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The Coastal-Tract (Part 1): A Conceptual Approach to Aggregated Modeling of Low-Order Coastal Change

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ABSTRACT

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Evolution of coastal morphology over centuries to millennia (low-order coastal change) is relevant to chronic problems in coastal management (*e.g.*, systematic shoreline erosion). This type of coastal change involves parts of the coast normally ignored in predictions required for management of coastal morphology: *i.e.*, shoreline evolution linked to behavior of the continental shelf and coastal plain. We therefore introduce a meta-morphology, the *coastal tract*, defined as the morphological composite comprising the lower shoreface, upper shoreface and backbarrier (where present). It is the first order-system within a cascade hierarchy that provides a framework for aggregation of processes in modeling low-order coastal change. We use this framework in defining boundary conditions and internal dynamics to separate low-order from higher-order coastal behavior for site-specific cases. This procedure involves preparation of a *data-model* by *templating* site data into a structure that complies with scale-specific properties of any given predictive model.

Each level of the *coastal-tract cascade* is distinguished as a system that shares sediments internally. This sediment sharing constrains morphological responses of the system on a given scale. The internal dynamics of these responses involve morphological coupling of the upper shoreface to the backbarrier and to the lower shoreface. The coupling mechanisms govern systematic lateral displacements of the shoreface, and therefore determine trends in shoreline advance and retreat. These changes manifest as the most fundamental modes of coastal evolution upon which higher-order (shorter-term) changes are superimposed. We illustrate the principles in a companion paper (The Coastal Tract: Part 2).

ADDITIONAL INDEX WORDS: *Shoreface, backbarrier, scale, coastal tract, coastal cell, coastal-tract cascade, templating, data-model, behavior-oriented models, sediment-sharing systems, morphological coupling, sea level, sediment supply, coastal management, sea-level rise, transgression, barrier, continental-shelf, sediments, accommodation space, numerical-model.*

INTRODUCTION

Coastal management and engineering requires predictions of low-order (large-scale) coastal change to determine whether shoreline and seabed movements involve systematic trends. Such trends may cause chronic problems that require long-term planning and major engineering interventions. Morphological change entailing temporary fluctuations may cause acute problems, but these can usually be remedied with local measures. Coastal management strategies are very different for the two types of problem, and usually involve very different levels of expense. If estimates of the long-term change cannot be quantified, then it seems unlikely that predictions will discriminate adequately between chronic and acute coastal change.

Aggregated-scale approaches have been developed to sidestep upscaling problems (*e.g.*, DE VRIEND, 1998) from which emerge inadequacies in conventional perceptions and definitions of the coast and coastal processes. In particular, the boundary conditions and internal dynamics are poorly de-

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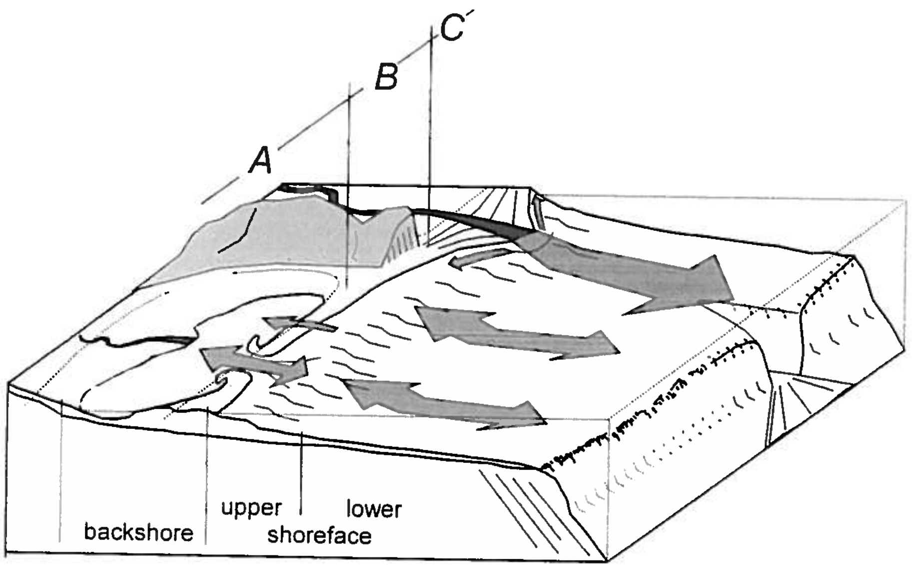

Figure 1. Physical morphology encompassed by the coastal tract (see text for explanation).

fined in site-specific analyses for long-term coastal management. The traditional focus on text-book morphologies (such as the beach, shoreface, dunes, estuaries, and deltas) has tended to promote a reductionist approach involving separate analysis of the classical morphologies. This approach has proved incapable of solving, or even properly addressing, large-scale coastal problems. We therefore need more appropriate concepts applicable to how the coast operates on large scales.

Consequently, we take a broader view of coastal processes by recasting traditional reductionist concepts about coastal morphology into a more unified framework. To establish this framework, we introduce a new, over-arching morphological entity that we term the *coastal tract* (Figure 1), borrowed from geological concepts on *depositional systems tracts* and related sedimentation processes (FISHER and MCGOWAN, 1967; BROWN and FISHER, 1977). The coastal tract is, for our purposes, a generalized term for the continuum of mutually dependent morphological units (surfaces and surficial deposits) on continental margins (Figure 1). Our new geomorphic feature therefore revises traditional ideas about the shoreface and, by emphasizing contiguity between the coastal morphologies, how these interact. More specifically, the revised concepts encompass sedimentary processes that occur not just on the beach and shoreface, but also well landward and seaward of the littoral zone (*i.e.*, to include the estuarine basin and continental shelf respectively).

The purpose of this paper therefore is to (a) introduce the coastal tract as a composite morphology that we can use in structuring problems of scale and aggregation in predicting coastal-change, and (b) propose a methodological framework for describing coastal behavior in nature with models that necessarily have a much lower dimensionality. That is, we introduce the coastal-tract concept to define the most fundamental (lowest-order) modes of morphological change on coasts. Then, based on the coastal-tract concept, a hierarchy of higher-order processes can be discriminated to organize and simplify methods of prediction and analysis. Our objectives therefore are to:

- establish physical principles of the coastal tract, especially regarding fundamental modes of coastal evolution that have been missing from previous approaches to quantitative prediction;
- provide a method to define the hierarchy of morphological processes that take place in the coastal tract (*i.e.*, the *coastal-tract cascade*); and
- provide a protocol, that we term *coastal-tract templating*, for mapping site-specific problems onto *data-models* (*i.e.*, to decompose real-world, measured complexity into constituents on the basis of scale and systematic distinction between external controls and internal responses).

In the companion paper, *Part 2* (COWELL *et al.*, 2003, this volume), we will illustrate the concepts through model applications that elucidate how the coastal tract operates.

METHODS

The new concepts presented here derive from a deeper understanding of our aggregate-scale behavior models (Part 2), forged through tensions over two aspects of work within the project PACE (Predicting Aggregate-scale Coastal Evolution). First, comparative modeling undertaken to determine where and how each of the models are best applied enabled us to formalize representation of process regimes for different elements within the coastal tract. Second, in comparing the models in terms of their alternative representations of aggregated processes (an exercise partly driven by our attempts to develop a hybrid 'super' model), we were forced to reconcile linkages between components of the different models. This led to a more complete picture of coastal-tract sub-systems and how these interact. Thus, paradoxically perhaps, the principles encapsulated formally within our models have been crucial in developing our concepts of the coastal tract and cascade templating.

