Approaches to long-term modelling of coastal morphology: a review

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ABSTRACT

Aspects of long-term mathematical modelling of coastal morphology are inventoried and discussed. They concern reduction techniques for input data, process descriptions and output data, as well as model concepts ranging from statistical extrapolation of the past coastal behaviour, via semi-empirical behaviour models, to formally integrated descriptions of the constituent small-scale processes.

All approaches have in common that they reduce the need for detailed descriptions in space and time of the underlying physical processes. They lead not only to more transparent and robust models which require less computational effort, but also to a better insight into which aspects of coastal behaviour are relevant from a long-term point of view and which are not ("signal" vs. "noise").

INTRODUCTION

Long-term prediction of sediment transport and morphological behaviour in the coastal zone, in response to human interference or to changing environmental conditions, is an increasingly important issue, now that we are becoming aware of the need for sustainable development and are facing threats such as accelerated sea level rise and climatic changes.

These predictions concern the behaviour of the coastline and its immediate surroundings, but also the evolution of more complex systems, such as estuaries, inlets and coastal lagoons. Besides, it is not only the topography of the bed that is important, but also the water and bottom quality (transport, storage and release of pollutants and nutrients which adhere to the sediment) and, more generally, the ecosystem.

Although some existing model concepts, such as coastline models, enable us to cover significant time spans, even at historical and geological scales (cf.

De Vroeg et al., 1988), long-term modelling of coastal behaviour is still in its infancy. Most of the physical understanding and the corresponding models concern processes at time scales which are basically hydrodynamical, i.e. orders of magnitude smaller than the scale of the phenomena in which we are interested.

Even if we would have enough computing power to run such small-scale models for a sufficiently long period of time, this might not be the best approach of long-term modelling. Since the natural system is nonlinear, its long-term behaviour may even be inherently unpredictable, and if this is not the case, the validity of long-term model predictions does not follow straightforwardly from the model's validity at the hydrodynamic time scale. Processes which are negligible at this smaller scale can have significant long-term effects and conversely. Our knowledge on which short-term processes are important in the long run is still insufficient.

On the other hand, the problem of nonlinear multiple-scale dynamics is not unique. It shows up in many areas of applied physics, such as meteorology, oceanography, or structural dynamics, and a good deal of pioneering research has already been done there (for instance, see Pedlosky, 1987). As far as morphodynamics is concerned, river dynamics is leading the way (Ikeda and Parker, 1989), and we can profit from this in the coastal field.

Some work has been done on long-term coastal modelling, but it is rather scattered and eclectic. The biggest and most consistent body of work concerns the coastline models (for instance, see Hanson, 1987), though too often they are utilised in a purely deterministic way, without bothering about the statistics and the reliability of the prediction. In fact, this way of utilizing this tool should be restricted to short-term applications.

Another important body of work concerns the equilibrium state of the system, given the extrinsic conditions. Examples are the equilibrium coastal profile (Dean, 1991) and the empirical equilibrium relationships for tidal inlets (O'Brien, 1969; Bruun and Gerritsen, 1960). Although they provide indispensable information, these concepts are lacking the dynamics of the coastal evolution, and it would certainly be an oversimplification to assume that the transition from an initial state to an equilibrium state is always described by an exponential decay of the difference.

The need for long-term modelling includes cohesive sediments. As compared with non-cohesive sediment, this involves additional complications. Consolidation, for instance, introduces another, quite variable time scale and a hysteresis effect which puts much more emphasis on the chronology of inputs (cf. Villaret and Latteux, 1992).

In the present paper we will not only describe the state-of-the-art in long-term coastal morphodynamic and sediment transport modelling, but we will also formulate a strategy for further development, and attempt to trigger discussions and new ideas on this issue.

DATA REDUCTION

Data play an important role in various phases of long-term modelling. Firstly, we need observations to trigger us to the phenomena and the mechanisms of interest, and to help us formulate working hypotheses and model concepts. After these have been elaborated to a prototype of the model, data are needed for validation. Once we have a validated model system, we need data for calibration and verification in every new application. Finally, we can update operational models by assimilating new data.

Since coastal behaviour is a highly stochastic process in one or two horizontal dimensions and time, with a wide range of space and time scales involved, the data are usually difficult to grasp. To that end, we need data reduction techniques which enable us to separate the relevant information from the "noise" and to map this onto a manageable number of parameters.

The same goes, to a certain extent, for model results, which usually represents one realisation of the stochastic process. Mapping these results and the corresponding data onto a small number of characteristic parameters facilitates their interpretation and verification.

In the following we will briefly indicate some of the data reduction techniques which may be of use to coastal dynamics.

Correlation

The principle of correlation originates from statistics. If two data fields, e.g. one from a model and another from field measurements, are identical, their correlation index is one, and the more they differ, the smaller the index.

Correlation is therefore an extreme form of data reduction: all information on the difference between the data fields is stored in a single index. This technique is widely used in various forms, but its application to coastal morphology requires special attention, because we are dealing with stochastic inputs. Hence uncertainty is part of the system, and even a perfect model and perfect data can be different, because they refer to different realisations of the same process. The correlation index should therefore be defined in a statistically averaged sense.

Statistical analysis

The temporal and spatial behaviour of the coast includes a variety of scales. Measurements of this behaviour usually consist of series of discrete samples, often concise, at different intervals and from different sources. Parameters which characterize the coastal evolution, e.g. the change in position of a depth contour, have to be estimated on the basis of these data.

Dolan et al. (1991a) describe various estimation methods for the rate of

change of the shoreline and indicate how, with sufficient temporal sampling, long-term (e.g. sea level rise) and short-term (e.g. storm or seasonal) trends can be separated and how the results are affected by the various sources of errors. The separation of time scales turns out to depend critically on the rate of interdependence of the underlying processes.

In a second paper, Dolan et al. (1991b) discuss the optimization of spatial sampling, which is important when collecting large volumes of data. Too few data lead to inconclusive results and oversampling is inefficient.

Empirical orthogonal functions

The basic idea of the Empirical Orthogonal Function (EOF) technique is to map a data set onto an orthogonal set of spatial functions, which are derived from the eigenfunctions of the covariance matrix (Preisendorfer and Mobley, 1988). Each spatial function has its own time-evolution function and contains a certain part of the variance of the data, indicated by the corresponding eigenvalue. If we order the eigenvalues according to their magnitude, taking the largest one first, the summation of the first n contributions (the product of the n th spatial eigenfunction, its time function and the square root of the corresponding eigenvalue) can be shown to give the best fit to the data, in the least squares sense, of any combination of n contributions. Hence we are reasonably safe to use truncated series of contributions, which enables us to approximate the whole data set by a limited number of space functions, time functions and multiplication factors.

Winant et al. (1975) were the first to apply the EOF technique to coastal profile evolution. This has become common practice now. Hsu et al. (1986) and Medina et al. (1992) describe extensions to three dimensions, with spatial eigenfunctions in longshore and cross-shore directions.

A disadvantage of real eigenfunctions is that they are fixed in space, whence they are less suited for propagating features such as bars. The extension to complex eigenfunctions (for instance, see Liang et al., 1992) provides better possibilities here.

A typical example of an EOF analysis is shown in Fig. 1. It concerns a section of the Holland coast, as monitored once per year from 1964 to 1990, and shows the first few spatial and temporal eigenfunctions. The weighted sum of products of corresponding spatial and temporal functions gives an approximation of the observed profile evolution.

The amplitudes of the EOF's can also be used to extract information on the attractor properties of the underlying physical system (Preisendorfer et al., 1988).

Fig. 1. Decomposition of a coastal profile evolution into spatial and temporal EOF's (K. Wijnberg, University of Utrecht, private communication).

Contour analysis

The amount of work involved in the above data reduction techniques increases significantly if we step from one to two horizontal dimensions. Besides, the amount of information is large, even after the reduction, and the interpretation of results is correspondingly difficult.

