0) \approx \frac{l^2}{2D} - \frac{x_0^2}{6D} , \quad \tau_r(x_0) \approx \frac{x_0^2}{2D} ,$$

voor respectievelijk de gemiddelde leeftijd van een waterpakketje in  $x_o$  dat bij het bekkeneinde is ingestroomd, en de gemiddelde tijd dat een waterpakketje in  $x_o$  nog in het

bekken verblijft. Hier is het rivierdebiet klein verondersteld:  $Q_R \gg \frac{DA}{I}$  .

De verversingstijd  $T_x$  van het bekken is gedefinieerd als de gemiddelde tijd die een waterpakketje nodig heeft om het gehele bekken te doorlopen. Uit de definitie van de dispersiecoefficient volgt

 $T_x = \frac{l^2}{2D}$  (voor een uniform bekken).

Dit is gelijk aan de gemiddelde verblijftijd in het bekken van een waterpakketje dat bij het bekkeneinde instroomt,  $\tau_r(l)$ , en gelijk aan de gemiddelde leeftijd van dit waterpakketje bij de zeerand,  $\tau_a(l,0)$ .

Door gebruik te maken van de dispersievergelijkingen kunnen ook uitdrukkingen worden verkregen voor de verschillende tijdschalen als het bekken niet uniform is. Bij kleine rivierafvoer geldt

$$\tau_r(x_0) \approx \int_0^{x_0} dx \frac{l}{D(x)A(x)} \int_x^l A(x') dx'$$

In figuur 4.19 zijn de verblijftijden in de Oosterschelde weergegeven die berekend zijn volgens deze formule, met experimenteel bepaalde dispersiecoëfficiënten.

In het theoretische geval van een zeer grote afvoer (  $Q_R >> DAA$ ) wordt al het zeewater uit het getijdebekken verdreven en geldt dus



Figuur 4.19: Lijnen van gelijke verblijftijd (in Getijden) in de Oosterschelde (voor gedeeltelijke afsluiting)

## 11. Literatuur

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# 4.11 NOTATIE- EN SYMBOLENLIJST

а	getijamplitude [m]
A	oppervlak geuldoorsnede [m²], A <sub>s</sub> stroomvoerend, A <sub>k</sub> kombergend
b	tijdgemiddelde breedte [m], bs stroomvoerend, bk kombergend
С	golfvoortplantingssnelheid [m/s], concentratie zwevend sediment [kg/m <sup>3</sup> ]
- Съ	wrijvingscoëfficient (kwadratische wrijving, D staat voor drag)
C C	sedimentconcentratie gemiddeld over de diepte [kg/m <sup>3</sup> ]
d	diameter sedimentdeelties. [m] grenslaagdikte [m] index voor depositie
	diameter scamentacolijes (m), grensladgante (m), maex voor appente
D E	onorgio gotiigolf [ 1/m <sup>2</sup> ]
	Carialia versaalling (s/1)
1	
Fr	Froude-getal u/c
g	zwaartekrachtversneiling [m/s <sup>+</sup> ]
n	tijagemiadelde waterdiepte [m]
Н	
i	$\sqrt{-1}$ , index
i	index
, K	aolfaetal [m <sup>-1</sup> ] in x-richting, index
ĸ	complex golfgetal [m <sup>-1</sup> ], diffusiecoëfficient [m <sup>2</sup> /s]
1	ienate [m], index
i i	aolfienate [m], lenate [m]
M	sedimentmassa per eenheid bodemoppervlakte [kg/m²], massa [kg]
N	turbulente viscositeit (eddy-viscosity) [m²/s]
0	Oppen/lakte [m <sup>2</sup> ]
	van de grootte-orde van
0[]	debiet $Im^3/s1$ , $\Omega_{-}$ = rivierdebiet, $\Omega_{-}$ = actiidebiet
r	coëfficient gelineeriseerde wrijving (ms <sup>-1</sup> )
, R	kromte-straal van geulbocht [m] rivierindex
Re	Revnolds-getal ud/v
Di	Richardson getal
C	sedimenttransport per eenheid stroombreedte [kg/ms] saliniteit [promille]
+	
	uju (5) getij, golfneriode (s) getijindev
1	schuifenanningssnelheid [m/s]
<i>u</i> •	schulispanningssheineld [11/5]
<i>u</i>	snelheid in v richting [m/o] u =stroomanolheid door riviorativer u =door getii
u 	shelled in x-nonling [m/s], $u_R$ -shoon shelled door twe avoid, $u_T$ -door get
V	sheliheld in y-nonling (m/s)
W	sheiheid in z-richting (m/s)
VV	
X	x-coordinaat [m]
У	y-coordinaat [m]
Z	z-coordinaat (verticaal omnoog, x,y,z rechisoraalehu assenstelsel) [m]
	impula diffusionsäffisiont (oddu diffusivitu) [m²/s]
ε	impuls-amasiecoemicient (eady amasivity) (m/s)
ς	vorticiteit (v <sub>x</sub> -u <sub>y</sub> ) [s <sup>-</sup> ]
η	tijdsafhankelijke waterstandsvariatie (H-h) [m]
κ	golfgetal [m <sup>-1</sup> ]
λ	golflengte [m]
μ	imaginair deel golfgetal, dempingscoëfficient [m-1]
ν	kinematische viscositeit [m²/s], frequentie [s-1]
ρ	dichtheid [kg/m <sup>3</sup> ]
Σ	bekken-oppervlakte (m²)
τ	schuifspanning [N/m <sup>2</sup> ]
о Ф	hoekfrequentie golven, getijden [rad/s]
	······································

Gebruik van index voor partiële afgeleide:

 $u_t = \partial u/\partial t$  $u_x = \partial u/\partial x$ etc.

Middelingstekens:

$$< f > = tijdgemiddelde = \frac{1}{T} \int_{I}^{I+T} f \, dt$$

$$\overline{f} = dieptegemiddelde = \frac{1}{H} \int_{-h}^{n} f \, dz \,, \qquad f' = f - \overline{f}$$

$$\overline{\overline{f}} = dwarsdoorsnede - gemiddelde = \frac{1}{A} \iint_{A} f \, dy \, dz$$

$$f_{0} = < \overline{f} > , \quad f_{1} = \overline{f} - f_{0} \,, \quad f = f_{0} + f_{1} + f'$$

Tussen accolades:

 ${f} = \text{grootte-orde schatting van } f,$  $f = f / {f} = \text{geschaalde grootheid, van grootte-orde } 1.$ 

## 5. PHYSICS OF TIDAL INLETS

## 5.1 Introduction

The foregoing parts of this course have focused on the tidal basin, i.e. the estuary or the back-barrier lagoon, where the tide is the principal hydrodynamical agent. In this chapter, we will focus on the inlet proper, the outer delta and the adjacent coast. In preparation of the chapters on how to model these areas, we will concentrate on the physical processes which play a role there.

## 5.2 Morphology

The morphology of tidal inlets has been discussed extensively in the previous chapters. E have also seen that there is no generally valid 'prototype-inlet': the morphology depends to a large extent on local conditions, such as the tidal amplitude, the longshore drift, the sediment properties, etc. Yet, in order to have a reference frame for the description of the physical processes, we will briefly focus on a specific inlet, viz. the Wichter Ee, one of the the meso-tidal inlets of the East-Frisain Wadden Sea (see Fig 5.1).



Figure 5.5.1. Aerial photograph of the morphology of the tidal inlet Wichter Ee, between the islands Norderney (left) and Baltrum (right) [from: Ehlers, 1988]

The littoral drift along this part of the North-Sea coast is from west to east, as can be seen from the sand abundance at the eastern tip of Northerney and the lack of sand at the western tip of Baltrum. Apparently, the sand is bypassed via a number of shoals which migrate on the outer delta and get welded with the coast of Baltrum somewhere east of the tip, where sand abundance is found, again. This pattern is typical of most of the Wadden Sea inlets.

Depending on the spatial scale of interest, we can distinguish a variety of morphological elements in such an inlet.

At the micro-scale, there are all sorts of small-scale bedforms, which are hardly visible in Fig 5.1. Fig 5.2 shows an example of such a bedform pattern. This particular case concerns wind-induced ripples, but current-induced ripples exhibit the same behaviour.


Fig. 5. 2. Evolution of a wind-induced ripple pattern. The pictures have been taken at 5 minute intervals [from: Ehlers, 1988]

At the next higher scale level, Fig 5.1 shows a variety of morphological features. Quite obious is the system of shoals on the outer delta. They tend to migrate from the updrift to the downdrift island and probably form the principal sediment bypass mechanism across the inlet.

The shoals are separated by distinct channels, which all branch off from the main channel. In this particular case, the latter is strongly deflected eastward, but other inlets show that the main channel can just as well have a different direction. Also, the main channel tends to migrate and change direction through time.

The main channel and its branches are probably all ebb-dominant. If this is true, the principle of conservation of water mass requires that other parts of the area are flood-dominant. Often there are more or less distinct flood-channels, e.g. around the tip of the updrift island. In the case of the Wichter Ee, at least at the time that the photos of Fig 5.1 were taken, these flood channels were not so well-developed.





Another striking meso-scale phenomenon, which does not appear from Fig 5.1, is the formation of the so-called *sawtooth-bars*, systems of long-crested and slowly migrating waves in the seabed, with a wave length of typically 100 m, and a crest orientation under a large angle with the coastline (Fig 5.3).

At a still higher scale level, the *macro-scale*, tidal inlets have a basin, sometimes with a distinct inner delta, a gorge, a more or less pronounced outer delta, two adjacent coasts and the adjacent shoreface and seabed. The basin consists of a system of channels, shoals and

marshes, as has been discussed before. The gorge is usually dominated by the main channel, but some are split by a shoal or a small island (cf. the Frisian Inlet, between the isles of Ameland and Schiermonnikoog in the West-Frisian Wadden Sea). The outer delta involves of a bypassing shoal system, an ebb-dominated main channel and a number of flood channels. Together they form a sediment circulation system, as schematically indicated in Fig 5.4.



Figure 5.5.4. Schematic diagram of the outer delta residual sediment transport

At the highest scale level which within our sight, the *mega-scale*, considers a barrier-island system as a whole. An individual inlet is one element in this system, a barrier island is another. At this scale, one usually considers phenomena such as barrier island retreat, inlet formation and closure, etc. In this part of the course, we will not focus too much on this scale-level, simply because we don't know how the small-scale processes we will discuss here influence the coastal behaviour this scale.

# 5.3 Waves

We can safely state that the morphology of the outer delta is highly complicated and variable. This means that waves and currents encounter a very complex bed topography, with length scales which are not much larger than the wavelength of wind waves or swell, for instance. Via refraction, diffraction and reflection, this can lead to complex wave patterns. Fig 5.5 gives an example for the relatively simple case of refraction on a sawtooth bar system.



Figure 5.5.5. Refracting and breaking swell waves on a sawtooth bar system [from Ehlers, 1988]

An obvious consequence of this is a strong spatial variability of the wave field. Hence it hardly makes sense to take point measurements of the wave field in a tidal inlet: wave measurements have to be synoptic, it is the overall pattern that matters, much more than the local wave parameters. The same goes for models: they may be able to give correct predictions of the overall wave pattern, but by no means the local wave properties at every point in the area.

As to the large-scale pattern of waves, there are a few points of consideration, such as the penetration of wave energy into the gorge and the sheltering from wave exposure in various parts of the system.



Figure 5.5.6. Exposure to normally incident waves

In the case of normally incident waves, the entire delta edge and the gorge are exposed to wave energy. Due to refraction, the side lobes of the delta with be less exposed than the front part, and due to breaking on the outer delta shoals, the gorge will also be less exposed (Fig 5.6).



Figure 5.5. 7. Sheltering and exposure to obliquely incident waves.

In the case of obliquely incident waves, one side lobe will be exposed, and the other will be sheltered. By implication, the wave climate one the side lobes will be different from that offshore or at the front edge of the delta. The western lobe in Fig 5.7, for instance, will be fully exposed to westerly waves, less exposed (due to refraction) to northerly waves and sheltered from easterly waves. As a consequence, wave-driven longshore currents in the vicinity of this western lobe are predominantly eastbound. Similarly, the currents near the eastern lobe are predominantly westbound.

Another aspect which deserves attention is the wave penetration into the inlet. Clearly, the

barrier islands provide considerable shelter to the basin, but wave energy can penetrate through the gorge. The latter, however, is sheltered to a certain extent from the open sea by the outer delta. Hence the wave energy which reaches the back of the gorge is much less than offshore. Subsequently, this energy usually radiates into the basin, whence the energy density and the wave height rapidly decay. Hence the wave energy which penetrates from the open sea into the basin is usually rather small and it is restricted to the area right behind the gorge. Wave energy further into the basin must is generally due to waves generated inside the basin, if the prevailing wind meets a sufficiently long fetch.

A drastically different situation may arise if 'channeling' of the wave energy occurs (Fig 5.8). This phenomenon is associated with trapping of the waves in the channel if the wave direction is almost parallel to the banks. In such a case, areas which are at first sight sheltered from sea waves can be exposed to reatively high wave energy and thus to much more erosion than expected.



Figure 5.5.8. Exposure of a sheltered area due to wave channeling

The complexity of the bed topography and the mixture of sea/swell waves coming from offshore and locally generated waves of a much shorter period inside the basin make wave modelling for tidal inlets particularly difficult. A model for long-crested monochromatic waves (i.e. wave fields with one direction, one period and one height in each point of the model domain) will not work here. Instead, one would need a spectral wave model which allows for irregular, short-crested waves of various periods and which includes wave generation. Such fully spectral wave models exist since some time for deep-water wave prediction, but this concept has been translated only recently to shallower water.

## 5.4 Currents

## 5.4.1 General

The currents in the vicinity of a tidal inlet are partly tidal, partly wave-driven, and partly winddriven. The tidal currents are primarily concentrated in the main channels, the wave-driven currents in areas where waves are breaking. Wind-driven currents occur mainly during storm events and are therefore rather episodic, but nonetheless important (though often forgotten!).

#### 5.4.2 Tidal residual circulation

An important physical mechanism to bear in mind when interpreting the tidal currents is inertia. The tidal flow in the gorge has so much momentum, that it cannot spread out fast enough when leaving the gorge: it forms a 'tidal jet'. The length scale of such an inertial jet is a few hundred times the water depth, so a few kilometres. On the other hand, the water flowing into the gorge has to accelerate and is therefore inertia-dominated, i.e. close to potential flow. This means that the highest velocities coincide with the shortest path through the gorge, i.e. just around the tips of the islands (Fig 5.9).



Figure 5.9. In- and outgoing tidal jet

The tidal residual current pattern in this highly schematized situation boils down to a quadruplet of gyres, two at either side of the inlet (Fig 5.10).



Figure 5.10. Schematized tidal residual currents around an inlet

In reality, the residual current picture is much more complicated than this. There is usually a distinct ebb-dominated current over the outer delta, and often there are flood channels near the tips of the islands. In the Wadden Sea, well-developed flood deltas are hardly found (the entire basin acts as a flood delta), and the corresponding flood-residual current is difficult to distinguish. Many inlets on the east coast of the USA, however, do exhibit such a feature and have a well-developed residual circulation inside the inlet.

Since the tidal current is coupled to the variation of the water level, there is another type of residual current, associated with the phase coupling between the horizontal and the vertical tide. This can be shown by the following example. Suppose *u* is the tidal current velocity in a 1-D channel, *h* the mean water depth and  $\zeta$  the water surface elevation above mean sea level, and let *u* and  $\zeta$  be given by

$$u = U\sin(\omega t - kx - \varphi)$$
 and  $\zeta = Z\sin(\omega t - kx)$ 

then the tidal residual flux of water is given by

$$q_{res} = U(h+Z)\frac{1}{T}\int_{0}^{T}\sin(\omega t - kx - \varphi)\sin(\omega t - kx)dt = \frac{1}{2}U(h+Z)\cos\varphi$$

So, if the horizontal tide (*u*) and the vertical tide ( $\zeta$ ) are 90° out of phase ( $\varphi = \pi/2$ ), like they are in deep water, there is no residual flux. But if they are more or less in phase ( $\varphi \approx 0$ ), there can be a considerable residual current. This becomes even more apparent for shoals bordering a tidal channel which are flooding and drying during the tide. There the largest part of the flooded stage coincides with the flood tide, so there must be a flood-dominant residual current. By implication, the residual current in the channel must be ebb-dominant.

In summary, there is a variety of mechanisms which lead to tidal residual currents, and

hence to residual sediment transports and moprphological changes. In shallow areas such as a tidal inlet, we have to give up thinking of the tide as a small-amplitude wave.

#### 5.4.3 Tidal asymmetry

Tidal asymmetry has been treated in the previous chapters as one of the mechanisms to import or export sediment to or from a tidal basin. It generally does not give rise to residual currents, but in wide channels it may, as a consequence of the coriolis effect (due to the earth's rotation). On the northern hemisphere, this effect gives every moving particle a tendency to bend to the right. If a water flow is laterally bounded, e.g. by a channel bank, this tendency is translated into a transverse water surface slope, which exerts an additional force on the particles, such that these keep moving straight on. Thus the tide at the right side of a tidal channel (when looking in the direction of tidal wave propagation) tends to be slightly higher than at the left side. If the tide is flood-dominant, it means that the inward residual current at the right side is somewhat stronger than at the left side. If, moreover, the tidal wave is reflected, the reflected wave will have induce a slightly stronger outgoing current at the other side of the channel. Thus the overall residual current pattern in wide channels can be as shown if Fig 5.11.

shoal			
low amplitude	weak residual		
INCOMING TIDAL WAVE			
high amplitude	strong residual		
sho	1		
shoal			
high amplitude	strong residual		
REFLECTED TIDAL WAVE			
low amplitude	weak residual		
shoa			
shoal			
<b>+ + + +</b> +	+ + +		
Snoar			

Figure 5.11. Tidal residual current due to a combination of tidal asymmetry and the coriolis effect

#### 5.4.4 Secondary flows

So far, we have considered the depth-averaged tidal current field. When considering this to be representative of the entire flow pattern, one makes the implicit assumption of vertical similarity of the velocity profiel, i.e. the velocity profile in every point in the horizontal has the same shape (e.g. logarithmic). In reality, however, the velocity field is more complex than this. Like in rivers, the curvature of the tidal current, as well as the coriolis effect, give rise to secondary flow components:

(1) 
$$v_{curv} = \frac{h\overline{u}_{tot}}{R_c} f_{curv} \left(\frac{z-z_b}{h}\right) \quad and \quad v_{cor} = -f_c h \frac{\overline{u}_{tot}}{|\overline{u}_{tot}|} f_{cor} \left(\frac{z-z_b}{h}\right)$$

in which v is the transverse velocity component (perpendicular to the depth-averaged flow, positive to the left), h is the water depth,  $R_c$  is the radius of curvature of the depth-averaged tidal flow (positive when turning to the right),  $f_c$  is the coriolis factor (1.2 10<sup>-4</sup> at our latitude), z is the vertical co-ordinate,  $z_b$  is the bed level, and  $f_{curv}$  and  $f_{cor}$  are the vertical distribution functions of these secondary flow components.

These secondary flows, though rather weak compared to the maximum tidal current, can have a significant residual effect, on the current, but also on the sediment transport and the bed topography. This goes especially for the curvature-induced secondary flow, which does not change sign as the tide turns: in the upper part of the water column it is always directed towards the centre of curvature of the flow, in the lower part always away from it. Thus the curvature-induced secondary flow contributes to the maintenance of shoals, for instance (see Fig 5.12).



Figure 5.12. Curvature-induced secondary flow: always directed towards the shoal

A striking example of the role of curvature-induced secondary flow on an outer delta is the case of the groyne which was built to protect the north part of the coast of Texel, one of the barrier islands of the Dutch Wadden Sea. The idea was to let this groyne interrupt the longshore drift into the inlet called Eijerlandse Gat. Studies with 2-D depth-averaged models revealed, that this would lead to the 'textbook-'pattern of updrift accretion and downdrift erosion, so to strong erosion north of the groyne. It was foreseen that the groyne would have to be connected by a hard structure to the existing revetment which protects the tip of the island.

Once the groyne had been built, however, accretion occurred at either side of it, to the effect that it is now largely buried in sand (so the project turns out to be extremely successful!). This must be attributed to a large extent to the curvature-induced secondary flow in the gully around the tip of the groyne, which consistently brings sediment towards the groyne (see Fig 5.13). New computations with a model which includes this secondary flow effect have reproduced the observed morphological evolution (further see Steijn et al., 1998).



Figure 5.13. Observed (left) and computed (right) sedimentation and erosion around the groyne at the north point of Texel during the first two years after construction [from Steijn et al., 1998]

#### 5.4.5 Wave-induced currents

As stated before, an outer delta is nowhere near a straight prismatic coast. This means that the traditional way of thinking about wave-driven currents (longshore and cross-shore) is hardly applicable here. In order to build up a more 3-D view on wave-driven currents, we consider the depth-averaged shallow-water equations

(2) 
$$\frac{\partial \overline{u}}{\partial t} + \overline{u}\frac{\partial \overline{u}}{\partial x} + \overline{v}\frac{\partial \overline{u}}{\partial y} = -g\frac{\partial z_s}{\partial x} - \frac{\tau_{bx}}{\rho h} - \frac{1}{\rho h}\left[\frac{\partial S_{xx}}{\partial x} + \frac{\partial S_{xy}}{\partial y}\right] + HDT$$

(3) 
$$\frac{\partial \overline{v}}{\partial t} + \overline{u} \frac{\partial \overline{v}}{\partial x} + \overline{v} \frac{\partial \overline{v}}{\partial y} = -g \frac{\partial z_s}{\partial y} - \frac{\tau_{by}}{\rho h} - \frac{1}{\rho h} \left[ \frac{\partial S_{xy}}{\partial x} + \frac{\partial S_{yy}}{\partial y} \right] + HDT$$

(4) 
$$\frac{\partial h}{\partial t} + \frac{u}{\partial x} \frac{\partial h}{\partial x} + \frac{v}{\partial y} \frac{\partial h}{\partial y} + h \frac{\partial \overline{u}}{\partial x} + h \frac{\partial \overline{v}}{\partial y} = 0$$

in which x and y are cartesian co-ordinates in the horizontal plane, u and v are the horizontal velocity components in the x- and y-direction, respectively, the overbars denote depth-averaging, h is the water depth, g is the acceleration due to gravity,  $z_s$  is the water surface level,  $\tau_{bx}$  and  $\tau_{by}$  are the components of the bed shear stress,  $S_{xx}$ ,  $S_{xy}$  and  $S_{yy}$  are the radiation stress components and *HDT* stands for the horizontal diffusion terms. The dependent variables in these equations are the velocity components and the water level. Therefore, additional information about the bed shear stress and the radiation stresses is needed before being able to solve this system. The radiation stress components can be evaluated on the basis of a wave propagation model, the bed shear stress is related to the velocity and a number of wave parameters (cf. Soulsby et al., 1993).

Theoretically, there is nothing wrong with this approach, but practically there is: it is not very transparent as far as the wave-induced forces are concerned. Therefore, we will consider an alternative formulation of these forces:

(5) 
$$\begin{pmatrix} F_x \\ F_y \end{pmatrix} = \frac{D}{c_w} \begin{pmatrix} \cos \theta \\ \sin \theta \end{pmatrix} + \rho h \begin{pmatrix} \frac{\partial \Phi}{\partial x} \\ \frac{\partial \Phi}{\partial y} \end{pmatrix}$$

in which *D* is the wave energy dissipation rate per unit area,  $c_w$  is the phase speed of the waves,  $\rho$  is the mass density of the fluid and  $\vartheta$  is the angle between the direction of wave propagation and the positive *x*-axis. The potential  $\Phi$  is given by

(6) 
$$\Phi = \frac{\left(n - \frac{1}{2}\right)E}{h}$$

in which *E* is the wave energy density, and *n* denotes the ratio between the group celerity and the phase celerity of the waves.

This way to compute the forces is essentially simpler and more transparent than differentiating the radiation stress components. Note that the first part in the RHS of Eq. (5) represents a rotational force field, which can drive circulation currents. This part of the force vector is proportional to the wave-energy dissipation rate and its direction coincides with the direction of wave propagation. Via these properties, wave-induced current fields are much easier to interpret than via the forces based on the radiation stress gradients. The second part of the forces according to Eq. (5) is irrotational and will generally yield just a

'landscape'of water surface elevations. Here we will ignore this part of the forcing. Substituting the dissipation-related force components into the momentum equations (2) and (3) yields

(7) 
$$\frac{\partial u}{\partial t} + \frac{\partial u}{\partial x} + \frac{\partial u}{\partial y} = -g \frac{\partial z_s}{\partial x} - \frac{\tau_{bx}}{\rho h} + \frac{D}{\rho h c_w} \cos \theta + HDT$$

(8) 
$$\frac{\partial v}{\partial t} + \frac{u}{\partial x} \frac{\partial v}{\partial x} + \frac{v}{\partial y} \frac{\partial v}{\partial y} = -g \frac{\partial z_s}{\partial y} - \frac{\tau_{by}}{\rho h} + \frac{D}{\rho h c_w} \sin \theta + HDT$$

In the case of steady flow near a prismatic coast along the x-axis, this reduces to

(9) 
$$0 = -\frac{\tau_{bx}}{\rho h} + \frac{D}{\rho h c_w} \cos \theta$$

$$0 = -g \frac{\partial z_x}{\partial y} + \frac{D}{\rho h c_w} \sin \theta$$

So here the longshore and cross-shore momentum equations get separated, the former describing the momentum balance for a uniform longshore current and the latter the set-up balance.

In the case of a complex topography with interrupted breaking, however, the situation is much more complicated. A simple example concerns a shoal on which waves are breaking (Fig 5.14).



Due to refraction, the waves will tend to converge toward the top of the shoal. As the wave break on the seaward slope of the shoal, they generate a current-driving force. At the top of the shoal, however, there is no closed boundary to stop the water. Hence the 'cross-shore' component of the force cannot build up a water level set-up. Instead, the water will flow in the direction of the force, until it reaches the channel behind the shoal, where water level gradients will deflect it and drive it to the sides of the shoal. There it has room to flow seawards again, thus closing the circulation.

net convergence of wave energy to the delta

Figure 5.14. Wave-induced forces and currents around a shoal



Figure 5.15. Yearly residual transport patern on the outer delta of the Frisian Inlet

Fig 5.15 gives an example of a residual transport computation for the Frisian Inlet. Apart from the longshore drift along the coast of Ameland, it clearly shows the above circulations around the shoals, always with the current at the top of the shoal in the direction of wave propagation. Apparently, the longshore drift and the transport by the wave-driven currents around the shoals on the outer delta are so strong, that they dominate the tidal residuals.

In principle, the wave-driven current in this type of situation is 3-D. The mechanisms which use to be called 'cross-shore' in the case of a prismatic coast, are actually acting in the direction of wave propragation.

The effective surface shear stress which drives the undertow is basically the same as the forcing of the whole wave-driven current. In a depth-averaged model, this force is treated as if it were distributed uniformly over the vertical, but in fact it is concentrated near the water surface. Since it acts in the direction of wave propagation and the wave-driven current or the set-up gradient does not necessarily have the same direction, the undertow-type of current which we know from prismatic coasts is probably rare here.

The cross-shore current induced by the mass flux is even more complicated here. In fact, the return current due to the Eulerian mass flux between the crest and trough levels of the waves is part of the primary flow, since the mass flux terms ought to be included in the depth-averaged flow eqautions. As long as the mass-flux contributions are small relative to the other current components, they can be included by replacing the velocity components by

(11) 
$$\left(\frac{\overline{u}}{\overline{v}}\right) \rightarrow \left(\frac{U}{V}\right) = \left(\frac{\overline{u}}{\overline{v}}\right) + \frac{1}{\rho h} \left(\frac{M_x}{M_y}\right) \quad with \quad \left(\frac{M_x}{M_y}\right) = \frac{E}{c_w} \left(\frac{\cos\theta}{\sin\theta}\right)$$

in which  $M_x$  and  $M_y$  are the components of the wave-induced mass flux. Thus the same numerical solver can be used as for Eqs. (3) through (5), but now with U and V as dependent variables and with slightly different the boundary conditions. The computed velocities (U, V) have to be corrected afterwards for the mass-flux contribution.

The implication is, that the magnitude and direction of the primary flow contribution of the mass flux are difficult to assess, since the extent to wehich this effect creates a set-up or a current velocity depends entirely on the situation. In any case, mass-flux contributions exactly opposite to the direction of wave propagation, like in the case of a prismatic coast, are not likely to be abundant in outer delta flow.