COASTAL TRACT AND CASCADE HIERARCHY

The Practical Imperative

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

Prediction of shoreline change adopts different approaches, depending on the space and time scale over which predictions are required. For short-term (sub-decadal) coastal change (event and synoptic-scale changes occurring over hours through seasons to years), the focus is generally on the local sediment dynamics. These affect the shoreline planform and the across-shore profile (*e.g.*, shoreline and profile models described by HANSON *et al.*, 2003, this volume) in response to fluctuations in environmental conditions (*i.e.*, the wave climate, littoral sediment budgets, sea level and the effects of anthropogenic activities). Theoretical and empirical approaches to these sub-decadal time scales generally focus on changes to the upper shoreface (defined loosely as the *active zone*; *cf.* STIVE and DE VRIEND, 1995), which correlate with shoreline movements. These changes are moderated by littoral sediment budgets and by sediment 'production' via shoreline erosion cutting into onshore sand reserves (*e.g.* eroding dunes or cliffs), or through artificial nourishment of beaches.

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

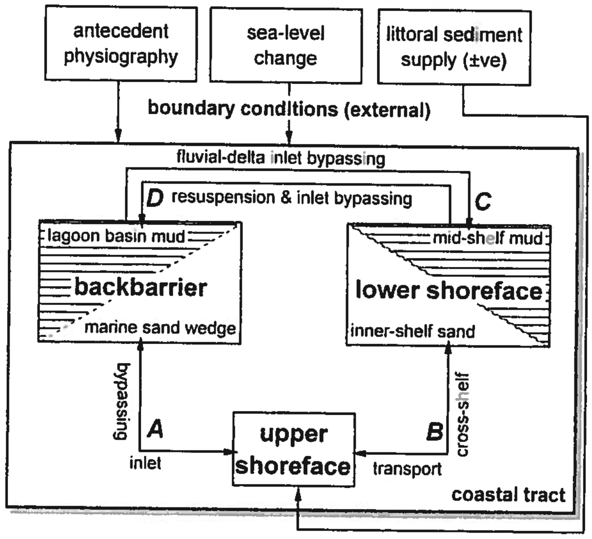

Figure 2. Schematic representation of mechanisms steering the location of the upper shoreface.

(Figure 1B) and fluvial deltas (Figure 1C). Each of these may be present or absent, depending on local conditions, especially the regional substrate slope (ROY *et al.*, 1994; COWELL *et al.*, 1995).

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

For sub-decadal prediction of horizontal movements in the upper shoreface, sand exchanges with the lower shoreface (Figure 2B) are usually ignored (HANSON *et al.*, 2003, this volume) because these fluxes are so small that resulting morphological change is negligible: *i.e.* the annual closure-depth concept (HALLERMEIER, 1981; NICHOLLS *et al.*, 1998). The fluxes of fine sediments (Figure 2, C and D) are not directly relevant to the upper-shoreface sediment budget because mud deposition there is negligible.

For long-term predictions however, none of the internal sediment exchanges depicted in Figure 2 can be ignored. This is because systematic residual fluxes, that are small on the sub-decadal time scale, eventually cumulate through time enough to produce non-negligible (*i.e.*, measurable) morphological changes. Moreover, the changes in morphology of the backbarrier, lower shoreface and upper shoreface cause these three zones to interact dynamically: *i.e.*, the sediment exchanges themselves become influenced by the morphological changes (see Low-Order Coastal Change below).

Definition of the Coastal Tract

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

We formally define the *coastal tract* in both physical and abstract terms. Under the physical definition,

the coastal tract is a spatially contiguous set of morphological units representative of a sediment-sharing coastal cell.

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

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

Coastal-Tract Cascade

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

Physical Tract Constituents

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

Only the lower shoreface departs from the conventional form, in that it is generalized to extend seaward to the edge of the continental shelf (Figure 3). This extended concept encapsulates the regions traditionally (albeit inconsistently) termed the shoreface, as well as the inner, mid, and outer continental shelf (COWELL *et al.*, 1999a). A lumped definition of the lower shoreface is essential if coastal change is to be understood across a sufficiently large range of time scales to enable practical prediction of low-order coastal behavior, and hence higher-order behaviors within a process cascade. Time scales for morphological change (Figure 3) decrease by several orders of magnitude between the upper shoreface and the shelf edge (NIEDORODA *et al.*, 1995). The extended lower-shoreface concept also admits the inclusion of fine sediments (*i.e.,* mud) into the problem through maximum aggregation of sedimentation processes: *i.e.,* across the entire contental-shelf surface.

The physical representation of the coastal tract as a composite feature is simple because the constituents are traditional text-book morphodynamic systems. These systems however collectively contain a large amount of complexity and are also numerous. Paradoxically therefore, the simple physical representation of the coastal tract is too complex as a basis for robust and transparent models of low-order change. Greater aggregation of the constituents is required to achieve this, for which reason we turn to the more abstract concept of the tract as a meta-morphology.

Tract Cascade Hierarchy

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

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

Based on observations of large-scale coastal behavior, for example throughout the Holocene (BEETS *et al.,* 1992; ROY *et al.,* 1994; COWELL *et al.*, 2001), we consider a coastal-tract system to form the first-order level in the hierarchy (Table 1). At this level, the coastal tract behaves as a single unit

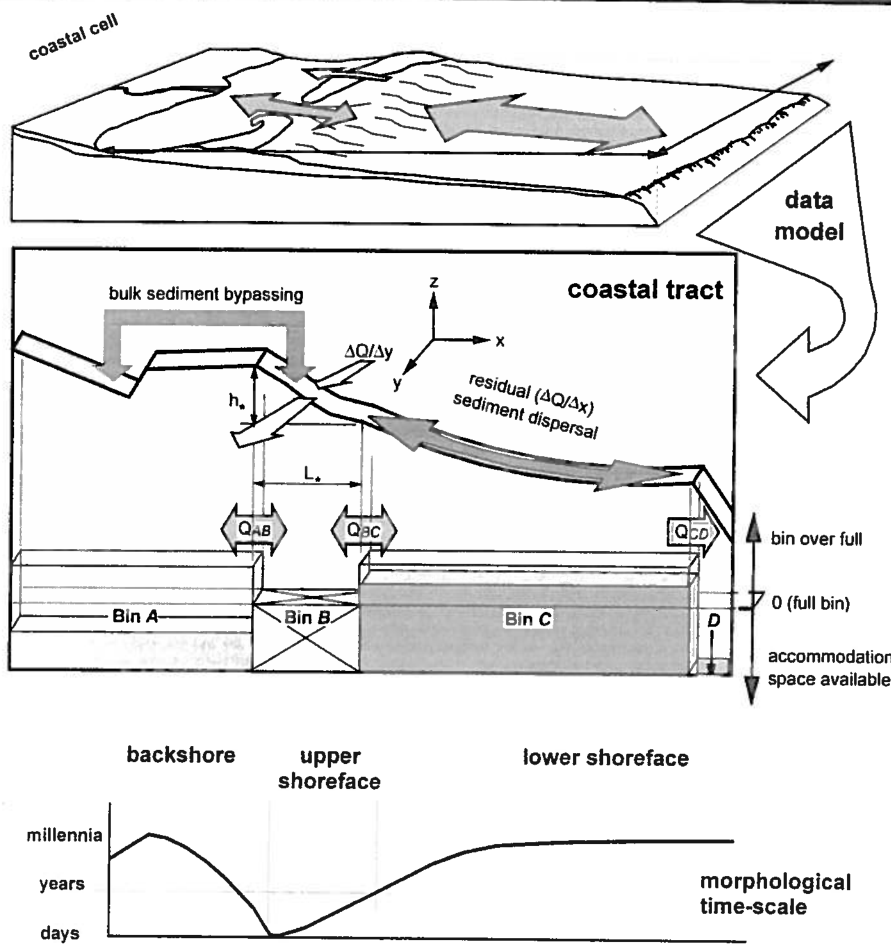

Figure 3. Definition sketch of the coastal tract and sediment-accommodation space as bins in a sediment-sharing system (see text for details).