On the other hand, when looking at the time evolution of a 3D sea bed topography, we often focus on certain features and try to follow them through time. In other words: we try to recognize certain topographical patterns and trace their time evolution. Image processing and pattern recognition are issues in various other disciplines, and the necessary techniques have been developed and implemented there. Morphodynamics can profit from that work.

One approach which complies well with the common practice in morphology is the analysis of bed level contours. Starting from a collection of elementary patterns, contour analysis enables us to assess their presence and their location in the bottom contour map, to characterize them in terms of invariant properties (Voss and Rothe, 1992) and to follow them through a sequence of contour maps, even if they evolve in time (Persoon, 1986). Thus a large amount of data can be reduced to a few pattern parameters. A hypothetical case is shown in Fig. 2. The elementary forms ("hump", "sine wave", "golfclub") which constitute the left part of this figure are still present in the right part, but each has evolved in number, size, position and shape. As the basic elements remain the same, the changes in these few parameters characterize the entire evolution from the left to the right pattern. Thus a significant data reduction is achieved.

Fig. 2. Displacement and evolution of a composition of three elementary contour patterns (hump, sine-wave, "golfclub").

LONG-TERM MODELLING APPROACH

Coastal behaviour is the result of a large number of processes and mechanisms which act and interact on a variety of space and time scales. The notion of scales is therefore a very important element in long-term modelling.

One could argue that the phenomena of interest (e.g. bar formation), if predictable, are likely to be related to aspects of the underlying physical processes (waves, currents, sediment transport) on similar space and time scales. The variation of inputs and processes on much smaller scales is only relevant as far as there are residual effects, whereas variations on much larger scales can be considered as concerning the extrinsic conditions (see Fig. 3; also see De Vriend, 1991).

This is the justification of long-term model concepts which neither go into details of small-scale processes, nor extend to scales far beyond those of interest. It also gives access to powerful mathematical techniques based on scale separation and perturbation methods.

One of the key elements in long-term modelling is reduction of information. It involves essentially four levels (see Fig. 4), which concern the input, the physical system or its model, the output and the interpretation or generalization. They are reflected in the various approaches to long-term modelling, viz.:

- input reduction, based on the idea that we can describe long-term residual effects (e.g. transport fields) with models based on the description of small-scale processes, if we can find the representative inputs to drive them;
- model reduction, based on the idea that, by using more or less formal anal-

Fig. 3. Scale interactions in coastal morphodynamics.

Fig. 4. Aggregate flow chart of long-term prediction process.

ysis and integration methods, the model can reformulated at the scale of interest without describing the details of smaller-scale effects;

- behaviour-oriented modelling, which attempts to describe the phenomena without going into the underlying processes.

Each of these approaches will be discussed in the following sections.

INPUT REDUCTION

Tidal input filtering

Although the astronomical tide is deterministic and can be predicted accurately, it exhibits significant longer-term modulation cycles (spring/neap, year, nodal cycle), which make the real-time representation of the tidal motion in a long-term model prohibitively expensive. Therefore, we need to find a way to represent the tidal motion by a small number of unmodulated tides.

The idea of tidal input filtering is to take a complete tidal record over the period of interest and compute the mean absolute sediment transport (or any other quantity which can be used as a reference for schematized computations) in a number of salient points of the model domain. Subsequently, a small number of representative tides (=time functions covering one tidal cycle) is selected, such that in these points the reference quantity is reproduced.

We have performed various model tests, for schematic as well as real-life cases, in order to gather insight into and experience with this tidal input filtering. The following questions were addressed:

- to what extent does the concept work?
- what is a practicable number of representative unmodulated tides (one? two? more?)?
- once this number has been chosen, how to find the proper representative tide(s) and, if necessary, the corresponding weight factors?

- what are reasonable values of the parameters which characterize the representative tides?

If we stick to a single representative tide (sometimes called “morphological tide”), we can require that this should reproduce the long-term average sediment transport in a number of salient points in the model domain. Given

- the long-term variation of the tidal range,
- the basic shape of the tidal curve,
- the relationship between tidal range and tidal current velocity, and
- the sediment transport rate as a function of the hydrodynamic parameters (velocity, water depth),

we can compute the transport rate per tidal cycle and the corresponding long-term average due to the actual tide.

The cycle yielding a transport per unit time which is equal to the corresponding long-term mean transport due the actual tide is taken as the representative one.

This approach was elaborated by Latteux (1987) for tides only and by Steijn (1989, 1992) for tides in combination with waves. They both find that the representative tide is slightly (7 to 20%) higher than the mean tide, depending on the sediment transport formula and the threshold for sediment motion. The higher the exponent in the formula, and the higher the threshold velocity relative to the tidal velocities, the higher the representative tide relative to the mean tide.

Clearly, this approach rests upon a number of more or less implicit assumptions, such as:

- the tidal range and the shape of the tidal curve are the most important properties of the tide from a morphological point of view, i.e. aspects such as ellipticity or phase are less important;
 - the reference quantity characterizes all morphologically relevant properties of the tide;
 - the representative tide reproduces both the intensity and the direction of the long-term mean transport;
 - the shape of the tidal curve (as opposed to the tidal range) is more or less uniform over the area of interest, such that the selected tide is representative for the whole area, not only for a single point;
 - tidal currents are the only cause of sediment motion, or, if waves are also important, tidal and wave climate schematizations can be treated separately.
- In the following we will summarize some investigations into the applicability of a single representative tide and the validity of these assumptions, for a number of hypothetical and real-life situations.

Real-life tests

Latteux (1987) investigated the applicability of a single representative tide for single points at three different sites in France (see Fig. 5), viz.

Fig. 5. Tidal input filtering for three isolated points: locations and observed tidal currents.

- (1) the Gulf of St. Malo (C), where the tidal ellipse is almost circular and the currents are always rotating,
- (2) the coast of Pays de Caux (B), in the eastern part of the English Channel, where the tidal ellipse is flat and the currents are always parallel to the coast, and
- (3) near Calais (A), in the southernmost part of the North Sea, where the currents are influenced by a complex topography.

Measured current data from these sites were put into a sediment transport formula and the flood total, ebb total and tidal residual transports were computed for each tidal range. Subsequently, these results for all tidal ranges were combined, to yield reference values for the yearly-averaged flood, ebb and residual transports. Representative tides were selected for each of these reference transports and applied to the same site. Various aspects of the resulting yearly-averaged transport (e.g. magnitude, direction) were compared with the corresponding reference quantities. This led to the following conclusions:

- If the topography is not too complex, a single representative tide can be found in every point in the area of interest. Whether this tide is more or less the same for all points remains to be investigated.
- If the topography leads to irregular variations of the current velocity, a sin-

gle representative tide does not make sense, because the topography will also affect the shape of the tidal curve and the residual current field. In that case, the effects of the variation of the tide have to be reproduced by a number of representative tides, one for each class of tidal ranges. In morphological computations, the results for each of these tides have to be weighted and summed up.

Hypothetical test case

In addition to these single-point tests, we have investigated a schematic 2-D case of an estuary on a straight coast (Latteux, 1992). This is a severe test, because the current patterns change significantly with the tidal range, as a consequence of the shape of the domain.

The long-term (19-yearly) cyclic variation of the tidal range was divided into 20 classes, each with its own frequency of occurrence. For each class, tidal and transport computations over one tidal cycle were performed. The resulting changes of the bed topography were then summed up, each weighted with the frequency of occurrence of the range class. This yielded the 1-year reference bed evolution (Fig. 6a).

Various tidal schematizations were tested against this result. For the transport formula which was used here, the single tide for which the topographic changes are closest to this reference leads to current velocities about 12% above those due to mean tide. If the relative error in the bed level changes is defined as the mean square deviation of the bed level changes from those for the reference tide, divided by mean square of the bed level changes for the reference tide, this quantity amounts to some 6% (also see Fig. 6b). If we use two representative tides, one near neap and one near spring tide, the error goes down to 1.5%, but if we take three tides (neap, mean and spring), the error hardly goes further down.