## 5.4.6 Wind-induced currents

An often forgotten type of currents in tidal inlets are the wind-driven currents, either directly, via the wind-induced shear stress on the water surface in the inlet area, or indirectly, via the set-up of the water level against the coast. The former effect emerges directly when including the wind shear stress components  $\tau_{wx}$  and  $\tau_{wy}$  into the depth-averaged momentum equations:

(12) 
$$\frac{\partial \overline{u}}{\partial t} + \overline{u}\frac{\partial \overline{u}}{\partial x} + \overline{v}\frac{\partial \overline{u}}{\partial y} = -g\frac{\partial z_s}{\partial x} - \frac{\tau_{bx}}{\rho h} + \frac{\tau_{wx}}{\rho h} + \frac{D}{\rho h c_w}\cos\vartheta + HDT$$

(13) 
$$\frac{\partial \overline{v}}{\partial t} + \overline{u} \frac{\partial \overline{v}}{\partial x} + \overline{v} \frac{\partial \overline{v}}{\partial y} = -g \frac{\partial z_s}{\partial y} - \frac{\tau_{by}}{\rho h} + \frac{\tau_{wy}}{\rho h} + \frac{D}{\rho h c_w} \sin \theta + HDT$$

In these equations, the wind stress terms are divided by the water depth, which means that the wind tends to be more effective in driving a current when it acts on shallower water. Due to the large variations in water depth which are inherent to a tidal inlet, the wind-driven current field will therefore strongly vary in space. In general there will be a tendency of the flow to follow the wind in the shallower parts, and to oppose it in the deeper parts, but this picture can be complicated greatly by spatial interactions via water level gradients.

As the wind forcing acts at the water surface, there will also be an effect on the vertical structure of the flow: the primary flow profile (e.g. logarithmic) will be disturbed by a 'secondary'flow component which follows the wind in the upper part of the water column and goes against it in the lower part. Note that this secondary flow has to be superimposed on the primary flow (the depth-averaged circulation with a primary flow profile). The result can be a complex 3-D flow pattern.

Probably even more important than the direct wind-induced forcing is the effect of the water level set-up during a severe storm. Although the peak of the wind speed usually does not last much more than a few hours, the water level set-up can last much longer. However, it takes

time for the back-barrier basin to follow the water level at open sea, as setting up such a basin by a few metres takes huge amounts of water, all of which has to be squeezed in through the inlets. Hence it sometimes occurs that the ebb current is entirely suppressed and that the there is a flood current in the inlet during a day or more. Clearly, such events may offset the inlet morphology considerably. It is also clear that new inlets and channels will preferably be created under such conditions, when there is a large head differences between the sea and the basin.

## 5.4.7 Summary

In summary, we have seen that the current field around an inlet is essentially more complex than on a uniform straight coast, and that all consituents of this field are essentially 3-D. So when modelling currents around inlets, we have to think 3-D, even if we decide to use a 2-D depth-averaged model. The interpretation and postprocessing of the results (e.g. bed shear stress to be put into the sediment transport model) is much less straightforward here.

## 5.5 Wave-current interaction

The tidal and wave-driven current pattern on the outer delta is largely concentrated in the deeper channels. Consequently, there can be strong currents which affect the wave propagation via the so-called current refraction. This may even go as far as wave blocking. When standing on the coast and overlooking a tidal inlet, one often sees a sharp distinction between areas with waves and areas with a flat water surface. This is simply because the current is strong enough to prevent the waves from entering this area. Moreover, where waves do occur, their patter is often quite irregular, again as a consequence of refraction on a strongly varying current field.

This form of wave-current interaction makes it particularly difficult to predict the wave field on the outer delta. Such a prediction should be based on a combined wave and current model, and both should be carefully calibrated in order to find the right pattern. In fact, this is an impossible task without synoptic wave data (also see Section 3).

A more straightforward form of wave-current interaction is the effect of waves on the bottom shear stress experienced by the current. The mechanisms underlying this effect are not essentially different from those in the case of a prismatic coast, though waves and near-bed currents can have arbitrary directions now. The result is a complex 3-D boundary layer with a strongly veering velocity vector and a non-trivial direction of the bed shear stress. In large-scale models like the ones we use for tidal inlets, however, this effect is usually ignored and the shear stress is assumed to be opposite to the mean current. What we do take into account, however, is the bed shear stress enhancement induced by the waves. This may have major effects on the current pattern: especially the tidal flow will tend to avoid shallow areas, where wave action and shear stress enhancement are strongest.

An anecdotic example concerns a model of one of the inlets of the Wadden Sea, which was run not only for the usual 'representative' – realtively low - wave conditions, but also for an extreme storm event. To our surprise, the longshore tidal current in the vicinity of the outer delta virtually vanished in the storm case! Our initial response, being unexperienced in storm modelling: this cannot be right, there must be something wrong with the model ....

## 5.6 Sediment transport processes

Waves and currents, separate and combined, are known to be able to transport sediment in various modes:

bed load transport, primarily via the migration of small-scale bedforms,

sheetflow transport, with the top layer of the bed fluidized during part of the wave cycle, suspended load transport, with the grains suspended in the water column.

The quantitative description of these transport modes is still largely empirical (e.g. Van Rijn, 1989; Fredsoe and Deigaard, 1992). There is a variety of transport formulae and models, each with its own strong points and weaknesses. This is not the place to discuss them in depth, since such a discussion is not typical for tidal inlets. For an overview we refer to

Horikawa (1988), Van Rijn (1989) and Fredsoe and Deigaard (1992).

For morphological computations, an important, but much neglected transport component is the one due to gravitation along a sloping bed: moving grains will rather go downhill than uphill. This gives a transport component which is directed downhill. A highly simplified transport model including this term reads

(14) 
$$\begin{pmatrix} q_x \\ q_y \end{pmatrix} = q_{tot} \left[ \frac{1}{\overline{u_{tot}}} \begin{pmatrix} \overline{u} \\ \overline{v} \end{pmatrix} - \beta \begin{pmatrix} \frac{\partial z_b}{\partial x} \\ \frac{\partial z_b}{\partial y} \end{pmatrix} \right]$$

in which  $q_{tot}$  is the total transport rate,  $z_b$  is the bed level and the factor  $\beta$  is a known scalar. The factor  $\beta q_{tot}$  basically measures the degree of sediment motion. Note that, if this motion is oscillatory, like in the case of combined waves and currents, this degree is much larger than one would tell from the wave-averaged transport rate. In that case, the factor in front of the slope term should rather be  $\beta < q_{tot} >$ , in which the brackets indicate wave averaging. If we subsitute Eq. (14) into the sediment balance equation, we find

(15) 
$$(1-\varepsilon_p)\frac{\partial z_b}{\partial t} = -\frac{\partial q_{bx0}}{\partial x} - \frac{\partial q_{by0}}{\partial y} + \beta q_{tot} \left(\frac{\partial^2 z_b}{\partial x^2} + \frac{\partial^2 z_b}{\partial y^2}\right) + h.o.t.$$

in which the suffix '0' refers to the situation with the same water motion on a horizontal bed and '*h.o.t.*' stands for higher-order terms, i.e. terms which are of higher order (quadratic or more) in terms of the bed slope. Clearly, the downhill gravitational transport component yields a diffusion term in the sediment balance equation, and thus has a smoothing effect on the bed topography. This effect is of great importance to the morphological stability of the bed and to the equilibrium state to which the bed topography tends (but which it probably never reaches, since this state is a function of the ever changing input conditions). For further reference: see De Vriend et al., 1993.

So far, little is known about the downhill gravitational effect on the transport under sheetflow conditions. Since this type of transport is typical of highly energetic wave conditions, when large amounts of sediment are brought into motion, the effect may well be considerable. The same goes for suspended load transport. At first sight, it may seem unlogical to assume that suspended load transport be subject to bed slope effects: the sediment grains are in suspension, and therefore not in contact with the bed, so how can the experience any effect of the bed slope? We have to realise, however, that the suspended state applies to the average sediment concentration, but not permananetly to each individual grain; each grain will get into contact with the bed from time to time, and thus experience the slope effect. There are indications from river research that this influences the bed topography, e.g. in river bends. The nature and the magnitude of this effect under coastal conditions, however, are still in the dark

Suspended load transport is a quite common condition in the coastal environment, maybe even more common than pure bed load. From a morphological modelling point of view, we must distinguish two types of suspended load: one which is determined entirely by the hydrodynamic conditions and the sediment properties at the point of consideration, and one which includes a 'memory effect' and responds to the conditions in all points it has come through in the past. The former type can be modelled with a sediment transport formula (e.g. the Bijker or the Bailard formula), or with a simple intra-wave model which describes the suspension process during a wave cycle (cf. Fredsoe and Deigaard, 1992). The sediment concentration associated with the second type of suspended load transport is described by an advection/diffusion equation of the type (e.g.Wang and Ribberink, 1986 for tidal currents, and Katopodi and Ribberink, 1992, for nearshore applications)

(16) 
$$T_{A} \frac{\partial \overline{c}}{\partial t} + L_{A} \left[ \frac{\overline{u}}{\overline{u_{tot}}} \frac{\partial \overline{c}}{\partial x} + \frac{\overline{v}}{\overline{u_{tot}}} \frac{\partial \overline{c}}{\partial y} \right] = \overline{c_{e}} - \overline{c}$$

in which  $T_A$  is a time scale of the order of magnitude  $h/w_s$  and  $L_A$  is a length scale of the order of magnitude  $u_{tot}h/w_s$ . The equilibrium concentration,  $c_e$ , corresponds with the uniform situation and is usually derived from a sediment transport formula, or from a simpler model which applies to uniform situations.

Note that Eq. (16) can also be considered as a decay equation of the type

(17) 
$$\frac{d\bar{c}}{dt} + \frac{\bar{c}}{T_A} = \frac{\bar{c}_e}{T_A}$$

in which the d/dt stand for the material derivative, i.e. moving along with the sediment flow. If the concentration at t=0 is given, the general solution can be witten as

(18) 
$$\overline{c}(t) = \overline{c}(0) e^{-t/T_{\rm d}} + \int_{0}^{t} \overline{c_e}(\tau) e^{-(t-\tau)/T_{\rm d}} d\tau$$

in which  $\tau$  is a formal time variable which runs from 0 to the actual time *t*. Apparently, all values of the equilibrium condition encountered in the time interval between 0 and *t* contribute to the forcing term (i.e. the second term in the right-hand part of the equation), but their contribution becomes smaller as they occurred longer ago. This shows that the memory of the system is not infinite, but has a certain time scale, viz.  $T_A$ .

This memory effect of the suspended load concentration can be rather important in a tidal inlet, with its strong spatial variations of the bed topography and the hydrodynamic conditions.



Figure 5.16. Residual uphill transport due to deformations of the velocity profile

The combination of suspended load transport with secondary flows has another important effect. The sediment concentration is usually not uniform ovr the vertical: most of the sediment is concentrated in the lower part of the water column. Consequently, secondary flows will have a net effect on the transport direction. The curvature-induced secondary flow, for instance, will give a net transport component in towards the centre of curvature, and the undertow-induced secondary flow will give a net transport component against the direction of wave propagation. Another potentially important contribution of this type is associated with the deformation of the velocity profile as the flow moves up or down a steep slope (see Fig 5.16). When averaged over the tide, this gives a net upslope transport. This may explain why steep banks of tidal channels can be stable.

#### Sediment transport patterns

The combination of wave, curerent and sediment transport mechanisms on the outer delta gives rise to a typical residual sediment circulation pattern. Fig 5.4 shows a highly schematised picture of it, reality is usually more irregular and diffuse (cf. Fig 5.17).



Figure 5.17. Residual transport pattern around the Eijerlandse Gat, as derived from field observations and results of various mathematical models

In general, the flood channels carry sediment from the adjacent coasts to the inlet. Depending on the demand of the basin, this sediment is transporte into the basin, or somehow it ends up in the main ebb channel, which brings it out towards the edge of the outer delta. There part of it is picked up by wave-driven and/or tidal longshore currents (storms!) and transported along the delta edge, ultimitely towards the downdrift island, or towards the updrift one, depending on the situation. Figure 5.18 gives an example of such a residual transport pattern.



Figure 5.18. Yearly residual transport pattern on the outer delta of the Eijerlandse Gat

Another part of the sediment from the ebb channel, as well as part of what is brought into the area by the longshore drift along the updrift island, end up in the shoal system on the outer delta. Due to the various hydrodynamic and sediment transport processes around these shoals, there is a long-term residual transport and a slow migration of the shoals in the direction of the longshore drift (Fig 5.19). The time needed for a shoal to cross the inlet may be tens of years (for the Borndiep, for instance, the inlet between Terschelling and Ameland, it is typically 40 years).



Figure 5.19. Sediment bypassing via shoal migration on the outer delta.

An example of what can go wrong when ignoring the nature of this sediment circulation refers to Nerang River, Australia, where an inlet to a tidal lagoon was made into a shipping channel by building groynes at either side of the inlet and bypassing the longshore drift via a pumping system. This scheme works well from the point of view of downdrift erosion, but it does a much worse job for the outer delta and the hydrodynamic conditions in the inlet are concerned. By artificially bypassing all the sediment which arrived at the updrift side of the inlet, the shoal bypassing system was cut off from its sand supply, the shaols rapidly degraded, and the outer delta disappeared. As a consequence, the tidal motion into and out of the lagoon was no longer hampered by the delta and increased considerably. This led to a dramatic increase of the flow velocities in the inlet, due to which its navigation function (the prime reason for carrying out the scheme) is heavily threatened.

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# 6. AGGREGATED MODELLING OF COASTAL INLETS AND TIDAL BASSINS

MATHEMATICAL MODELLING OF MESO-TIDAL BARRIER ISLAND COASTS. PART 1: EMPERICAL ANS SEMI-EMPERICAL MODELS

## 6.1 Introduction

A large part of the world's coastline consists of more or less elongated barrier islands. They are separated from the mainland by a shallow tidal basin and from each other by relatively narrow inlets to that basin. For various reasons, these barrier island coasts are important issues in coastal zone management.

Barrier islands tend to be morphologically quite active, whence they are difficult to be controlled by engineering measures. Yet, many of these islands are inhabited, and the houses and properties often require protection measures. This is a continuous source of problems at the engineering level.

The outer deltas in front of the inlets store large amounts of often well-sorted sand, which is attractive to be mined for all sorts of purposes (for instance beach nourishment). As the delta is part of a coherent morphological system, however, this sand mining may have detrimental effects on the other parts of this system (viz. the adjacent island coasts and the basin).

The sheltered borders of estuaries and lagoons are attractive places for economical and recreational activities related to the sea. This explains why many inlets are important navigation channels. As inlet channels tend to be morphologically even more active than the barrier islands, it is tempting to try and stabilize them with engineering works like jetties and revetments. The connection of such works with the mobile natural system is rather problematic and costly.

Many tidal lagoons, especially the intertidal zones, are of paramount importance to the ecosystem as feeding and breeding grounds for many species throughout the food chain. Thus these areas have a high nature-preservation status. At the same time, the system is quite vulnerable, because the intertidal area is very sensitive to changes in the physical system (e.g. accelerated mean sea-level rise).

Since tidal inlet systems respond to the mean sea level, the tide and the wave conditions, they are not only threatened by sea level rise and subsidence. Also changes in the tidal regime (amplitude, asymmetry) and the wave climate (mean wave height and direction, storm climate, chronology of events) can have a major impact.

The ability to understand and predict the morphological behaviour of barrier island coasts, including inlets and basins, is therefore an important issue in coastal zone management. Mathematical models are particularly attractive here, because of their flexibility and because the complexity of the system makes physical scale modelling very difficult, especially when it concerns long-term morphological evolution.

In recent years, tidal inlets have come within the reach of mathematical modellers. It has become possible to model many of the complex physical processes (waves, currents, sediment transport, morphological changes) which take place in the vicinity of an inlet. Such models enable researchers to develop and test hypotheses on how an inlet system works, and to identify the most important knowledge gaps. They also enable practitioners to analyse the response of a given inlet system, for instance to proposed engineering measures.

In spite of the computer power which is available at present, these process-based simulation models are as yet unable to cover time spans which are much larger than the principal inherent morphological time scales. Much of the empirical knowledge on the equilibrium state of tidal inlets, for instance, cannot yet be reproduced by these models. This has boosted the formulation of semi-empirical long-term models which are based on elementary physics, but include the empirical knowledge.

As it is hardly possible to cover the entire range of coastal inlets, this chapter focuses on what Hayes (1979) calls *mixed-energy tide-dominant* inlets. This is the type of inlets which is common along the Northwest European coast, for instance. The tidal range is typically a few metres and the yearly average significant wave height is typically 1 m. In their natural state, these inlets are rather wide, with a large dynamic outer delta and drumstick-shaped barrier islands.

After a brief outline of the practically relevant phenomena which inlet models should be able to cover, the various model types are classified. Except for the process-based simulation models, which are treated in a companion chapter, each of the classes is described and its potentials, shortcomings and perspectives are discussed.

## 6.2 Phenomena to be modelled

In the present state of development it is not realistic to claim that one model type can provide the necessary morphological information for all possible practical problems. In other words: the all-purpose model for tidal inlet morphodynamics does not exist and is not likely to emerge in the near future. This means that the present models are linked to classes of phenomena.

Barrier islands, outer deltas, inlet channels and flood basins are all elements of one coherent morphodynamic system (Fig. 6.1). The nature of this "tidal inlet system", however simple at first sight, is very complex. It is forced by a partly unpredictable input signal (waves, storm surges), and its behaviour involves a wide range of space and time scales. Part of this behaviour is imposed by variations of the forcing signal, part is inherent to the system. An example of forced behaviour is the 18.6-yearly variation of the inlet geometry due to the nodal cycle of the tide (Oost et al., 1993). An example of inherent behaviour is the formation and migration of channel and shoal patterns in the flood basin and on the delta fringe (for instance, see Ehlers, 1988).

The complexity and the scale range of this inherent behaviour make tidal inlets particularly difficult to model. Models which perform well in describing small-scale phenomena are not necessarily suited for long-term predictions, even if there is ample computer time available. The phenomena which a model is able to describe are therefore classified relative to the inherent scale of the corresponding morphodynamic process. The latter represents both the space and the time scale, because these are assumed to be coupled: larger-scale phenomena will proceed more slowly than small-scale ones (cf. De Vriend, 1991; Fenster et al., 1993). This leads to the following classification:



Figure 6.1: Larg-scale tidal inlet system

- 1. Micro-scale (*process-scale*) phenomena take place at an essentially smaller scale than the corresponding morphodynamic behaviour. It concerns primarily the constituent processes (waves, currents and sediment transport). Examples of practically relevant information are
  - wave conditions in channels, on shoals and near structures; relevance: navigation, habitat, coastal defence;
  - current speed in channels, on shoals and near structures; relevance: navigation, habitat, local scour;
  - patterns of water mass exchange; relevance: water guality, import of nutrients, larvae, etc.;
  - suspended sediment concentration and turbidity; relevance: habitat, mobility of channels and shoals;
  - sediment transport rate; relevance: mobility of channels and shoals; import of adhering nutrients and pollutants.
- 2. Meso-scale (*dynamic scale*) phenomena concern the "primary" morphodynamic behaviour, due to the interaction of the constituent processes and the bed topography. The scales involved are of the order of magnitude of those inherent to this interaction, which is generally much larger than those of the constituent processes. Examples of practically relevant information are
  - bedform (ripples, dunes, sandwaves) dimensions, propagation and patterns; relevance: influence on current and transport patterns, depth for navigation, risk of denudation of pipelines.
  - shoaling and migration rates of natural channels; relevance: navigation, coastal defence.
  - reside time of bottom pollutants; relevance: water quality, habitat.
  - morphological changes around man-made works; relevance: stability of structures, (un)burial of pipelines, shoaling of dredged navigation channels, filling-up of sand-mining pits.
  - response to human interference, such as land reclamation, persistent sand mining,
    - repeated channel dredging, etc.;

relevance: to avoid detrimental effects elsewhere in the system.

- 3. Macro-scale (*trend-scale*) phenomena concern slow trends at scales much larger than those of the "primary" morphodynamic behaviour. These trends can be due to secular effects in the system's inherent behaviour, or to gradual changes in the extrinsic forcing or the system parameters. The cyclic migration of channels in a wide inlet (Bakker and Joustra, 1970; Oost and De Haas, 1993; Oost, 1995) may be considered as a secular effect. Gradual changes in the extrinsic forcing may be due to mean sea-level rise, subsidence, tidal range evolution, etc. Erodibility is an example of a system-parameterwhich is is affected by changes in the cohesive sediment content, the intensity of mussel fishing, and the marshland vegetation (due to pollution, for instance). Examples of practically relevant phenomena are:
  - evolution of channel/shoal patterns in the basin and on the delta; relevance: habitat and ecosystem, coastal defence at barrier islands, navigability, stability of pipelines;
  - evolution of the intertidal zone (area, level, duration of flooding and drying); relevance: habitat and ecosystem, (indirectly:) coastal defence at barrier islands;
  - inlet instability; relevance: coastal defence, ecosystem, navigation.

The above classification is relative, so the scale classes cannot be expressed in absolute terms of metres or years. Fig. 6.2 illustrates that the three classes exist for a wide range of phenomena at different scales.

# 6.3 Classification of mathematical models

The statement that the all-purpose model does not exist is substantiated by the wide variety of tidal inlet models which has been proposed. The above classification of phenomena can be used to structure this set.

Another axis of classification is the model approach. In broad outline, the following model types can be distinguished:



Figure 6.2: Phenomena and scales

- *data-based models*, making use of measured data only and aiming directly at the phenomena to be described;
- *empirical relationships*, also observation-based, but establishing a relationship between input and output of the system;
- process-based models, which describe waves, currents, sediment transport and bed level changes via a set of mathematical equations based on first physical principles (conservation of mass, momentum, energy, etc.);
- **formally integrated long-term models**, which are derived from a process-based model by formal integration over time (and space), with empirical or parametric closure relationships;
- semi-empirical long-term models, which describe the dynamic interaction between large elements of the system, using empirical relationships to represent the effects of smaller-scale processes.

Fig. 6.3 shows that these two axes of classification are not entirely orthogonal. Henceforth, the various model types will therefore be distinguished by approach only.

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Figure 6.3: Mathematical models and phenomena

## 6.4 Data-based models

Geostatistical models belong to the class of data-based models. The simplest form of geostatistical modelling is the extrapolation in time of a certain parameter of the system's state (for instance the minimum depth of a certain channel) via a linear regression analysis of observed values. The model essentially consists of the coefficients which result from the regression analysis, together with the assumption that the parameter to be predicted will continue its past evolution. The underlying assumption is that the processes which determine the trend in this evolution will remain constant.

Fenster et al. (1993) decribe a more sophisticated version for the multi-scale nonlinear system of an uninterrupted coast. Although inlet systems are essentially more complicated, the technique should be transferable to tidal inlets. A necessary condition for validity as a prediction tool is that the processes which lead to the described behaviour will remain unaltered in the future.

Another form of data-based modelling is "translation" of observations at one site to predictions at another. This approach may be applicable if the system is to undergo changes (due to engineering works, for instance) which have taken place before in a well-monitored similar system. The underlying assumption is that the two systems will exhibit a similar response to this interference.

Empirical Orthogonal Function (EOF) analysis (Preisendorfer and Mobley, 1988) is a form of data-based modelling which separates various modes of variation in space and time. The weighted sum of products of these modes constitutes the complete data set, and for each of these products the content of measured information can be assessed. Thus EOF-modes can be ordered according to their content of information on the observed behaviour. In the case of uninterrupted coasts, these series use to converge rapidly, which means that the behaviour can largely be mapped onto a few EOF-modes (cf. Medina al., 1992; Wijnberg and Wolf, 1994; also see De Vriend et al., 1993a). As yet, aplications to tidal inlet systems are scarce, so it is unclear whether this favourable property also applies there.

In general, EOF-analysis allows for a certain degree of physical interpretation per component. Hence extrapolation into the future can be done per component, based on physical considerations. At any future time, the state of the system can be composed of the various components.

Data-based models of tidal inlets typically concern macro-scale phenomena, such as the historical or geological evolution of the system as a whole. Their concept ("let nature tell what it is like") may seem attractive for complex systems, because it seems to evade the necessity to analyse the system's behaviour. It is exactly this complexity, however, which rules out such a "black-box" approach. In simple cases it is relatively easy to predict whether or not the relevant conditions and processes are likely to remain unaltered. The system's behaviour is transparent and it is determined by one or a few predominant mechanisms. The relevant input and process parameters are not many, which makes it easier to predict whether the system's behaviour is likely to change in the future. In a system as complex as a tidal inlet, however, there are many input and process parameters to take into account, and the system's behaviour is difficult to analyse. Hence it is difficult to asses which mechanisms determine which aspect of the system's evolution. Utilizing data-based models in such a situation therefore requires an equally thorough insight into the mechanisms at work as any physics-based model.

## 6.5 Empirical models

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## Equilibrium-state relationships

O'Brien (1931, 1969) was the first to suggest a linear relationship between the crosssectional area of an inlet and the tidal prism, defined as the total amount of water which moves in and out the inlet during one tide,

$$A_c = c_A P$$
 1

in which  $A_c$  = cross-sectional area of the inlet channel below MSL [m<sup>2</sup>],

 coefficient of proportionality; Eysink (1990) suggests for the Dutch Wadden Sea a value of 7.0×10<sup>-5</sup> m<sup>-1</sup>, to be used in combination with the mean tidal prism,

P = representative tidal prism [m<sup>3</sup>].

Qualitatively speaking, a positive correlation between  $A_c$  and P makes sense, because channels are likely to be larger if they have to convey more water. An obvious disadvantage of the above form is that the constant of proportionality is dimensional. Measured data from many inlets all over the world have confirmed that such a relationship exists (Fig. 6.4), although Jarrett (1976) claims that the exponent of P should be 1.1 for the inlets on the U.S. coasts. The validity of Eq. (1) has also been shown for the Dutch and German Wadden Sea inlets (cf. Eysink, 1990; Schroeder et al., 1994). Moreover, a similar relationship was shown to exist over a very wide range of scales for channels inside the flood basin (Gerritsen, 1990; Eysink et al., 1992). In that case, P stands for the total amount of water conveyed by the channel during one tide.

An alternative version of this relationship expresses  $A_c$  in terms of the so-called "stability shear stress", which is defined as the shear stress needed to maintain a zero net transport gradient along the channel (Gerritsen, 1990; Friedrichs, 1995). This shear stress is not necessarily equal to the critical shear stress for sediment motion. In systems terms, Eq. (1) represents a typical "black-box" input/output relationship. Unlike a geostatistical model, it describes the state of the system (represented by  $A_c$ ) as a function of the forcing (represented by P). Clearly, a widely valid empirical relationship like this must concern the equilibrium state, otherwise site-specific history-effects would have led to a much larger scatter in the  $A_c$ -P diagram.



Figure 6.4: Emperical relationship between cross-sectional area and tidal prism [from: Jarrett, 1976]

Later work on inlet data has revealed a number of other relationships between equilibriumstate properties and the tidal prism, such as

- the flood-basin volume below MSL (Eysink, 1990), found to be proportional to P<sup>1.5</sup>. This volume is largely equal to the sum of volumes of the flow-conveying channels. Since the cross-sectional area of these channels is proportional to P and their length probably scales with the square root of the basin area, the proportionality with P√P seems to make sense.
- the sand volume of the outer delta (Walton and Adams, 1976; Eysink et al., 1992), which is proportional to P<sup>1.23</sup>. As the outer delta is mainly formed by the ebb-jet out of the inlet, a relationship with P is to be expected. The weak dependency on the wave conditions, however, is rather surprising. Apparently, the system's sand budget is almost closed and the waves only deform the body of sand;
- the protrusion of the outer delta (De Vriend et al., 1994), which turns out to be proportional to *P*<sup>0.6</sup>, at least for the Dutch and the East-Frisian Wadden Sea. Note that the effect of the wave conditions cannot be identified here, because the wave cllimate is largely the same throughout the area.

This may give the impression that any large-scale equilibrium-state property of tidal inlets can be expressed in terms of the tidal prism. Clearly, this is only true for the properties which are directly related to the amount of water which moves through the system during the tide.

A very important aspect of the morphological system, viz. the level of the intertidal flats, is mainly correlated to the mean high-water level (MHW) and the local wind/wave climate. In broad outline, shoals tend to be built up by the tide during calm weather and their top is eroded by waves during rough weather (cf. De Vriend et al., 1989).

Another important property which has a rather weak correlation with the prism is the flats' area. By definition, this quantity is equal to the difference between the basin area and the channel area, and the latter is equal to the channel volume divided by some characteristic channel depth. This leads to a relationship of the type

$$A_f = A_b - A_{ch} \approx A_b - \alpha \frac{P\sqrt{A_b}}{D_{ch}} \approx A_b - \beta \frac{H_m}{D_{ch}} A_b^{3/2}$$

2

in which  $A_r$ 

 $A_b$ 

 $A_{ch}$ 

lpha $D_{ch}$ 

β H<sub>m</sub> the flats' area, i.e. the area above MSL [m<sup>2</sup>],
the gross basin area (flats and channels) [m<sup>2</sup>],
the horizontal area covered by all channels [m<sup>2</sup>],
constant of proportionality [m<sup>-1</sup>],
characteristic channel depth [m],
constant of proportionality [m<sup>-1</sup>],
mean tidal range [m].

Renger and Partenscky (1974) suggest, on the basis of data from the inlets in the German Bight,

$$A_{f} = A_{b} - 0.025 A_{b}^{3/2}$$

which would mean that the flats' area depends exclusively on the basin area. According to Eq. (2), this can only be the case if  $D_{ch}$  is proportional to  $H_m$ . Eysink (1990) shows that if a channel-depth relationship exists, it contains a power of P (or  $H_m$ ) which is significantly less than 1. Apparently, Eq. (3) is an oversimplification freality.