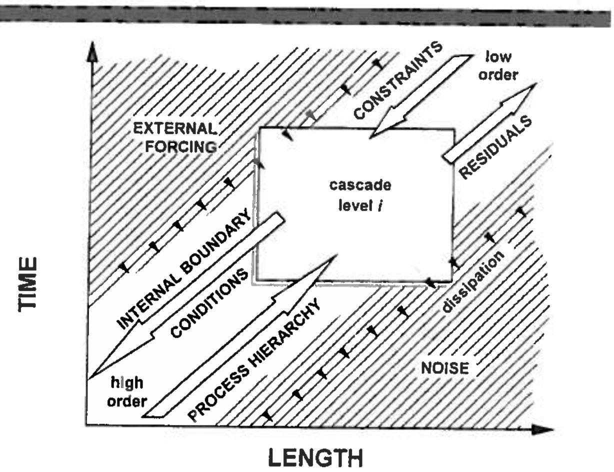

Figure 4. Coastal-tract cascade. Sub systems in the process hierarchy and internal and external constraints. Lower order systems have the largest length and time scales.

that adjusts internally in response to a) gross environmental factors, such as relative sea-level rise, coastal ocean climatology and external sediment sources (such as river input) or sediment sinks (such as submarine canyons), and b) lower level constraints, such as a geologically inherited substrate and tectonic movements (the zero-order system).

Table 1. *Coastal-tract systems and system scales.*

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

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

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

- for beaches—the beach face and surfzone bars;
- for a tidal inlet—the gorge plus the bars and channels on the ebb- and flood-tide delta; and
- for the inner tidal basin—the channels, lower and higher tidal flats, plus fringing salt marshes or mangroves.
- Whereas here we consider first, second and third order systems with decades as the smallest time scale, the fourth-order system displays relevant behavior on the sub-decadal time scale. On smaller time scales (days to seasons), fifth-order systems may be distinguished, such as beach states on the upper shoreface. For the low-order processes (first to third), these fifth-order systems can be considered as noise (Table 1).

These distinctions provide a formal basis for aggregation methods used in development of models of coastal change in general, and for low-order change models in particular (as illustrated in Part 2).

Sediment-Sharing System

Heterogeneity of sediment-flux gradients implies mutual dependence of morphological units in coastal evolution: morphologically coupled behavior is mediated by sediment-volume exchanges. Thus, heterogeneity distinguishes the coastal tract as an internally *sediment-sharing system.* Sediment sharing constrains the internally coupled morphological behavior and is quantified by mass continuity in gross sediment exchanges between morphological units (Figure 2). In applying the coastal-tract concept to predict first-order shoreline displacement (upper-shoreface movements), the morphological coupling is represented through the volumetric effects of sediment transfers between sediment-accommodation bins (Figure 3).

Accommodation space is the volume between the actual bottom shape and the shape that would develop if the short-term physical processes continued to operate uniformly long enough for the morphology to cease changing. Zero accommodation within any of the bins occurs if the associated morphological sub system (second-order system) is in a state of equilibrium with respect to the sedimentation regime. Gross equilibrium does not exist if the coastal tract experiences external forcing (such as a change in relative sea level), or if there is a time lag between an earlier forcing and the attainment of equilibrium. For example, radiocarbon dates in SE Australia show that time lags for lower-shoreface and flood-tide deltas are of the order of 10^3 years (COWELL and THOM, 1994).

The significance of identifying a system that shares a common pool of sediments is that any morphological change in one sub system must cause corresponding changes in the other sub systems. That is, these morphological changes are coupled, with the overall system dynamics governed by flux rates of sediment exchanges between sub systems. This principle is simply the continuity constraint for sediment mass but, for the coastal tract, it takes a defining role in the partitioning and aggregation of processes within the cascade hierarchy. The sediment-sharing concept thus permits us to identify, simplify and thus analyse low-order coastal change.

Coastal-Tract Representation of a Coastal Cell

While the tract-cascade introduces concepts about the aggregation of processes, the cell part of the tract definition relates to aggregation of spatial dimensions. In many cases there is sufficient alongshore homogeneity to permit generalization of a littoral cell into a cross-sectional sediment-control volume of unit width (Figure 3). This is a special case of the more general concept outlined in the Discussion (where spatial and process dimensions are less readily aggregated). For present purposes, however, the special case provides a simpler illustration of coastal-tract principles. Aggregation procedures may involve transformation of process and morphological representations to meet the alongshore homogeneity assumption, if possible. This transformation is one function of templating procedures to construct the data-model.

The littoral cell defines the alongshore extent of the coastal tract. Natural boundaries of the cell may coincide with points of convergence, divergence or topographic barriers in the littoral transport system (CARTER, 1988). Alternatively, the littoral cell may be of arbitrary extent (*e.g.,* defined by administrative boundaries), depending on the site-specific problem. In general however, the littoral cell need not be closed to alongshore sediment transport, but net gains and losses of sediments across cell boundaries must be estimated. All of the boundaries of the coastal tract must be established where the sediment fluxes are extrinsic to all components within the coastal-tract cell.

Alongshore homogeneity is a fundamental assumption when defining a coastal tract. This does not mean however that alongshelf components of flows and sediment flux can be ignored. They are usually one or two orders of magnitude larger than their across-shelf counterparts (ROELVINK and STIVE, 1991; WRIGHT, 1995). Rather, the homogeneity assumption justifies treating alongshelf sediment fluxes as boundary conditions (*i.e.,* as a net input or output of the tract), provided that flux gradients can be assumed to be uniform. Thus, even when we can aggregate the coastal tract into a cross-shore profile of unit width, as in Figure 3, we implicitly include alongshore phenomena (such as littoral transport, and alternating shoreline features such as beaches, tidal inlets, promontories and their attendant processes). That is, although these phenomena are spatially aggregated into non-spatial forms, their effects are retained.

Overall therefore, reduction of the coastal cell from a planimetric- to a profile-representation depends upon assumptions of homogeneity in time-averaged morphology and processes parallel to the shore within the coastal cell (see below under *Cell Templating*). If the alongshelf homogeneity cannot be assumed, or if transformation of process and morphological representations to meet this assumption is not possible, then the *tract* must be treated as a surface with defined alongshore as well as cross-shore dimensions (see Discussion). Clearly, homogeneity does not exist perpendicular to the shore because of the systematic across-shore variation in water depths, morphological units, and associated differences in depth-dependent processes.

LOW-ORDER COASTAL CHANGE

The schematic representation of the coastal tract in Figure 3 generalizes the most fundamental modes of coastal behavior on the decadal to millennial time scales (*e.g.,* within the Holocene period). Behavior of the separate morphological complexes within the coastal tract (*i.e.,* the upper and lower shoreface, and the backbarrier sub systems) can also be treated separately in terms of smaller-scale components (such as surf-zone bars, shoreface-connected ridges, and tidal inlets and channel networks) through analysis at higher levels in the cascade hierarchy. For first-order coastal change however, the coastal tract not only constitutes the framework in which its sub-systems function, it also defines boundary conditions for the sub-systems.