Another way of tidal schematization is to multiply the bed evolution due to a single tidal cycle by a factor, such that the yearly bed evolution, averaged over the domain, is reproduced. This approach was tested in the above hypothetical case for three typical tides, mean neap, mean and mean spring tide. The multiplication factor turned out to be 12 years, 1.7 years and 5 months, respectively. The corresponding relative errors in the bed level change were 24%, 11% and 3%, respectively. The latter value is much lower than that for a single "morphological tide" (see the previous section). Apparently, the tide which provides the most accurate transport rate is not necessarily the most accurate in terms of morphological evolution patterns. This difference must be attributed to the morphodynamic interactions during the simulation period, i.e. the feedback effects of the topographic changes on the water and sediment motion.

The above approach can be based on a single tide if

Fig. 6. Tidal input filtering for area model tests (Latteux, 1992). (a) Bed topography changes in one year, according to the reference computation; (b) errors in bed topography changes when computed with a single representative tide.

(a) it produces the right direction of the residual transport, and
 (b) the fraction of the transport intensity which it produces is approximately constant across the model domain.

If these conditions are not met, two or more tides have to be used.

In summary, we can state that automatic procedures for tidal input filtering are not yet available. Much depends on the specific properties of the tide in the area of interest. The availability of sufficient data on the tidal velocities in that area is therefore of vital importance.

Wave input filtering

The overall wave conditions in an area vary randomly with time (occurrence of calms and storms, overall mean wave height, period and direction). Moreover, the local sea state for a given overall condition can be more or less irregular (local variation of wave height, period and direction).

At the scale of the morphological evolutions of interest, overall randomness has to be taken into account via the boundary conditions, local irregularity should be included in the wave model concept. In this context, wave input filtering concerns only overall randomness.

The objective of wave input filtering is to represent the wave climate by a small number of representative conditions, with which we can drive the wave module in our deterministic model system in order to estimate long-term mean transports and bed evolutions.

So far, two approaches have been developed, which we shall call the multiple and the single representative wave approach (MRW and SRW), respectively.

The MRW approach

The MRW approach is described extensively in Steijn (1989, 1992). It starts from measured wave heights and directions (wave or wind data) and reduces those to a limited number of combinations of wave height and direction inputs at the model boundary, plus the weight factor to be applied to the results for each of them in the calculation of long-term mean transports and bed evolutions. Fig. 7a gives a schematic representation of this idea.

The criteria which are applied to determine these representative conditions and the corresponding weight factors refer to the two most prominent effects of waves on the transport:

- the bulk longshore drift, for which the longshore power, P_L , is a useful yardstick (SPM, 1984), and
- the stirring of sediment, which is then carried along with the ambient current; here we use the “stirring parameter”, τ_w , as a yardstick.

Being a surf zone parameter, P_L can be evaluated using the shallow water approximation, to yield a proportionality with H_s and $\sin 2\phi$ in the point under consideration, in which H_s is the significant wave height and ϕ is the angle of wave incidence with respect to the shore normal. The stirring parameter actually indicates the magnitude of the wave contribution to the bottom friction.

Fig. 7. Multiple (a) and single (b) representative wave approach (ISE=initial sedimentation/erosion).

Both the longshore power and the stirring parameter have to be reproduced by the set of representative conditions. This means that we have to meet two requirements simultaneously, for which we need two degrees of freedom. Steijn (1989, 1992) suggests a procedure in which the direction sectors are preset, whilst the representative wave height per sector and the weight factor in the transport computation are the degrees of freedom.

This procedure, which was tested in various practical cases, works reasonably well in rather simple situations with not too wide a variety of transport mechanisms.

The SRW approach

Chesher and Miles (1992) describe a different approach, in which the results of wave computations for a number of sectors are combined, to yield a single set of representative wave parameters which is put into the flow and sediment transport modules. Fig. 7b gives a flow chart of this "single representative wave" approach. Clearly, this would lead to a considerable saving in time and computational expense, since the flow and sand transport models would need to be run once only. Therefore, the approach is worth investigating.

As was stated before, waves affect sediment transport in two main ways: they generate and affect currents and they stir up the sediment which is sub-

Fig. 8. Poole Bay test results, with the multiple (a) and the single (b) representative wave approach.

sequently advected away by the flow. The latter mechanism is exploited in the SRW concept, since the stirring is independent of direction.

The schematization procedure comprises the following steps:

- definition of wave direction sectors,
- selection of a representative wave height, H_ϕ , for each sector, e.g. on the basis of the weighting algorithm

$$H_\phi = [\sum (fH^n) / \sum f]^{1/n} \quad (1)$$

in which H is the wave height, f is the frequency of occurrence and n is the exponent of the wave height in the sand transport formula,

- wave model runs for each chosen wave height from each sector, and
- weighted averaging over the sectors, with suitable weighting factors.

Since the wave conditions have a strong influence on the character and the magnitude of the sediment transport, the choice of these weighting factors depends on what is considered as the most relevant type of transport (e.g. rippled-bed transport or sheet flow). The decision on which type is the most relevant is not always clear.

In the following test case, we will use weighting factors based on the occurrence and the wave height for each sector, under the assumption of a linear response.

Comparison of the MRW and SRW approaches

The SRW and MRW schematization concepts were applied to an open coastal area at Poole Bay in S. England, in which the sedimentation regime is strongly influenced by both currents and waves. Model sensitivity tests with and without waves showed an increase of the transport by an order of magnitude upon inclusion of average-sized waves.

As the waves are largest from the south and the east, and the effects of low waves were ignored, only three direction sectors were used, viz. 90–120, 120–150 and 150–180 degrees (N). Figs. 8a and 8b show the net sedimentation patterns under spring tide conditions using the MRW and the SRW approach, respectively. Clearly there are differences due to the chosen schematization, but these have to be considered in relation to the many other deficiencies involved even in this rather elaborate system. If we take the MRW results as a standard, the errors due to the SRW approach should also be considered against the cost reduction (the flow model is by far the most expensive module to run) and the objective of the study.

In quantitative terms, the comparison of the two approaches yields the following results:

- in less than 4% of the model cells (524 out of 14490) the sign of the predicted bed level changes is different,

- the maximum difference corresponds with a bed level change of about 10 cm/tide, and
- the overall mean absolute value of the bed level changes is some 20% lower with the SRW method.

The latter difference could be reduced by adjusting the weighting factors.

In summary, it seems as though the two approaches for wave input reduction are not too far apart in this case. However, one should bear in mind that this is a tide-dominated case, without strong wave-driven currents. In more complex cases with a variety of transport mechanisms, the applicability of either technique is less clear. The SRW approach rests exclusively upon the stirring effect of the waves, so its applicability to situations with strong wave-driven currents is not obvious. Steijn (1992) has shown that the MRW approach involves the risk of overlooking transports which only occur under extreme conditions, e.g. in the deeper parts of the model domain. Besides, the various transport mechanisms exhibit a different response to the wave conditions.

This illustrates the inherent limitation of these approaches: if the schematization is correct for one point or one transport mechanism, it is not necessarily so for all others. In order to achieve a more general applicability, it may be necessary to use more wave height classes, or separate schematizations for mean conditions and events. The latter seems recommendable, anyway, in view of the difference in morphological response time between mean conditions (e.g. beach recovery) and extreme events (e.g. dune erosion).

Concomitance of tides and waves

If both tides and waves are important for the transport, it is not always possible to consider them separately. Firstly, there will be interactions, e.g. via the effect of the water level variation on the wave-induced sediment stirring. Secondly, the relative importance of waves and tides will vary in space (deep channels vs. shoals) and time. Both can lead to rather intricate residual effects.