These properties of the intertidal zone are not only important for the functions of the inlet system (ecosystem, mussel banks), but also for the large-scale morphodynamics. The bathymetric variation in a tidal basin gives rise to a tidal rectification which can contribute significantly to the residual transport (Van de Kreeke and Robaczewska, 1993; Bakker and De Vriend, 1995), and hence to the net import or export of sediment. Via the so-called "hypsometry effect" (Pethick, 1980; Boon and Byrne, 1981; also see Van Dongeren and De Vriend, 1994), the flats also influence the tidal asymmetry in the gorge: the smaller the flats' area relative to the basin area, the more positive the asymmetry. Thus the flats regulate the import or export of sediment and enable the system to reach an equilibrium state.

#### Example of application

The empirical relationships for the large-scale equilibrium state can be utilized to give offhand predictions of the impact of factors which affect the tidal prism, the basin area, or the MHW-level. An example of such a case is "The Friesche Zeegat", one of the inlets of the Dutch Wadden Sea (Fig. 6.5) with had following characteristics until 1969:

$$H_{mo} = 2 \text{ m}$$
;  $A_{bo} = 2.1 \ 10^8 \text{ m}^2$ ;  $P_o = 3.0 \ 10^8 \text{ m}^3$ ;  
 $A_{co} = 1.9 \ 10^4 \text{ m}^2$ ;  $V_{bo} = 3.5 \ 10^8 \text{ m}^3$ ;  $V_{\Delta o} \approx 1.8 \ 10^8 \text{ m}^3$ ;  $A_{ro} = 1.6 \ 10^8 \text{ m}^2$ 

In 1969, the Lauwerszee, with an area of 90 10<sup>6</sup> m<sup>2</sup>, was separated from the basin by a dam and reclaimed. This means that in 1969 the above values changed abruptly into

$$H_m = 2 \text{ m}$$
;  $A_b = 1.2 \ 10^8 \text{ m}^2$ ;  $P = 2.0 \ 10^8 \text{ m}^3 \sim 0.65 P_0$ ;

$$A_{c} = 1.9 \ 10^{4} \ \text{m}^{2}$$
;  $V_{b} = 2.7 \ 10^{8} \ \text{m}^{3}$ ;  $V_{a} \approx 1.8 \ 10^{8} \ \text{m}^{3}$ ;  $A_{f} \approx 0.75 \ 10^{8} \ \text{m}^{2}$ 



Figure 6.5: The Frisian inlet after closure of the Lauwerszee

Obviously, the system has responded to this major interference. According to the above relationships, it should seek a new equilibrium state in which

$$H_m = 2 \text{ m}; A_b = 1.2 \ 10^8 \text{ m}^2; P = 2.0 \ 10^8 \text{ m}^3;$$
$$A_c \approx 0.65 A_{co} = 1.2 \ 10^4 \text{ m}^2; V_b \approx (0.65)^{1.5} V_{bo} = 1.9 \ 10^8 \text{ m}^3;$$
$$V_\Delta = (0.65)^{1.23} V_{\Delta 0} \approx 1.1 \ 10^8 \text{ m}^3; A_f \approx 0.85 \ 10^8 \text{ m}^2$$

Note that the relative flats' area is expected to increase after closure, because the value of this parameter in the remaining basin is initially too small (the Lauwerszee used to be the back part of the basin, and hence contained a relatively large portion of flats). In reality, a rapid narrowing of the inlet gorge has taken place, by the formation of a large sandy hook at the east side of the inlet (Fig. 6.6). The actual amount of sediment import into the basin was initially 7  $10^6$  m<sup>3</sup>/yr, and between 1970 and 1987 it is estimated at 34  $10^6$  m<sup>3</sup> (Oost and De Haas, 1992). Clearly, the adjustment process has not yet come to an end. Although the delta evolution is rather dependent on long-term meteorological variations, its volume has decreased by 26  $10^6$  m<sup>3</sup> and its seaward protrusion has decreased roughly by 1.5 km in these 17 years (De Vriend et al., 1994). So the volume decrease is of the same order of magnitude as the sediment demand in the basin. Apparently, the delta serves a buffer for the basin. Yet, the equilibrium value of  $V_b$  is proportional to  $P^{1.5}$  and that of  $V_{\Delta}$  is proportional to  $P^{1.23}$ , which means that the basin ultimately demands more sediment than the delta can supply. This sediment has to be provided by the sea bed or the adjacent coasts.



Figure 6.6: Frisian Inlet: accretion and erosion between 1970 and 1987 (the darker the shade, the stronger the accretion)

## Transient empirical models

Transient empiricial models (O'Connor et al., 1990; Eysink, 1990; Eysink et al., 1992) are one step further along this line. Given the intital state and the equilibrium state, they describe the evolution of the discrepancy between the actual state and the equilibrium state as an exponential decay process. In mathematical terms, this is described by

$$\frac{dA}{dt} = \frac{A_e - A}{\tau}$$

4

in which A = morphological state parameter (e.g.  $A_c$ ,  $V_b$ , or  $A_t$ ), t = time,  $A_e$  = equilibrium value of A,  $\tau$  = characteristic time scale of the decay process.

If A<sub>e</sub> is a constant, the corresponding solution reads

$$A(t) = A_{e} - [A_{e} - A(0)] \exp(-\frac{t}{\tau})$$
5

If  $A_e$  exhibits a slow linear variation in time, for instance in the case of the flats' level responding to mean sea-level rise, the solution reads

$$A(t) = A_e(t) - \tau \frac{dA_e}{dt} - \left[A_e(0) - \tau \frac{dA_e}{dt} - A(0)\right] \exp(-\frac{t}{\tau})$$
6

The example in Fig. 6.7 shows that, after the effect of the initial condition has damped out  $(t \gg \tau)$ , the actual state keeps on lagging behind the equilibrium state. This should be borne in mind when trying to find empirical flat-level relationships in areas with a significant rate of sea-level rise.



Figure 6.7: Flats 'level' evolution for rising mean high-water level

Transient models of this type concern primarily macro-scale phenomena. They rest upon the implicit assumption that each system element behaves independent of the others. In other words: the element would exibit this behaviour if sufficient sediment would be

available. In a coherent system like a tidal inlet, this may be true for some elements, but presumably not for all of them. If, for instance, Eq. (4) is utilized to assess the sediment demand of a basin under the influence of sea level rise, the interaction with the outer delta and the adjacent island coasts is ignored.

# 6.6 Semi-empirical long-term models

## 6.6.1 General approach

Although a theoretical approach based on formal integration of the constituting equations (see the companion chapter) is probably the most accurate, it is a long shot. The identification of the relevant physical processes and the understanding of their nonlinear interaction requires much further research, which is going to take its time. Therefore, it seems wise to pursue a more pragmatic line of development at the same time: semi-empirical long-term modelling. This approach tries to make use of all information available, from the field via the empirical equilibrium-state relationships, and from theory via the large-scale balance equations. Wherever needed, parameterized results of more detailed models are included.

Since the empirical information is not available at a very detailed scale, it is important to think at a high level of aggregation, for instance in terms of the principal system elements (basin, gorge, outer delta, island coasts). In the following a brief outline will be given of a number of models for various of these elements.

## 6.6.2 Stability of the gorge

Wide-open inlets like those in the Wadden Sea are not likely to have a strong reducing effect on the tide inside the basin. The tidal amplitude there is closely similar to that at open sea, apart from possible shallow-water effects. If the inlet is narrower, however, it will hamper the tide to enter the basin, with a reduction of the amplitude and the tidal prism as a consequence. But if the tidal prism is reduced, the cross-sectional area of the inlet will also be reduced, which leads to a further reduction of the prism, etc. This phenomenon of positive feedback, which ends up in the closure of the inlet, is usually called inlet instability. Escoffier (1940) was the first to point out and model this instability phenomenon. He compared the maximum tidal velocity through the inlet,  $V_{max}$ , with the critical velocity for sediment transport in the inlet channel,  $V_{cr}$ ,  $V_{max}$  is a function of the cross-sectional area of the channel, A<sub>c</sub>. If A<sub>c</sub> is small, this function will be increasing, because the growth of the tidal prism will be predominant. If  $A_c$  is large, the function will be decreasing, because the the tidal prism has become independent of A<sub>c</sub> and the same amount of flow through a wider channel leads to smaller velocities. Fig. 6.8 outlines the curves of  $V_{max}$  and  $V_{cr}$ against  $A_c$ . If  $V_{max} = V_{cr}$ , which is the case in Fig. 6.8 for two values of  $A_{cr}$  the system is in equilibrium. The stability of the system in these "stationary points" can be estimated via

$$\Delta(V_{\max} - V_{cr}) = \frac{dV_{\max}}{dA_c} \Delta(A_c - A_{cs})$$
7

In the stationary point *SP1* the derivative of  $V_{max}$  with respect to  $A_c$  is positive, so  $V_{max} - V_{cr}$  increases if  $A_c - A_{cs}$  increases. On the other hand,  $A_c$  tends to increase with  $V_{max}$ , which means that around this stationary point  $V_{max} - V_{cr}$  and  $A_c - A_{cs}$  are positively coupled and the system is unstable. Conversely, the system is stable around the other stationary point, *SP2*. All inlets of the Wadden Sea are within this stable domain.

Van de Kreeke (1990 a,b) uses a similar technique to analyse the stability of a multipleinlet system and arrives at the conclusion that a two-inlet system cannot be stable. He suggests that the same conclusion holds for a multiple-inlet system.



Figure 6.8: Stability diagram for tidal inlets

6.6.3 Di Silvio's basin models

Di Silvio (1989) proposes a simple box model for the basin, consisting of one deep part which represents all channels and one shallow part which represents all flats (Fig. 6.9). The total cross-sectional area of the channels and the level of the flats are the dependent variables. The model works with a long-term representative sediment concentration in each part of the model (sea, channel, flat). This concentration is assumed to depend only on local parameters, such as the water depth, but not on the tide. The concentration at sea is assumed to be independent of the state of the basin, so it is a given constant.



Figure 6.9: Principle of Di Silvio's (1989) box model

The sediment fluxes between the model parts are assumed to be purely advective, which means that they can be written as

$$S_1 = Q_{ch,ebb} C_{ch}$$
;  $S_2 = Q_{ch,flood} C_{sec}$ 

8

$$S_3 = Q_{fl,ebb} C_{fl}$$
;  $S_4 = Q_{fl,flood} C_{ch}$ 

in which  $C_{ch}C_{fl}$  = representative sediment concentration in the channel and the flats part, respectively,

- C<sub>sea</sub> = representative concentration in the sea part of the model,
- $Q_{ch,ebb}$  = total water mass which flows from the channel part to the sea during ebb,

# $Q_{fl,ebb}$ = total water mass flowing from the flats part to the channel part during ebb, etc.

Clearly, the conservation of water mass requires

$$Q_{ch,ebb} = Q_{ch,flood}$$
;  $Q_{fl,ebb} = Q_{fl,flood}$  9

The sediment balance equations for the channel and the flat parts of the model read

$$B_{ch} \frac{dA_{ch}}{dt} = S_1 - S_2 - S_3 + S_4$$
  
=  $Q_{ch,ebb} C_{ch} - Q_{ch,flood} C_{sea} - Q_{fl,ebb} C_{fl} + Q_{fl,flood} C_{ch}$  10

$$B_{fl} \frac{dz_{fl}}{dt} = S_{4} - S_{3} = Q_{fl,flood} C_{ch} - Q_{fl,ebb} C_{fl}$$
 11

in which  $A_{ch}(t)$  denotes the total cross-sectional area of the channels and  $z_{fl}(t)$  the flats' level. The constants *B* contain geometrical properties (channel length, flats' area) and the sediment content per unit volume of bed. Combination with Eq. (9) shows that the system is in equilibrium if  $C_{fl} = C_{ch} = C_{sea}$ .

The sediment dynamics of the system are reflected by a relationship between the concentration and the dependent variable in the relevant model part. In a linear approximation in  $A_{ch}$  and  $z_{fl}$ , Eqs. (10) and (11) can be elaborated to morphological evolution equations of the form

$$\frac{dA_{ch}}{dt} + X_{ch} A_{ch} = Y_{ch} Z_{fl} + Z_{ch}$$
<sup>12</sup>

$$\frac{dz_{fl}}{dt} + X_{fl} z_{fl} = Y_{fl} A_{ch}$$
<sup>13</sup>

in wich X, Y and Z are constants. These coupled linear decay equations form a secondorder system, as can be shown by reducing them to

$$\frac{d^2 A_{ch}}{dt^2} + (X_{ch} + X_{fl}) \frac{dA_{ch}}{dt} + (X_{fl} X_{ch} - Y_{fl} Y_{ch}) A_{ch} = 0$$
14

If  $X_{ch}$ ,  $X_{f}$  and  $Y_{f}Y_{ch}$  are positive, the solution of this equation is of the type

$$A_{ch}(t) = A_{ch,e} + P_1 \exp(-\lambda_1 t) + P_2 \exp(-\lambda_2 t)$$
**15**

in which  $A_{ch,e}$  is the equilibrium value of  $A_{ch}$ ,  $P_1$  and  $P_2$  are constants which follow from the initial conditions, and  $\lambda_1$  and  $\lambda_2$  are the (real and positive) roots of the characteristic equation of Eq. (14).

Apparently, the dynamic interaction between the flats and the channel parts give rise to a behaviour which differs from the exponential decay which would be found if the model parts were mutually independent (cf. Section 5).

The above model was used with reasonable success to hindcast the morphological evolution of the Venice Lagoon over the last 180 years (Di Silvio, 1991). The same author describes a 2-D finite-difference version of the concept, in which each grid cell represents a channel or a flat element. The sediment transport in this model is described with an

advection/diffusion model, still in terms of the long-term representative concentration *C*. In either of these models, the constituting equations are reduced to a set of tide-integrated morphological evolution equations which can be solved at the morphological time scale. Thus the same line of thought is followed as in the formally averaged models which are described in the companion chapter.



Figure 6.10: Hindcast of the morphological evolution of the Chioggia basin (Venice Lagoon) between 1810 and 1990 with Di Silvio's (1989) box model,

a. flats' level, b. cross-sectional area of the channels (from Di Silvio, 1991; courtesy author)

6.6.4 Van Dongeren's basin model

Van Dongeren and De Vriend (1994) suggest to divide a channel/shoal system into a chain of consecutive boxes, aligned according to their distance from the gorge. In each box there is a channel part and a flats part (Fig. 6.10). The dependent variables in each box are the channel cross-sectional area,  $A_{ch}$ , the flats' area,  $A_{fl}$ , and the flats' level,  $z_{fl}$ . Empirical relationships are used to determine the equilibrium values of these variables. The sediment balance for the channel part of each box involves the exchange with the neighbouring channel sections and the adjacent flats. The sediment balance of the flats part involves only the exchange with the adjacent channel part (there is no sediment excange between flats parts).



Figure 6.11: Principle of Van Dongeren's basin model

In contrast to the models in Section 6.3, the constituting equations are not reduced to one or more morphological evolution equations. In fact, this element of the model is replaced by introducing empirical decay equations for  $A_{ch}$  and  $A_{fh}$ , which apply if there is enough sediment supply to (or export from) the relevant part of the box:

$$\frac{dA_{ch}}{dt} = \frac{A_{ch,e} - A_{ch}}{\tau_{ch}} \quad and \quad \frac{dA_{ff}}{dt} = \frac{A_{ff,e} - A_{ff}}{\tau_{ff}}$$
16

where  $A_{ch,e}$  and  $A_{fl,e}$  denote equilibrium values and  $\tau_{ch}$  and  $\tau_{fl}$  are characteristic time scales.

The former two quantities follow from empirical relationships, the latter two are input parameters, to be estimated from measured data or otherwise. If there is not enough sediment supply or export, the available supply/export used for morphological changes, with a fixed sequence of priority (see Fig. 6.12).



Figure 6.12: Flow chart of Van Dongeren's basin model

The boxes are handled consecutively, in a sequence which corresponds with the direction of propagation of the morphological information ("sand wave"). Sometimes this sequence is rather obvious, like in the case of the Friesche Zeegat (see Section 5), but in other cases this is much less clear (cf. Goldenbogen et al., 1994).

The system is essentially driven by the hypsometry effect (see Section 5): if the flats' area is too small, the tidal asymmetry in the gorge is positive and sediment is imported. Conversely, sediment is exported if the flats' area is too large. The dynamic response of the basin actually boils down to a tendency of the channel cross-section to adjust rapidly to the hydrodynamic conditions, if necessary at the expense of the flats. In the longer run, when the sand (or sand-deficit) wave from the gorge has reached the relevant section, part of the incoming sediment is used to restore the flats.

The spatial resolution along the channel provides the possibility to model bifurcating systems, and to reproduce something of the wave-type behaviour of the morphological changes. Yet, the concept is essentially large-scale, and the model results should be treated that way.

This model has been applied with reasonable success to the Friesche Zeegat (Van Dongeren and De Vriend, 1994), but in other cases it yields somewhat poorer results,

especially for the evolution of the flats' level (Goldenbogen et al., 1994). Clearly, the interaction of flats and channels needs further investigation. Since the concept is limited to basins which are small compared with the tidal wave length (the water surface elevation has to be in phase throughout the basin), the model in its present form does not work in larger basins. Bakker and De Vriend (1995) describe an alternative network model for such basins.



Figure 6.13: Hindcast of channel evolution in the Frisian Inlet with Van Dongeren's (1994) basin model, at (a) 3km, (b) 7 km, © 9 km from the gorge. (From: De Vriend et al., 1994; courtesy Elsevier)

6.6.5 Karssen's basin model

Karssen (1994a, 1994b) takes a somewhat similar approach as Di Silvio in a 1-D network model concept for estuaries and tidal basins. The basis is a 1-D network model for unsteady flow, in which the cross-section of each channel consists of a deep channel part, a low-lying flats part and a high flats part. The results of a tidal computation with this network model are used to determine a number of equilibrium-state parameters via empirical relationships.

The sediment is assumed to be transported primarily as suspended load, which means that a representative sediment concentration in each of the three parts of the channel cross-section can be used as a dependent variable. The transport mechanisms in the channel are advection and longitudinal diffusion. The transport on the flat sections is diffusive in the cross-channel direction, i.e. there is no longitudinal transport over the flats. Upon integration of these diffusive fluxes in the cross-channel direction, the net lateral sediment flux between the various parts of a cross-section becomes proportional to the difference in concentration:

$$F_{lc} = D_l h_l \frac{C_l - C_c}{L_{lc}} \quad and \quad F_{hl} = D_h \frac{h_h}{4} \frac{C_h - C_l}{L_{hl}}$$
 17

in which

 $F_{lc}$ 

net sediment flux from the low-lying flats to the channel
 net sediment flux from the high to the low-lying flats

 $F_{hl}$  = net sediment flux from the high to the low-lying flats  $D_{h}D_{h}$  = lateral diffusion coefficients for the low-lying and high flats,

 $D_h D_h$  = lateral diffusion coefficients for the low-lying and high flats, respectively,

 $h_{b}h_{b}$  = effective water depth above these flats,

 $C_{\mu}C_{\mu}$  = representative sediment concentration above these flats,

C<sub>c</sub> = representative sediment concentration in the channel,

*L<sub>lc</sub>* = the distance over which the cross-channel integration takes place, so a typical distance from the low-lying flats to the channel, and

 $L_{hl}$  = the same from the high to the low-lying flats.

Thus the concentration evolution equations for the high and the low-lying flats become

$$\frac{\partial (A_h C_h)}{\partial \tau} = W_h w_s (C_{he} - C_h) - F_{hl}$$
<sup>18</sup>

$$\frac{\partial(A_lC_l)}{\partial \tau} = W_l w_s (C_{le} - C_l) - F_{lc} + F_{hl}$$
<sup>19</sup>

in which  $A_{\mu}A_{h}$  = cross-sectional area on the high and the low-lying flats part, respectively,  $W_{\nu}W_{\nu}$  = settling coefficients on these flats.

 $\begin{array}{ll} \mathcal{W}_{\mu}\mathcal{W}_{h} &= \text{settling coefficients on these flats,} \\ \mathcal{C}_{he}, \mathcal{C}_{le} &= \text{equilibrium concentration on these flats,} \\ \tau &= \text{"slow" time co-ordinate, which varies at the morphological time scale,} \\ \mathcal{W}_{s} &= \text{sediment settling velocity.} \end{array}$ 

The corresponding sediment balance equation describes the evolution of the crosssectional areas of the three channel parts in terms of the net sediment flux from the bed:

$$\frac{\partial A_i}{\partial \tau} = W_i W_s (C_{ie} - C_i)$$
<sup>20</sup>

in which the suffix i (i=c,l,h) refers to each of the parts.

Unlike Di Silvio's (1989) model, these sediment balance equations are not elaborated to morphological evolution equations by expressing  $C_i$  in terms of  $A_j$  (*j*=*c*,*l*,*h*). Instead, Eqs. (20) are solved in a simulation loop, together with the concentration equations and the empirical equilibrium-state parameters computed from the tidal model results. After the topography has changed enough to influence the tide, another hydrodynamic model run is made. To that end, the cross-sectional areas which result from the morphological computation are translated into a 7-point cross-sectional shape per channel section. A critical point is this model is the definition of the nodal point relationships for the transport (concentration). Depending on how these are chosen, a bifurcated channel



system can persist or not (Wang et al., 1993). If the transport is distributed according to the flow rates through the channel sections connected to the node, for instance, part of the channels will fill up.

A first application of this model to the "Friesche Zeegat" case (Section 5) was rather succesful in hindcasting the evolution of the channels' crosssectional area throughout the basin (Karssen, 1994b). Whether the model performs equally well for the tidal flats remains to be investigated.


Figure 6.14: Hindcast of channel evolution in the Frisian Inlet with Karsten's (1994) basin model, at (a) 2.5 km, (b) 7 km, (c) 8 km from

the gorge. (From: Karsten, 1994; courtesy DELFT hYDRAULICS)

## 6.6.6 De Vriend et al.'s delta model

The morphological evolution of the basin has its effects on the outer delta. In the "Friesche Zeegat" case (Section 5) the delta volume decreased with about the same amount as was deposited in the basin. Although there is no evidence that it was the sediment from the delta which was deposited in the basin, the amounts of sediment involved are so large (appr. 30 10<sup>6</sup> m<sup>3</sup>) that this must have been the case to a large extent.

The behaviour of the outer delta is determined by tide and sea waves. The tide generally tends to build out the delta seawards, the waves tend to squeeze it shorewards (bulldozereffect). The evolution of the outer deltas in front of the Delta-works, in the southwest of The Netherlands, is a spectacular illustration of what happens when one of these agents is eliminated. After closure of the inlets, the deltas were pushed up against the shore and formed large systems of shoals, which seem to be the onset of barrier islands (cf. Stive, 1986; De Vriend and Ribberink, 1988; Steijn et al., 1989). Also the closure of the Lauwerszee led to a spectacular morphological response of the delta of the Friesche Zeegat, viz. the formation of a major sandy hook at the tip of the isle of Schiermonnikoog (see Fig. 6.6).

These observations illustrate that there are two very powerful, but largely counteracting agents at work on the outer delta. Any morphodynamic model of the outer delta should therefore take each of these agents into account. A large-scale concept which seems to contain them both is a two-line model in which there is a forced transfer of sediment from the beach line to the inshore line (Fig. 6.12). This forced cross-shore transport represents the ebb-dominated tidal transport on the outer delta. The "normal" longshore and cross-shore transport components represent the wave-induced effects. The beach and inshore lines respond by taking such an orientation that the wave-induced transport is exactly big enough to feed the forced cross-shore transport at the shoreward end, and to remove the sediment which it delivers at the seaward end. If the forced cross-shore transport is cut off, the system will return to a straight coast.



De Vriend et al. (1994) describe a model of the outer delta and the adjacent coasts which is based on a refined version of this concept (Fig. 6.13). It includes not only the main ebb channel, but also the flood channels near the island tips. The equilibrium state of the delta (volume, seaward protrusion) is described using empirical relationships. The sediment balance is maintained throughout the evolution, and the sediment demand or supply by the basin is taken into account as an input parameter. The model output consists of the cross-shore positions of the beach line and the inshore line as a function of time and the longshore co-ordinate, but these results should be interpreted at a higher aggregation level (for instance, in terms of the total sand volume changes in a certain coastal section).



Figure 6.16: Principle of De Vriend et al. 's (1994) outer delta model

The model was validated against measured data from the Friesche Zeegat (cf. Section 5), of which the delta evolution after closure could be reproduced reasonably well. Further analyses of the results made clear that the delta acts as a mid-term sediment buffer for the basin, but that in the long run the sediment surplus or deficit of the delta has to be compensated by the adjacent island coasts. This indicates that sand mining from the delta can have detrimental long-term effects on the island coasts.



Figure 6.17: Hindcast of the outer delta protusion of the Frisian Inlet with De Vriend et al.'s (1994) model

6.6.7 Steetzel's model of the entire Wadden Sea coast

Steetzel (1994) uses a similar model to hindcast and predict the large-scale evolution of the entire 150 km long multi-inlet coast of the Dutch Wadden Sea, under the influence of accelerated sea level rise, climatic change and subsidence due to oil and gas mining. The response of the basin volume to the relative sea level rise is assumed to be described by an exponential decay model as given by Eq. (4). The resulting time-evolution of the sediment demand from the basin is determined for each of the inlets. With these time-functions as input conditions at the basin end, the model computes the positions of the island coastlines and the edge of the outer delta.

The basic concept of Steetzel's model is the same as described by De Vriend et al. (194), but it is based on 5 lines instead of 2, in order to be able to deal with the variety of transport mechanisms in the various depth zones. The layers concern the dune front, the subaerial beach, the surf zone, the middle shoreface and the lower shoreface. They cover the zone between O.D. + 7 m and O.D. - 20 m.

Information from detailed process-based models of the Eierlandse Gat and the Friesche Zeegat is used to set some of the model parameters. The model is validated against 25 years of data from the JARKUS data set (yearly profiles of 800 m length, taken at alongshore distances of 250 m).

model boundary	sediment flux into the model domain
Callantsoog (SW end of the model)	+ 1.1 10 <sup>6</sup> m <sup>3</sup> /yr alongshore
Schiermonnikoog (NE end of the model)	- 2.1 10 <sup>6</sup> m³/yr alongshore
O.D 20 m (seaward model boundary)	+ 0.8 10 <sup>6</sup> m <sup>3</sup> /yr cross-shore
O.D. + 7 m (dune top = landward model boundary)	- 0.3 10 <sup>6</sup> m³/yr cross-shore
inlet gorges (demand from basins)	- 12.7 10 <sup>6</sup> m <sup>3</sup> /yr cross-shore
nourishments (sand mined offshore)	+ 2 to 3 10 <sup>6</sup> m <sup>3</sup> /yr source

Table 5.1: Overall sediment balance for the Dutch Wadden Sea coast [from: Steetzel, 1994; courtesy DELFT HYDRAULICS]

Table 5.1 summarizes the overall sediment balance for the model area, which extends from Callantsoog at the SW end to the easternmost tip of Schiermonnikoog at the NE end. Clearly, the demand from the basins is of predominant importance to the evolution of this coast.

A further analysis of JARKUS and other bathymetric survey data has revealed that 75% of the net sediment loss in the area goes at the expense of the outer deltas, and that 50 to 60% takes place at the middle and lower shoreface. Apparently, the erosion is concentrated at the delta edges.

These findings are confirmed by the model. Fig. 6.14 shows a typical result: the predicted accretion/erosion per coastal section over the next 50 years. The spatial resolution of the presented results is much smaller than that of numerical model, which computes the cross-shore position of each line as a function of time and longshore position. This aggregation of output is in line with the strong geometrical schematization in the model.





### 6.6.8 Compound models

The semi-empirical models which were discussed so far concern either the basin or the coast (including the outer delta). In both cases, the long-term residual transport through the gorge is prescribed, either as a function of the basin state (basin model), or as a given quantity (delta model). A logical next step would be to couple basin, delta and island coasts, to yield a model for the entire inlet system.

This job is less trivial than it may seem at first sight. Since the transport through the gorge is modelled differently in either case, it is not just a matter of patching. In fact, the processes in the gorge which link the basin to the outer coast should be reanalysed, taking the influence of all elements of the system into account (for instance, the combined effect of the outer delta and the intertidal flats on the residual transport through the gorge). Process-based simulation models can be of help to such an analysis.

One possible way to couple the basin and the outer coast is to extend the basin model through the gorge and let it interact with the outer coast model. Bakker and De Vriend (1995) suggest to use a network model for the basin morphology, and to let one or more branches of this network stick through the inlet into the open sea. It must be possible to let the sediment which is transported through these braches be picked up or deposited near the delta edge, thus feeding a multi-line coastal model like the one described in the foregoing. If these channels are allowed to migrate, die out and regenerate, for example under the influence of the longshore drift, it may even be possible to achieve an essentially higher resolution than in the present models of the outer coast.

At this moment, these is little more than ideas which need to be substantiated, implemented and tested against reality. Detailed process-based simulation models and frequent (remote-sensingl) topographic surveys can be of great help here.