The three morphological complexes of coastal tract (Figure 3) interact dynamically: a morphological change in one necessarily means a corresponding change in the others (*morphological coupling*). Morphological change of lower shoreface and backshore are thus mediated by sediment fluxes between bins in Figure 3: *i.e.,* these fluxes are internal variables within the first-order sediment-sharing system. Furthermore, if we were to focus separately on the lower shoreface and backshore, evolution of these second-order morphologies is not only governed by their internal sediment dispersal mechanisms. Their evolution also is constrained by the between-bin fluxes. Thus, whereas the between-bin fluxes are internal processes at first-order, they constitute boundary conditions to the second-order sub-systems.

Sediment Exchanges in the Coastal Tract

The backshore coupling (Figure 2*A*) involves either bypassing of sand from the shoreface to the lagoon behind the barrier (if a barrier exists), or the net supply of sediment to the overall coastal sediment budget. The latter can take place through deposition of muds in the lagoon, or through supply of terrestrial sands to the shoreface by estuaries that act as net-sediment sources rather than sinks (*i.e.,* fluvial deltas). Alternatively, the backshore may be closed to sediment bypassing to and from the shoreface if inlets and river mouths are absent. These conditions occur in coastal cells where the shoreface abuts the mainland, or where high dunes run uninterrupted along the coastline, isolating the coastal lowland from the sea. The mainland can act as a sand source during landward translation of the upper shoreface (Figure 1*C*) if the backshore is steeper than the shoreface (COWELL *et al.,* 1999b), and provided that the mainland is composed of soft sediments rather than bedrock. The mainland may act as a sand sink if landward-migrating transgressive dunes outrun the horizontally translating shoreface.

The lower-shoreface coupling (Figure 2*B*) involves long-term adjustments which tend to steer the upper shoreface in a landward or seaward direction, depending on whether evolution of the lower shoreface causes it to act as a net sink or source for sand-sized sediments. The actual direction of upper-shoreface translation depends upon combined effects of the upper-shoreface coupling with the lower-shoreface and with the backshore. For example, if the lower shoreface is a net sand source for the upper shoreface, but the volume of sand bypassing to the backbarrier exceeds that supplied from the lower shoreface, then upper-shoreface translation will be landward. This becomes manifest to coastal managers as chronic shoreline recession. In the case of the central Netherlands coast on the other hand, the sand supply from the lower shoreface appears to balance littoral losses to the Wadden lagoon (STIVE *et al.,* 1991).

A direct coupling between the lagoon and lower shoreface exists (Figure 2*C*) due to the possibility that fine sediments, discharged at the fluvial delta, bypass through the tidal inlet and upper shoreface to settle out of suspension on the lower shoreface as mid-shelf mud deposits. Conversely, upon resuspension on the lower shoreface (Figure 2*D*), these sediments may find their way back into the same or other lagoons where they contribute to reduction of backbarrier accommodation capacity (WOODROFFE *et al.,* 1989; BEETS *et al.,* 1992). The resulting decrease in the potential of lagoons to act as sediment sinks (*i.e.,* less accommodation space) also reduces the potential for sand loss from the upper shoreface to the lagoon. More generally, and of significance to coastal management, mud from any source, deposited in the central basin of the lagoon, reduces the rate of long-term recession of the upper shoreface.

Overall therefore, Figure 2 summarises the processes governing the fundamental modes of coastal behavior upon which all other coastal changes are superimposed. Further details of these processes and illustration of resulting low-order coastal evolution are presented in our companion paper (Part 2).

Morphological Coupling

The morphological coupling mediated by the various sediment exchanges outlined above is schematized in Figure 3. For long-term (hyperdecadal) predictions we can ignore the changes that occur within synoptic time scales, so that the upper shoreface assumes a time-averaged profile of invariant form, or a time-averaged form that evolves with secular changes in wave climate or sediment grain size (*e.g.*, due to changes in sediment provenance). The first-order, coastal-tract problem then reduces to prediction of time-averaged shoreline positions that move only with vertical and horizontal translation of the time-invariant (or slowly evolving) upper-shoreface profile. The vertical translations are governed totally by long-term movements in sea level. The horizontal translations are steered by two sets of gross processes. These processes involve a coupling of the upper shoreface to a) the backshore environment, and b) the lower shoreface.

Within the schema of Figure 3 therefore, the backbarrier (Bin A) may be absent (mainland-beach case) but, if present (barrier-beach case), sediment sharing between the backbarrier and the upper and lower shoreface is governed by sediment accommodation capacity, aggregated volumetrically as a set of storage bins for each complex in the tract. As illustrated here, the backbarrier (Bin *A*) has positive accommodation capacity and exerts a demand for more sediment to attain tidal-basin equilibrium, whereas the lower shoreface (Bin *C*) has a sediment surplus (*i.e.*, the continental shelf is shallower than its equilibrium elevation) and therefore acts as a sediment source for the continental slope (Bin *D*) and the upper shoreface (Bin *B*).

Changes in relative sea level alter the accommodation space of all bins. Under conditions of stable sea level, the upper shoreface (Bin *B* in Figure 3), where the morphological time scale is shortest, is permanently full (*i.e.*, in equilibrium). Since this bin can then undergo neither net gains, nor losses of sediments, transfers from adjacent bins (*A* and *C*) bypass the upper shoreface to adjacent bins. Depending on the balance between net sediment fluxes from both adjacent bins and external inputs from littoral sources (Q_{AB}, Q_{BC} and $\Delta Q/\Delta y$ respectively in Figure 3), the upper shoreface (Bin *B*) must advance or retreat (*progradation* and *retrogradation* respectively).

The accommodation of Bins *A* and *C* may be positive (unfilled), negative (over-full) or zero (full or equilibrium). Positive accommodation space means that the bin demands sediment, causing retreat of the upper shoreface and eating into the backbarrier (or mainland-beach) sediments to supply the demand and to return to equilibrium. Negative accommodation space exists if either Bin *A* or *C* are overfull: *i.e.*, they are sediment sources. Such a state implies that the lower shoreface (the continental shelf in general) is shallower than equilibrium under the prevailing shelf flow climatology (Roy *et al.*, 1994). The tendency then is for the shelf to degrade. Bin *D* is the continental slope and therefore is a deep hole in terms of accommodation space on time scales of millennia or less. Seaward sediment fluxes into *D* are one-way accordingly, provided that an ocean basin exists: ocean basins are absent from 'buttressed' continental margins (*e.g.*, the European North Sea coast).

The morphological coupling of the coastal tract complexes not only causes lateral (shore-normal) movements of the upper shoreface. Each of the bin boundaries in Figure 3 can undergo such lateral migration as a result of sediment transfers between the bins. Sediment transfers beyond the continental-shelf break (Q_{CD}) cause the coastal tract to extend itself seawards (ocean basin-margin fill). Feeding of the backbarrier from the upper shoreface involves a sediment recycling processes: barrier rollover (Leatherman, 1983). That is, the upper shoreface preserves its equilibrium bin-volume by retreating into the marine-sediment wedge at the seaward end of the backbarrier, thereby reworking this sediment-wedge further landward (a landward shift of *AB* in Figure 3). The backbarrier bin volume tends toward equilibrium through this process since a reduction occurs in the width of Bin *A* if there is no sea-level rise. With rising sea level, the processes are similar, except that the reduction of the backbarrier accommodation due to barrier rollover is offset by flooding of the backbarrier and retreat of the inland shoreline of the lagoon. This tends to maintain the width of Bin *A*. Conversely, falling sea levels cause a seaward migration of all bin boundaries (*regression*). Seaward translation of the upper shoreface and shoreline (a coastal *progradation*) also occurs if $Q_{BC} - \Delta Q/\Delta y > Q_{AB}$ in the absence of sea level rise, or if the positive difference of this inequality is sufficient to offset upper-shoreface accommodation demand created by a rising sea level.