In order to take such effects into account, a single representative tide in combination with a single representative wave per sector may not be enough. For complex topographies, we should rather think of two tides (neap, or mean, and spring) and three representative wave conditions (calm, mean, storm) per sector.

Steijn and Hartsuiker (1992) and Van Overeem et al. (1992) give an example of what can happen if only the rather moderate wave which represents the whole wave climate is taken into account. They show that the intermittence of storms and calmer periods plays an essential role in the sediment circulation pattern around the outer delta of a tidal inlet.

Example of input reduction

The idea of input reduction is to describe the essence of the phenomena of interest with a model which is driven by reduced inputs. Apart from data reduction techniques, this requires a clear definition of which phenomena are of interest.

If we base the filter algorithm for a tidal morphological model on the representation of the long-term mean transport, we assume implicitly that the phenomena of interest are determined by this aspect of the transport, not by other statistical properties of the transports which occur in nature (unless, of course, these are coupled to the long-term mean).

In order to illustrate the sensitivity of the results to a proper choice of the standard for input reduction, we take one of the simplest morphodynamic models, viz. the propagation of a hump on the bed in a 1D rigid-lid channel with quasi-steady flow (see Fig. 9a). If we ignore the sediment transport due to downslope gravitational forces, the mathematical system for this case reduces to

Fig. 9. Random motion of a bottom hump, and its representation by a model with steady representative inputs. (a) Example at time evolution; (b) correlation of crest height and displacement; (c) steady model based on mean displacement; (d) steady model based on mean transport rate.

$$\frac{\partial z_b}{\partial t} + \frac{C_b(z_b)}{1 - \epsilon_p} \frac{\partial z_b}{\partial x} = 0 \quad (2)$$

in which z_b is the bed level, t is time, x is the spatial coordinate, ϵ_p is the porosity of the bed. The celerity C_b in this non-linear simple-wave equation is given by

$$C_b(z_b) = \frac{\partial q_s}{\partial z_b} \sim \frac{\partial}{\partial z_b} (u^n) \sim \frac{\partial}{\partial z_b} (h^{-n}) \sim n \frac{q_s}{h} \quad (3)$$

in which q_s is the sediment transport rate, u is the current velocity, n is the exponent in the transport formula, and h is the water depth. This model is driven by an input velocity which is a block-function of time. The magnitude of the velocity and the duration of each block are independent random variables, the former with a Gaussian distribution, the latter with a uniform distribution between 0 and an upper bound. Due to the non-linearity, bottom waves will not only propagate, but also deform, with a steepening front and a flattening back (Whitham, 1974). Initially, this deformation is reversible when the flow direction turns, but after some time of continuous propagation in the same direction a shock wave will form. At that moment, the non-linear wave solution starts to overtop, whereas a sand wave will form a sliding front and maintain the shockwave character. Therefore, we abandon the formal non-linear wave solution there and assume a vertical shock front, the propagation speed of which follows from its height and the sediment transport over its top. Mass conservation then requires a decay of the wave height.

Fig. 9a shows a typical result of a model run with a solver which is free of diffusion. The hump is moving randomly up and down the x -axis and gradually loses its height, depending on the time it has spent in the shock wave mode.

If we now try to find a single representative input velocity for which a single model run yields the right average result, the value depends strongly on the model aspect we wish to reproduce. Fig. 9b, for instance, shows that there is a rather poor correlation between the height and the displacement of the hump. The latter depends much more on the chronology of the input (via the continuous duration of the shock-wave state). This means that an input schematization which is appropriate to predict the displacement is likely to be wrong for the height.

Figs. 9c and 9d show that, even if we consider only the displacement, the schematization is not trivial. If we take the mean displacement as a reference for a representative steady flow velocity, there is a very good correlation between "observed" displacements after a given period of time and those computed with the schematized input (Fig. 9c). However, if we take the mean transport as a reference, the correlation is significantly worse (Fig. 9d).

Though very simple and run with rather extreme parameters, this model represents a very common morphodynamic phenomenon. Moving bed features occur almost everywhere, and most of them are driven by currents which have a significant random component. Also shock fronts (foresets) are rather common. Hence the observations from this experiment are relevant in a practical sense.

MODEL REDUCTION

General

From a coastal morphology point of view, “mean” flows differ from turbulence and waves by their length scale, which is usually of the same order of magnitude as the topographic scale of interest. Non-linearities due to advection, bottom friction and propagation therefore act over the whole model domain, such that we have to use a fully non-linear flow module in morphodynamic models (cf. De Vriend, 1987a).

Especially in horizontally 2-D morphodynamic models, the flow module is usually by far the most expensive constituent in terms of computer time, to the extent that, even on top-of-the-line computers, the practical application of models in which the flow module is called after each morphological time step is often prohibitively time-consuming. Reduction of the computational effort is therefore a worthwhile research issue.

The principal approaches to achieve this reduction are

- reduction of the number of calls of the expensive modules (flow, waves, sometimes sediment transport), via methods to increase the effective morphological time step,
- tabulation of constituent model outputs as functions of the input parameters (e.g. for intra-wave sediment transport models),
- introduction of simplified methods to update the flow and wave fields after a small change in bed topography,
- formal model reduction and analysis,
- semi-empirical averaging over the smaller space- and time-scales, and
- parameterization of the model as a whole.

One way or the other, all reduction methods make use of the difference in time scale between the hydrodynamic and the significant morphodynamic processes.

We will discuss some of these methods in further detail in the next sections, first for tidal systems, later more in general.

Increase of the effective morphological time step

The simplest method to reduce the number of calls of the flow model in a

tidal morphodynamic computation process is to extrapolate the bottom changes over one tidal period. The most common approaches are:

- straightforward extrapolation: the bottom changes due to a single tidal cycle are multiplied by a factor N , which is chosen such that the changes are not too large; this method is very simple and cheap, but, being a forward extrapolation, it easily leads to instability; besides, it only extrapolates in time, which means that it does not cover the propagation of bed features;
- a predictor/corrector approach, with straightforward extrapolation over N tides as a predictor; the result is used to compute sediment transports and bed level change rate at this new time level; the average of this change rate and the one at the old time level is used to correct the extrapolation from the old to the new time level; this method is usually more accurate and more stable than straightforward extrapolation.

Higher-order predictor/corrector methods, similar to those which are commonly used in numerical computations, can also be used.

Another way of reducing the number of calls of the expensive flow module is to replace it by a simpler model. The most common of these techniques is the “continuity correction”

$$Uh = \text{invariant in each point of the model domain} \quad (4)$$

in which U is the flow velocity and h is the water depth. This approach is based on the assumption that the flow direction is less sensitive to the bed topography than the velocity, which turns out to make sense for a wide variety of situations. A more refined technique (Hauguel, 1978) is to solve the velocity perturbations from a Poisson equation, assuming irrotationality (i.e. small perturbations) and invariance of the free surface elevation. In tidal situations, however, this approach appears not to work better than the previous one.

In the case of a tidal model, there are various ways to apply the continuity correction in order to increase the time step, such as

- updating residual transports and bed level changes: the bed topography is assumed to change abruptly after one or more tidal cycles, instead of continuously; after such a change, the flow velocity is updated and the residual transport is recomputed; subsequently, the sediment balance yields another bed topography, etc.;
- lengthening of the tide: corresponding phases of N successive tidal cycles are lumped together in the transport and bottom evolution modules; N is chosen such, that the bed evolution per time step is no longer negligible, and that N times the hydrodynamic time step can be taken as the morphological time step; if after each time step the flow is adjusted to the bed level changes, this method includes the propagation of bed features; besides, it is more stable and more accurate than straightforward extrapolation of the result for one tide.

The key element in the latter approach is the assumption of linearity, which allows for the superposition of the various contributions to the bed level change in an arbitrary sequence.