## 6.6.9 Conclusion

Empirical and semi-empirical models have been shown to be of great value to tidal inlet modelling, although (or maybe because) there is still a long way to go to their understanding in terms of elementary physical processes. The empirical knowledge of inlet systems is based on data from many inlets all over the world. This, together with the vast scientific effort stored in these models, makes that anomalies have probably been removed from most of the empirical equilibrium-state relationships. Thus these relationships provide valuable knowledge which should not be ignored, but used to bridge the present gap between process-based simulation and long-term prediction of tidal inlet behaviour. The semi-empirical model approach tries to combine the empirical knowledge with basic physical principles, such as the conservation of sediment and water mass. Thus it can be a powerful tool to predict the long-term dynamic behaviour of tidal inlet systems, as has been shown in various instances. The further development of this approach, together with process-based simulation models, is certainly worth pursuing.

An important research issue in this area is the theoretical substantiation of the empirical equilibrium-state relationships, starting from first physical principles. Furthermore, the semi-empirical model concepts need further improvement and validation. This requires consistent bathymetric records over a sufficiently long time (relative to the morphological time scale of interest). Longlasting monitoring programmes, as well as historical and sedimentological reconstructions are therefore absolute necessities for further progress. Moreover, better use should be made of process-based simulation models and formally averaged models to improve and refine semi-empirical models.

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# 7. Network Modelling

# 7.1 Introduction

Mathematical modelling of morphological development is applied for evaluating impacts of natural developments e.g. sea level rise and human interference e.g. dredging / dumping, land reclamation, etc. Examples of practise problems in the Netherlands are the deepening of the navigation channels in the Western Scheldt and opening of the Haringvliet sluices.

Two types of network models will be dealt with, i.e. process based models and semiempirical models. By process based models we mean morphodynamic models based on mathematical descriptions of the relevant processes, i.e. flow, sediment transport, and bed level change, using physical laws. Semi-empirical models are models based on combination empirical relations and description of physical relations.

This chapter consists of two parts. In the first part, section 7.2 and section 7.3 discussing the two types of models respectively, contains the obligatory material of the lecture. The second part, section 7.4 through 7.6, contains three recently published papers containing additional information on the subjects considered in the first part.

# 7.2 Process based models

One-dimensional network morphodynamic models for tidal basins can be considered as an extension of 1D morphological models for rivers, which are discussed in the lecture ctwa3340 and ctwa5311. Therefore the basic theory on 1D morphological modelling will not be discussed here. Only some specific aspects for tidal regions are discussed. The lecture mainly deals with application rather than design of models (software development).

## 7.2.1 A comparison to river case

First a comparison with the non-tidal river case is made in order to highlight the special aspects of morphological modelling in tidal regions.

## Morphological equilibrium

For the river case morphological equilibrium state is well defined. If the upstream discharge, the upstream sediment transport and the river width is given the equilibrium water depth and the equilibrium bed slope can be determined. Human interference usually implies a new morphological equilibrium at the long-term. The influence of the various model parameters on the morphological equilibrium is well known. This makes morphological modelling relatively easier. The computation starts at a known (initial) state and end at the long term to a known state if the hydrodynamic condition remains the same.

In estuaries and tidal inlets the driving forces for morphological development is much more complicated than in the case of (non-tidal) rivers. In addition to the upstream discharge the other relevant components of the driving forces are tidal motion, wind, (short) waves, density flow, etc. We restrict here to the simplest situation, i.e. only tidal motion is considered. For most tidal basins in the Netherlands tidal motion is the dominating driving force for sediment transport and morphological development. Even with this simplification the morphological equilibrium cannot be derived from the equations describing the water motion and sediment transport. The relation between the morphological equilibrium and the driving forces, i.e. the upstream discharge and the tide in the open sea cannot be made explicit. It is not even clear whether or not a morphological equilibrium theoretically exists. This makes the morphological modelling much more complicated: the computation starts at a known state but goes to an unknown end.

## Inter-tidal area

Inter-tidal area, or the tidal flat, is the area, which is dry at low water and drowned at high water. It may be compared to the floodplains of rivers, which are only drowned during flood waves. However, the inter-tidal area plays a much more important role for the morphology in estuaries and tidal inlets than the floodplain for rivers. Take for example a short (relatively to the length of tidal wave) tidal inlet. The morphology of the inter-tidal area determines the amount of water flowing in and out during flood and ebb, i.e., the discharge and the tidal volume. This means that the morphology of the inter-tidal area itself influences the driving force of the morphological development. This complicated interaction is one of the reasons why morphological equilibrium in tidal regions is difficult to be determined. It has also the consequence that the impact of human interference in the inter-tidal area is usually not restricted locally. Consider e.g. land reclamation, its impact can immediately (speaking on the morphological time scale, the exact propagation velocity is that of the tidal wave) be felt at the entrance of the tidal basin.

## Morphological tide

Morphological modelling for rivers makes often use of the quasisteadiness assumption. An extension of this assumption for regions is the quasitidal periodicity assumption. The basic idea is that the time scale of morphological development is much larger than the tidal period. The morphological time step can be much larger than the tidal period. This makes it possible to adopt a computational procedure as shown in Fig. 7-1



### Fig. 7-1: Computational procedrure

At each morphological time step the tidal flow and the sediment transport during a whole tidal period are calculated. Then the net sediment transport is used to calculate the morphological change.

A simplification usually made is replacing the ever-changing tide (e.g. due to spring-neap variation) by a single periodic tide. This single periodic tide is called the morphological tide. This is often defined as the tide representing the averaged net sediment transport, which is usually between the averaged tide and the spring tide. However, if this definition is followed strictly, the morphological tide is usually different at each location of the model area. This is the reason that another definition of the morphological tide has been introduced recently. The tide, for which the relative spatial variation of the ratio between the net transport due to this tide and the net transport in reality (e.g. due to the spring-neap tidal cycle) is minimal, is chosen as the morphological tide. A constant correction coefficient is introduced to reproduce the realistic averaged transport.

At present the consequence of the application of morphological tide on the computed morphological development is not clear. A comparable concept for rivers is the dominant discharge, which is used, in the early days of morphological modelling for rivers. However, it is now known that a single discharge can never represent all the relevant morphological features of a river.

### Residual sediment transport

The net transport or residual transport determines the morphological development in tidal regions. It is a small difference between the large ebb transport and flood transport. This makes the accurate calculation of the residual transport very difficult.

Tidal flow velocity can be written into the sum of a residual flow velocity and many harmonic components.

$$u(t) = u_0 + u_1 \sin(\varpi_1 t) + u_2 \sin(\varpi_2 t) + \dots$$
(7.1)

Using a simple power-law as sediment transport formula one can show that the residual sediment transport is not determined by the main tidal component  $u_1$  only, but it is mainly determined by the small residual flow  $u_0$ , and the small higher frequency component (tidal asymmetry). This put a special requirement to the flow module in the morphological model. During the set up and the calibration of the flow module this should already be kept in mind. Morphological modelling is not simply post-processing of flow modelling.

It is noted that during the calibration of flow models a calibration parameter like the Chezy coefficient is some times used as a "garbage bascket". Be aware that this may have consequences for the sediment transport modelling. As an example, for reproducing the tidal asymmetry and the residual circulation in 1D network tidal flow model it may be necessary to use direction-dependent roughness coefficient, with a distinction between the ebb channels and flood channels. However, if the so resulted Chezy coefficient is also used in the sediment transport formula it may produce unrealistic sediment transport pattern.

## 7.2.2 Nodal point relations and stability of bifurcation's

In a network model for a tidal area each nodal point represents a confluence as well as a bifurcation since the flow during ebb is in the opposite direction as during flood. For the case of bifurcation a user defined nodal point relation is required for the sediment transport. This nodal point relation has important influence on the stability of the bifurcation and thereby it determines to a large extent the behaviour of the morphodynamic model.

For the river case this subject is extensively discussed in the paper enclosed in 7.4. In the practice it has been proven that all the conclusions drawn for the river case also apply for the estuarine tidal flow case.

Here the stability problem is illustrated with a simple example of river case: a river bifurcates into two identical downstream branches with half the width of the opstream branch. Take for the upstream branch the following properties:

Width	B <sub>0</sub> =	100 m
Discharge	Q <sub>0</sub> =	400 m³/s
Sediment transport	S <sub>0</sub> =	0.04 m³/s
Chezy coefficient	C =	50 m <sup>0.5</sup> /s

Suppose that the following sediment transport formula applies:

$$S = BMu^5$$

(7.2)

Herein

B =	River width
M =	Constant coefficient (= 0.0004 here)
u =	Flow velocity

It can then be shown that the equilibrium depth is  $a_0 = 4$  m and the equilibrium slope is  $i_0 = 0.0001$ .

The two downstream branches have the same length and width ( $B_1 = B_2 = 0.5 B_0$ ). Further it is assumed that the Chezy coefficients in the two branches are the same as in the upstream branch. It is then obvious that the equilibrium state with the two branches open is

$a_1 = a_2 = a_0 =$	4 m
$i_1 = i_2 = i_0 =$	0.0001
$Q_1 = Q_2 = 0.5Q_0 =$	200 m³/s
$S_1 = S_2 = 0.5S_0 =$	0.02 m³/s

To examine whether this equilibrium state is stable, consider a small disturbance of 0.1 m:

a <sub>1</sub> =	4.1 m
a <sub>2</sub> =	3.9 m

The discharge distribution among the two branches is governed by the following equations:

$$Q_1 = B_1 C a_1^{\frac{3}{2}} i_1^{\frac{1}{2}}$$
(7.3)

$$Q_2 = B_2 C a_2^{\frac{3}{2}} i_2^{\frac{1}{2}}$$
(7.4)

 $Q_1 + Q_2 = Q_0 \tag{7.5}$ 

$$i_1 = i_2$$
 (7.6)

Solving these equations yields:

Q <sub>1</sub> =	207.5 m³/s
Q <sub>2</sub> =	192.5 m³/s
i <sub>1</sub> = i <sub>2</sub> =	0.00009955

Suppose the following nodal point relation is applied:

$$\frac{S_1}{S_2} = \frac{Q_1}{Q_2}$$
(7.7)

It follows then that at the initial state of the disturbed situation:

S <sub>1</sub> =	0.02075 m³/s
S <sub>2</sub> =	0.01925 m <sup>3</sup> /s

However, the flow velocities and the sediment transport capacities of the two branches are:

u <sub>1</sub> =	1.0122 m/s	S <sub>1e</sub> =	0.02125 m <sup>3</sup> /s
u <sub>2</sub> =	0.9872 m/s	S <sub>2e</sub> =	0.01875 m <sup>3</sup> /s

Thus, the disturbance causes overloading in branch 2 ( $S_2 > S_{2e}$ ) and underloading in branch 1 ( $S_1 < S_{1e}$ ). This means that branch 2, which is shallower, will become even shallower and that the deeper branch 1 will become even deeper. This indicates that the equilibrium state ( $a_1=a_2=4$  m) is unstable.

One can do the same exercise, but now with the following nodal point relation instead of (5.6):

$$\frac{S_1}{S_2} = \left(\frac{Q_1}{Q_2}\right)^k \tag{7.8}$$

And try different values of k. By doing this one will find out that as long as k<5/3 the equilibrium is unstable, and the equilibrium becomes stable when k>5/3.

Mathematical analysis of the stability for the general river bifurcation case is presented in 7.4. It is important to notice that there is not one, but at least three physically realistic equilibrium states, representing the following situations:

- Both branches are open; each branch transports a part of the water and sediment.
- Branch 2 is closed and all the water and sediment is transported through branch 1.
- Branch 2 is closed and all the water and sediment is transported through branch 1.

It is also important to know that the nodal point relation

$$\frac{S_1}{S_2} = \left(\frac{Q_1}{Q_2}\right)^k \left(\frac{B_1}{B_2}\right)^{1-k} \tag{7.9}$$

For small values of k the equilibrium with both branches open is unstable and for large values of k it is stable. The critical value of k depends on the sediment transport formula. If the sediment transport is proportional to the nth power of the flow velocity then the critical value of k is n/3.

In the analysis described in 7.4 the simple mass-balance equation for sediment are used for

describing the morphodynamic development of the branches:

$$\frac{da_1}{dt} = \frac{S_{1e} - S_1}{B_1 L_1}$$
(7.10)

$$\frac{da_2}{dt} = \frac{S_{2e} - S_2}{B_2 L_2} \tag{7.11}$$

As demonstrated in the example above, it is possible to calculate the sediment transport and the sediment transport capacities for a given combination of the water depths in the two branches. It is thus relatively easy to construct a table like the following using a simple computer program.

aı	a <sub>2</sub>	da <sub>1</sub> /dt	0a <sub>2</sub> /01

With some graphical presentation software it is then easy to draw a picture as shown in Fig.



7-2 The cross points of the isolines  $da_1/dt=0$  and  $da_2/dt=0$ represent the equilibrium states and the vectors ( $da_1/dt$ ,  $da_2/dt$ ) show then the stability of the equilibrium states. Such a figure gives the same information as the phase diagram presented in 7.4.

This technique can easily be used for more complicated sediment transport formulas, like that of Meyer-Peter-Müller as used in Fig. 7-2 By including an extra source and/or sink term in equations (7.9) and (7.10), the technique can also be used to investigate the influence of dredging and/or dumping on an originally stable bifurcation, as shown in Fig. 7-3 and. 7-4

Fig. 7-2.







Fig. 7-3: Equillibrium states and their stability, Symmetric case under influence of dredging (10%)



Fig. 7-4: Equillibrium states and their stability, Symmetric case under influence of dredging (30%)

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# 7.3 Semi-empirical models

## 7.3.1 Principles

As mentioned in the previous section, one of the difficulties in morphodynamic modelling for tidal regions is that it is usually impossible to derive a morphological equilibrium for given hydrodynamic conditions from the equations in the model. On the other hand data from the practise show that morphological equilibrium in tidal region can be defined and the equilibrium state is well related to the hydrodynamic parameters. A well-known example of such empirical relations is the linear relation between the cross-sectional area of a tidal channel and the tidal volume through the channel. The semi-empirical models combines such empirical relations describing morphological equilibrium and process descriptions. The basic principle of such models can be summarised as follows

- a) At each time step, a morphological equilibrium state exists if the hydrodynamic condition is given.
- b) The net sediment transport can be described by advection-dispersion equation based on residual flow field. A consequence is then that there exists an overall equilibrium sediment concentration that is present in the whole model area when the whole system is in equilibrium.
- c) When an area is out of equilibrium, a tendency of sedimentation or erosion exists. This is expressed in the locally adjusted equilibrium concentration.
- d) The exchange between the bottom and water, thus also the morphological change, is determined by the difference between the local concentration en de local equilibrium concentration.

In the following the theory is explained for a network system, and illustrated with simple examples. In section 7.5 and 7.6 more extensive description of this type of models are given.

### 7.3.2 Basic equations

Consider a network consisting of morphological elements connected to each other. For each element a variable describing the morphological state, e.g. the wet volume *V*, is defined. Further per element a sediment concentration c, and an equilibrium concentration  $c_{e}$ . The following equations applies per element:

Morphological equilibrium

$$V_{e} = f(hydrodynamic \ parameters)$$
(7.12)

Herein: V<sub>e</sub> = equilibrium volume.

Equilibrium concentration

$$c_e = c_E \left(\frac{V_e}{V}\right)^n \tag{7.13}$$

Herein:  $c_E$  = overall equilibrium concentration,

n = constant.

Mass-balance in the water phase

$$\sum_{i} T_{i} = w_{s} A(c_{e} - c)$$
(7.14)

Herein: T<sub>i</sub> = outgoing transports along the boundaries of the element,

A = horizontal area of the element,

Morphological change

$$\frac{dV}{dt} = w_s A(c_e - c) - I + A \frac{d\zeta}{dt}$$
(7.15)

Herein I = dredging and dumping (dumping = positive);

 $\zeta$  = sea level.

The transport in equation (7.13) is defined per connection between two elements. This consists of a diffusive part and an advective part (due to residual flow):

$$T = \delta(c_1 - c_2) + Q(c_1 + c_2)/2$$
(7.16)

Herein T = transport from element 1 to element 2,

- δ = horizontal exchanging coefficient between the two elements due to dispersion,
- Q = discharge of residual flow from element 1 to element 2.

At an open boundary the transport is determined in the same way. The concentration outside the model area should be given as boundary condition and it is usually assumed to be equal to the overall equilibrium concentration. At a closed boundary the transport equals zero.

### 7.3.3 Computational procedure

The computational procedure in the model is as shown in Fig.7-5 At each time step the equilibrium volumes of the element are determined first from the hydrodynamic parameters (7.11). Then the equilibrium concentrations per element are determined with equation (7.12). The concentrations in the elements are determined by solving a system of equation, which is set up by substituting (7.15) into equation (7.13). Then the morphological change can be determined from equation (7.14).

## 7.3.4 Illustrative examples

## A single element model

In order to illustrate the working of the model we consider first a simple example: a network existing of only one single element. For this single element we have then the following equations:

Morphological equilibrium

$$V_e = f(hydrodynamic \ parameters) \tag{7.17}$$

Equilibrium concentration

$$c_e = c_E \left(\frac{V_e}{V}\right)^n \tag{7.18}$$

Mass-balance in the water phase

$$T_o = w_s A(c_e - c) \tag{7.19}$$

There is only one transport  $T_{o}$ , i.e. the transport from the element to the outside world. Suppose that there is no residual flow, then

$$T_o = \delta(c - c_E) \tag{7.20}$$

Morphological change (no dredging / dumping, no sea level change)

$$\frac{dV}{dt} = w_s A(c_e - c) \tag{7.21}$$

Combination of the equations (7.17) through (7.20) yields:

.

$$\frac{dV}{dt} = \frac{\delta w_s A c_E}{\delta + w_s A} \left[ \left( \frac{V_e}{V} \right)^n - 1 \right]$$
(1)

Obviously when  $V=V_e$  no morphological change takes place. Consider a small disturbance from this equilibrium state  $V=V_e + \Delta V$ , and linearize equation (7.20) it can be shown that the disturbance  $\Delta V$  is decaying with a time scale (time in which the  $\Delta V$  is decreased by a factor e)

$$T = \frac{1}{nc_E} \left( \frac{V_e}{w_s A} + \frac{V_e}{\delta} \right)$$
(7.23)

The part outside the brackets shows that the morphological time scale is smaller if more sediment is in motion. The time scale is also smaller if the power in the equilibrium concentration equation is larger, which is similar as in the process based models where the time scale is dependent on the power flow velocity in the sediment transport formula. The two terms between the brackets represent respectively a time scale for the vertical exchange (between water and bottom) and for the horizontal exchange.

## A two elements model

Consider now a network consisting of two elements, with a connection between them, and with one of the elements connected to the outside world. Following the same procedures as given above it can be shown that the morphological development is described by a system of two equations:

$$\frac{dV_1}{dt} = f_1 \left( \frac{V_{e1}}{V_1}, \frac{V_{e2}}{V_2} \right)$$
(7.24)

$$\frac{dV_2}{dt} = f_2 \left( \frac{V_{e1}}{V_1}, \frac{V_{e2}}{V_2} \right)$$
(7.25)

It is now important to notice that there are now two morphological time scales. A disturbance  $(\Delta V_{1}, \Delta V_{2})$  with respect to the equilibrium state  $(V_{e1}, V_{e2})$  still decays, but the decaying rate depends on the disturbance itself. It is obvious that the decay is much faster for the case  $\Delta V_{1}=-\Delta V_{2}$  than for the case  $\Delta V_{1}=\Delta V_{2}$ .

# 7.4 Stability of River Bifurcations in 1D Morphodynamic Models

#### 7.4.1 Introduction

Nowadays, 1D morphodynamic models for single-channel rivers, based on water flow and sediment transport equations, have become useful tools for engineering practise. However, 1D morphodynamic models for river networks have rarely been reported in the literature, despite of the common use of 1D network hydrodynamic models. This is possibly due to the difficulty in representing bifurcations in such models.

At a bifurcating node in a 1D model a user-defined nodal-point relation is needed, which determines the sediment distribution over the downstream branches. In contrast to other equations in the model, this relation lacks a theoretical basis. Physically, the sediment distribution into the downstream branches is determined by the geometry and the (three-dimensional) current structure at the bifurcation. Obviously, it is impossible to represent this accurately in a 1D model. Because of this uncertainty of the nodal-point relation it is very important to understand its influence on the behaviour of the model.

The influence of the nodal-point relation on 1D network morphodynamic models has been analyzed mathematically in this paper. The conclusions from the analysis are verified by numerical simulations using SOBEK, a general 1D network package for flow, sediment transport and morphological development.

#### 7.4.2 Nodal-point Relations

S

Q

а

В

Consider the simple river network as shown in Fig.7-6: a river bifurcates into two branches ending in a lake. In a 1D model, it is reasonable to assume that the nodal-point relation at the bifurcation depends on dimensionless quantities, as follows

 $\frac{S_1}{S_2} = f\left(\frac{B_1}{B_2}, \frac{Q_1}{Q_2}, \frac{C_1}{C_2}, \frac{a_1}{a_2}, \dots\right)$ 

= sediment transport

= water discharge = water depth

= channel width

C = Chézy coefficient





The properties of the sediment in all three branches are assumed to be the same in this paper.

Model-technically, the function f should satisfy the following requirements:

(i) It is symmetric, i.e. the computed sediment transport into a branch does not depend on

the numbering of the branches, or mathematically,

$$f\left(\frac{B_1}{B_2}, \frac{Q_1}{Q_2}, \frac{C_1}{C_2}, \frac{a_1}{a_2}, \dots\right) = \frac{1}{f\left(\frac{B_2}{B_1}, \frac{Q_2}{Q_1}, \frac{C_2}{C_1}, \frac{a_2}{a_1}, \dots\right)}$$
(2)

(ii) It allows the model to represent all possible physically realistic situations.

A simple function satisfying the first requirement is the power function:

$$\frac{S_1}{S_2} = \left(\frac{B_1}{B_2}\right)^{k_B} \left(\frac{Q_1}{Q_2}\right)^{k_Q} \left(\frac{C_1}{C_2}\right)^{k_C} \left(\frac{a_1}{a_2}\right)^{k_a}$$
(3)

Note that a linear relation such as

$$\frac{S_1}{S_2} = \alpha \frac{Q_1}{Q_2} + \beta \tag{4}$$

does not fulfil requirement (i) unless  $\alpha=1$  and  $\beta=0$ .

In particular, the following case is considered.

$$\frac{S_1}{S_2} = \left[\frac{Q_1}{Q_2}\right]^k \left[\frac{B_1}{B_2}\right]^{l-k} \quad (k > 0)$$
(5)

This equation implies that the ratio between the specific sediment transport rates (transport per unit of width) is related to the ratio between the specific discharges. It will be

shown in this paper that this relation satisfies requirement (ii).

Measurements in scale models and/or in nature are required to verify the physical validity of the nodal-point relation (5). Unfortunately, measurements of sediment distribution at bifurcations are scarce. Here data for the bifurcation at Pannerden (The Netherlands) of the River Rhine are presented. The sediment distribution over the downstream branches has been measured, in nature as well as in a scale model (Van der Zwaard, 1981). The result of these measurements is shown in Fig. 7-7.

The least squares fit of a power of  $Q_1/Q_2$  for the results of the scale-model is:

$$\frac{S_1}{S_2} = 0.51 \left(\frac{Q_1}{Q_2}\right)^{2/2}$$
(6)

The least squares fit for the site measurements is:

$$\frac{S_1}{S_2} = 0.02 \left(\frac{Q_1}{Q_2}\right)^{6.0}$$
(7)

The values of the exponent k in Eqs (6) and (7) are clearly different, but both equations fulfil Eq. (5) if  $B_1/B_2$  is equal to 2.2, which is close to the ratio in reality.



Fig. 7-7 : Measurments of sediment distribution at Pannerden bifurcation

Observe that the variation of the ratio  $Q_1/Q_2$  is relatively small for the site measurements. It is also noted that the accuracy of the site measurements is much worse than that of the scale model measurements. Further the scale model was <u>distorted</u>. Hence <u>in detail</u> the current pattern at the bifurcation may not be similar, resulting in a different *k* value.

## 7.4.3 Equilibrium States

The river network system as shown in Fig.1 is in equilibrium if the flow is uniform in all branches and if the sediment transport capacity of each branch is equal to the sediment supply at the upstream end of the branch. The equilibrium state of the two downstream branches is determined by the following equations.

Mass-balance for water

$$Q_1 + Q_2 = Q_0 \tag{8}$$

Water motion described by the Chézy formula

$$Q_j = B_j C_j a_j^3 i_j^4 \qquad j = 1,2$$
 (9)

Geometric relation

$$i_1 L_1 = i_2 L_2$$
 (10)

Mass-balance for sediment

$$S_1 + S_2 = S_0$$
 (11)

Sediment transport described by a power-law

$$S_j = B_j m \left(\frac{Q_j}{B_j a_j}\right)^n \quad j = 1,2$$
(12)

In these equations *i* denotes the slope. Together with the nodal-point relation (1) these equations can be solved to determine the eight variables  $Q_{j}$ ,  $S_{j}$ ,  $a_{j}$ ,  $i_{j}$  (for j=1,2) as follows. From Eqs (9) and (10) it follows

$$\frac{Q_1}{Q_2} = \frac{B_1}{B_2} \frac{C_1}{C_2} \left(\frac{a_1}{a_2}\right)^{\frac{3}{2}} \left(\frac{L_1}{L_2}\right)^{\frac{1}{2}}$$
(13)

Together with Eq. (12) this gives a relation between the sediment transport ratio and the depth ratio:

$$\frac{S_1}{S_2} = \frac{B_1}{B_2} \left(\frac{C_1}{C_2}\right)^n \left(\frac{a_1}{a_2}\right)^{\frac{n}{2}} \left(\frac{L_1}{L_2}\right)^{\frac{n}{2}}$$
(14)

Substituting Eq. (13) into the nodal point relation (1) yields another relation between these two variables:

$$\frac{S_1}{S_2} = f\left(\frac{a_1}{a_2}\right) \tag{15}$$

From Eqs (14) and (15)  $S_1/S_2$  and  $a_1/a_2$  can be solved. With the known sediment transport ratio  $S_1$  and  $S_2$  can be solved from Eq. (11). With the discharge ratio determined by Eq. (13)  $Q_1$  and  $Q_2$  can be solved from Eq. (8). The depth values  $a_1$  and  $a_2$  then follow from Eq. (12).

It is important to note that there are not one but three equilibrium states, representing the following situations:

- Both branches are open, each branch transports a part of the water and sediment.
- Branch 2 is closed (*Q*<sub>2</sub>=*S*<sub>2</sub>=*a*<sub>2</sub>=0) and all the water and sediment is transported through branch 1.
- Branch 1 is closed  $(Q_1=S_1=a_1=0)$  and all the water and sediment is transported through branch 2.

All three possibilities are physically realistic. For the symmetric case  $(B_1=B_2=B_0/2, C_1=C_2, L_1=L_2)$  and a nodal point relation satisfying the symmetry requirement (i) the three equilibrium states are  $(a_0, a_0)$ ,  $(2^{(n-1)/n}a_0, 0)$  and  $(0, 2^{(n-1)/n}a_0)$ , where  $B_0$  is the width and  $a_0$  is the equilibrium water depth of the upstream branch.

Remark: Nodal point relations leading to not uniquely determined equilibria with both branches open are not considered because they seem physically not realistic. When a simulation of the morphodynamic development is carried out for a network system, one of these equilibrium states will be reached. Which one of the three equilibrium states will be reached depends on the stability of the equilibrium states and on the initial conditions. In the section below, the stability of all three equilibrium states is analyzed for the general nodal-point relation (1).

# 7.4.4 Stability of the Equilibrium States

In order to analyze the stability of the equilibrium states the morphological development of the system out of equilibrium is considered. To ease the analytical treatment, it is assumed that the bathymetry of each branch can be represented by a single depth value, even if the system is not in equilibrium. Under this assumption the morphological development is described by a system of ordinary differential equations instead of a system of partial differential equations.

With a uniform depth, the mass-balance for sediment in each branch yields

$$\frac{d a_j}{d t} = \frac{S_{je} \cdot S_j}{B_j L_j} \quad (j = 1, 2)$$

$$L_j = \text{length of the branch}$$

$$S_j = \text{sediment transport as supplied by the main channel}$$

$$S_{je} = \text{sediment transport at the downstream boundary}$$

$$t = \text{time}$$

$$(16)$$

With Eqs (8), (9) and (10) the discharges in the two downstream branches can be expressed as function of the water depths.

$$Q_{1} = \frac{\beta_{1}a^{\frac{3}{p}}}{\beta_{1}a^{\frac{3}{p}} + \beta_{2}a^{\frac{3}{2}}} Q_{0}$$
(17)

$$Q_{2} = \frac{\beta_{2} a_{\overline{2}}^{3}}{\beta_{1} a_{\overline{p}}^{3} + \beta_{2} a_{\overline{2}}^{3}} Q_{0}$$
(18)

Herein  $\beta_j = B_j C_j L_j^{-1/2}$ , *j*=1,2. With the general nodal-point relation (1), the ratio between the sediment transports at the upstream end of the two branches can be expressed as a function of the ratio between the depths.

$$S_{1} = \frac{f(a_{1}/a_{2})}{1 + f(a_{1}/a_{2})}S_{0}$$
(19)

$$S_2 = \frac{1}{1 + f(a_1/a_2)} S_0$$
(20)

The sediment transport at the end of a branch is determined by the transport capacity of the branches defined by Eq.(12).

$$S_{je} = \frac{B_j m C_j^n a_j^{n/2} L_j^{n/2} Q_0^n}{(B_l C_l a_l^{3/2} L_l^{1/2} + B_2 C_2 a_2^{3/2} L_2^{-1/2})^n} \quad j = 1,2$$
(21)

Substitution of Eqs (19), (20) and (21) into Eq. (16) gives a system of ordinary differential equations for  $a_1$  and  $a_2$ :

$$\frac{d a_1}{d t} = \frac{Q_0^n}{B_1 L_1} \left[ \frac{\gamma_1 a_1^{n/2}}{(\beta_1 a_1^{3/2} + \beta_2 a_2^{3/2})^n} - \frac{f(a_1/a_2)}{1 + f(a_1/a_2)} \frac{m_0}{B_0^{n-1} a_0^n} \right]$$
(22)

$$\frac{d a_2}{d t} = \frac{Q_0^n}{B_2 L_2} \left[ \frac{\gamma_2 a_2^{n/2}}{(\beta_1 a_1^{3/2} + \beta_2 a_2^{3/2})^n} - \frac{1}{1 + f(a_1 / a_2)} \frac{m_0}{B_0^{n-1} a_0^n} \right]$$
(23)

The parameters  $\gamma_j$  for j = 1 or 2, are defined as  $\gamma_j = mB_jC_j^nL_j^{-1/2}$ .