This rationale broadly follows the qualitative concepts of Curray (1964) on coastline displacement due to the combined effects of relative sea-level changes and sediment supply, which in a narrower sense underlies the Bruun Rule (Dean and Maumeyer, 1993); although neither includes an interaction with the lower shoreface. Where net sources or net sinks for sediment are absent, 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. A net source of sediment shifts the balance toward regressive behavior for any given change in relative sea level, whereas net sediment losses to a sink shifts the balance toward transgression. Swift (1976) extended this qualitative tract model by specifying sources and sinks in terms of alongshore or cross-shore processes. He related alongshore processes to deltaic sediment sources and estuarine sinks, while the cross-shore processes depended on the relative importance of storm and fair weather conditions.

COASTAL TRACT TEMPLATING

We introduce the idea of *coastal-tract templating* as a protocol that not only guides design of computer experiments, but also helps clarify just what is and what is not being modelled in any given application. The term templating was introduced recently as a formal approach that guides the configuring of a geological concept-model to a given data set in explaining coastal evolution for a specific site or region (Zhang *et al.*, 1997; Parsons *et al.*, in press). We prefer to define templating in an opposite sense, as the procedures involved in preparation of a data-model (*cf.* Rhoads and Thorne, 1996, p.127).

Modeling and analysis have always required preparation of a data-model, but hitherto this generally has been a subjective and tacit exercise. Templating therefore is unavoidable since site-specific morphology must be generalized into the form required by the particular tools used in prediction: a requirement relevant to coastal analysis on all scales. However, formal treatment of this step in modeling procedures has been lacking until now, with the consequence that ambiguity and disagreement has often surrounded modeling objectives, limitations, and interpretation of predictive results (*e.g.*, PILKEY *et al.*, 1993). Thus, recognition of templating as a formal step in any modeling (or analysis) exercise is more likely to guarantee that a) documentation of analysis and modeling methods includes an explicit statement of templating procedures, and b) coastal managers take scale-dependence into account when defining site-specific problems.

Definitions

We define the *data-model* as the product of synthesizing raw data into a structure that complies with the scale-specific form and properties (*e.g.*, dimensions and parameters) of predictive models applied to a given problem. Any practical exercise in modeling or analysis of low-order coastal change must begin with procedures to define the tract. These procedures involve templating the observations about a specific case (or coastal region) in nature onto a data-model that is of dimensional relevance to the modeling or analysis method to be applied. The coastal-tract cascade is the template that guides these procedures. More specifically, templating involves any procedures that a) distinguish boundary conditions from internal dynamics on a specific level (cascade templating), b) delimit the spatial extent of the coastal cell (cell templating), and c) transform the data set so that the tract has the same spatial dimensions as the predictive model and contains comparable variables or parameters (data-set templating).

Templating procedures must be performed in the order listed above. Transformation of the data set is pointless until the scale of interest has been specified, the associated morphodynamic (sediment-sharing) systems isolated, and the coastal cell identified. Cell templating follows completion of cascade templating because alongshore extent of cells generally differs at each level in the scale hierarchy: *i.e.*, lower-order systems tend to encompass larger cells. This tendency exists because the homogeneity assumption, upon which cell definition depends, becomes less restrictive at lower order. That is, with a greater the degree of aggregation, the spatial detail (in flow fields and morphology) is progressively averaged out (lumped together).

Cascade Templating

Cascade templating is undertaken to determine a) the level of aggregation, and b) which of the nested morphodynamic systems in the cascade hierarchy must be included in the analysis. Both considerations relate to the practical problem and management objectives (*e.g.*, long-term planning or short-term remediation). Selection of the predictive model also depends upon the analysis objectives. Thus, both the analysis objectives and properties of the predictive model form the templates for the cell definition and the data-model.

In the context of the predictive modeling procedures specifically, the purpose of cascade templating is to discriminate components of the system that can be regarded as boundary conditions from those we are forced to retain as internal system components. In principle however, each of the internal components of a sediment-sharing system on a given scale has nested within it the dynamics of all higher-order systems (Figure 4). Formal discrimination between which internal components are primary and which are nested higher-order systems involve aggregation issues that are yet to be fully resolved (see Discussion).

In practice therefore, a pragmatic approach is simply to define cascade relationships on the basis of the morphological-scale classification in Table 1. The specifics on how this should be done will vary from case to case (both by site and management focus) and depend upon the predictive model for which the templating is undertaken. Generally however, the result will involve spatial averaging of morphology and lumping of process details within a sediment-sharing system (Table 1).

Cell Templating

For simpler cases, cell templating follows well established conventions for the littoral-cell concept: *i.e.*, that natural cell boundaries coincide with points of convergence, divergence or topographic barriers in the littoral transport system (CARTER, 1988). Alternatively, arbitrary boundaries are established for the purposes of sediment budget analysis (usually for engineering or planning projects), often coinciding with administrative boundaries or notable landmarks that otherwise do not interrupt littoral transport (*e.g.*, GELFENBAUM *et al.*, 1999). For this type of cell boundary, the non-zero estimates of alongshore sediment fluxes must be obtained. In the context of the coastal tract, cells that have natural boundaries with respect to alongshore surfzone-sand transport are unlikely to have corresponding boundaries with respect to alongshelf sediment transport far offshore. So cell templating generally involves contrived boundaries to some degree.

The Homogeneity Assumption

Cell templating requires the assumption of alongshore homogeneity in morphodynamics. The assumption is more readily satisfied at lower order because of aggregation of morphodynamics implicit in the tract-cascade concept. The homogeneity assumption relies upon the absence of strong systematic variations in morphology and transport processes parallel to the coast. In particular, variation in alongshore sediment-transport gradients must be weak enough for the flux difference across the coastal cell to be regarded simply as a net gain or loss at the cell boundaries. If this is the case, then we can treat deltas and estuaries outside the coastal tract as exogenous sources and sinks. The nature of these sources or sinks does not matter: they could equally be anthropogenic or due to up-drift coastal erosion or down-drift shoreline progradation, respectively. At first approximation therefore, the existence and magnitude of a source or sink adjacent to the coastal tract is considered as an independent variable, forcing the first-order system: one of the *Sloss variables* referred to by THORNE and SWIFT (1991) and outlined briefly in Part 2.

An example of an alongshore source is that of a hypothetical outbuilding river-delta (Figure 5a), while an example of