If the bed level changes are small enough to assume linearity, the sediment transport can also be linearized about the result of the latest full computation step. Making use of the continuity correction concept, we can express the sediment transport rate as a function of the water depth only. In the case of combined waves and currents, this is still possible if we have a relationship between the bed level changes and the changes of the wave parameters which affect the transport. In general, we can write

$$q_s(x, y, t + NT) = q_s(x, y, t) - \frac{\partial q_s}{\partial h} \bigg|_t \{z_b(x, y, t + NT) - z_b(x, y, t)\} \quad (5)$$

in which q_s is the volumetric sediment transport rate, x and y are coordinates in the horizontal plane, t is time, T is the tidal period, h is the water depth and z_b is the bed level.

Substitution of Eq. (5) into the sediment balance equation

$$(1 - \epsilon_p) \frac{\partial z_b}{\partial t} + \text{div } q_s = 0 \quad (6)$$

(ϵ_p is the porosity of the bed) and integration over the time interval $[NT, (N+1)T]$ yields

$$\zeta|_{(N+1)T} - \zeta|_{NT} = \zeta|_T - T \text{div} [c_b \zeta|_{NT}] \quad (7)$$

in which ζ denotes the change in bed level with respect to the initial state, and

$$c_b = \left(\frac{\partial q_s}{\partial h} \right)_{\text{mean}} \quad (8)$$

Note that the approximation in Eq. (7) is based on the assumption that c_b varies much faster than ζ .

According to Eq. (7), the bed level changes after the $(N+1)$ th tide can be computed straightforwardly from those after the N th tide and the quantities c_b and $\zeta|_T$, which can be derived from the computations for the first tidal cycle. Thus we can achieve an increase of the time interval between two subsequent full hydrodynamic computations, up to a multiple of the morphological time step.

Four of the methods described herein (straightforward extrapolation, predictor-corrector approach, tide lengthening and linearization of the transport changes) were tested for a hypothetical case of a 1D progressive tide over a pronounced bed topography (Fig. 10a). The latter two methods, and especially the transport linearization, worked far better than the others (Fig. 10b)

Fig. 10. 1-D numerical experiment on tidal filtering. (a) Hypothetical initial bottom profile; (b) relative error in bed topography changes as a function of CPU-time.

and led to very small errors with respect to the reference case ($N=1$), even for large values of N .

A formal approach to model reduction

Because of the difference in response time, the hydrodynamic and the morphodynamic processes can be decoupled, in that the bed can be kept fixed during every hydrodynamic computation. This opens up perspectives for model reduction, as we have shown in the previous section (also see the quasi-steady approach which is often used in river dynamics; cf. Jansen, 1979).

The reduction methods which we described so far are all intuitive, as they have not been derived rigorously from the complete mathematical system. Formal methods should enable us to further exploit the time-scale difference and to understand more of the behaviour of the morphodynamic system. Therefore, another more formal line of thought has been developed, originally in river morphodynamics (see Ikeda and Parker, 1989, for a review), later also in coastal and estuarine morphodynamics. It uses formal mathematical techniques to analyse and reduce the complex mathematical system which constitutes a morphodynamic model. Without pretending to be exhaustive, and without going through the mathematics, we will discuss some of these techniques.

Since we are trying to make use of differences in scale between various processes, scaling the constituting equations is a logical first step. Using physical arguments, we introduce characteristic scales for all variables, which enables us to reformulate the model equations in a non-dimensional form. The resulting system contains a number of dimensionless parameters which indicate the relative strength of the various terms in the equations (= effects in the model). Depending on the class of situations, some of these parameters are small and allow us to apply perturbation methods to derive simplified equations. These can be expected to describe the phenomena of interest without dealing with the unnecessary details which would result from the original model.

Thus we have a formal procedure to derive a quantitative simplified model on the basis of scale arguments concerning the phenomena to be described. However, since this procedure cannot be made rigorous, consistency must always be checked by verifying the original assumptions in the simplified model results.

Note that the scaling can hardly be universal: it will depend on the situation. This means that a loss of generality is the price we pay for model reduction.

Formal averaging methods

The time scale of morphological evolution is often much larger than the typical response time of the water motion to a sudden change in the bed topography. In river models, this is the justification to neglect the time derivatives in the flow model (quasi-steady approach). In coastal morphological systems the non-transient water motion is often oscillatory (sea waves, tides), but we are not so much interested in the details of this short-term motion, rather in the residual effects which cause the morphological evolution.

This intuitive notion is used in analyses of various rhythmic features of the sea bed, e.g. Huthnance (1982a,b) for tidal ridges, Boczar-Karakiewicz and Bona (1986) for ridge systems due to long gravity waves, Boczar-Karakiewicz et al. (1991) for those due to internal waves, and Blondeaux and Vittori (1990) for ripple patterns due to sea waves. In these analyses the bed topography is kept constant in the flow computation, and the transport rate in the sediment balance equation is averaged over the hydrodynamic time scale.

This method can be made rigorous by applying the averaging theorem (Sanders and Verhulst, 1985), which guarantees asymptotic validity. Krol (1990) was the first to apply this to a morphological model of a simple tidal basin. Other applications of this kind are described by Fokkink (1992). Fig. 11 gives a typical equilibrium topography for a rectangular tidal basin which is short relative to the tidal wave length.

The application of the averaging theorem requires the explicit expression of the response of the "fast" flow variables to the "slow" morphological changes. In simple models this can be achieved via series expansions in a small parameter (e.g. the ratio of the vertical scale of the morphological features and the undisturbed water depth). However, this is not possible in more complicated numerical systems, such as the 2-D models (Lattaux, 1992) or $2\frac{1}{2}$ -D models (Wang et al., 1991) which are utilized in tidal situations. Rigorous averaging of these models therefore requires further research.

Fig. 11. Equilibrium bed topography in a narrow rectangular tidal basin of which the length is much smaller than the tidal wavelength.

Linear analysis

As stated before, a proper understanding of the mechanisms behind large-scale morphological evolutions is vital to model reduction in any form. It is necessary, therefore, to analyse the complicated mathematical system which constitutes a morphodynamic model.

A particularly successful approach investigates free instabilities of the morphodynamic system, in order to explain rhythmic morphological features. This approach is supported by the observation that in many cases there is no direct correlation between the scales of these features and those of the various types of water motion which cause the sediment to move.

In early studies, the analysis consists of three steps, viz.:

- (1) model simplification for the specific class of situations under consideration,
- (2) description of a basic state, which is a solution of the simplified model in the absence of the morphological features which are to be investigated, and
- (3) analysis of the linear stability of this basic state by investigating the evolution of infinitesimally small perturbations with a priori arbitrary wave numbers and orientations.

This results in a dispersion relationship for the complex frequency as a function of the wave number and the model parameters. The real part of this frequency gives the exponential growth rate of the perturbations. If this is posi-

Fig. 12. Growth rate of Fourier-type bottom disturbances with wavenumber components k_x and k_y , for a plane tidal current along the x -axis (after Hulscher et al., 1993).

tive for a certain class of wave numbers, the basic state is unstable with respect to perturbations which belong to that class. The imaginary part is the key to the propagation properties of the perturbations (period, phase speeds).

Fig. 12 shows a typical result of a linear stability analysis, for tidal ridges on an initially horizontal bed (see Hulscher et al., 1993). As the system will initially select the fastest growing mode, this should be detectable in observations from the field.

The first applications of this method were reported for unidirectional flows in open channels. Kennedy (1963) used a potential flow model to explain the generation of bed ripples and dunes. Callander (1969) found large-scale features like alternate bars from a linear stability analysis of a system based on the 2D shallow water equations. The linear dynamics of river bends and meanders were studied by Ikeda et al. (1981). In all of these studies, the results were in satisfactory agreement with laboratory and/or field data.