The set of differential equations (22) and (23) is of the general form

$$\frac{d a_1}{d t} = \phi_1(a_1, a_2)$$

$$\frac{d a_2}{d t} = \phi_2(a_1, a_2)$$
(24)
(25)

where  $\phi_1$ ,  $\phi_2$  are functions in two variables. The depths at the equilibrium states satisfy  $\phi_1 = \phi_2 = 0.$ 

The stability of an equilibrium state depends on the eigenvalues of the Jacobian. If the eigenvalues have negative real parts, perturbations of the equilibrium decay exponentially in time and the equilibrium is stable. If one of the eigenvalues has a positive real part, however, the equilibrium is unstable.

For general values of the parameters, the eigenvalues of the Jacobian cannot be determined by straightforward computation. Only in the special case with symmetric branches, it is easy to compute the eigenvalues. The Jacobian at  $(a_0, a_0)$  is

$$J(a_{0},a_{0}) = \frac{S_{0}}{B_{0}a_{0}L_{1}} \begin{bmatrix} -\frac{n}{4} - \frac{2a_{0}}{(1+f)^{2}} \frac{\partial f}{\partial a_{1}} & -\frac{3n}{4} - \frac{2a_{0}}{(1+f)^{2}} \frac{\partial f}{\partial a_{2}} \\ -\frac{3n}{4} + \frac{2a_{0}}{(1+f)^{2}} \frac{\partial f}{\partial a_{1}} & -\frac{n}{4} + \frac{2a_{0}}{(1+f)^{2}} \frac{\partial f}{\partial a_{2}} \end{bmatrix}$$
(26)

The eigenvalues of the matrix are

$$-n\frac{S_0}{B_0 a_0 L_1}, \quad \left(\frac{n}{2} + \frac{2a_0}{(1+f)^2} \left(\frac{\partial f}{\partial a_2} - \frac{\partial f}{\partial a_1}\right)\right) \frac{S_0}{B_0 a_0 L_1}$$
(27)

The stability of the equilibrium  $(a_0, a_0)$  depends on the sign of the second eigenvalue. If this sign is positive, the equilibrium is unstable; if this sign is negative, the equilibrium is stable. The equilibrium is stable if, and only if, the following condition is satisfied

$$\frac{\partial f}{\partial a_1} - \frac{\partial f}{\partial a_2} > \frac{n(1+f)^2}{4a_0}$$
(28)

For the proposed nodal-point relation (5), the equilibrium is stable if k > n/3 and unstable if k < n/3. For k = n/3 Eq. (5) is not a good nodal point relation since then the equilibrium with both branches open is not well defined. If Eq.(4) is used as nodal-point relation the equilibrium is unstable, independent of the parameters. The Jacobians at the other two equilibria  $(2^{(n-1)/n}a_0, 0)$  and  $(0, 2^{(n-1)/n}a_0)$  are

$$J(2^{\frac{n-1}{n}}a_{0},0) = \frac{S_{0}}{B_{0}a_{0}L_{1}} \begin{bmatrix} -\frac{n}{2^{\frac{n-1}{n}}} & 0\\ 0 & 0 \end{bmatrix}$$

$$J(0,2^{\frac{n-1}{n}}a_{0}) = \frac{S_{0}}{B_{0}a_{0}L_{1}} \begin{bmatrix} 0 & 0\\ 0 & -\frac{n}{2^{\frac{n-1}{n}}} \end{bmatrix}$$
(29)
(29)
(30)

The eigenvalues of these matrices are independent of the nodal-point relation, because these equilibria consist of one channel only. One of the eigenvalues is negative and the other one is zero. So, the equilibria can be stable or unstable. The stability depends on the stability of the equilibrium at  $(a_0, a_0)$ . If  $(a_0, a_0)$  is a stable equilibrium, then  $(2^{(n-1)/n}a_0, 0)$  and  $(0, 2^{(n-1)/n}a_0)$  are unstable. If  $(a_0, a_0)$  is unstable, then  $(2^{(n-1)/n}a_0, 0)$  and  $(0, 2^{(n-1)/n}a_0)$  are stable.

For general parameters the stability of the equilibria remains the same. This follows from a continuity argument. The general case can be seen as a deformation of the symmetric case: start from the symmetric case and slowly deform the geometry of the system. The stability of an equilibrium can only change if a mathematical bifurcation occurs, i.e., if the number of equilibria changes. This happens possibly when Eqs. (24) and (25) degenerate. For our nodal-point relation, this does not happen unless k=n/3.

The qualitative behaviour of the system can be represented in a phase diagram, which shows the evolution of  $a_1$  and  $a_2$  in time (see Fig.7-8). In case k > n/3, the system evolves to a unique stable equilibrium, independent of the initial position of  $a_1$  and  $a_2$ . In case k < n/3, one of the two branches silts up, depending on the initial conditions.



Fig. 7-8: Phase diagram of the differential equations

The absolute values of the eigenvalues determine the time scales of the decay or growth of a perturbation of the equilibrium. They are the reciprocals of the time in which a perturbation grows or decays with a factor *e* (relaxation times). In case 0 < k < n/3, the eigenvector (1,-1) has a positive eigenvalue, and the eigenvector (1,1) has a negative eigenvalue. It follows that a difference between  $a_0$  and the average of  $a_1$  and  $a_2$  decays, but a difference between  $a_1$  and  $a_2$  grows. Moreover, in absolute value the positive eigenvalue is much smaller than the negative eigenvalue. It follows that a difference between  $a_0$  and the average of  $a_1$  and  $a_2$  decays, but a difference between  $a_1$  and  $a_2$  grows. Moreover, in absolute value the positive eigenvalue is much smaller than the negative eigenvalue. It follows that a difference between  $a_0$  and the average of  $a_1$  and  $a_2$  decays rapidly compared to the growth of a difference between  $a_1$  and  $a_2$ . This is illustrated by the numerical simulations. A more detailed description of the stability analysis with Eq. (5) as nodal point relation is given in Wang et al (1993).

#### 7.4.5 Numerical Verification

The analysis above shows that the nodal-point relation has a crucial influence on the behaviour of a 1D network morphodynamic model. It shows also that relation (5) is able to represent both the stable and unstable case. These conclusions have been verified by numerous computations (Fokkink and Wang, 1993; Den Dekker and Van Voorthuizen, 1994), which all agree with the theoretical analysis.

The results of two simulations are presented here. The numerical simulations consider the same problem as the theoretical analysis: a channel bifurcates in two branches. The sediment transport formula of Engelund-Hansen is applied which is equivalent to Eq.(12) with n=5.

The parameters used in both simulations are

model parameters	main	branch 1	branch
	channel		2
length	100 km	10 km	10 km
width	100 m	50 m	50 m
relative density of sand	⊿ = 1.65		
representative grain diameter	D <sub>50</sub> = 0.2 mr	n.	
roughness	C = 50 m/s <sup>17</sup>	2	
total discharge	Q <sub>0</sub> = 1000 m	า <sup>3</sup> /ร	
bed slope	0.00004		
total transport	S <sub>0</sub> = 0.02345	5 m³/s	
spatial step	⊿x = 1000 m		
Morphological time step	<i>∆t</i> = 1 day		

The water flow in the model is quasi-steady.

The first simulation is carried out with the nodal-point relation (5) using k=1 and the second one with k=3.

According to the analysis, k=1 should lead to one branch silting up and k=3 should keep both branches open. This is confirmed by the computations. Figure 7-9 shows the water depths of the two branches at the

nodal-point as a function of time and Fig.7-10 shows the evolution on the  $a_1$ - $a_2$  plane which gives a kind of phase diagram.



Fig 7.9: Computed water depth of the two branches

7.10 Computed phase diagrams

Clearly, for k=1 the bifurcation is unstable. A very small difference between the initial depths in the two branches leads to the situation that the shallower branch silts up. Note that there is also a difference between the averaged value of the two initial depths and the equilibrium depth. In the first period of the simulation this difference decreases quickly. When both depths  $a_1$  and  $a_2$  are almost equal to the equilibrium value, the difference between the two depths starts to grow significantly. As remarked in the previous section, this is explained by the fact that the absolute value of the negative eigenvalue (for the development along  $a_1=a_2$ ) is five times as large as the positive eigenvalue (for the system, such results can be very frustrating.

For k=3, the bifurcation is stable. Even a considerable large difference between the initial depths in the two branches is damped out on the long-term.

The theoretical analysis predicts that the equilibrium with both branches open is stable for k>5/3 and unstable for k<5/3. In reality, this transition can occur between 1.5 and 2, depending on the geometry of the branches in the model (Fokkink and Wang, 1993; Den Dekker and Van Voorthuyzen, 1994). This is due to the fact that the hydraulic radius of a branch is not equal to the water depth, as is assumed in the analysis.

## 7.4.6 Conclusions

A nodal-point relation is proposed, which relates the ratio between specific sediment transport rates to a power of the ratio between the specific discharges. The scarce data available for sediment distributions at bifurcations support this power relation. However, the exponent seems to vary from case to case.

Unlike a single-channel system more than one physically realistic morphological equilibrium state is possible in a network system. The morphological development of such a system is strongly influenced by the stability of the equilibrium states.

In a simple model of a bifurcation, the influence of the proposed nodal point relation on the behaviour of 1D network morphodynamic models is analyzed. The exponent in the relation appears to have crucial influence on the stability of the equilibrium states. For small values of the exponent the bifurcation is unstable: only one of the downstream branches tends to remain open at long-term. For large values of the exponent the bifurcation is stable: all branches tend to remain open.

The conclusions from the theoretical analysis have been verified by the results of numerical simulations using a general 1D network package.

## 7.4.7 References

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# 7.5 ESTMORF model

A DYNAMIC-EMPIRICAL MODEL FOR ESTUARINE MORPHOLOGY

## 7.5.1 INTRODUCTION

Morphological development is the result of interactions between flow and alluvial bed. The relevant processes for morphological development are flow, sediment transport and bed level change. A mathematical model for morphological development can be set-up by simulating each of these processes using physical laws. Such a process-based morphological model is called a dynamic model.

At present (1996), the dynamic type of models, one-dimensional as well as two-dimensional, are already commonly used in engineering practice for rivers. For estuaries, however, the long-term morphological development under the influence of natural processes, e.g. sea level rise, and/or under the influence of human activities, e.g. land reclamation, dredging and dumping, can still not be predicted by dynamic models. In practice, prediction of long-term morphological development in estuaries still relies to a large extent on empirical relations, in which the morphological variables, such as the cross-sectional area of channels, are related to integrated hydrodynamic parameters, e.g. the tidal volume.

Recently, ESTMORF is developed by DELFT HYDRAULICS and Rijkswaterstaat, a mathematical model for predicting the long-term (decades) morphological development of estuaries and tidal lagoons, under the influence of natural processes and human interference. The model is partly based on process descriptions and partly on empirical relations which combines the advantages of the dynamic type of models with those of the empirical ones.

# 7.5.2 MODEL DESCRIPTION

## General model concept

The most important hypothesis used in the ESTMORF model concept is that an equilibrium state can be defined for each morphological element of the system depending on the hydrodynamic conditions. An empirical relation is required for each element to define the morphological equilibrium state. This kind of empirical relations are commonly used for tidal systems.

Various morphological models for estuaries and tidal inlets are based on such relations in combination with a transient model describing the evolution of the actual state with respect to its equilibrium as an exponential decay process (O'Connor et al., 1990, Eysink, 1990, Eysink, 1992). When this model concept is applied to a system consisting of more than one element, linked with each other, extra assumptions are required to guarantee the mass-balance of sediment (Allersma, 1988, Van Dongeren and De Vriend, 1994). This can make the model results sensitive to the sequence in which the various elements are dealt with in the computation: e.g. computations starting from the sea side give different results than computations starting from the land side.

In the tidal basin model of Di Silvio (1989) this problem is overcome by introducing a characteristic sediment concentration in each element of the model. The concentration field is governed by the advection-diffusion equation based on a residual flow field, which guarantees the fulfilment of the sediment mass-balance. This concept is adopted in the ESTMORF model.

The basic philosophy is as follows. If all elements in the morphological system are in equilibrium, there is no accumulation of sediment or water anywhere in the area. If the sediment is mainly transported in suspension, the sediment flux field is therefore likely to be proportional to the flow rate. The ratio between sediment flux and flow rate can be considered as a sediment concentration and is called in this paper the overall equilibrium concentration  $c_E$ . For each element in the system a local equilibrium sediment concentration  $c_e$  is defined such that it equals  $c_E$  if the element is in morphological equilibrium, that it is larger than  $c_E$  if a tendency for erosion exist (e.g. the cross-sectional area of a channel is smaller than the equilibrium value), and smaller than  $c_E$  if a tendency for sedimentation

exists. Morphological changes occur when the local sediment concentration deviates from its local equilibrium. Erosion occurs when the sediment concentration is smaller than its equilibrium value and sedimentation occurs if it is larger than its equilibrium value.



The ESTMORF model described here belongs to the semi-empirical type according to the classification due to De Vriend (1996). An important difference between the present model and the process-based models is that the equilibrium concentration is not directly computed from the hydrodynamic parameters, but via the equilibrium morphological state (see Fig.7-11). This makes the model always converging to a state of which the equilibrium relations are satisfied.







## Geometrical schematisation

The model uses an existing one-dimensional network flow model to simulate the flow. Therefore, the modelling area is schematised into a network consisting of branches. The cross-sections of the branches are schematised as shown in Fig. 7.12 (Note that in the figure the tidal flat is shown only at one side of the channel). It is divided into three parts: the channel part (under the low water level, MLW), the low tidal flat between MLW and the mean water level (MSL), and the high tidal flat between MSL and the high water level (MHW). The channel part is assumed to have a trapezoidal cross-section. The two parts of the tidal flat are described by their widths as well as their heights. This schematisation of the cross-section is adopted from Eysink (1992). Morphological changes in the three parts are calculated separately, and can be due to sedimentation/erosion and to changes in the relative water level e.g. sea level rise/land subsidence.

The equilibrium state

The model uses three variables, the cross-sectional area of the channel, the height of the low tidal flat and the height of the high tidal flat. For each of these variables an empirical relation is required defining the morphological equilibrium state of the system. The equilibrium cross-sectional area of the channel is related to the tidal volume. For the Wadden Sea such a relation is given by e.g. Eysink (1992).

$$A_{ce} = f_1(P_v) \tag{1}$$

Herein

 $A_{ce}$  = equilibrium cross-sectional area under MSL  $P_v$  = tidal volume.

The equilibrium heights (measured from MLW) of the two parts of the tidal flat are related to the tidal range and the total area of the basin (see e.g. Eysink 1992).

$$H_{he} = f_2(H, A_b) \tag{2}$$

$$H_{le} = f_3(H, A_b) \tag{3}$$

In these two equations

 $H_{he}$  = equilibrium height of the high tidal flat measured from mean low water,  $H_{he}$  = equilibrium height of the low tidal flat

measured from mean low water,

- H =tidal range,
- $A_b$  = total area of the basin.

In equations (1) through (3) no specific relations are given. They have to be chosen on the basis of field data when the model is applied for a specific area. This part is kept as flexible as possible to guarantee a wide applicability of the model. The coefficients in the equations may also vary spatially such that an existing equilibrium state can be represented throughout the system.

## Morphological development

When the system is not in equilibrium, morphological changes will occur. The sedimentation and/or erosion rate is assumed to be proportional to the deficit/excess of the local sediment concentration, following the formulation of Galappatti and Vreugdenhil (19985). The bed level change is then described by the following simple formulation:

$$\frac{\partial z}{\partial t} = w_s (c - c_e) - \alpha_s \qquad (4)$$
Herein
$$z = \text{bed level},$$

$$w_s = \text{coefficient having the dimension of}$$

velocity, to be called vertical exchange velocity,

- t = morphological time,
- c = sediment concentration,

*c*<sub>e</sub> = local and instantaneous equilibrium

sediment concentration,  $\alpha_s$  = rate of land subsidence.

Applied to the three parts of the cross-section, the following equations are derived:

$$\frac{\partial A_c}{\partial t} = W_c \Big[ w_s \left( c_{ce} - c_c \right) + \alpha_s + \alpha_r \Big]$$
(5)

$$\frac{\partial A_l}{\partial t} = W_l \Big[ w_s \left( c_{le} - c_l \right) + \alpha_s + \alpha_r \Big]$$
(6)

$$\frac{\partial A_h}{\partial t} = W_h \Big[ w_s \left( c_{he} - c_h \right) + \alpha_s + \alpha_r \Big]$$
(7)

In these three equations we define (see Fig.1):

- $A_c$  = cross-sectional area of the channel par (below MSL),
- $A_h$  = cross-sectional area of the high tidal flat,
- $A_{l}$  = cross-sectional area of the low tidal flat,
- $c_c$  = sediment concentration in the channel part,
- $c_h$  = sediment concentration in the high tidal flat part,
- $c_i$  = sediment concentration in the low tidal flat part,
- $c_{ce}$  = local equilibrium sediment concentration in the channel part,
- $c_{he}$  = local equilibrium sediment concentration in the high tidal flat part,
- $c_{le}$  = local equilibrium sediment concentration in the low tidal flat part,
- $W_c$  =width of the channel part,
- $W_h$  = width of the high part of the tidal flat,
- $W_1$  = width of the low part of the tidal flat,
- $\alpha_r$  = rate of sea level rise.

## Sediment Transport

The equilibrium concentrations are determined as follows:

$$c_{ce} = C_E \left(\frac{A_{ce}}{A_c}\right)^{n_c} \tag{8}$$

$$c_{le} = C_E \left(\frac{H_l}{H_{le}}\right)^n$$

$$c_{he} = C_E \left(\frac{H_h}{H_{he}}\right)^{n_h}$$
(9)
(10)

In these equations:

 $C_E$  = overall equilibrium sediment concentration,

 $A_c$  = cross-sectional area of the channel,

 $H_h$  = height of the high tidal flat measured from MLW,

 $H_{l}$  = height of the low tidal flat measured from MLW,

- $n_l$  = constant coefficient,
- $n_h$  = constant coefficient,
- $n_c$  = constant coefficient

When the whole system is in equilibrium, the sediment concentration will be equal to its overall equilibrium value. When for example the cross-sectional area is larger than the equilibrium value it means in fact that the local flow velocity will be lower by the same factor as the ratio  $A_{ce}/A_c$ . The formulation (8) of the local equilibrium sediment concentration is thus similar to the sediment transport capacity formula using power law. The actual sediment concentration is governed by an advection-diffusion model. It is further assumed that the tidal flat can only exchange sediment with the channel in the same section. In other words, there is no longitudinal sediment transport over the intertidal

tidal flats.

The mass balance equation for sediment in the channel may be written as:

$$\frac{\partial A_c c_c}{\partial t} + \frac{\partial A_c u c_c}{\partial x} - \frac{\partial}{\partial x} \left( A_c D_c \frac{\partial c_c}{\partial x} \right) = W_c w_s (c_{ce} - c_c) + F_{lc}$$
(12)

where:

Cc	= sediment concentration by volume in the channel,
$D_c$	= horizontal dispersion coefficient in the channel,
F <sub>lc</sub>	= exchange rate of sediment between the channel and the low tidal flat,
t	= time,
и	= residual flow velocity.

The sediment flux from the tidal flat to the channel  $F_{lc}$  is elaborated as follows:

$$F_{lc} = D_l \ \overline{\Delta h_l} \ \frac{c_l - c_c}{L_{lc}}$$
(13)

Herein:

$D_l$	= diffusion coefficient,
LIC	= distance between the centre of the channel and that of the
	low part of the tidal flat,
<u> </u>	

$$\Delta n_l$$
 = time-averaged water depth at MLW

The mass balance for sediment at the low part of the tidal flat is given by the following equation:

$$\frac{\partial (A_l c_l)}{\partial t} = W_l w_s (c_{le} - c_l) - F_{lc} + F_{hl}$$
(14)

And at the high part of the tidal flat the mass-balance reads

$$\frac{\partial(A_h c_h)}{\partial t} = W_h w_s (c_{he} - c_h) - F_{hl}$$
(15)

The exchange rate between the two parts of the tidal flat is formulated as

$$F_{hl} = D_h \,\overline{\Delta h_h} \, \frac{c_h - c_l}{L_{hl}} \tag{16}$$

Herein:

 $D_h$  = diffusion coefficient,

- $L_{hl}$  = distance between the centre of the high flat and that of the low flat,
- $\overline{\Delta h_h}$  = time-averaged water depth at MSL.

# 7.5.3 Comparison with dynamic models

The behaviour of the model can be analysed by considering some simplified cases. Here two extreme cases are considered. It will be shown that in both cases the behaviour of the model is very similar to a process-based model.

The first case concerns the unidirectional steady flow in a river without tidal influence (hence

no tidal flats exist). In this case the tidal volume is proportional to the river discharge, so the equations describing the sediment transport and the morphological development used in the model become identical to those in a process-based one-dimensional model using depthaveraged model for suspended sediment transport (see e.g. Galappatti and Vreugdenhil, 1985). The only difference is that the morphological equilibrium state in the present model is merely determined by the river discharge, whereas in a 1D process-based model it is determined by the discharge, the upstream sediment supply and the river width. It is obvious that the empirical relations used in the model should be site-specific.

The second case concerns a rectangular tidal basin, with one end closed and the other end connected with the sea. This case has been used for extensive sensitivity analysis with respect to the various model-parameters (Karssen, 1994). A typical example of the results is shown in Fig. 7-13. The simulation start with a horizontal bed and ends with a linear bottom. which is in full agreement with the commonly used empirical relation that the cross-sectional area is proportional to the tidal volume through that cross-section (note that the diffusion coefficient at the end of the basin is not set to zero). Also in this case the simulated morphological development is very similar to that simulated with a process-based model (see Wang, 1992, Schuttelaars and De Swart, 1996). From the sensitivity analyses it is further concluded that the vertical exchange velocity, the overall equilibrium concentration, the dispersion coefficient and the power in the equilibrium concentration equations are the most important parameters influencing the simulated morphological development.



### 7.5.4 Morphological time scale

Another way of examining the behaviour of the model is by analysing how the morphological time scale is related to the relevant length scale. Knowledge of how the morphological time scale is determined by the model parameters is very important for the calibration of the model. For this purpose a linear analysis is carried out for a small disturbance on a system in equilibrium. The time scale defined is the time in which the disturbance decreases its amplitude with a factor e. For the sake of simplicity only the channel is considered, and the residual flow is neglected. The time scale in this case is:

$$T = \frac{T_s + T_D}{c_E n_c}$$
Herein  $T_s = \frac{a}{c_E}$ 
(17)

Herein

$$T_D = \frac{w_s}{k^2 D}$$

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and *k* is the wave number of the disturbance. It is clear that the time scale consists of two parts, one related to the vertical exchange of sediment and the other related to the horizontal exchange of sediment. It is also clear that the parameters D,  $w_s$ ,  $c_E$  and n are the important model parameters determining the morphological time scale. More extensive and detailed analysis of the morphological time scale related to the model parameters is given by Fokkink (1996).

# 7.5.5 Applications

# Het Friesche Zeegat

"Het Friesche Zeegat" is a tidal inlet in the Wadden Sea (Fig.7-14). Since 1969, this tidal inlet system, with an area of about 450 km<sup>2</sup>, has undergone significant morphological changes due to the reclamation of a large part, the Lauwerszee. As a consequence the in- and out-going tidal volume has been reduced by about 34%. The response of this tidal inlet system to the closure of the Lauwerszee is well documented. The deeper parts of the outer delta are eroding, large bars are forming, the tidal channels are shoaling and the eastern watershed is shifting eastward.



Fig.7-14: Het Friesche Zeegat

The ESTMORF model is applied to simulate the morphological development induced by the closure of the Lauwerszee in 1969. The simulation focuses on the morphological development of the tidal basin, especially the channels.

The model is set-up based on an existing 1D network model for this inlet. The schematisation is also shown in Fig. 7-15. It must be pointed out that the schematisation was originally designed for flow simulation purposes only. From the morphological modelling point of view it is not an ideal schematisation. Nevertheless the application has been rather successful in hindcasting the evolution of the channels' cross-sectional areas throughout the basin as shown in Fig.7-17, which presents the morphological evolution of some of the channel cross-sections. As can be observed in Fig.7-15, there is a channel bifurcation at the entrance of the Lauwerszee. The closure has cut off one of the branches of this bifurcation. After the closure sedimentation has occurred in the channel west of the bifurcation and erosion took place east of the bifurcation. The sedimentation in the western part (cross-sections 3 and 4) are well reproduced (Fig.7-17).

## The Western Scheldt

The Western Scheldt is an estuary located in the South-west of The Netherlands. It is the entrance to Antwerp Harbour. Dredging operations maintaining and improving the navigation conditions can have significant impact on the morphological development, which in turn can influence the ecological conditions in the estuary. In the near future a significant deepening of the navigation channel is planned. It is very important to predict the morphological and ecological impacts of the capital dredging and of the increased maintenance dredging in the future.

An ESTMORF model is set-up for the estuary to simulate the morphological developments under influence of human activities, like dredging and dumping, and under influence of natural processes, like sea level rise. Also in this case the model is based on an existing 1D-network tidal flow model. The schematisation of the model is shown in Fig.7-15. As an example of the model results, the simulated bed level change after 25 years is shown in Fig.7-16. The results have also been used for ecological modelling. For this purpose a two-dimensional interpretation is made of the simulated morphological changes in order to evaluate the changes of relevant habitats.

At present the study is still continuing. More detailed discussion on the morphological modelling for the Western Scheldt is given by Fokkink et al. (1996).



Fig.7-15: Schematisation of the Westernscheldt estuary



Fig.7-16: Computed bed level change after 25 year



Fig.7-17: Computed and measured cross-sectional areas

### 7.5.6 Conclusions

A semi-empirical model for morphological changes in estuaries and tidal lagoons is developed on the basis of a one-dimensional network tidal flow model. The model uses well-established empirical relations between the morphological equilibrium state and integrated tidal flow parameters. The sediment is assumed to be transported primarily as suspended load.

Tidal flow is considered as the major force causing morphological changes, but the influences of other factors like sea level rise and human activities (e.g. dredging and dumping) can also be included. The model can be used for predicting morphological changes on a time scale of decades.

The model makes a distinction between the morphological changes in the channel and those on the intertidal flats. Morphological changes on these flats are important for the ecosystem, and morphological changes in the channel for navigation.

For two extreme cases, a river flow without tidal action and a prismatic tidal basin without river input, the model behaves the same as the process-based morphodynamic models. This gives the confidence that the model is widely applicable to estuaries. In addition to the empirical relations defining the morphological equilibrium, the most important parameters determining the behaviour of the model are the overall equilibrium concentration, the vertical exchange velocity of sediment, the horizontal diffusion coefficient for sediment and the power in the equilibrium concentration formula.

The model has been applied to "Het Friesche Zeegat", a tidal inlet in the Wadden Sea, and The Western Scheldt, an estuary in the Southwest of The Netherlands. In the first case the morphological changes, especially in the channel, caused by the closure of a part of the tidal basin, have been successfully simulated by the model. Good agreement between the computed and observed changes of the cross-sectional areas of the channels has been obtained. In the second case the study is still going on but encouraging results have already been achieved in simulating the impact of dredging and dumping activities in the navigation channel.
# 7.5.7 References

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# 7.6 ASMITA model

## 7.6.1 Introduction

The motivation for the present work is derived from studies into the response of the Mediterranean deltaic plains of Ebro, Po and Rhone rivers to climate and other humaninduced changes, with an emphasis on the larger time- and space-scales (see e.g. Jimenez et al., 1996; and Capobianco et al., 1996). A specific part of those studies is concerned with the land-sea interface of the deltaic plains. A common element of this interface, which we term the deltaic fringe, is that of sediment starvation due to the upstream regulation of water resources and to artificialisation of the deltaic plains. This implies that the three deltaic fringes are nowadays fairly wave-dominated in as far as it concerns their morphological evolution and they have been progressively reducing their extension because of sea forcings on the one side and human reclamation activities on the other side. Typical geophysical elements which the deltas have in common are, beside the rivers' mouth regions, lagoons and bays which interrupt the more continuous coastal stretches, which latter exhibit strong local curvatures. The lagoons and bays provide an important sediment accommodation space for a situation in which relative mean sea level has been dramatically rising in the past and is still rising in a significant way nowadays.