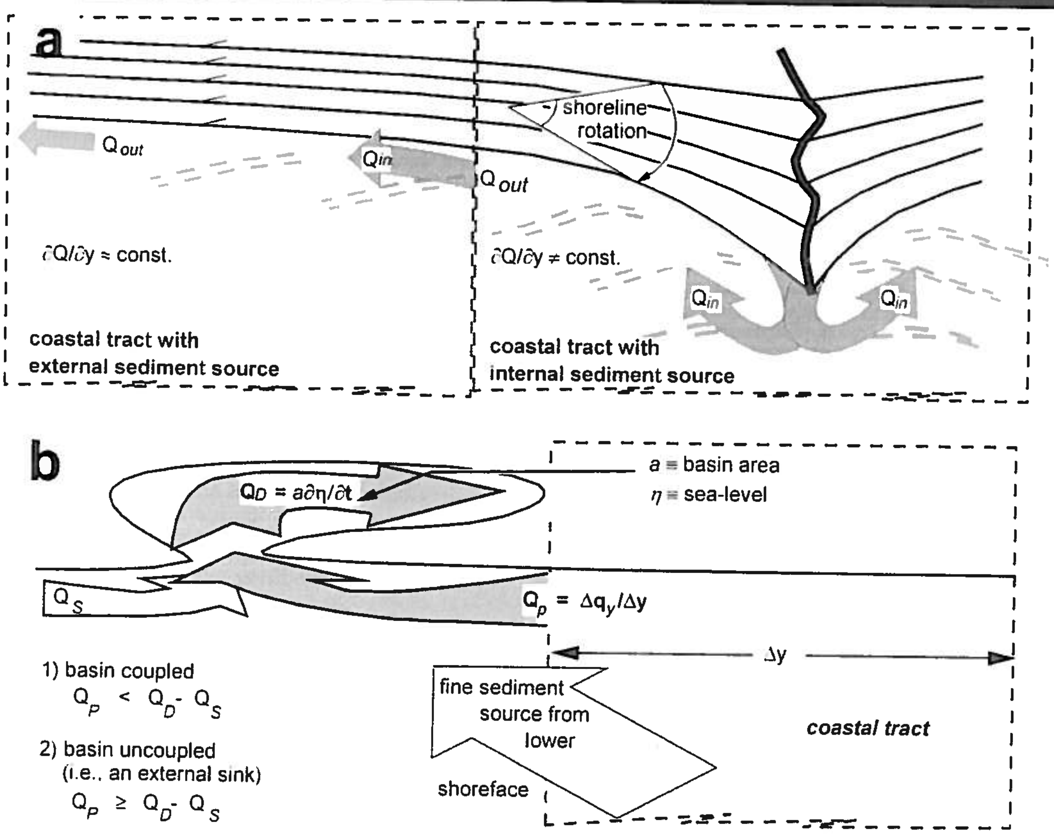

Figure 5. External and internal sources and sinks: a) Successive shorelines supplied by a river delta; and b) Estuarine basin as 1. dynamically-coupled sink, vs 2. external sink.

a sink is that of a hypothetical down-drift tidal basin providing accommodation space (or sediment volume demand, Q_D) due to relative sea-level rise (Figure 5b). The source is external for those parts of the coast where uniform gradients in littoral transport rates prevail ($\partial Q/\partial y \approx$ const.), causing parallel shoreline displacements (indicated by tick marks on successive shorelines in the left panel of Figure 5a). If the river sediment supply is greater than the littoral transport capacity, then shorelines will rotate (right panel, Figure 5a) and the wave-induced transport will be modified accordingly (*cf.* KOMAR, 1973). The source is then internal to the tract rather than an external forcing; so it no longer can be considered a Sloss variable. The tract and river delta co-evolve and are thus coupled.

The sink is external if the size of the sink does not effect the littoral transport rate within the tract: littoral transport represents a potential supply volume (Q_p) from the tract to the sink (Figure 5b). We adopt the model concept that the infill capacity per unit time (*i.e.*, Q_D) is proportional to the theoretical accommodation space of the basin. This proportionality is determined (empirically) by the deviation from an equilibrium between the tidal prism and basin volume relative to mean sea level (*e.g.*, WANG *et al.*, 1996). Homogeneity exists if $Q_p \geq Q_D$ since, under these circumstances, there is no tendency for the sediment demand of the sink to grow through time, and we may view the sink magnitude as a Sloss variable.

If the sink and source are coupled ($Q_p < Q_D$), the demand of the basin increases through time, which feeds back into Q_p through the effects of shoreline rotation: the sink-zone shoreline recedes more quickly than the source-zone shoreline, which in turn causes Q_p to grow due to increasing wave-incidence angles. The result is shoreline rotation and increasing alongshore transport gradients closer to the inlet (*i.e.*, $\partial Q/\partial y \neq$ const.). Under these circumstances, even if segments of coast containing the source shoreline and the estuary are geographically offset, they nevertheless constitute internal components of the same tract. That is, the two zones recede in unison so that the estuary effectively constitutes a backbarrier basin for the updrift shoreface. They are coupled intrinsically through their sediment sharing within a barrier rollover process (LEATHERMAN, 1983) on a tract width-averaged basis.

Two further complications exist. First, due to other littoral sources (Q_S), the effective demand imposed by the sink on the tract is reduced to $Q_D - Q_S$ (Figure 5b). Second, the sink demand also can be satisfied in part by supply of fine marine sediments resuspended on the lower shoreface (*e.g.*, WOODROFFE *et al.*, 1989) and transported toward the coast (Figure 2D). Since this direct exchange of fine sediments between the lower shoreface and the backbarrier can also operate in reverse, the linkage constitutes an internal coupling.

Data-Set Templating

Data-set templating involves (a) transformations of the tract and its dimensions into the same dimensional form as the predictive model or to allow the homogeneity assumption to be satisfied, and (b) distillation of the data to estimate values of variables and parameters used to calibrate the predictive model. Parameter estimates required from site-specific data include values for process parameters, morphological dimensions, and boundary conditions (*i.e.*, inputs and outputs such as sediment fluxes, plus the sea-level curve and energy inputs).

Transformations include spatial aggregation of processes and morphologies to suit the predictive model. Existing models for low-order coastal change (Part 2) are framed as a one-dimensional (1D) cross-shore profile (Figure 3). For the simplest cases in which alongshore homogeneity of processes and morphologies can be assumed in the coastal tract, spatial-aggregation generally involves a simple averaging alongshore of morphology, flow fields and sediment-transport rates. A cruder but common approximation is simply to take a representative, cross-shore transect through the tract. If the homogeneity assumption is strong, then such an approach is reasonable.

Transformation and aggregation of tract morphodynamics can be taken even further by representing the tract in a spatial (zero dimensional) form. For instance, the second-order system involving the tidal inlet, ebb and flood-tide deltas can be reduced to a representation involving only aggregate volumes for each of the morphological units and corresponding sediment-flux averages (STIVE *et al.*, 1998). These quantities can then be incorporated into the first-order system (*e.g.*, rates of sediment exchange between the shoreface and backbarrier complexes). This example illustrates that parameter-estimation aspects of data-set templating are closely related to the cascade templating.

DISCUSSION: UNRESOLVED ISSUES OF AGGREGATION

The concepts presented here on the coastal tract, cascade hierarchy and templating distil an epistemology implicit in our various approaches to modeling low-order coastal change (Part 2). Quantitative prediction for this type of coastal change (*i.e.*, on time scales of 10^2 to 10^3 years) has moved only recently beyond a simple Bruun-type analysis (DEAN and MAURMEYER, 1983). It is worth emphasizing that much of the controversy regarding that type of analysis, and coastal models in general (*cf.* PILKEY *et al.*, 1993; THIELER *et al.*, 2000), stems in part from the absence of a formal rationale on aggregation. Although we seek to address this problem, the issues run deeper than outlined in the previous sections. In particular, we are yet to resolve (a) adequate methods to deal with sites for which the alongshore homogeneity assumption cannot be justified (although work is in progress on this aspect), and (b) a comprehensive formalism for process aggregation down through the successive levels of the coastal-tract cascade.

Intractible Alongshore Heterogeneity

Cases for which alongshore heterogeneity cannot be ignored force us to increase tract dimensionality and require an unwelcome expansion in the level of model parameterization accordingly. Although from our collective experience we think that, under most circumstances, the data-model can be reduced to an alongshore-averaged cross-shore profile, situations such as those shown in Figure 5 illustrate that such spatial generalization is not always possible. Another example here is the Columbia Cell in NW USA (GELFENBAUM *et al.*, 1999), where a band of mud deposition running diagonally across the continental shelf northwest from the Columbia River mouth (STERNBERG, 1986) indicates a systematic along-shelf component of advection (idealized hypothetically in Figure 6). Major point-source rivers like the Columbia show an alongshelf logarithmic decrease in sediment accumulation rates, and also typically have alongshelf trends in both grain size (progressive fining with distance from source) and sediment structure (higher physical stratification close to source and greater bioturbation with distance) (NITTROUER and STERNBERG, 1981).