Hino (1974) was the first to report on a linear stability analysis of a near-shore morphological model. He considered the generation of nearshore bars using a depth-integrated and wave-averaged nearshore current model. Since the basic state is non-uniform in space, even a linear analysis becomes quite complicated here and numerical techniques have to be used. The bar patterns predicted by the model look rather realistic.

Huthnance (1982a) used a linear stability analysis to explain tidal sand banks in off-shore coastal seas, starting from a depth-integrated shallow water model. The wave length and the orientation of the fastest growing mode he found are in good agreement with observations. Successive extensions of this work are presented by Huthnance (1982b), De Vriend (1990) and Hulscher et al. (1993).

Hulscher et al. (1993) also give an extension of the flow model with information on the vertical structure of the flow, including velocity veering due to secondary flows in vertical planes. In addition to the sand banks, this leads to the generation of sand waves, with an essentially smaller wave length and with their crests perpendicular to the tidal flow.

Blondeaux and Vittori (1990) used a linear stability analysis to study the formation of sea bed ripples due to wind waves. They compute the vorticity in the viscous boundary layer near the bed, in the case of a purely oscillatory flow outside this layer. Here, too, the model results agree well with observations in nature.

In general one can state: the smaller the morphological features, the more detailed the information on the flow structure which is required to explain them.

Another method, the characteristics analysis (Courant and Hilbert, 1962), refers to the hyperbolic aspects of the mathematical model system. The analysis of the characteristics (in 2D: characteristic planes and bicharacteristics) for a 2D morphological model based on the shallow water equations (De

Vriend, 1987a,b) showed that the morphological information can cross the streamlines while the sediment cannot. This explains the tendency of large-scale morphological features in a tidal environment to elongate under a sharp angle with the tidal flow (cf. Huthnance, 1982b). Besides, the analysis showed that a proper representation of the flow of morphological information requires a flow model which includes advection terms, even if these terms are initially unimportant.

Weakly non-linear stability analysis

Linear stability analyses show morphological systems to allow for the exponential growth of certain infinitesimally small perturbations of the basic state. Once these features have reached a finite size, however, non-linear effects can no longer be neglected. The long-term behaviour of finite-amplitude perturbations therefore requires a non-linear analysis.

In general, there will be a broad spectrum of initially growing perturbations, not just a single mode. Nonlinearity gives rise to complicated interactions between the various modes in the spectrum. Consequently, the fastest growing mode according to the linear analysis is not necessarily predominant in the long run.

If there is a domain in parameter space where the basic state is linearly stable for all perturbations (as outlined in Fig. 13), it is often possible to find the parameter setting for which the first unstable mode occurs: the critical conditions and the critical mode. If we move a little bit further into the unstable domain, there will be a narrow spectrum of unstable modes, all with small growth rates, because we are still close to the stable domain. In this case the bed topography can be represented by

Fig. 13. Typical neutral stability curve, separating areas of growth and decay of linear perturbations in the k - r -plane (k =wave number, r =model parameter). For $r < r_c$, the basic state is linearly stable, for $r = r_c$ the "critical" mode, k_c , is the first to grow.

$$z_b = A(X, T) \exp[i(k_c x - \omega_c t)] + \text{c.c.} \quad (9)$$

in which $A(X, T)$, k_c and ω_c are the complex amplitude, wave number and frequency of the critical mode, respectively. Here the space and time scales of the amplitude evolution are much larger than wave length and the period of the critical mode. Hence we can apply multiple-scale techniques to derive the so-called Ginzburg–Landau equation

$$\frac{\partial A}{\partial T} = \alpha_1 A + \alpha_2 \frac{\partial^2 A}{\partial X^2} + \alpha_3 |A|^2 A \quad (10)$$

in which the coefficients α_1 and α_2 follow from linear theory, whereas α_3 , the Landau coefficient, follows from a non-linear perturbation analysis (cf. Newell, 1974).

Note that Eqs. (9) and (10) represent the evolution of a slowly modulated critical bottom wave. Apart from the values of the coefficients, this system describes the behaviour of marginally unstable waves at near-critical conditions in many non-linear physical systems. If the basic model is conservative, like in the case of gravity waves, all coefficients can be shown to be purely imaginary, which yields the non-linear Schrödinger equation (cf. Mei, 1983).

Colombini et al. (1987) were the first to perform a weakly-nonlinear analysis of a morphodynamical system, viz. a model of alternate bar formation in rivers. As they restricted their analysis to the critical wave number, k_c , they obtained a reduced version of Eq. 10, with $\alpha_2 = 0$ (the Landau equation), which describes the evolution of perturbations to a stable finite-amplitude periodic bed topography. The results were in rather good agreement with measured data. Seminara and Tubino (1992) describe how this method was applied to analyse various models in river morphology.

Schielen et al. (1993) extended the analysis to the full Ginzburg–Landau equation. They demonstrate that the periodic behaviour predicted by the Landau equation can become unstable due to side-band interaction. This leads to a more complicated behaviour of the bed in space and time, such as quasi-periodicity and possibly even deterministic chaos. This suggests that there may be limits to the long-term predictability of morphology.

Following this development in river dynamics, Vittori and Blondeaux (1990, 1992) have performed weakly non-linear stability analyses to describe the finite-amplitude behaviour of ripples at the sea bed due to water waves. The former paper concerns the 2-D case (water and bed waves in the same vertical plane). The resulting saturation amplitudes for the ripples are in good agreement with observations. The extension to two horizontal dimensions, in the 1992 paper, shows that for realistic parameter settings three-dimensional perturbation patterns (brick-pattern ripples; see Fig. 14) may dominate two-dimensional ones, as a consequence of non-linear triad interactions.

So far, amplitude equations of the Ginzburg–Landau type have not been

Fig. 14. Brick-pattern ripples (from Vittori and Blondeaux, 1992; courtesy J. Fluid Mech.).

studied in the context of coastal morphodynamics, though Hulscher et al. (1993) indicate that such an analysis is being undertaken for offshore tidal features such as sand banks and sandwaves.

Semi-empirical averaging

The major difficulty of formal averaging is the specification of the resulting non-linear closure terms. Instead of a formal elaboration, we can also try to model these terms empirically, much like in turbulence modelling. Since the other terms in the averaged equations are formally correct, we can call this a semi-empirical approach.

If we average the suspended load sediment transport equation over a long period of time, say a year, the residual transport consists of two parts, viz. the product of the mean velocity and the mean concentration (advection term) and the mean product of the velocity and concentration fluctuations (dispersion term). Like in many smaller-scale models, we can attempt to model the dispersion term via a gradient-type relationship with the mean concentration (diffusion analogon; cf. Di Silvio, 1991).

In tidal cases, the Eulerian residual velocities are usually rather small, whence the advection term is not very critical in the model. All complexity is stored in the dispersion term, and modelling this with a diffusion term cannot always be substantiated rigorously. Yet this approach has been applied successfully to tidal morphodynamic systems, such as the Venice Lagoon (Di

Silvio and Teatini, 1993). For the time being, it should therefore be considered as empirical, although the theoretical analysis of intertidal dispersion is on its way (Heemink, 1993).

If we want to use this approach in a morphodynamic model, another crucial step has to be made, which relates the local hydrodynamic parameters and the local mean concentration to the local bed level, z_b . Via this step the long-term sediment balance can be brought into a closed form, with z_b as a dependent variable. This constitutes the long-term morphodynamic model. When we use formal procedures, this step can become rather complicated, even for simple systems (cf. Di Silvio, 1989). Therefore, an empirical approach may be expedient here.

BEHAVIOUR-ORIENTED MODELLING

As was stated before, we have to try and meet the demand for long-term morphological predictions with the knowledge we have available at this moment, viz. short-term process-knowledge based on first physical principles and empirical knowledge based on observations in a large number of similar systems.

For a long time, these two sources of information were separated: a study was approached either from the empirical end or from the process-based modelling end. In either case, important resources were ignored, which is not quite recommendable in an area like this (cf. Terwindt and Battjes, 1990).