Obviously, morphological interactions take place between the adjacent coastal stretches, the lagoons and their submerged deltas at different spatial and temporal scales, according to intrinsic dynamics, that we would like to characterise, and to external forcing factors. The work presented here focuses on the "residual" interaction occurring on a long term scale. Based on existing field information and on generic modelling knowledge regarding tidal basins and lagoons on the one hand and wave-dominated coastal stretches on the other hand, we have formulated a modelling approach which would allow to determine the possible impact of increased, or decreased, relative sea-level rise, of changes in wave climate and human-induced regulation of the lagoons and the adjacent coastal areas. Because of the time-scales of interest we consider the geophysical elements of the coastal fringe at an aggregated scale, e.g. single-inlet lagoons are schematised into two or at most three spatial units such as channel area, and (high and low) flats area. The new element in our approach is that we study the non-linear dynamic interaction between the lagoon elements, the submerged delta and the adjacent coastal sections.

#### 7.6.2 Model formulation

From our aggregated scale model perspective we view a tidal inlet system as consisting of three major morphological elements, viz. the lagoon or tidal basin, the ebb tidal delta and the directly adjacent coast. Each of these elements is primarily influenced by the basin related tidal prism flows and secondarily by wave related hydrodynamics. However, within these elements the relative importance of the tidal prism flow and wave related hydrodynamics for their morphological development is different. For the coast, the additional effects of short waves and wave driven flow may be assumed strongest. Within the lagoon it is the tidal flow that is most important. For the ebb-tidal delta wave action as well as tidal action are important.

Like in many other classical situations in coastal morphodynamics, here we enter the problem of definition of possible equilibrium conditions. Particularly in situations where the relative role of the wave forcing with respect to tides is significant, it could happen that the ebb-tidal delta, as a distinct morphological unit, is nonexisting. An inlet in equilibrium is due to the balance between the wave energy which tends to close the inlet and the tidal energy which maintains the opening. For the theory of stability of the lagoon inlets, we refer to Bruun (1978). We hereby just recall some elementary concepts. Stability and equilibrium must be used in relative terms. No absolutely "stable" or "in equilibrium" inlet exists in a situation of significant longshore sediment transport; it is always subject to changes in its planform as well as in its cross-sectional area and geometry. Our approach is based on the use of morphological equilibrium relations and on equilibrium assumptions that, however, are still subject of active research investigation.

In our integrative effort, the most delicate point is represented by the ebb-tidal delta which is in fact the interfacing component by excellence. Apparently, such a sedimentary environment can undergo a self-organisation leading to a (quasi-) equilibrium morphology for each of the elements under relatively constant hydrodynamic forcing conditions. Our conceptual starting point is that we accept the existence of such a (quasi-) equilibrium state, and that we are specifically interested in the response of the system to disturbances of the equilibrium state. While earlier work has studied this for individual elements (see e.g. De Vriend, 1996), we find that the different elements cannot be isolated from each other when their morphological development is considered. The interactions between the different elements through sediment exchange play an important role for the morphological development of the whole system as well as of the individual elements.

Our approach can be considered as an aggregation on the one hand and an extension on the other hand of a model formulation for tidal basins (the so-named ESTMORF model, see Wang et al., 1996; and Fokkink et al., 1996). The aggregation concerns the fact that we characterise the system elements by only one state variable (viz. a total wet or dry volume). The extension concerns the incorporation of formulations for the ebb tidal delta and directly adjacent coast as well, without modifying the basic concepts. This applicability is based on the idea that also there tidal basin related flows are of sufficient importance to apply the ESTMORF concepts for equilibrium concentrations, sedimentation and transport. These concepts are:

- under constant (in a long term average sense) external (hydrodynamic) conditions the equilibrium situation for the morphological state variables is known as a function of those conditions or it is assumed to be known under the given scenario;
- in the equilibrium situation an overall long-term averaged equal sediment concentration in each system element exists;
- in case of external forcing of the elements' state variables new local equilibrium concentrations are defined. This leads to net sediment exchanges between the elements which are described by a diffusion-type transport, where the diffusion is assumed to take into account all the possible transport processes.

First, the state variables describing the system need to be defined. An ESTMORF model uses a network schematisation for spatial elements covering the tidal basin. In practice, this network can be as detailed as one wishes; the more detailed the network is the more the model would resemble a bidimensional finite element schematisation of the lagoonal area. Within the network each element or branch is divided into three parts: the channel part, the low tidal flat and the high tidal flat. However, the degree of schematisation in the present case is determined by that element of our system which delivers the lower boundary to the relevant spatial scale. This concerns typically the ebb-tidal delta, for which we presently have no other option than to consider its volume as an integral state variable, implying that the ebb tidal delta is modelled as a single element. It is then not very sensible to model the tidal basin and/or the adjacent coasts in more detail, and we logically start with the same level of schematisation, i.e. using the same spatial scales, for all the elements in the system. Later the model may be extended by using more detailed schematisation for the elements. Therefore the following basic elements are included in the model (Fig.7-18).

- the ebb tidal delta as a whole;
- the total inter tidal flat area in the basin;
- the total channel volume in the basin;
- the adjacent coast at one side;
- the adjacent coast at the other side.

It is assumed that sediment exchanges only occur between the tidal flat and the channel in the basin, between the channel in the basin and the ebb tidal delta, between the adjacent coasts and the ebb tidal delta, and between the coasts and the adjacent boundaries. The sediment exchange with the surrounding zones, viz. the more offshore shoreface and the coastal stretches further away, is assumed to not play a role in the morphodynamic interactions considered. This assumption looks reasonable in the presence of undisturbed situations where we could have a "migrating inlet", but this is probably not the case in the presence of artificially disturbed situations, where a morphological modification could occur. In such a case it would certainly be important to implement a more process-based integrative

formulation between the ebb-tidal delta dynamics and the adjacent coastal area longshore dynamics.









# Side view



As described above each of these elements is described by one morphologic variable representing its bathymetry:

- for the ebb tidal delta the total volume of the delta V<sub>d</sub> above a fictive sea bottom, which would be there if there was no inlet (sand volume);
- for the tidal flat the total volume of the flat Vf between MLW and MHW (sand volume);
- for the channel the total channel volume under MLW V<sub>C</sub> (water volume)
- for the two coast elements the volume below MSL and above a certain depth line  $V_{C1}$  and  $V_{C2}.$  (water 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. An empirical relation is required for each element to define the morphological equilibrium state. These relations are

derived from literature (cf. Eysink, 1990) and, even if derived for situations where tideinduced sediment transport is more significant than wave-induced sediment transport, we assume to be of general enough validity under the assumption that extreme waves are limited in such a way not to open and close new inlets and the occurrence of extreme wave events is "regular enough".

In the above equilibrium conditions two hydrodynamic parameters are the tidal range (H) and the tidal prism (P):

$$V_e = f(P, H) \tag{2}$$

In the present study a very simple hydrodynamic model is used. The size of the tidal basin is assumed to be small compared with the tidal wave length. Spatial variation of the water level in the basin is neglected. The tidal range is assumed to be given as function of (morphological) time. The tidal prism can thus be calculated from:

$$P = H \cdot A_{b} - V_{f}$$
(3)

where Ab equals the total basin area.

A key element in the modelling concept is the equilibrium concentration. The definition thereof is based on the following arguments. When all the elements in the morphological system are in equilibrium a constant sediment concentration is present in the whole system. This constant concentration is called the overall equilibrium concentration  $c_E$ . For each element in the system a local equilibrium sediment concentration  $c_e$  is defined such that it is equal to  $c_E$  if the element is in morphological equilibrium, larger than  $c_E$  if tendency of erosion exist (e.g. the volume of the ebb tidal delta is larger than the equilibrium value), and smaller than  $c_E$  if tendency of sedimentation exists. The local equilibrium concentration depends on the actual volume V<sub>c</sub>. To represent this behaviour a simple power relation is used for the equilibrium concentrations, e.g. for the channel element this reads:

$$c_{ce} = c_E \cdot \left(\frac{V_{ce}}{V_c}\right)^n \tag{4}$$

where the power n is larger than one, must commonly taken as 2 in compliance with a third power for the sediment transport as a non-linear function of the mean flow velocity. According to the modelling concept morphological changes occur when the local sediment concentration deviates from the local equilibrium sediment concentration. Erosion occurs when the sediment concentration is smaller than its equilibrium value and sedimentation occurs if it is larger than its equilibrium value. Again, for the channel element this reads:

$$\frac{\partial V_{c}}{\partial t} = W_{s} \cdot A_{c} \cdot (c_{ce} - c_{c})$$
(5)

where: w<sub>S</sub> is the vertical exchange coefficient and A<sub>c</sub> is the horizontal area of the element.

The sediment concentration and sediment transport is governed by the advection-diffusion equation with the source/sink term representing the sediment exchange with the bottom. The general form of this equation for the channels as used in ESTMORF reads:

$$\frac{\partial A_c \cdot c_c}{\partial t} + \frac{\partial A_c \cdot u \cdot c_c}{\partial x} + \frac{\partial}{\partial x} \left( A_c \cdot D_c \cdot \frac{\partial c_c}{\partial x} \right) + \frac{\partial}{\partial y} \left( A_l \cdot D_l \cdot \frac{\partial c_c}{\partial y} \right) = W_c \cdot w_s \cdot (c_e - c_c)$$
(6)

where: u is a residual flow velocity,

D<sub>C</sub> and D<sub>I</sub> are the transversal and lateral diffusion coefficients, A<sub>c</sub> and A<sub>I</sub> are an effective transversal and lateral cross-section respectively, W is the width of the element.

For the present case only individual elements of constant cross-section in transversal sense are considered, which means that a discretized form of the equations is used. Furthermore, it is assumed that the morphological time scale is much larger than the time scale for the sediment concentration adjustment to its equilibrium value and to its surrounding environment, so that the first term averages out. It is also assumed that advection transport only occurs along the coast due to longshore current with a constant discharge Q, so that the second term applies along the coast only. The third term applies to the tranversal sediment exchanges between the coast elements and the delta and between the delta and the channel, while the fourth term applies to the lateral sediment exchange between the channel element and the flats. The above equations applied to all elements result in a system of five linear equations for the concentrations in the five elements in the system. Considering the neighbouring elements flat and delta, the diffusion based sediment exchange yields:

$$\delta_{cd} \cdot (c_c - c_d) + \delta_{cf} \cdot (c_c - c_f) = W_s \cdot A_c \cdot (c_{ce} - c_c)$$
(6)

Here  $\delta$  represents the horizontal exchange coefficient.

Every element has a morphological time-scale, depending on the magnitude of the disturbance and the geometry of the elements. For a single element the morphological time-scale  $\tau$  equals:

$$\tau = \frac{1}{c_{E} \cdot n} \cdot \left( \frac{V_{e}}{W_{s} \cdot A} + \frac{V_{e}}{\delta} \right)$$
(7)

### Simplified "process-based" computation of the equilibrium concentration

The concept of equilibrium concentration was used extensively, even if still in a rather experimental way, for the evaluation of the morphological evolution of the inlets in the Lagoon of Venice (Di Silvio, 1989). The equilibrium concentration represents the fundamental link between hydrodynamics, sediment transport and morphology at the entrance of the lagoon. We can say that  $c_E$  is the average annual concentration induced by waves and currents. In principle we could say that it is possible to obtain a formulation for  $c_E$  by integrating over a long period of time any sediment transport formula, and obtaining something like:

 $c_E = c_E$  (local wave climate, tidal currents, water depth, grain diameter) (8)

Further, it should be noted that if a constant  $c_E$  is used<sup>1</sup>, than  $c_E$  by itself does not have any influence on the final equilibrium state of the system but it is an important parameter determining the time scale of the morphological development together with the dispersion coefficients and the fall velocity. Therefore  $c_E$  could be used as one of the calibration parameters in the model.

We could for instance write a formulation for c<sub>E</sub> by simply considering the Acker & White (1973) formula:

$$c = \frac{\alpha}{h} \cdot \left(u - U_0\right)^m \tag{6}$$

where:

c is the average concentration on the depth h, u is the current velocity, (9)

<sup>&</sup>lt;sup>1</sup> This is not a logical incongruence; simply, as we will further state in the following, the equilibrium concentration can be made dependent on variables that are subject to change.

 $U_{o}$  is the critical value for inducing the movement of a single particle.

The coefficient  $\alpha$ , the exponent m and the threshold value U<sub>o</sub> are a function of the grain size (d). Experiences from the lagoon of Venice (Di Silvio & Teatini, 1995) give values such as summarised in Table 7-1.

d	α	т	U <sub>o</sub> (m/s)
(µm)			
0.06	209 10-4	6.58	0.24
0.15	6.5 10-4	3.64	0.28
0.50	3.17 10-4	2.25	0.42

Tab. 7-1) Calibrated Coefficient for Various Grain Sizes

The equation could be applied not only to sediment suspended by tidal currents but also, with some approximation, to sediment suspended by waves by assuming the instantaneous velocity close to the bottom. The equilibrium concentration value is then obtained by averaging the equation over the whole year and by considering both currents and waves. Considering tidal currents, we obtain:

$$c_{Et} = \frac{1}{T_a} \cdot \int_{0}^{t_0} c_t(t) dt$$
 (10)

where:

 $c_{t}(t)$  is the instantaneous concentration computed from the instantaneous tidal current  $U_{t}(t)$ ,

 $t_0$  is the time, on the period  $T_a$ , when the tidal current is higher than  $U_0$ . In the lagoons of the northern Adriatic sea we can consider an almost linear curve for  $U_t(t)$ :

$$U_{t}(t) = U_{tm} \cdot \left(1 - \frac{t}{T_{a}}\right)$$
(11)

where  $U_{tm}$  is the maximum annual velocity in a certain point. Simplifying, we obtain:

$$\mathbf{c}_{\mathrm{Et}} = \frac{\alpha \cdot \left(\mathbf{U}_{\mathrm{tm}} - \mathbf{U}_{0}\right)^{\mathrm{m}+1}}{\mathbf{h} \cdot \mathbf{U}_{\mathrm{tm}} \cdot (\mathrm{m}+1)} \tag{12}$$

As far as the wave motion is concerned, the equilibrium concentration is still obtained by integrating the instantaneous concentration obtained from the instantaneous velocity:

$$c_{Ew} = \frac{1}{T_a} \cdot \int_0^{t_0} c_w(t) dt$$
 (13)

In such a case, however, the bottom velocity is obtained by waves of different height coming from different directions and related to wind strength, wind direction and fetch. In a certain point of the lagoon at depth h, the maximum velocity at the bottom  $U_w(\alpha)$  induced by the wave of height  $H(\alpha)$ ,generated by the wind  $W(\alpha)$  in the direction  $\alpha$ , can be approximately computed as (shallow water approximation):

$$U_{w}(\alpha) = \frac{\sqrt{g}}{2 \cdot \sqrt{h}} \cdot H(\alpha) = \phi(\alpha) \cdot W^{w}(\alpha)$$
(14)

where  $\varphi(\alpha)$  depends on the length of the fetch and the average depth while w is a coefficient approximately equal to 3/4.

We now assume that the curve of duration of the wave height H(t), once aggregated to include all the waves conditions coming from all directions, can be defined with an exponential function:

$$\frac{U_{w}(t)}{U_{wm}} = \frac{H(t)}{H_{m}} = e^{-\frac{\gamma \cdot t}{T_{a}}}$$
(15)

where  $U_{wm}$  is the maximum velocity at the bottom, corresponding to the maximum wave  $H_m$  that may occur in the period  $T_a$ . The ratio  $T_a/\gamma$  represents the "persistence" of the wave motion and is of course site dependent.

If we now consider that during an event the instantaneous velocity varies linearly from the maximum value to zero, the curve of duration of the instantaneous velocity results:

$$U_{w}(t) = U_{wm} \cdot e^{-\frac{2\gamma \cdot t}{T_{a}}}$$
(16)

or, approximately:

$$U_{w}(t) = U_{wm} \cdot \left(1 - \frac{2\gamma \cdot t}{T_{a}}\right)$$
(17)

Simplifying:

$$\mathbf{c}_{\mathrm{Ew}} = \frac{1}{2 \cdot \gamma} \cdot \frac{\alpha}{h} \cdot \frac{\left(\mathbf{U}_{\mathrm{wm}} - \mathbf{U}_{0}\right)^{m+1}}{\mathbf{U}_{\mathrm{wm}} \cdot (m+1)}$$
(18)

The equilibrium concentration is given by the annual average for both waves and tidal currents, i.e. the sum of the contributions given by waves and by tidal currents:

$$\mathbf{c}_{\mathrm{E}} = \mathbf{c}_{\mathrm{Et}} + \mathbf{c}_{\mathrm{Ew}} \tag{19}$$

It should be noted that, given the fact that U<sub>tm</sub> tends to be very small in the shallow areas while W<sub>tm</sub> tends to be very small in the channels, c<sub>et</sub> and c<sub>ew</sub> may be respectively neglected when the equilibrium concentration in the two areas is computed. The value of  $c_{et}$  may be computed in the channels for various sets of values  $(U_{tm} = q_m/h, h)$  i.e. for those sets  $(q_m, h)$ for which q<sub>m</sub> represents the maximum annual value of tidal flux per unit of channel width and h is the depth. Similarly, the value of  $c_{Ew}$  could be computed in the shallow water areas for various combinations ( $U_{tm} = H \cdot \sqrt{g}/(2\sqrt{h})$ , h,  $\gamma$ ) i.e. for various combinations  $H_m$ , h,  $\gamma$ , where  $H_m$ is the maximum annual wave height and  $1/2\gamma$  the persistence of wave motion in that area. The interesting aspect is that the values of  $q_m$ ,  $\gamma$ , h can be computed through an hydrodynamic tidal model and with a wave generation model (see Ruol and Sclavo, 1996). It is in addition interesting to consider that (differently from U<sub>m</sub> and U<sub>wm</sub> such values are not much dependent on local variations of depth determined by erosion or accretion and by sea level rise or subsidence, being much more related to the large scale characteristics of the lagoon, i.e.: dimension of tidal basins, width of channel subject to tidal motion (as far as qm is concerned); wind speed, extension and depth of the fetch area (as far as  $H_m$ ,  $\gamma$ , are concerned). The computation of  $H_m$ ,  $\gamma$ ,  $q_m$  should not be undertaken very often, unless we want to consider such drastic human induced modification of the lagoon extension. The computation of c<sub>F</sub> through formulas may be subject to various simplifications. Such formulas, on the other hand may be calibrated by using morphological and granulometric information about the lagoon.

## 7.6.3 Applications

## Simple normalised case

To illustrate the working of the model a simple normalised case is considered: all the system variables have the equilibrium value unity, all the relevant coefficients have the value unity. The ASMITA model consists of the five elements which are described in the above. The system is disturbed from the equilibrium by giving one or more of the variables an initial value different from the equilibrium value unity. Also the case considered is symmetric, i.e. the two coast elements are identical. The results of a number of simulations are discussed below. The following observations have been made:

If only one of the elements in the system is disturbed from equilibrium, the system reacts quickest if the disturbed element is the coast (Fig.7-19) and slowest if the disturbed element is the tidal flat (Fig.7-120). The disturbance for the coast equals a decrease of 10% and for the flat a decrease of 20%. A simple explanation for this is that the coast is the nearest to the outside world and the tidal flat is furthest from the outside world.



Simple normalized case; Vc1 +10%

Fig 7-19) Initial volume of coast1 is decreased with 10%





Fig 7-20) Initial volume of tidal flat is decreased with 20%

More than one time scale can be identified in the response of the system to the disturbance. The first reaction of the system with the smallest time scale is related to the diffusion of the disturbance to the other elements. This causes disturbances in all other elements in the system. Later all the disturbances are damped out with a much larger time scale. In fact it can be shown mathematically that the number of time scales in the system is at least equal to the number of elements in the system. For the present case there are thus at least four

different time scales<sup>2</sup>. This behaviour does not agree with the assumption in many empirical models that the disturbance in an individual element is damping out exponentially. The system reacts quickest when two neighbouring elements have the opposite disturbance. For instance the case where the channel and the ebb tidal delta have been disturbed in the opposite direction (Fig.7-21). Both the volumes of the ebb-tidal delta and the channel are enlarged with 10 million cubic metres. The dominating reaction of the system in this case is the compensation of the disturbances in the two elements between each other. The two undisturbed elements, the tidal flat and the coast, almost do not feel the disturbances. It is also noticed here that the reaction of the ebb tidal delta is faster than that of the channel in the basin although the magnitude of both disturbances are similar. It should be noted that such a case occurs when the tidal basin is made smaller by e.g. reclamation or closure. The decreased basin area causes a decrease in the tidal volume which means that the equilibrium values of the channel volume and the ebb tidal delta tends to be eroded.



# Simple normalized case; Vd +10E6, Vc -10E6 m3

# Friesche Zeegat

A first realistic application of the model is the application of the model for the Friesche Zeegat along the Dutch Wadden coast. The five-element model ASMITA is calibrated for the development of the Zoutkamperlaag Inlet (Fig.7-22) since the closure of the Lauwerszee in 1969. As a result of the closure both the channel and the delta are too large.



Fig 7-22) The Friesche Zeegat consists of the Zoutkamperlaag and the Pinkegat Inlet

For the calibration we used the data set of Biegel (1993) for flat and channel and the data set of Oost (1995) for the ebb-tidal delta. Table 7-2 shows some characteristics of the Zoutkamperlaag Inlet before and directly after the closure and in the calculated equilibrium state.

Fig 7-21) Initial volume of both delta and channel enlarged with 10 million cubic metres

<sup>&</sup>lt;sup>2</sup> In the linear approximation, i.e. for small disturbances around the equilibrium, there are exactly four time scales; under non linear regime, i.e. for large disturbances around the equilibrium, "additional" time scales appear.

	before closure	equilibrium
basin data		
lagoon area at MHW A <sub>b</sub> (10 <sup>6</sup> m <sup>2</sup> )	230	123
exchanged volume V per tidal cycle (tidal prism or flood volume) (10 <sup>6</sup> m <sup>3</sup> )	320	193
tidal range H (m)	2.25	2.25
channel data:		
channel area A <sub>ch</sub> (10 <sup>6</sup> m²)	57	41.2
channel volume V <sub>ch</sub> (10 <sup>6</sup> m³) <i>flat data:</i>	230	143
tidal flat area (10 <sup>6</sup> m <sup>2</sup> )	150	81.6
tidal flat volume (above MLW) (10 <sup>6</sup> m <sup>3</sup> )	170	83
delta data:		
delta area (10 <sup>6</sup> m <sup>2</sup> ) delta volume V <sub>d</sub> (10 <sup>6</sup> m <sup>3</sup> )	78 145	78 117
coast1 or coast2:		
coast area (10 <sup>6</sup> m <sup>2</sup> )	25	25
coast volume (10 <sup>6</sup> m <sup>3</sup> )	150	150
Tab. 7-2) Characteristic va	lues for the Zoutkamperlaag Inl	et of the Friesche Zeegat

The equilibrium volumes are calculated with the equilibrium relations for the flat between MLW and MHW (Eysink, 1990):

$$V_{fe} = \alpha_f \cdot A_b \cdot H \tag{20}$$

for the channel below MLW (Biegel, 1993):

$$V_{ce} = \alpha_c \cdot P^{1.55} \tag{21}$$

and for the delta (Biegel, 1993):

$$V_{de} = \alpha_d \cdot P^{1.23} \tag{22}$$

The coefficient  $\alpha_f$  depends on the tidal basin area at MHW. The original values and the used values of these relations are mentioned in Table 7-3.

Coefficients	original	used
α <sub>f</sub> (-)	-	0.3
$\alpha_{\rm c}  ({\rm m}^{-1.65})$	16 10 <sup>-6</sup>	20.5 10 <sup>-6</sup>
$\alpha_{\rm d}  ({\rm m}^{-0.69})$	<u>63.6</u> 10 <sup>-4</sup>	75 10 <sup>-4</sup>

Tab. 7-3) Coefficients of the equilibrium relations

It is assumed that the coefficients of the used relations fall within the ranges of uncertainty.

For the calibration the basic five-element model is used. We assumed a net discharge of  $100 \text{ m}^3$ /s along the coast in eastern direction. The developments in time of the five elements after the Lauwerszee closure are presented in Fig.23 to Fig.26. The model was calibrated to reproduce the observed changes in the volumes of the channels, the flats and the delta.



Friesche Zeegat - channel

Fig 7-23) Development of the channel below MLW

Friesche Zeegat - flat



Fig 7-24) Development of the flat between MLW and MHW

Friesche Zeegat - delta









Fig 7-26) Development of coast1 and coast2

Our interpretation of the results is as follows. The ebb-tidal delta is too large, and much sediment is transported to the channel, which accretes. The larger ebb-tidal delta causes the coasts to accrete. The difference between the two graphs in Fig.7-26 is the result of the longshore net discharge.

Some remarkable observations of the results may be made, which are in good accordance with observations:

The time scale of the first reaction of the system is less then 10 year, corresponding with a rapid erosion of the ebb tidal delta and a rapid sedimentation in the channel.

After 20 years, the system is still not in equilibrium although the disturbances in the ebb tidal delta and the channel have almost been damped out.

The slow extension of the tidal flat in the basin has the largest time scale.

The first reaction of the adjacent coast is sedimentation and later followed by a much slower erosion.

## Scardovari lagoon

In the framework of the earlier mentioned climate change project (see the Acknowledgements) a model application was defined for one of the lagoons in the southern part of the Po Delta, viz. Scardovari Lagoon (Table 7-4). Available data for Scardovari Lagoon in the Po Delta were rather scarce. The inner channel network and the recent morphological evolution of the inlet was derived from bathymetric information, see Table 7-4. In Fig.7-27 a plan view of the Scardovari Lagooon is shown. The equilibrium state was calculated with the Wadden Sea relations, with  $\alpha_c$ =65 10<sup>6</sup> m<sup>-1.5</sup> and  $\alpha_d$ =6.57 10<sup>6</sup> m<sup>-0.69</sup> (Eysink, 1990).

Scardovari	initial	equilibrium
basin data		
lagoon area at MSL $A_b$ (10 <sup>6</sup> m <sup>2</sup> )	29.0	29.0
exchanged volume V per tidal cycle	15.7	15.7
(tidal prism or flood volume) (10 <sup>6</sup> m <sup>3</sup> )		
tidal range H (m)	1.28	1.28
channel data:		
total length (10 <sup>6</sup> m)	18.0	18.0
average depth (m)	1.75	1.75
mean width (m)	150	150
channel area $A_{ch}$ (10 <sup>6</sup> m <sup>2</sup> )	2.7	2.7
channel volume V <sub>ch</sub> (10 <sup>6</sup> m <sup>3</sup> )	4.7	4.0
flat data:		
lagoon area at MHW (10 <sup>6</sup> m <sup>2</sup> )	30.0	30.0
lagoon area at MLW $(10^6 \text{ m}^2)$	28.0	28.0
tidal flat area (10 <sup>6</sup> m <sup>2</sup> )	27.3	27.3
tidal flat volume (above MLW) (10 <sup>6</sup> m <sup>3</sup> )	22.7	22.7

delta data:		
active base over a slope of 1:100 (m)	7.0	7.0
longshore extension (km)	2.0	2.0
delta volume $V_d$ (10 <sup>6</sup> m <sup>3</sup> )	4.9	4.7
coast:		
active base over 500 m cross-shore (m)	5.0	5.0
longshore extension (km)	2.0	2.0
coast volume (10 <sup>6</sup> m <sup>3</sup> )	7.5	7.5

Tab. 7-4) Characteristics of the Scardovari lagoon

From comparing the Wadden Sea equilibrium results with those of our crude estimates, we thus conclude that the equilibrium data are applicable to our case and are therefore adopted. Lacking validated information about the equilibrium flat volume we assume that the flats are in equilibrium in the case of the Scardovari lagoon.

The basic ASMITA model was applied to the Scardovari lagoon (Fig.7-28, Fig.7-29 and Fig.7-30). In this case the effects of longshore transport are neglected. In Fig.7-28 it is shown that the channel and delta volumes in the initial states are a little larger than in their equilibrium states.

It is management practice to dredge the channels for navigation and flushing purposes. To maintain its initial volume we adapt a scenario in which the channel is dredged with an interval of about 4.1 years. In Fig.7-28 it can be seen that the flat suffers mostly from the dredging. Much sand is transported directly from the flat to the channel. Because of the import of sand across the boundaries the small ebb-tidal delta starts to grow. As can be seen all elements develop to a new equilibrium.



Fig.7-27) Plan view of the Scardovari Lagoon



Fig.7-28) Applying the Wadden Sea relations to the Scardovari lagoon



Fig.7-29) Dredging the channel of the lagoon Scardovari with an interval of 4.1 years



Fig.7-30) Dredging and dumping of sediment in the Scardovari lagoon

Another possible management practice is to dump the dredged sediment onto the tidal flats (Fig.7-30). As we can see in Fig.7-30 the delta is negatively affected by the dumping of sediment onto the flats. Because the import of sediment across the boundary is zero, the demand of sand of the channel affects also the delta. The volume of the flat increases as a result of sand-storage.