Such conditions require extension of the unit-width coastal-tract concept to provide a more general approach to mapping reality onto the data-model. Figure 6 schematically illustrates the problem: the river mouth corresponds to a divide in the net littoral transport of sand (indicated by the arrows). This transport varies in direction on shorter time scales (events to seasons), but there is a net direction to the flux over the long term (decades and longer). In this hypothetical case, estuaries exist with (Cell B) and without (Cell A) the complications of couplings referred to in the previous section.

Various possibilities for extending the unit-width coastal-tract concept are canvassed in Figure 6 for the different segments of coast. The equilibria and rates of change in segment labelled "A" can be analyzed along a simple cell-width-averaged offshore profile (Tract A). The effects of alongshore heterogeneity in the segment labelled "B" could be dealt with crudely by aggregating the river delta and Bay 1 into two separate sub tracts (Tracts B1 and B2) represented using two profile models: one to simulate progradation of the delta, and the other to simulate the recession caused by the sediment demand of Bay 1. We would need to couple the two profile models via their littoral sediment exchanges. These exchanges would depend on the relative displacement of the shorelines in the two sub tracts: the decreasing negative gradient in net fluxes away from the delta produces a shoreline rotation (*cf.* KOMAR, 1973). Thus, estimation of the sand exchanges would also require application of a shoreline model (*e.g.*, HANSON *et al.*, 2003, this volume) to simulate the effects of shoreline rotation on the wave field and littoral transport rates. As far as we know, a formalized hybrid model of this type is not yet available.

Both Tracts A and B assume that advective fluxes responsible for the fine-sediment band running diagonally across the shoreface can be ignored in modeling low-order behavior.

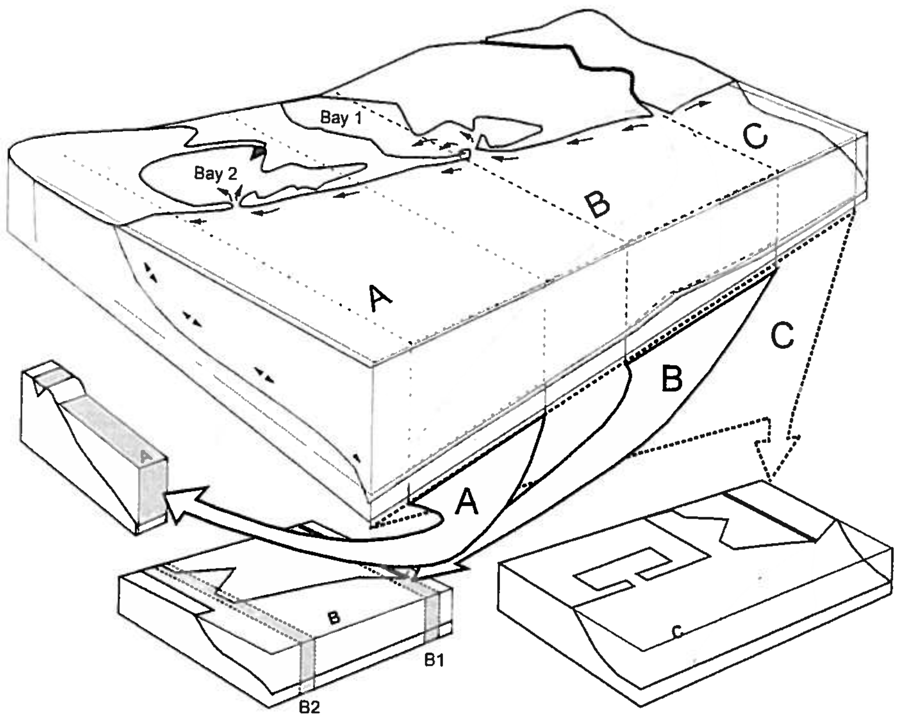

Figure 6. Extended coastal-tract generalizations for segments of coast a) without dynamic coupling between the shoreface and sediment sinks or sources, b) with time-dependent shoreline rotation due to alongshore variation in littoral transport, and c) with non-homogeneity in coast-parallel sediment transport due to advective processes on the shoreface.

If this assumption cannot be justified, then representation of the tract must include alongshore as well as across-shore components explicitly. Tract C in Figure 6 therefore would need to incorporate the entire length of coast. The processes responsible for the fine-sediment band involve deposition in water depths that vary systematically along the coast. Although the mud band looks discrete in Figure 6, in nature such deposits probably represent the modal track of otherwise stochastic plume dynamics. Unless we find some new trick for aggregating processes involved in this type of problem, we are stuck with the need to incorporate an undesirable amount of detail into modeling the low-order behavior. This is the case especially if we must include the shoreline-source/sink coupling as well. This type of problem is a subject of our ongoing research.

Aggregation Issues

Although we have mapped out the concepts and a pragmatic approach for defining the coastal tract to model low-order coastal change for any practical purposes (Table 1), the need to develop a formal aggregation theory for the tract remains. It is not enough to distinguish clearly between scales and orders of behavior within the cascade. We also must decide which higher-order effects constitute useful information. These effects need to be represented in some way at the next lower order: *i.e.*, the question is how to carry useful sub-scale information to the scale of interest. We have formal aggregation methods for some aspects of the coastal tract: *e.g.*, for the processes involving components of the tidal inlet and tidal basin (channels and flats) that have been aggregated and mapped onto the next low-order coupled shoreface-backbarrier behavior (STIVE *et al.*, 1998). We need however to formalise such methods for other aspects and levels of the tract cascade.

Adequate representation of sub-scale processes and effects requires application of a closure concept in aggregating processes into a simplified lower order model. This aggregation may be straightforward if we can assume that the dynamic interaction between higher and lower order processes is weak enough to be ignored, which may not always be possible. We have adopted the following viewpoints and definitions to schematize these aggregation concepts. By definition, the first-order system is a sediment-sharing coastal tract of such a spatial and temporal scale that its first-order evolution is determined by external boundary conditions and constraints alone. These external conditions are assumed to be known and to remain known on the time scale of the coastal change to be predicted. This implies that the integrated sediment mass balance of the first-order system is known. Note that the spatial and temporal scales of a first-order system depend upon site specific-conditions. To assess these scales requires a careful analysis of the system and in practice may require several iterations.