A further step on the way to integration is formed by the diagnostic utilization of process-based models, to hindcast and analyse a morphological system's past behaviour. Combined with empirical knowledge from similar cases, these models can then be used to predict the future evolution (for instance, see De Vriend et al., 1989; Steijn et al., 1989).

Another way to integrate physical knowledge and measured data is data assimilation. De Vroeg et al. (1988), for instance, describe the application of a Kalman filter to assimilate measured data into a coastline model.

Recently, another line of thought has come forward: behaviour-oriented modelling. The idea is to map the behaviour of a coastal system, as observed in the field or from process-based model runs with real-life input conditions, onto a simple mathematical model which exhibits the same behaviour. This model does not need to have any relationship with the underlying physical processes: it is essentially phenomenological. In some cases, however, it is possible to keep a relationship with the underlying processes, for instance via semi-empirical averaging (see the previous section).

Note that the term "behaviour" has a restricted meaning in this context: it does not refer to the morphological evolution in all its complexity, but only to certain aspects, e.g. within a given range of space and time scales. For instance, if we consider long-term behaviour, this implies that short-term fluc-

tuations as such (not their residual effects!) are ignored. A behaviour-oriented model therefore represents only these aspects of the coastal behaviour, whence it is less generally applicable than a process-based model.

In essence, the approach is not entirely new. Multiple coastline models, for instance (see Bakker, 1968, or Perlin and Dean, 1983), make use of an empirical relationship between the cross-shore transport and the profile slope. In a continuum representation, this boils down to a diffusion model, which is rather popular as a basis for behaviour-oriented coastal profile models (for instance, see Stive et al., 1991).

Behaviour-oriented models are usually much simpler than process-based models of the same phenomenon. As a consequence, they are more suited for model fitting: instead of computing the behaviour straightforwardly from a mathematical system with preset parameters, the parameter values, and maybe even some aspects of the mathematical system, are derived from the behaviour which we want the model to exhibit (usually the observed behaviour). This means that we can make use of the theoretical framework and the gamut of techniques which have been developed for parameter identification and model fitting in other fields.

Behaviour-oriented coastal profile modelling

In order to be able to build up experience with behaviour-oriented modelling and the techniques associated with it, we start with a rather "simple" morphological system, viz. the profile of a coast which is uniform alongshore. Even the long-term behaviour of this system can be simulated with a process-based model (see the companion paper by Roelvink and Brøker Hedegaard, 1993 in this issue), provided that this is driven with real-life inputs. Hence we can use a process-based model to create detailed and complete data sets to work on, before moving to field data sets, which are usually less detailed and complete.

On the basis of a variety of indications, including the experience with multiple coastline models and a number of numerical experiments with process-based models, we have chosen a diffusion concept as a starting point of the model. This model can be formulated as follows (also see Fig. 15 and Stive et al., 1991)

$$\frac{\partial(\delta X)}{\partial t} - \frac{\partial}{\partial z} \left[D(z) \frac{\partial(\delta X)}{\partial z} \right] = S(z) \quad (11)$$

in which z is the vertical coordinate, $\delta X(z)$ is the cross-shore position of the profile with respect to either the undisturbed profile (in the case of a disturbance), or the "equilibrium" profile (in the case of a natural evolution), $D(z)$ is the diffusion coefficient and $S(z)$ is a source term. The corresponding

Fig. 15. Behaviour-oriented coastal profile models: definition sketch for $\delta X(z)$ -formulation of diffusion model.

boundary conditions are a zero gradient at the top of the profile (representing the observed invariance of the upper profile shape) and $\delta X = 0$ at the seaward boundary.

Clearly, all process-based and empirical knowledge has to be stored in the diffusion coefficient and, to some extent, the boundary conditions. The source term enables us to introduce effects of random forcing, longshore transport gradients and human interferences, such as nourishments and sand mining.

In the above formulation, which is close to that of multi-line models, we use the horizontal position of the profile with respect to the equilibrium profile, $\delta X(z)$, as a dependent variable. Alternative formulations can also be used, e.g. with the deviation of X from the actual profile before disturbance, or the vertical position $Z(x)$ as a dependent variable. The latter has the advantage of not requiring a monotonically descending profile, whence it allows for phenomena such as bars. On the other hand, D remains a function of z , which makes the equation non-linear in z .

A typical result of a diffusion-type profile model is shown in Fig. 16, for the long-term evolution of a subaqueous profile nourishment (also see De Vriend et al., 1993). For this case two runs were made with the process-based model UNIBEST (see Roelvink and Brøker Hedegaard, 1993), one to describe the evolution of the undisturbed profile and the other with the nourishment. The difference between the two is considered as the quantity to be predicted. Fig. 16a shows the evolution of the profile which results from a UNIBEST-run driven by a realistic input time-series of seven years. The changes with respect to the initial profile, $x_a(z)$, are shown in Fig. 16b. Fig. 16c shows the corresponding result from the diffusion model (Eq. 11). In this case, the distribution of the diffusion coefficient, $D(z)$, contained three free parameters, which

Fig. 16. Evolution of a subaqueous profile nourishment around MSL -3 m. (a, b) According to UNIBEST; (c) according to a diffusion model for $\delta X(z)$; (d) diffusion coefficient after fitting the diffusion model to the UNIBEST-results.

were determined by minimizing the difference between the two model outputs (De Vriend et al., 1993). Comparison of Figs. 16b and 16c shows that the diffusion model performs reasonably well in this case.

The diffusion concept was tested against the results of process-based models at various time scales, ranging from seasons to decades. In all cases, a rather good agreement was found for roughly the same spatial distribution of the diffusion coefficient: large in the active zone and decreasing to almost zero at the shoreface. The transition coincides more or less with the edge of the active zone, whence it moves seawards as the timescale increases (Stive et al., 1992). In shorter-term applications, where there is more emphasis on bar migration, it may be necessary to add an advection term with a separately adjustable advection velocity.

Behaviour-oriented large-scale modelling of a coastal stretch

Cowell and Roy (1988) describe a behaviour-oriented model for coastal evolution at geological timescales, which is based on the assumption that the various parts of the coastal profile have a fixed shape and act as rigid bodies.

When dealing with time scales in the order of decades to centuries, e.g. in order to evaluate maintenance policies in the light of sea level rise, we may attempt to take a representation of the coastal profile which is somewhere between Cowell and Roy's approach and the diffusion model.

At this timescale, the upper part of the profile, the "active zone", is highly dynamic, but its long-term average shape tends to be invariant relative to the mean sea level (Dean, 1991). This means that the upper part of the profile tends to move as a rigid body along with the mean sea level, with the cross-shore position as a degree of freedom to satisfy the sediment balance (cf. Bruun, 1962).

The lower part of the shoreface is much less dynamic, though not immobile at this timescale (cf. De Vroeg et al., 1988 and Van Alphen et al., 1990). If we assume the inner shelf to be at rest, it seems reasonable to fix the position of the lower end of the shoreface.

Thus we have the following cross-sectional units (see Fig. 17):

- the dune area, which is assumed to be morphologically separated from the other parts of the profile, and to act as a sink of sediment transported by wind over the top of the dune front,
- the active zone, from the top of the dune front to approx. 5 m waterdepth, which moves as a rigid body along with the sea level,
- the middle shoreface, a transition zone between 5 m and 12 m water depth, which is "hung up" between the active zone and the lower shoreface,
- the lower shoreface, from 12 m to 18 m waterdepth, which tilts about its foot point, and

Fig. 17. Large-scale coastal behaviour model: subdivision of the profile in panels (U=active zone, T=middle shoreface, L=lower shoreface).

– the inner shelf, below 18 m waterdepth, which is assumed to be morphologically at rest at these timescales.