As regards the model-results we can conclude the following:

The smaller the lagoon the quicker the response time of the system to disturbances. The time-scales of the system of the Scardovari lagoon are less then 10 year. In the case of dredging only the time-scales are larger then in the case of dredging and

#### dumping.

The attempt to enlarge the flat volume in small lagoons is cancelled out by the quick response of the small lagoon.

#### 7.6.4 Discussion and conclusion

Since the Netherlands' Coastal Defence Policy Study published in 1990, the morphological interaction between tidal basins and their ebb-tidal deltas and directly adjacent coasts (e.g. as encountered along the Wadden barrier island systems) has been identified as a basically unknown physical phenomenon. The most accepted working hypothesis has been that the tidal channel and flat elements obey equilibrium or empirical transient relations (of a relaxation character) strongly depending on the tidal prism strength and independent of the ebb-tidal delta and adjacent coast morphological behaviour. The idea behind the latter assumption is that the morphological adjustment process of the basin overrules that of the other system elements, thereby demanding a time-varying net influx (in case of sea-level rise driven creation of sediment accommodation space), which is finally delivered by the adjacent coast delivery capacity is temporally constrained the ebb-tidal delta acts as a sediment buffer. These hypotheses are based on sediment budget observations and interpretations over decadal time scales.

On a relatively smaller spatial scale and in a relatively more wave-dominated environment like the Po Delta fringe, under the above mentioned working hypothesis, the observation of the conditions of the ebb-tidal deltas would allow us to assess the balance of on-going processes for long term protection and management strategies. As a necessary prerequisite we should probably define the ebb-tidal delta more accurate as a distinctive morphological feature in wave-dominated environments.

Here we have presented an initial formulation of an aggregated-scale behaviour model for the interaction between tidal basin and its adjacent coastal environment, without adopting the a-priori assumptions described above. Necessarily so, the model combines observations and findings resulting from analogous model applications. It is based on existing field information and on generic modelling knowledge regarding tidal lagoons and basins on the one hand and wave-dominated coastal stretches on the other hand. Because of the timescales of interest we consider the geophysical elements of the coastal fringe at an aggregated scale, e.g. single-inlet lagoons are schematised into two or at most three spatial units such as channel area, and (high and low) flats area. This leads to a box-type or lumped-parameter model, which is not new in tidal inlet processes modelling (Mayor-Mora, 1974). The new element in our approach is that we study the non-linear dynamic interaction between the lagoon elements, the submerged delta and the adjacent coastal sections. This modelling approach should allow to determine the morphological response of the system to any imposed modification within the lagoon or along the coasts: for example new channel dredging or marshes creation within the lagoon as well as shore protection intervention along the outer region can be simulated and analysed. By means of the model application it is also possible to determine the impact of increased (or decreased) relative sea-level rise, of changes in wave climate and human-induced regulation of the tidal basin, the tidal inlet and the adjacent coast. The important aspect to consider is that the approach allows us to produce reasonable results with relatively poor initial information. A more complex modelling framework would not have produced better or more precise results. What remains to be done is to quantify, to the maximum possible extent, the uncertainty that, in our opinion, is a very time-scale related concept for such applications.

7.6.5 References

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<u>.</u>

# 8.1 Introduction

Mathematical modelling of tidal inlets can be based on either of the following strategies (cf. Terwindt and Battjes, 1990):

- down-scaling, which starts from large-scale field information on the system's behaviour and tries to unravel the underlying physical processes only as far as needed for good predictions. This approach is followed in most of the empirical and semiempirical models which are discussed in the companion chapter.
- up-scaling, which starts from the observation that tidal inlet behaviour, however large-scale, is the effect of water and sediment motion. Hence it must be possible to simulate this behaviour with a model based on descriptions of the elementary hydrodynamic and sediment transport processes. The constituting equations of such a model are derived from conservation laws for wave energy, wave number, water mass, mean flow momentum and sediment mass. This process-oriented approach is the key element of the models which will be discussed hereafter.

Present-day computer capacity and the development of powerful software systems for the mathematical modelling of waves, currents and sediment transport have brought process-based mathematical simulation models of morphological evolutions within the reach of tidal inlet research. Yet, the literature on applications of this type of models to tidal inlets is far from abundant.

A possible explanation is that tidal inlets are extremely complex dynamic systems with a variety of processes and mechanisms. At this moment, they are insufficiently understood to be sure that the elementary physics incorporated in a simulation model are sufficient to reproduce the phenomena of interest. Any simulation model, at one scale level or another, has to call upon "black-box" empirical or parametric submodels (cf. turbulence modelling, or wave-driven current modelling). This implies simplifications and approximations of which the long-term effects on the morphodynamic behaviour can hardly be overseen. Process-based simulation model, and that it takes only a big computer to simulate large-scale behaviour.

# 8.2 ISE-models vs. morphodynamic models

Process-based simulation models consist of a number of modules which describe waves, current and sediment transport, respectively. Depending on whether the dynamic interaction of these processes with the bed-topography changes is taken into account, these modules are used in series or in a time-loop (see Figs. 1a and b, respectively). Models of this type essentially work at the hydrodynamic time scale (typically one tidal period). Larger time scales are reached by continuing the computations over a longer time-span.

Models which take the interaction with the topographic changes into account are called *Medium-Term Morphodynamic* (MTM) models. In principle, they are able to describe the meso-scale dynamic behaviour of a morphological system, such as the formation and migration of morphological features (channel shoaling or deepening, bar formation and migration, sandwaves, etc.). Models which don't take this interaction into account only describe micro-scale phenomena such as the sedimentation/erosion rate on a given bed topography. In the time-evolution of the system's state, this represents the initial changes. Therefore, this type of models is called *Initial Sedimentation/Erosion* models (ISE-models).

The above terminology is relative to the inherent morphodynamic time scale of the phenomena of interest. The term "initial" therefore does not imply that the results of such a model only provide a snapshot at the hydrodynamic time scale. In the case of large-scale evolution of an inlet system, for instance, the morphological time scale can be of the order

of decades or centuries. Morphological changes over a few years at this large spatial scale may therefore be considered as "initial". The essence of ISE-models that the feedback of morphological changes on the water and sediment motion is negligible.

This may explain why, in spite of the inherent limitations of ISE-models, there is a demand for their practical application, especially to 3-D cases. Most of the practical tidal inlet problems belong to this category. The model is utilized to investigate wave, current and sediment transport fields in the area, and their response to projected interferences (engineering works, sand mining concessions, land reclamation projects, etc.). An example of such a case will be discussed in a later section.

Both the ISE- and the MTM-modelling process consists of a number of steps, such as problem definition, formulation of objectives, analysis of physical processes, model set-up, model composition, input schematization and definition of runs, processing and visualization of results, verification, interpretation and reliability assessment of results, etc. In the following sections, some of these steps will be discussed in further depth, with the emphasis on specific aspects of tidal inlet modelling.





Figure 8.1: Aggregated flow charts of proces-base simulation models for tidal inlets, (a) Initial Sedimentation/Erosion models, (b) Medium-Term Morphodynamic models

# 8.3 Model composition

# 8.3.1 General

Both ISE- and MTM-models require a careful selection of the constituent models. Clearly, each of the relevant constituent processes has to be modelled adequately. This adequacy, however, is not only a matter of the process description per se, but also of the combination of modules which form the model as a whole. If, for instance, the wave model gives reasonable predictions of the wave height distribution, and the tidal model has been shown to reproduce water levels and current velocities within acceptable bounds, their combination does not necessarily give good results for the near-bed velocity which is put into the sediment transport model.

Especially in complex situations such as tidal inlets, the model composition ought to be the result of an analysis which starts from the objective of the study and its translation into phenomena to be modelled. This analysis should identify the relevant underlying processes and mechanisms, and assess the type of constituent models which is needed to describe them. The combination of constituent models which is ultimately chosen should be balanced from an efficiency and accuracy point of view, and it should not lead to spurious interactions or a blow-up of inaccuracies (cf. De Vriend, 1987).

Tidal inlets involve a number of specific aspects which need due consideration when

modelling the constituent processes. In the following sections, the most important of these aspects will be discussed.

## 8.3.2 Aspects of wave modelling

The complex topography of the outer delta, with one or more deep ebb-channels and often a well-developed channel/shoal system on the fringe, leads to a much more complex wave field than an uninterrupted piece of exposed coast. The channels tend to convey the wave energy, and the shoals give rise to complex refraction patterns. Wave breaking and the associated current-driving force (radiation-stressgradient) are quite variable in space and drive complex circulation patterns (Fig. 8.2). In order to describe these phenomena, a spectral or semi-spectral (single-frequency) refraction model with directional spreading (the HISWA-model, for instance; see Holthuysen et al., 1989) will usually do for wide natural inlets. The computation of the wave-induced forces via the radiation stresses requires caution, in order to avoid large numerical errors (cf. Dingemans et al., 1987).

Inlets which are fixed by engineering works such as jetties and revetments are usually narrower (see many inlets at the east coast of the U.S.A.) and more reflective. This can lead to very complex wave patterns which can only be described if reflection and diffraction are included (cf. Kirby et al., 1994).



Figure 8.2: Residual circulation pattrens on the outer delta of the Frisian Inlet. (From: Steijn and Hartsuiker, 1992; courtesy DELFT HYDRAULICS)

Wave penetration into the basin mainly occurs via the deeper channels (Fig. 8.3), where the waves experience less bottom friction and where they tend to be trapped by refraction. For the wide inlets considered herein, these phenomena are described reasonably well with the aforementioned type of (semi-)spectral refraction model.

The top level of the tidal flats inside the basin is the result of two opposing agents, viz. tide (accretive) and wave action (erosive; see De Vriend et al., 1989). Since this level, as well as the wave conditions on the flats, are important factors to the ecosystem, the ability to adequately predict the waves in the intertidal zone will be a requirement in many

applications. This means that the model has to include wave generation. Depending on the situation (exposure to sea waves, fetch inside the basin), it may also be necessary for the wave model to describe sea waves and locally generated waves together, i.e. to allow for a double-peaked spectrum.



<sup>8.3.3</sup> Aspects of current modelling

As sediment transport responds nonlinearly to the current velocity, the net transport through an inlet is quite sensitive to asymmetries in the tidal velocity and stage curves (Fig. 8.4). One type of asymmetry concerns higher harmonics (overtides), which can be due to shallow-water effects, or to asymmetric flooding and drying of the intertidal zone (hypsometry-effect; cf. Pethick, 1980; Speer and Aubrey, 1985; Friedrichs and Aubrey, 1994). Another type concerns tidal rectification (subharmonics, residual currents), for instance due to inertial effects (ebb-jet), coriolis effects (ebb- and flood-branches of a channel), and topography-induced rectification (Bakker and De Vriend, 1995).

The latter phenomenon is related to the phase coupling between velocity and water level, due to which the residual current above low water tends to be flood-dominated and that below low water ebb-dominated. As the transport of sand through the channels is determined mainly by the near-bed velocity, this mechanism tends to export sand through the channels.

Also storm surges can give rise to important asymmetries of either type. Although current measurements under storm-surge conditions are scarce, observations by fishermen suggest that surge-induced currents through the inlet can be so strong, that the tide does not even turn. Bathymetric surveys before and after such an event show large morphological changes, but if the inlet is sufficiently stable, these effects tend to be annihilated within a few months (Niemeyer, 1994, private communication). If the inlet is less stable, a major event can bring it into an essentially different state from which it does not return.





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The relative importance of these various sources of asymmetry depends on the situation and the time-scale which is considered. Van de Kreeke and Robaczewska (1993) show for the tide-dominated Ems-Dollard estuary that the contribution of the residual current to the long-term residual transport is even larger than that of the overtides.

A carefully calibrated 2-D depth-integrated model for combined tidal and wave-driven currents generally suffices to describe these phenomena, provided that it contains a flooding and drying procedure which takes due account of the hydrodynamics involved (i.e. not only based on mass conservation). In formulae, such a model reads

$$\frac{DU}{Dt} - f_c V h^2 = -gh \frac{\partial \zeta}{\partial x} + \frac{\tau_{wx}}{\rho} + \frac{F_x}{\rho} - \frac{\tau_{bx}}{\rho} + HDT$$

$$\frac{DV}{Dt} + f_c U = -gh \frac{\partial \zeta}{\partial y} + \frac{\tau_{wy}}{\rho} + \frac{F_y}{\rho} - \frac{\tau_{by}}{\rho} + HDT$$

$$\frac{\partial \zeta}{\partial t} + \frac{\partial}{\partial x} \left( Uh + \frac{M_x}{\rho} \right) + \frac{\partial}{\partial y} \left( Vh + \frac{M_y}{\rho} \right) = 0$$

in which t = time,

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x,y =	horizontal cartesian co-ordinates,
<i>U,V</i> =	x- and y-component of the depth-averaged velocity,
h	<ul> <li>water depth below MSL,</li> </ul>
ζ	<ul> <li>water surface elevation above MSL,</li> </ul>
D/Dt=	material derivative, moving along with the depth-averaged flow,
$f_c$	= coriolis parameter,
g	= acceleration due to gravity,
ρ	<ul> <li>mass density of the fluid,</li> </ul>
$\tau_{wx}, \tau_{wy}$	<ul> <li>x- and y-component of the wind shear stress,</li> </ul>
$F_{x}, F_{y}$	= x- and y-component of the wave-induced effective stress,
$\tau_{bx}, \tau_{by}$	<ul> <li>x- and y-component of the bottom shear stress,</li> </ul>
$M_{x}, \dot{M_{y}}$	= x- and y-component of the wave-induced mass flux.
•	

Note that the wave effects are included via the wave-induced effective stress, the bottom shear stress and the wave-induced mass flux. The latter usually has a rather small contribution to the total velocity. It has to be included, however, if the wave-induced undertow is taken into account, e.g. via a 2½-D approach (further see below).

Various types of 3-D currents are common in tidal inlets. Channel curvature and coriolis effects give rise to secondary circulations in the vertical, and also wave-induced mass flux and undertow play a part. These secondary flow components usually have a relatively weak contribution to the instantaneous sediment transport, but some of them are not changing sign during the tide, whence their relative contribution to the residual transport can be significant. Moreover, if these residual transport components have to be compensated by the downslope gravitational transport, their morphological impact can be large. An example of such a case is the formation of a point bar near the inner bank of a curved channel, to compensate for the net centripetal transport component due to the curvature-induced secondary flow.



Figure 8.5: MTM-model of the Frisian Inlet Wang et al. (1991) investigate the role of secondary flow effects in a medium-term morphodynamic model (tide only) of the inlet "Het Friesche Zeegat" (Frisian Inlet) in the Dutch Wadden Sea (see Fig. 8.5). They find that secondary flows, curvature-induced as well as coriolis-induced, cannot be disregarded in such a situation. The morphological evolution near the tip of the updrift barrier island (Ameland), for instance, is predicted essentially wrong if the curvature-induced secondary flow is disregarded. Also channel migration is likely to be influenced by secondary flow effects.

A rather efficient way to include these effects is a "2½-D" approach, in which the results of a 2-D depth-averaged flow computation are used in 1-D vertical profile models for the main velocity and the secondary flow (Kalkwijk and Booij, 1986). Depending on the input needed for the sediment transport module, this yields the total bottom shear stress or the total near-bed velocity as a function of space and time. This approach only works if the secondary flow velocity is much weaker than the main velocity. A more elegant and robust version is the "quasi-3D" approach, in which the 1-D profile models and the 2-D depthaveraged model are intrinsically coupled (see Zitman, 1992, for tidal currents; also see De Vriend and Stive, 1987, for wave-induced currents).). Examples of 2½-D and quasi-3D model applications to tidal inlets are described by Wang et al. (1991, 1995), for the basin and the gorge, and De Vriend and Ribberink (1988), for the outer delta.

#### 8.3.4 Aspects of sediment transport modelling

Sediment transport can occur in various modes, roughly divided into bed load and suspended load. Sediment grains which are transported in the bed-load mode are in more or less continuous contact with the bed (sliding, rolling, jumping, fluidized), and they often move intermittently, between periods of being trapped in bed forms. In the suspended-load mode, the grains are held in suspension by turbulent diffusion.

Modellers usually make a slightly different distinction, viz. between transport formulae and advection/diffusion models. Each of the largely empirical transport formulae relates the transport rate (bed load and some forms of suspended load) in a point to local parameters such as near-bed velocity, bed shear stress, critical shear stress, grain size, etc. Horikawa (1988) and Van Rijn (1989) give reviews of transport formulae for fluvial, estuarine and coastal applications.

In the case of a horizontal bed, the transport direction is assumed to coincide with the direction of the bed shear stress. If the bed is sloping, a downslope gravitational transport component deflects the transport direction. This component usually forms only a small part of the total transport. Nevertheless it can have strong morphological effects, as stated before. Even if this is not the case, it is worthwhile to take this component into account in morphodynamic models, since it represents a natural smoothing effect on bottom perturbations (cf. De Vriend et al., 1993b).

The current-induced transport rate according to a transport formula is an odd nonlinear function of the near-bed current velocity. Thus a periodic velocity can give rise to a residual transport if there are higher harmonics. This residual transport has very little to do with the residual current velocity, it may even be in the opposite direction (Fig. 8.4). If in a point all tidal constituents are given, and the M<sub>2</sub>-component is predominant, it can be shown that only M-constituents contribute significantly to the long-term residual transport (Van de Kreeke and Robaczewska, 1993). All other constituents have a frequency which differs from an exact multiple of the M<sub>2</sub>-frequency, whence their interaction with the M<sub>2</sub>-constituent leads to time-modulations and a zero mean contribution in the long run.

Yet, it is not sufficient in a tidal inlet model to impose the net current, the predominant tidal constituent and its overtides at the open boundaries. Due to the complex flooding and drying topography, each constituent generates residual currents which do contribute to the residual transport.

Advection/diffusion models for suspended load transport usually concern the sediment concentration. Their basic form reads

$$\frac{\partial c}{\partial t} + u \cdot (\nabla c) - w_s \cdot (\nabla c) = \nabla \cdot (D \nabla c)$$

4

5

in which c = sediment concentration, by weight or by volume,

- $\nabla$  = gradient vector,
- u = current velocity vector,
- $w_s$  = settling velocity vector,
- D = turbulent diffusion coefficient.

These models reflect that in the horizontally uniform equilibrium state (zero gradients in time and horizontal space) the upward diffusive flux is balanced by the downward settling flux. Away from the equilibrium state, an observer who moves along with the current sees a tendency of the concentration to adjust to the equilibrium state. This means that the transport has a non-local character in space and time: the concentration in a given point depends on its history further upstream. The characteristic length and time scales involved in these lag effects can be estimated by (cf. Wang, 1992; Katopodi and Ribberink, 1992)

$$L_c = O\left(\frac{|u|h}{w_s}\right) \quad and \quad T_c = O\left(\frac{h}{w_s}\right)$$

in which  $L_c$  = characteristic decay-length,  $T_c$  = characteristic decay-time, h = water depth, |u| = a representative measure of the current velocity,  $w_s$  = magnitude of the settling velocity.

The characteristic time scale for sand with a settling velocity in the order of 10<sup>-2</sup> m/s is usually less than 15 min., so much smaller than the the tidal period. The corresponding length scale is of the order of a kilometre, which is not small relative to the dimensions of channels and shoals. Spatial lag-effects should therefore be taken into account a suspended-sediment model for such a case.

This spatial lag-effect may explain how the tide can build up a shoal (Fig. 8.6). When the flats are flooded, sediment-laden water from the channels moves up the shoals, where the velocity drops and the sediment tends to settle. Due to the lag-effect, the sediment is brought further up the flats. When the tide turns and the flats are drying, the water which comes from the flats contains little sediment when it reaches the channels. There the concentration tends to grow, but due to the lag-effect less sediment is brought back into the channels during ebb then has been carried up the flats during flood. This picture reverses in the presence of waves, which are much more effective in stirring sediment on the flooded flats than in the deep channels. In that case, the ebb flow from the flats contains more sediment than the flood flow towards them and the flats are eroded. Since accretion and erosion occur at different times, flats can only be in dynamic equilibrium.



Figure 8.6: Shoal formation by tidal transport

The settling velocity of fine cohesive sediment is much smaller than that of sand. Hence the characteristic time scale of the adjustment process is of the order of magnitude of the tidal period or larger, and the length scale is of the order of the basin size or larger. An advection/diffusion model is therefore essential to describe the transport of this type of sediment. As a consequence of these large characteristic scales, the residual transport is determined by other properties of the tidal motion than in the case of sand (cf. Dronkers, 1986).

Apart from this, cohesive sediment behaves quite different from sand (for instance, see Berlamont et al., 1993, and Teisson et al., 1993). One specific complication of cohesive sediment is consolidation, which makes erodibility a function of the time elapsed since deposition. In a tidal environment, this introduces a time-hysteresis which can lead to residual deposition (cf. Fritsch et al., 1989). In order to take this into account, a model should include an administration of the time since deposition of different bottom layers.

Lag-effects in suspended sand transport are sometimes taken into acccount via the depth-averaged equation

$$\frac{\partial(\bar{hc})}{\partial t} + \gamma \frac{\partial(\bar{u}\bar{hc})}{\partial x} + \gamma \frac{\partial(\bar{v}\bar{hc})}{\partial y} = -\alpha (\bar{c} - \bar{c_e})$$

in which the overbars denote depth-averaging and

x,y = horizontal cartesian co-ordinates,

u,v = velocity components in x- and y-direction, respectively,

 $\gamma$  = ratio between the depth-avaraged sediment flux and the

product of the depth-averaged velocity and the depth-averaged concentration,  $\alpha$  = decay coefficient,

 $c_e$  = equilibrium concentration, given as a function of flow and sediment parameters.

This is a very crude approximation which does not do much justice to the physical processes involved. A quasi-3D approach yields a physically much better founded equation of the type (Wang, 1992; Katopodi and Ribberink, 1992)

$$T_{c}\frac{\partial \overline{c}}{\partial t} + \frac{L_{c}}{\left|\overline{u}\right|}\left[\overline{u}\frac{\partial \overline{c}}{\partial x} + \overline{v}\frac{\partial \overline{c}}{\partial y}\right] = \frac{\alpha}{h}(\overline{c} - \overline{c_{e}})$$

6

7

in which the decay parameters  $T_c$ ,  $L_c$  and  $\alpha$  follow from a 1D-vertical model of the sediment concentration profile. An alternative to this quasi-3D approach is to solve the

complete 3-D concentration equation (4). Van Rijn et al. (1991) describe such a model for current-borne suspended sediment.

The above models concern the wave-averaged current-borne transport. They may include the net stirring effect of waves, but they ignore typical wave-related mechanisms such as the residual wave-borne transport due to the asymmetry of the wave orbital motion, and the transport due to the wave-induced boundary layer streaming. These mechanisms play a part in wave-dominated environments like to outer delta and the adjacent island coasts. Too often, they are lumped under the heading "cross-shore transport mechanisms" and ignored as being too difficult to model.

In recent years, coastal morphological modellers have started to open this "black box", trying to simulate coastal profile evolution (cf. Roelvink and Broker, 1993). A fairly crude way to take these effects into account is to apply an intra-wave transport formula (e.g. Bailard, 1981). A more sophisticated approach is to model the suspended sediment concentration at the intra-wave level (Deigaard et al., 1986; also see Fredsoe and Deigaard, 1992). Obviously, it is not efficient to simulate tidal inlet morphodynamics at an intra-wave time scale. This problem can be overcome by running the intra-wave transport rates and directions as functions of input parameters such as wave height, wave direction, wave period, current velocity, etc. This table replaces the transport module in the inlet simulation model (cf. Broker Hedegaard et al., 1993).

# 8.4 Input schematization and definition of model runs

The natural variability of input conditions such as waves, wind stress and storm surges poses a special problem to the practical operation of process-based simulation models. Firstly, running the model with real-time input is seldomly feasible, or even impossible if it concerns the future. In the case of ISE-models, running input time-series other than the tide does not make much sense. In the cae of MTM-models, which can have long intervals between two consecutive hydrodynamic updates with the complete wave and current models, the "pace" of the computation process seldomly coincides with the frequency of variation of the input conditions.

Therefore, the randomly varying input conditions have to be replaced by representative steady conditions (waves, wind), or a representative storm surge. This input schematization leads to a manageable number of input sets, each with its own probability of occurrence. The model is run for each set, and the results are weighted according to the probability of occurrence of the input,  $p_i$ . In the case of sediment transport, the nonlinearily of the system leads to weight factors which differ from  $p_i$  (cf. De Vriend, 1994). Criteria to determine these factors are, for example, the correct representation of the longshore drift on a given part of the coast, or the correct representation of the tidal transport capacity in another point of the model domain (cf. De Vriend et al., 1993a).

Note that the method of input schematization, the model runs to be made and the weighted composition of the results of the runs to predictions are mutually coupled. If the schematization principle changes, the weights will also change.

# 8.5 Validation

Validation is a continuous process during the entire course of a modelling project. It involves not only the verification of model results against measured data, but also other activities which contribute to the user's control of the model results, such as analyses of the system's behaviour in reality and according to the model, analyses of the effects of model simplifications, execution of test-runs on simple and verifiable cases, sensitivity analyses, etc.

In the case of long-term predictions, the empirical verification (i.e. against measured data after realization of the situation for which the predictions were made) is often impossible within a reasonable time. Hindcasts may help, but they are not always

appropriate, since the model usually concerns future situations which are different from the past (otherwise, the model could just as well be replaced by a geostatistical extrapolation technique). Validation therefore depends to a large extent on other activities than empirical verification.

The possibility of empirical verification via hindcasts depends to a large extent on the quality of the available data. This goes for the wave, current and sediment transport fields and for the morphological evolutions. Process data should provide information of the wave, current or sediment transport *patterns* on the complex inlet topography. Due to the strong spatial variability, isolated point values are of little use. Remote sensing techniques can possible offer new perspectives here, also for measurements under rough conditions, when it is impossible to send out vessels. Depending on the time scale of interest, it may be more useful to have a quick RS-scan of the bed topography every month than vary accurate depth soundings every five years.

( not)

# 8.6 Example of an ISE-model application

### Problem definition

The northern tip of the isle of Texel, in the Dutch Wadden Sea (Fig. 8.7), is subject to an erosion of some 5 10<sup>5</sup> m<sup>3</sup>/yr. So far, the sand deficit has been replenished via beach nourishments, which releaves the symptoms of the problem, but does remove its causes. A historical reconstruction of the inlet evolution (Ribberink and De Vroeg, 1991) revealed that the sediment budget of the island tip is determined to a large extent by the course of the main ebb-channel, which used to changed more or less continually between NE-SW and SE-NW. When the main ebb-channel had a NE-SW orientation, large amounts of sediment-starved when the channel had a SE-NW orientation. Since in the last 50 years various engineering works have been made to protect the tip of the island, the main ebb channel has stopped changing direction and has kept a SE-NW orientation. As a consequence, the isle of Texel keeps on eroding, and large amounts of sediment keep on being deposited on the vast beach plane which forms the tip of the next island, Vlieland.



Figure 8.7: Texel case: situation

As an alternative to an everlasting beach nourishment scheme, a long groyne near the tip of Texel was proposed (Fig. 8.8). This groyne would trap the north-going longshore drift and thus cause accretion of the coast at its south side. The corresponding erosion at the north side could be dealt with by connecting the groyne with the exisiting revetment around the northernmost tip of the island. DELFT HYDRAULICS has carried out an extensive mathematical model study (Ribberink and De Vroeg, 1992) to investigate the effectiveness of various forms and locations of the groyne, the pattern and depth of scour around the tip, and the nature and extent of the effects on the inlet system.

### Model composition

In order to predict the response of the inlet system to the cut-off of a sediment input of 5  $10^5$  m<sup>3</sup>/yr, horizontally two-dimensional models are needed which describe the wave and current fields on a complex topography, including flooding and drying areas.

The wave field is assumed to be in equilibrium with the topography at any moment during the tide. The wave model has to include depth-refraction, shoaling and breaking. Since the spatial variations in wave height are strong, but not abrupt, diffraction is probably less important. Because of the strong spatial variations of the current velocity, current refraction has to be included. Wind-induced growth and friction-induced decay are included in view of the large model area. In order to include some natural smoothing of wave-height variations, directional spreading is taken into account. The wave model which was used in this study is based on the HISWA-package (Holthuysen et al., 1989) and meets these requirements.



This wave model is run at 20 tidal stages, which are chosen such that they provide a maximum of information on the time-distribution of the wave properties. The water level and the tidal current velocity at each stage are derived from a larger-scale tidal model which covers the entire Dutch Wadden Sea and part of the North Sea and which is run without taking wave effects into account.

The current field is computed with a 2-D depth-integrated current model based on TRISULA-package (Stelling and Leendertse, 1992). It is basically a shallow-water tidal model, extended with (see Eqs. 1 through 3)

the wave-induced effective stresses, computed from the rate of wave-energy dissipation (Dingemans et al., 1987) via

$$F = D \frac{k}{\omega}$$

in which

- wave-induced effective stress vector [N/m<sup>2</sup>], F =
  - wave-energy dissipation rate [N/ms], D Ξ
  - k = wave number vector [m<sup>-1</sup>],
  - peak frequency of the wave field [s<sup>-1</sup>], ω =
- wave-induced enhancement of the bed shear stress, according to a parameterized version of Bijker's (1966) bed shear stress model with  $\xi = 1$  (cf. Soulsby et al., 1993).