These units are assumed to act as a system of hinged rigid panels. Note that the transition levels refer to the Holland coast, the central coastal stretch of The Netherlands. They will generally depend on the wave climate.

The motion of the panels is related to large-scale transport processes, part of which are “autonomous” (i.e. possibly dependent on the local water depth, but not on the shape of the entire profile) and part of which depend on the morphological state of the profile. For the time being we assume that scale-interactions are weak, i.e. large-scale coastal evolution is not a residual effect of smaller-scale behaviour such as bar or ridge formation. In that case, we can

TABLE 1

Large-scale coastal profile behaviour and underlying mechanisms

Profile zone	Behaviour and displacement
Active zone	<p>Yearly averaged profile shape remains unchanged</p> <p>Upward displacement along with sea level rise</p> <p>Shoreward displacement due to:</p> <ul style="list-style-type: none"> – upward displacement with sea level rise (“Bruun rule”) – aeolian transport into the dunes – sediment losses via longshore transport gradients – wave- and wind-induced downwelling transport on the shoreface <p>Seaward displacement due to:</p> <ul style="list-style-type: none"> – wave-induced onshore transport – sediment supply via longshore transport gradients – upwelling transport on the shoreface, due to wind and/or density gradients
Middle shoreface	<p>Inclining or declining “passively”, depending on:</p> <ul style="list-style-type: none"> – displacement of the active zone – rotating of the lower shoreface
Lower shoreface	<p>Declining and eroding in case of:</p> <ul style="list-style-type: none"> – predominance of wave asymmetry and upwelling transports – sediment losses to the inner shelf – sediment losses via longshore transport gradients <p>Inclining and accreting in case of:</p> <ul style="list-style-type: none"> – predominance of downwelling transports – sediment supply from the inner shelf – sediment supply via longshore transport gradients
Inner shelf	<p>Invariant mean level, with undulations due to bed forms (ripples, megaripples, sand-waves, ridges, sand banks)</p>

Fig. 18. Long-term evolution of depth contours for the coast of Holland (data from Knoester, 1990).

estimate the residual transports over the panel boundaries and determine the panel displacement from the sediment balance.

Table 1 gives an overview of the general behaviour and the possible displacements of the various cross-shore units. It also gives an indication of the underlying processes.

Since we disregard the dynamics of the active zone, we omit the main reason why we can often assume longshore uniformity when studying coastal profile evolution, viz. the difference in timescale between cross-shore and longshore dynamics in the active zone. Therefore, we include longshore gradients in this model, by defining a number of longshore units, in each of which we have the above cross-shore system of panels. Each of these panels is connected with its neighbours in the adjacent units.

The longshore sediment transport is a function of the orientation of the panel in the horizontal plane, much like in the usual coastline models.

The above ideas originate from an overall study of the long-term evolution of the Dutch coast, in the framework of the 1990 Coastal Defence Policy Analysis (see, among others, Stive et al., 1990), and they were developed further in research programmes such as the Dutch "Coastal Genesis" programme and the EC-funded project "G6 Coastal Morphodynamics". The latest version, when applied to the Holland coast, yields a good hindcast of the profile evolution over the last century (see Fig. 18).

Behaviour-oriented modelling of complex coastal systems

So far, we have considered rather simple coastal systems, such as a longshore uniform coast and an uninterrupted coastal stretch. This covers only part of the practical demand. Many large-scale coastal problems are associated with much more complex systems, such as river mouths or tidal inlets. The amounts of sediment which circulate in these systems are often much larger than those on the uninterrupted coast.

A variety of process-based models has been applied to such systems (for instance, see the section on input reduction; also see Wang et al., 1991, and Steijn, 1992), with very useful results from a diagnostic point of view. Their predictive value, however, is still limited, either because they only describe the residual transport field and ignore morphodynamic interactions, or because they include only part of the relevant physical processes (e.g. tidal effects only).

On the other hand, there is a wealth of field information on tidal inlets and well-mixed estuaries, in the form of well-established empirical relationships between the equilibrium state of large-scale morphological units and properties of the tide as a forcing agent. Hence there must be a basis for behaviour-oriented modelling of these systems.

A very early form of this approach is Escoffier's (1977) stability analysis

for tidal inlets (also see Van de Kreeke, 1990): the tidal flow through the inlet as a function of the cross-sectional area is estimated on the basis of first physical principles, whereas its the cross-sectional area as a function of the tidal prism is estimated from an empirical relationship. The combination of these two relationships yields a coupled non-linear system of which the stable states can be determined.

Recently, various authors have proposed behaviour-type models to describe the transient evolution of tidal basins, in terms of the equilibrium state and the time-scale of evolution of large-scale morphological units (O'Connor et al., 1990), or in terms of the time-evolution of these units (Di Silvio, 1989; Van Dongeren and De Vriend, 1993). These concepts can be extended to more refined models, on the basis of 1D network models or even 2D models for the tide (cf. Di Silvio and Gambolati, 1990), but a lot of work remains to be done before these are reliable predictive tools. One of the principal issues will be to try and link these basically heuristic models with the underlying physical processes, for instance via formal or semi-empirical averaging (cf. Di Silvio, 1991).

All behaviour models described so far are essentially deterministic, though at a high aggregation level. In some cases, however, we will have to accept a limited predictability at the aggregation level of interest, which excludes a fully deterministic approach. A possible example of such a case is the behaviour of channels and shoals on the outer delta of a tidal inlet.

If the number of possible evolutions is limited, or can be divided into distinct classes, a probabilistic approach may be an alternative. Klaassen et al. (1993) describe such an approach for the prediction of the planform changes of a braided river.

DISCUSSION AND CONCLUSION

Clearly, long-term coastal modelling is still in an early stage of development, although the fact that knowledge and models based on first physical principles are being brought together with empirical knowledge is certainly an important achievement.

A key element in the analysis and prediction of long-term coastal behaviour is reduction, of inputs, models, outputs and measured data. We have shown that a reduction of models and data does not make sense, unless it is intrinsically coupled to the formulation of an objective. This means that we have to specify beforehand which aspects of the coastal behaviour we wish to focus on, and which we consider as unimportant.

The present input reduction techniques are still closely related to a deterministic model approach: making use of statistical methods, we try to define a manageable number of deterministic model runs which provide information on long-term average transports and morphological changes. Research

on tidal and wave climate filtering has yielded a number of useful techniques and indications of their validity, but there remain numerous pitfalls and uncertainties. The possibility of moving towards models formulated in statistical terms seems worth investigating.

Formal model reduction and analysis techniques are potentially powerful, though mathematically complicated. They provide the possibility to identify the inherent behaviour of morphodynamic systems, which is non-trivial, even though we know the response of each constituent process to a given bathymetry. Non-linear techniques make it possible to identify scale interactions and predictability limits. So far, attention has been focused on deterministic aspects (including deterministic chaos), whereas stochastic aspects (forcing, filtering effects) have received little attention. The links which have been established with Applied Mathematics and Systems Theory must enable us to cover this area in the near future.

Data reduction and empirical modelling are vital to the link with real-life coastal behaviour. We may not expect our models to give exact predictions of the bottom topography as a function of place and time, if it were only because we are unable to predict the driving forces exactly. Therefore, if we want to validate models or generalize observations, we will have to translate measured data and model results into aggregated terms. This is the exactly what data reduction and empirical modelling do.

In broad outline, we have two sources of knowledge on coastal dynamics, one based on first physical principles, which concerns mainly on short- and medium-term processes, and the other one based on field observations, which includes long-term behaviour. Behaviour-oriented modelling must help us to integrate these two resources, to yield quantitative large-scale and long-term predictions. Although the first results look promising, we have to bear in mind that in general this concept brings existing knowledge into a manageable form, but does not fill any gaps. Therefore, behaviour-oriented modelling can never be an excuse to neglect process-based research. Rather should we try to improve the physical basis of behaviour-oriented models.

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