The vertical structure of the wave-induced current (mass-flux compensation, undertow, boundary layer streaming), as well as the mass flux in the depth-averaged equation of continuity, is ignored because cross-shore transport mechanisms are also ignored. A first estimate of the importance of curvature-induced secondary flows led to the conclusion that these can be left out of consideration.

Estimates of the relevant sediment transport properties revealed that it must be possible to use a transport formula here. This formula should apply to situations with combined waves and currents. For reasons of feasibility, "cross-shore" transport mechanisms due to wave asymmetry, undertow, streaming, etc. are left out of consideration. The remaining current-borne transport (the waves and the current stir the sediment, the current transports it) is modelled with Bijker's (1971) formula. This formula has to be applied in each run at each of the 20 tidal stages and in each point of the computational grid. Clearly, this requires an efficient algorithm, which is achieved by parameterizing the formula.

8

Model domain



The model domain is outlined in Fig. 8.9a. Inside the basin, the boundary follows the watershed, which encloses an area of nearly 200 km<sup>2</sup>. The island coasts form closed boundaries, which extend alongshore until "far away" from the inlet.

Figure 8.9a: Texel case: model domain. current and sediment transport models



This longshore extent is generally chosen such, that expected the changes in the inlet are not felt at the open boundaries. This means boundary that the conditions can be the same before and after of the construction As groyne. а consequence, the seaward part of the model covers an area of some 240 km<sup>2</sup>.

Figure 8.9b: Texel case: model domain wave models

The version of the HISWA-model which is utilized here works on a rectangular grid of which the orientation of the grid lines in one direction coincides more or less with the direction of wave incidence. Since various angles of wave incidence are investigated in this study, there are various wave model grids, each of which encompasses the current-model domain (Fig. 8.9b). In principle, the model works with three grids for each run, viz. an input-grid on which the bed topography is given, a computational grid, and an output-grid on which the output parameters are provided. Although the interpolations are fully automized, the variety of model domains requires extra administration, and thus it is a source of errors. Later applications of this model therefore work on the same domain as the current model.

#### Geometrical schematization

The wave computations are made on a rectangular grid, but the output can be given on any arbitrary set of points. Therefore, the sediment transport computations are made on the boundary-fitted curvilinear grid of the current model (Fig. 8.10). This grid contains 117 x 121 = 14157 points, out of which some 4200 are permanently inactive because they are located on land. The grid size varies in space, with a minimum of 40 x 125 m in the area of

interest. Criteria for the grid generation are orthogonality (required by the current model), smoothness and spatial resolution.

Resolution-demanding parts of the model are the main tidal channels and the surf zones. Via the CFL-criterion, the small grid size in the surf zones leads to small time steps (30 s, which corresponds with a CFL-number of maximally 15).

The bed topography as represented by the bathymetric maps available was translated to the curvilinear grid using a discretizer. Another possibility would be a digital terrain model, provided that the criteria for grid and topography optimization be adjusted to the demands of a tidal model.



Figure 8.10: Texel case: computated grid of the current model

#### Boundary conditions

The boundary conditions for the wave model concern the wave height and direction at the upwind boundary. The wave energy is spread over the spectral directions according to a *cos*<sup>2</sup>-function. No boundary conditions are imposed at the lateral boundaries, which are assumed to be more or less parallel to the direction of wave propagation. Thus they are treated as wave rays, across which there is no net energy flux. All energy is supposed to be dissipated before the waves reach the landward boundaries, so there is no reflection and there is no need to impose boundary conditions there. This is consistent with the refraction approximation which underlies the model.

Tidal current models of this size are driven by the conditions imposed at the open boundaries. The present model contains basically two open boundaries, one at the basin side and in the open sea. At the basin boundary and the NE part of the seaward boundary the waterlevel variation is prescribed, and at the SW and NW parts of the seaward boundary the tidal current velocity is imposed.

The wave-induced current is not driven at the boundaries, but by the effective waveinduced stress, F, inside the domain. This means that this part of the current should be disturbed as little as possible by the boundary conditions. In practice, this means that a cross-shore boundary which intersects an area with a strong wave-induced current or setup is usually treated as an axis of symmetry, with zero normal derivatives. The waveinduced contribution to the velocity or the water level distribution along such a boundary can be computed from a longshore uniform current or set-up model.

#### Representation of natural conditions

The model input has to be chosen such, that a reasonable estimate is obtained of the long-term residual transport field. Unfortunately, the relative importance of the transportgenerating mechanisms varies through the model area, so an input schematization which is adequate in one point of the model domain is not necessarily adequate in all other points.

In order to tackle this unresolved problem, the pragmatic "multiple representative wave" approach (cf. De Vriend et al., 1993a) is used. This leads to the set of representative wave conditions which in Table 7.1.

name	compass angle	H <sub>s</sub>	Τ <sub>ρ</sub>	% occurrence
westerly	255°	1.48 m	6.7 s	39.2
northerly	345°	1.33 m	6.6 s	21.4
storm	300°	4.60 m	10.0 s	1.0
tide only				38.4

Table 7.1. Representative wave conditions

The method also yields a representative tide, with an amplitude some 10% above the mean. A tide of this amplitude which had already been simulated with the ambient tidal model for the entire Wadden Sea is the astronomical tide of 21 March 1976. Therefore, this tide was chosen as the morphologically representative one.

For each situation, viz. without groyne, or with one out of various alternative forms of groyne, the number of wave-model runs is 60 (3 sectors x 20 tidal stages), and the number of current-model runs is 4 (tide only and 3 wave conditions).

#### Parameter setting and validation

Some of the modelparameters are chosen on the basis of professional judgement, others are calibrated against known properties of the system. Examples of the former kind are the coefficients in the wave-energy dissipation model, the bottom friction factor in the wave model, and Manning's n (0.027) and the eddy viscosity (2 m<sup>2</sup>/s) in the current model. Note that the latter value has little to do with the physical eddy viscosity: it mainly accounts for effects of the horizontal discretization.

Important elements of validation are the long-term residual total transport of sediment through the inlet gorge, and the gap of 5 10<sup>5</sup> m<sup>3</sup>/yr in the nearshore sediment balance near the tip of the island. Other validation material is provided by a coastline model which is run with a very detailed representation of the wave climate, and by additional experiments with the present model with a detailed representation of the wave climate (Negen, 1994). These experiments made clear that the wave schematization technique yields reasonably good estimates of the residual transport throughout the model domain.

In order to have the long-term residual transport through the gorge right, the tidal asymmetry effect had to be enhanced by amplifying the flood transport and reducing the ebb transport, both by 20%. The computed longshore drift is somewhat smaller than that computed with the coastline model. This difference is attributed to the wavle-climate schematization, and it is compensated by enhancing the weight of the transport due to westerly winds by a factor 1.7 and that due to northerly winds by a factor 2.6<sup>1</sup>.

<sup>&</sup>lt;sup>1</sup> These factors are not alarming, since the original weights are proportional to the percentage of occurrence of the wave height. In view of the the nonlinearity of the transport formula, this proportionality is not quite realistic.

## Large-scale processes without groyne

The first model application concerns the analysis of the hydrodynamic and sediment transport processes at work around the inlet. Fig. 8.11 shows a typical result, based upon the weighted sum of results from the various model runs. These results are interpreted in terms of sediment drift patterns, so at an essentially higher aggregation level than the individual transport vectors.

Fig. 8.11 shows a pronounced ebb-dominated transport in the main ebb channel. The sediment is deposited at the edge of the outer delta, from where it is removed understorm conditions. Furthermore, there are distinct flood-dominated transports near the island tips, in the case of Texel somewhat offshore. Between this flood "channel" and the coast there is a weakly developed ebb channel and a pronounced longshore drift into the inlet.



One reason why this longshore drift is so pronounced is the orientation of the coastline, which is by no means perpendicular to the predominant incoming wave direction. Another reason is the presence of the outer delta, which shelters the island tip from high northerly waves. Thus the westerly waves and the associated NE-bound longshore drift are predominant there. Once the drift enters the inlet, the tidal current takes over and brings the sediment via the ebb-dominated main channel towards the edge of the delta and from there mainly to the tip of the other island.

The storm-induced transport pattern is quite different from the weighted average one

The representative wave heights in each sector are meant to yield a good approximation of the longshore drift at a straight exposed coast. The result is a rather moderate wave height (see Table 1), and waves which only break on the beaches and on the shoals of the outer delta. Storm waves, however, break on the delta edge and drive a strong onshore and longshore transport there (Fig. 8.12). Clearly, these storm waves contribute significantly to the onshore "bulldozer effect", which counteracts the tendency of the tide to extend the delta further seawards.



### Large-scale groyne effects

A long groyne perpendicular to the shore will interrupt the longshore drift. As a consequence, sediment will pile up at the updrift side, and erosion will occur at the downdrift side. At a fully exposed coast without a tide, this process would continue until the coastline has built out so far that the amount of sediment which bypasses the groyne is equal to the undisturbed longshore drift. In the present situation, also the tidal current moves around the tip of the breakwater and forms another bypass mechanism. As a consequence, the coastline is likely to build out less far. An ISE-model is unable to predict how much less, but it gives an indication of the bypass-potential of the tidal current. Fig. 8.13 shows that this potential is significantly less than the longshore drift. This means that a substantial build-out of the coastline can be expected.

This is corroborated by the accretion/erosion patterns in Fig. 8.14. Figs. 13 and 14 also show that at the downdrift side the longshore transport picks up again, which leads to severe erosion. This effects is stronger the further the groyne lies from the island tip.



Figure 8.13: Texel case: transport around the groyne

At the scale of the inlet as a whole, Fig. 8.14 shows that the effects of the groyne are not restricted to its immediate vicinity, but that they extend all over the inlet. The nature of the effects does not become very clear, because the picture shows a sequence of erosion and deposition areas.



Figure 8.14: Texel case large-scale ISE-pattern
The latter phenomenon is typical for ISE-models. Morphological features have a strong tendency to migrate, but since ISE-models lack the morphodynamic interaction, they only show the onset of this migration, which is a coupled pair of erosion and deposition areas (Fig. 8.15). Moreover, experience with morphodynamic model applications shows that initially the model tends to remove irregularities from the given initial bed topography. Thus the topography is moulded into a generally smoother form which suits the model. From then on, a much slower and physically more realistic dynamic process takes place (cf. De Vriend et al., 1993b). Clearly, ISE-models represent the initial moulding process.



Figure 8.15: Migration of morphological features in ISE-models and MTM models

#### Local groyne effects

The principle near-field effect of the groyne is the formation of a scour hole near its tip. Again, the ISE-model results ought to be interpreted with care here. Instead of a single scour hole, Fig. 8.14 shows a pair of erosion and deposition areas, with the transition exactly in front of the groyne. This might lead to the erroneous conclusion that the most severe erosion occurs at some distance from the tip, and thus to an underestimation of its threat to the structural stability.

Fig. 8.16 explains what actually happens. Initial erosion occurs where the flow accelerates, deposition where it decelerates. In the longer run, however, the topography tends towards a state where the sediment flux through a streamlane is constant. In this situation the scour hole coincides more or less with the area of maximum velocity.



Figure 8.16: Morphological evolution around a groyne

## 8.7 ISE-modelling practice

The concept of ISE-models is rather simple, at least if adequate wave, current and sediment transport solvers are available. Yet, the above example shows that their practical operation involves points of attention, such as

- software infrastructure (steering programme, data transfer system, pre- and postprocessing facilities, etc.). Efficiency and structure are important issues here, because many runs have to be made in order to cope with the natural variability of conditions.
- model composition, i.e. which combination of process modules should be applied in order to be able to achieve the objectives of the model study. ISE models pose specific problems here, because a limited number of "snapshot" runs has to provide sufficient information on the morphological evolution to be expected. On the other hand, unnecessary refinements lead to inefficiency and sometimes even to interpre tation problems.
- geometrical schematization (model boundaries, grid definition, topographical schematization, etc.). Many choices and trade-offs have to be made here. One example is grid resolution vs. computer costs.
- input schematization, in order to have a manageable number of runs representing the whole spectrum of possible combinations of relevant input conditions (mean water level, deep-water wave height/period/direction,tidal constituents, etc.). This schematization is closely linked to the number of model runs to be made and to the derivation of quantitative predictions from the results of these runs.
- data transformation, if the various modules operate on different grids and in different domains (cf. the wave models in the above example). Unless special techniques are applied, data transformation gives rise to a loss of accuracy.
- interpretation of results. This is a particularly challenging part of ISE-modelling, as may have become clear from the above example (also see Section 9).

In practice, the main use of ISE-models lies in the analysis of residual tansport patterns under various conditions, and in the prediction of their reponse to large disturbances such as engineering works, land reclamation, sand mining, subsidence due to oil/gas-mining, etc. ISE-models are not suited to analyse the subtle balance of processes in an inlet system in (near-)equilibrium.

### 8.8 Specific aspects of MTM-modelling

MTM-models of tidal inlets will often be used to predict the effects of accelerated mean sea-level rise, enhanced subsidence due to oil and gas mining, land reclamation, sand mining, coastal nourishments, engineering structures, etc. Many of these disturbances boil down to source or sink terms in the sediment balance equation:

$$(1-\varepsilon_p)\frac{\partial z}{\partial t} + \frac{\partial S_x}{\partial x} + \frac{\partial S_y}{\partial y} = -\frac{dMSL}{dt} - \frac{dz_{ref}}{dt} + \Phi_{dep} - \Phi_{er}$$

9

in which MSL= mean sea level [m/yr],

 $z_{ref}$  = reference bed level (subsiding) [m],

 $\Phi_{er}$  = deposition flux (e.g. nourishment) [m<sup>3</sup>/m<sup>2</sup>/yr],

 $\Phi_{dep}$  = ersoion flux (e.g. mining) [m<sup>3</sup>/m<sup>2</sup>/yr].

These terms give rise to a forced response of the system, on top of the inherent behaviour which is present, anyway.

The numerical integration of the sediment balance equation may seem to be simple and

straightforward, and so it is if the transport field is known. In an MTM-model, however, the equation is included in a time-loop, due to which there is a feedback of z on S. As a consequence, the numerical problem becomes much more complicated and instability-prone. One-step explicit schemes are the most convenient in this case, and schemes of the Lax-Wendroff type turn out to be the most robust (cf. De Vriend et al., 1993b; also see Peltier et al., 1991). Like other explicit schemes, they involve a CFL-type stability criterion.

As the tidal computation is usually the most time-consuming, minimization of the number of its calls is an effective way to reduce computer costs. One rather robust technique to achieve this is the "continuity-update" (see Fig. 8.18 for its incorporation in the flow chart), which reads in its simplest form

$$(\overline{u}h)_{(+\Delta)} = (\overline{u}h)_{(+\Delta)}$$

10

More sophisticated versions make use of the full equation of continuity. Tests with horizontally 2-D morphological models have shown that very significant savings of computer time can be achieved with this technique (Latteux and Peltier, 1991; De Vriend et al., 1993b).

The above technique does not include the momentum equations, so it does not take the dynamic interaction between the current and the bed topography into account. Thus the propagation character of morphological changes is ignored. This introduces systematic errors in the solution, which increase with the number of consecutive updates with this technique. However, analyses have shown that these errors tend to disappear as soon as the full hydrodynamic model is applied again. Thus the intermittent application of the full model and a number of continuity-updatesyields acceptable results, also in the long run.

### 8.9 Examples of MTM-model applications

### Texel case

The model system described in the foregoing has recently been extended to a morphodynamic simulation system. In essence, this boils down to feeding the bed level changes from the sediment balance equation back into the hydrodynamic and sediment modules (Fig. 8.1b). This creates a time-loop which simulates the meso-scale morphodynamic behaviour of the inlet system.

Negen (1994) describes the application of this simulation technique to the Texel-case, be it with the groyne at a slightly different location. Fig. 8.17 shows a typical result, after 1 year's simulation, all the time with the westerly waves (Table 1). The results generally confirm the effect of the groyne on the coastline position. Striking differences from the ISE-model results are:

- the formation of a channel along the downdrift side of the groyne,
- the deepening of the weakly developed NE-SW ebb channel near the tip of the groyne, and shoaling of the main ebb channel where this crosses the course of the NE-SW channel,
- the onshore migration of the shoals on the outer delta.



Figure 8.17: Texel case: morphological evolution after one year, according to an MTM-simulation with westerly waves (from: Negen, 1994: courtesy DELFT HYDRAULICS)

The first point indicates that further study is needed to design a stable groyne. The second finding may indicate that the groyne tends to promote the reorientation of the main ebb channel, back to the NE-SW position which is favourable for the sediment budget of Texel. The third observation shows that the present simulation is probably not represent full reality, because the mechanisms which tend to move the shoals offshore are underestimated.

In summary, this application shows the potentials of fully dynamic simulations of tidal inlet morphodynamics. On the other hand, the present case definitely requires further attention before definitive conclusions can be drawn from the simulation.

#### Frisian Inlet case

Another, less complicated application of a morphodynamic simulation model concerns the Frisian Inlet (Wang et al., 1991, 1995; see Section 5 of the companion chapter for a description of the situation). The objectives of this study were

- to hindcast the system's response to the closure of the Lauwerszee,
- to assess residual transport patterns, and
- to investigate channel migration mechanisms,

especially in the basin and the inner part of the gorge. This explains why this study could be undertaken with a model which did not include waves.

Fig. 8.18 outlines the flow chart of this model, including a "continuity-update" technique to save computer time (see Section 8). The tidal module is essentially the same as in the Texel case, but extended to  $2\frac{1}{2}$ -D in order to take curvature- and coriolis-induced secondary flows into account. The sediment transport module is based on a  $2\frac{1}{2}$ -D advection/diffusion model for the concentration, with the equilibrium concentration derived from Van Rijn's (1984) transport formula (although other options are open). The choice of these modules was made on the basis of a number of orientating computations.

The model domain covers not only the area of interest, but also the outer delta and part of the open sea, in order to have the seaward boundary far enough away from the dynamic area to allow for the same tidal boundary conditions throughout the simulation.

The schematization of the inlet geometry turned out to be critical for the system's morphodynamic stability. If the emerging shoal in the centre of the inlet, the Engelsmanplaat, is omitted, the system becomes unstable. One might claim that the model should be able to reproduce the formation of this feature if that is absent from the initial state.

Closer investigation of the geological structure of the Engelsmanplaat (Oost and De Haas, 1993), however, revealed that it consists mainly of highly resistant pleistocene clay, i.e. it is not the result of a recent morphodynamic process. This feature therefore has to be included in the model as a permanent element.

As a consequence, the Frisian Inlet becomes a double inlet, with the Pinkegat subsystem in the west and the Zoutkamperlaag subsystem in the east. The large difference in size between these subsystems is reflected in a large difference in morphodynamic time scale. The model's morphological time step has to be based on the fastest evolving subsystem, the Pinkegat. Thus it is much smaller than necessary for the Zoutkamperlaag. Within the amount of computer time allotted to this model, the response of the Zoutkamperlaag could therefore only be simulated if the topography of the Pinkegat area was fixed. Clearly, such a drastic measure is bound to cause problems at the transition to the mobile-bed part of the model.

.



Figure 8.18: Frisian Inlet morphodynamic simulation model: flow chart

The seaward boundary conditions (tide, sediment transport) form another critical element. Disturbances due to errors or inconsistencies in these conditions are translated into bed level perturbations which tend to migrate into the model domain and sometimes even blow up. In general, a morphodynamic simulation requires one boundary condition, either to the transport or to the bed level, at every inflow boundary. This poses a special problem to tidal models, in which the open boundaries keep swapping between the inflow and the outflow type. This problem is largely unresolved, as yet, and definitely requires further attention.

The model results obtained so far concern mainly the shorter-term phenomenon of channel formation and migration, especially in the Pinkegat area. The model, as well as field observations (Oost and De Haas, 1993; Oost, 1995) show a cyclic behaviour of the channel pattern (Fig. 8.19), though at different time scales. This discrepancy is probably due to the absence of waves from the model. This leads to far too small transport rates, and hence to an overprediction of the morphological time scale. Yet, the character of the migration pattern is reproduced, which indicates that the model contains the right channel migration mechanism. Further analysis should reveal the details of this mechanism.

Another indication of the model's potential concerning channel formation follows from an experiment in which the initial bed topography was a strongly smoothed version of the measured topography. It turns out that the model tends to restore a channel/shoal system of a similar nature as before.



Figure 8.19: Frisian Inlet: channel pattern evolution, (a) initial conditions in the model, (b) predicted evolution, (C) observed cyclic evolution. (courtesy Z.B. Wang (a and b), A.P. Oost (c) )

# 8.10 MTM-modelling practice

The above examples have shown to some extent that MTM-modelling involves a number of specific complications, on top of those encountered in ISE-modelling. In brief summary:

- The software infrastructure becomes even more critical now, because the coupling of the modules has to be completely automized.
- The model domain has to be the same for the process-modules and the sediment balance module, in order to avoid discontinuities in the bed topography. In the case of nested models for waves or tide, this concerns only the smallest model domains.
- The model composition should be such, that errors cannot build up to an unacceptable level during the simulation.
- The boundary conditions need to be defined with care. In general, errors in the boundary conditions tend to be reflected in an amplified form in the bed topography, and thus propagate into the model domain (Wang et al., 1991; also see De Vriend et al., 1993b).
- The time-stepping procedure in the morphodynamic loop deserves special attention, not only because it is instability-prone, but also because there is a large amount ofcomputer time involved, whence clever time-stepping may lead to significant savings.
- In spite of these time-saving techniques, there can be situations where the morphological time scales vary strongly through the model area (cf. Wang et al., 1991, 1995). In that case, the shortest time scale determines the maximum allowable morphological time step, which may imply that large amounts of computer time are needed to cover the largest time scale of interest
- Since the simulation covers a certain period of time, it is possible to mimic the natural variation of conditions via input-scenarios for waves, water levels, tides and surges. Such a scenario is only one realization of a stochastic process, so the simulation should be repeated a number of times in order to get a feel for the variability of the output.

The same goes for verification against field observations, which are the result of 0 only one realization of the input conditions. Therefore, such a verification cannot validate the variation of the output due to different input scenarios.

In general, it is very important to clearly identify the objectives of the model study. Allpurpose simulation models are not yet available, so the model composition and set-up have to be derived from the objectives of the study, via an analysis of the relevant phenomena.

MTM-models, especially those including waves, are still in an early stage of development. Now that the numerical and software problems have largely been overcome, the focus of attention should return to the physical processes and their interactions, including the equilibrium state to which the system tends. Key issues of research are the tenability of the 2-D depth-integrated model concept and its 21/2-D or quasi-3D alternatives, and ways to deal with the natural variability of the input conditions.

### 8.11 Process-based long-term models

One way to achieve a long-term morphological model concept for tidal inlets is to formally avarage the constituting mathematical-physical equations over the space and time scales which are smaller than those of the phenomena of interest. Since these constituting equations are nonlinear, this involves closure terms which have to be modelled in terms of the large-scale dependent variables (cf. the Reynolds-stress terms in turbulent flow models, and the radiation-stress terms in wave-driven current models). This can be done empirically, or based upon a theoretical analysis of the relevant interaction processes.

Krol (1990) describes a method of formal averaging for a simple 1-D morphodynamic model of a tidal estuary. This work has been followed up by various other investigators, (Fokkink, 1992; Schuttelaars and De Swart, 1994), lately also in two dimensions (Schuttelaars, 1994, private communication).

A simple example of formal averaging concerns a narrow basin of length L and width B (cf. Van Dongeren and De Vriend, 1994). L is much smaller than the tidal wave length, and B is constant, whence it can be taken equal to 1 without loss of generality. The constituting equations are:

(1) the equation of continuity of the tidal motion

$$\frac{\partial \zeta}{\partial t} + \frac{\partial \left[ (\zeta - h) U \right]}{\partial x} = 0$$

in which

ζ

water surface elevation above MSL, =

bed level with respect to MSL (so usually h < 0), h =

дr

cross-sectional average velocity, U =

х = spatial co-ordinate along the basin.

(2) the dynamic equation for the tide, which reduces to

$$\frac{\partial \zeta}{\partial x} = 0$$
 12

11

(3) the sediment transport formula

$$S(x,t) = a \ U^b$$

in which a and b are constants.

(4) the sediment balance equation

$$\frac{\partial h}{\partial t} + \frac{\partial S}{\partial x} = 0$$
14

Integration of Eq. (11) over the length of the basin yields for  $\zeta \ll h^2$ 

$$U(x,t) \approx \frac{L-x}{-h(x,\tau)} \frac{d\zeta}{dt}$$
 15

in which  $\zeta$  denotes a time co-ordinate at the slow time scale of the morphological changes. Substitution into Eq. (13) and averaging over the tidal cycle yields

$$\overline{S}(x,\tau) = a \left[ \frac{L-x}{-h(x,\tau)} \right]^b Z_T \quad with \quad Z_T = \frac{1}{T} \int_0^T \left( \frac{d\zeta}{dt} \right)^b dt$$
16

Formal averaging of the sediment balance equation (14) over the tidal cycle yields

$$\frac{\partial h}{\partial \tau} + \frac{\partial S}{\partial t} = 0$$
 17

or, with Eq. (16),

$$\frac{\partial h}{\partial \tau} + \frac{b \overline{S}}{-h(x,\tau)} \frac{\partial h}{\partial x} = \frac{b \overline{S}}{L - x}$$
18

This equation describes the evolution of  $h(x,\tau)$  at the morphological time scale. Apparently, this equation is of the forced nonlinear advection type, with the advection speed bS/h

<sup>&</sup>lt;sup>2</sup> This assumption is made for explanatory reasons. Clearly, it does not hold in the shallow parts of the basin. Yet the resulting linear distribution of  $h_e(x)$  turns out to be a reasonable approximation (Schuttelaars and De Swart, 1994).

which is found in many other morphological systems with a transport formula of the same type as in Eq. (13). Note that for b > 1 the source term vanishes towards the back end of the basin. This means that the evolution process proceeds ever slower as x approaches L.

 $\frac{\partial h_e}{\partial x} = -\frac{h_e}{L-x}$  whence  $h_e = L - x$ 

19

The equilibrium state which follows from Eq. (18) is given by





As long as no part of the basin falls dry at low water, the tidal prism at any point x is also proportional to L-x. Hence the equilibrium cross-sectional area in this case is proportional to the tidal prosm. This corresponds with O'Brien's (1931, 1969) empirical findings.

Fig. 8.20 shows the result of a fully non-linear analytical model of the morphological evolution in a dead-end tidal channel without flats (Schuttelaars and De Swart, 1994). It shows that the above simplified linear model is a reasonable approximation. Note that the evolution process becomes infinitely slow towards the end of the channel, because the current velocity and the transport rate go to zero there (also see Van Dongeren and De Vriend, 1994).

The derivation of the above linearized model shows the essence of many long-term morphological model concepts, viz. to express the residual transport, and hence the timederivative of the bed level, in terms of the bed elevation. The result is an evolution equation in h. The same approach is taken by Bakker (1995) to derive a bed evolution equation for a tidal channel in a network, i.e. with two open ends. Bakker and De Vriend (1995) use the result in a network model of a near-resonant basin. One of the research questions which they address is to what extent a small interference at one place in the basin can have large effects elsewhere.

This theoretical approach of tidal inlet morphodynamics is quite complex on the one hand, and quite powerful on the other. Analyses of the mathematical system, once it has been formulated, are able reveal the inherent morphodynamic behaviour, such as channel

pattern formation and stability of channel/shoal systems. This information can not only be of use for the validation of numerical models, it may even indicate to what extent numerical models will ever be able to handle these nonlinear phenomena.

### 8.12 Conclusion

Computer-intensive multi-dimensional model types are now available for application to tidal inlets. ISE-model systems have reached a more or less operational stage (Latteux and Peltier, 1992; Broker Hedegaard et al., 1993; Roelvink et al., 1994; Chesher et al., 1993) and have proven their use to describe physical processes, such as residual transport patterns, under a variety of input conditions. This information can be used to analyse how waves, tide, currents and sediment transport influence each other in an area as complex as a tidal inlet, and to estimate the impact of human interference in the system (engineering structures, land reclamation, sand mining, etc.).

Process-based morphodynamic simulation models are still in an earlier stage of development. It is technically possible to run such models and pilot applications are being made, but the stage of routine practical application has not yet been reached. This requires further investments into a more thorough understanding of the processes and mechanisms at work, and a further build-up of know-how, experience and confidence in these models.

One objective of further research should be a better insight into the relevance of phenomena at the process level to larger-scale morphological behaviour in a given situation. Thus it must become clearer what physics should be included in the constituent models in order to simulate these evolutions.

Another research issue is how to deal with the natural variability of the conditions which drive a tidal inlet model, and how to interpret predictions which result from an assumed future input scenario. Conversely, it should become clear which aspects of the input time series are important to the morphological evolution of the system.

Finally, the formal averaging approach and the correponding theoretical analyses should be pursued further, if it were only for the large amount of insight which they provide.

At the development level, knowledge and tools for the adequate selection of model components would be welcome. Also, the insight into accuracy and reliability of these compound dynamic models should be improved, in conjunction with practical methods to deal with the limited predictability of inputs.

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