The internal dynamics of the first-order coastal tract are determined by the cascade of higher-order systems interacting within the first-order system (Figure 7). The concept

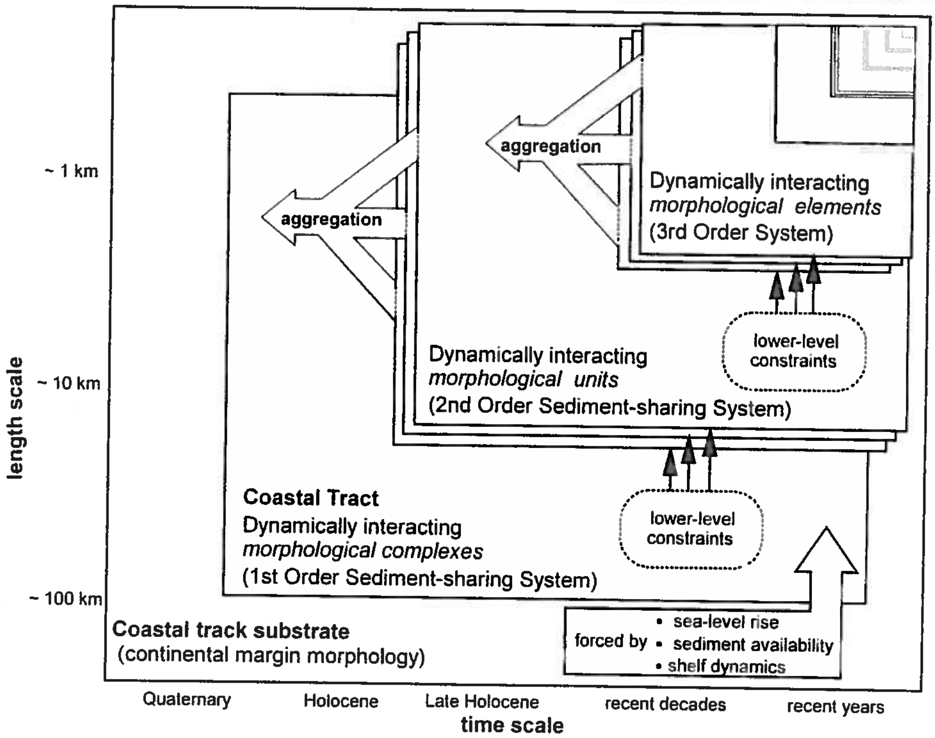

Figure 7. Nested systems of the coastal-tract cascade.

adopted here (*cf.* COWELL and THOM, 1994; DE VRIEND, 1998) is that the space and time scales of systems on each order are mutually coupled and confined, such that as the order increases the scales decrease (Figure 4). On larger space scales outside the dynamic interaction range, we find external conditions, while on smaller space scales we find 'noise'.

The first fundamental issue that we face is to define the order and scale of each higher-order system, including the boundary conditions and constraints that drive these systems. Commonly, a lower-order system will act as, provide or filter a boundary condition to the system on the next level up in the heirarchy (*i.e.*, one order higher). Ideally, multi-order systems have clearly different temporal and spatial scales at each level, so that we can discriminate between them. In spectral terms this implies that there exists a separation between their respective energy contents at the specific wave number and frequency ranges of their morphological features. In the less ideal cases, the different order systems have no clear spectral separation, which makes discrimination more difficult and sometimes arbitrary.

Morphological examples of ideal cases are wave- and current-induced bed ripples, which are known to be determined by local variables such as water depth, shear stress or mobility number and bed properties (NIELSEN, 1992). These higher-order dynamics can be separated through a local evaluation from a lower-order phenomenon such as shoreline and upper shoreface evolution due to alongshore-transport gradients. An example of a less ideal case are surfzone bars, which may interact with shoreline evolution, so that a separate evaluation is not obvious. In any specific case, higher-order effects that provide essential information to the lower-order scales must be identified.

The above reasoning implicitly suggests that we can describe the morphological state, behavior and identity of the system at each order over the full temporal scale of interest, given the external forcings and constraints. In practice, a complete description may not be possible due to a variety of limitations, such as lack of process knowledge, insufficient or uncertain information on the forcing, inherent unpredictable behavior, or practical limits such as insufficient computer power. Whatever of these limitations, either theoretical or practical, we still need to aggregate the knowledge of the dynamics of the higher-order systems into 'useful' information for the lower-order systems. That is, we need to find methods that allow us to sidestep theoretical and practical problems so that this useful information can be represented 'in some way' at the next lower order. Such pragmatism underlies the philosophy behind the behavior-oriented approach to modeling (STIVE *et al.*, 1995).

SUMMARY AND CONCLUSIONS

In an ideal world we would use knowledge of sediment dynamics based on first principles of physics to predict long-term change of shoreline position, in particular, and evolution of coastal morphology in general. Net sediment transport is, after all, the cause of all change in coastal morphology. Unfortunately we are unlikely ever to succeed with such an approach, for reasons that have been clear now for some time (DE VRIEND *et al.*, 1993; COWELL and THOM, 1994; DE VRIEND, this volume). We had at our disposal from the outset of this study, numerical tools (so-called behavior-oriented models) that provide an alternative approach to the prediction of long-term coastal change. These models encapsulate various elements of aggregated sediment dynamics and morphological behavior for coasts evolving over decades, centuries and millennia. What was lacking, however, was a systematic framework of where and how to apply these models, the relationships between the models themselves, and how each represents various aspects of low-order coastal behavior. Consequently, application of these models has required considerable expertise, but this expertise has been largely subjective.

We have addressed this methodological dilemma by devising three related concepts: (1) the coastal tract; (2) the coastal-tract cascade; and (3) coastal-tract templating. Implementing these concepts for any site-specific problem requires resolution of three issues:

(1) Distinguish which morphological features belong to which order, and discriminate their internal dynamics from external variables (cascade templating).
(2) Delimit the spatial extent of the coastal cell (cell templating).
(3) Transform the data set so that the tract has the same spatial dimensions as the predictive model and contains comparable variables or parameters (data-set templating), which involves
   (a) aggregation of higher-order dynamics into useful information for the lower-orders, which involves either a theoretical, or, if process knowledge is insufficient, a practical choice; and
   (b) mapping (representation of) the useful sub-scale information onto the scale of interest.

The practical benefits of these concepts for coastal management on time scales of decades to centuries are as follows.

- The *coastal-tract* concept provides a necessary framework for linking coasts in nature to predictive models. The concept does this by explicitly recognizing the fundamental modes of coastal change. This is a first and necessary step in prediction of low-order modes separately from higher frequency changes (*i.e.,* changes related to short-term disturbances such as storms or the evolution higher-order features such as surf-zone bars, dunes and tidal-channels). Characterization of low-order processes not only permits a physical explanation of long-term trends. It also places higher-order changes into an overall context. Apart from providing a more coherent basis for prediction of long-term change, recognition and understanding of coastal-tract behavior also should assist coastal managers to distinguish acute from chronic coastal-change problems, and to select more appropriate remedial or planning strategies in any specific case accordingly.
- The *coastal-tract cascade* formalizes concepts for separating coastal processes and behavior into a scale hierarchy, with the coastal tract as the most fundamental mode at the base of the hierarchy. The purpose of the cascade is to provide a systematic basis for distinguishing, on any level in the hierarchy, those processes that must be included as internal variables in modeling coastal change, from those that constitute boundary conditions (at lower-order), and yet others that may be regarded as unimportant 'noise' (at higher order). The criterion for internal variables is that they relate to features and processes that form a system sharing a common pool of sediment. Apart from its importance in guiding model application, the cascade concept discriminates between cause and effect on the basis of scale. This also should be a practical benefit for problem definition: *i.e.,* by explicitly separating components of the coastal-change problem, there should be less confusion and greater transparency concerning what is, and what is not being modelled (through adequate documentation of templating procedures and general recognition in coastal management that it is seldom possible to model or analyse all aspects of coastal change simultaneously).
- *Coastal-tract templating* provides a protocol for defining a site-specific problem (in coastal management, engineering or research). The protocol is a procedure by which data are interpreted or transformed into a *data-model.* The procedures give the data on the site (or region) the same structure and dimensions as the process model(s) to be used in predictive and explanatory simulations. The data-model thus defines the extent of coastline represented in any simulation-modeling exercise. It also defines the level in the cascade hierarchy that the modeling addresses. Templating procedures therefore generally involve spatial averaging (to reduce dimensions consistent with the process model), and identification of internal, external and unimportant variables (hence coastal-tract cascade templating). No universal recipe for templating can exist however, since each new problem requires its own templating. Because of this, scientific expertise remains an unavoidable requirement in coastal modeling, and the validity of modeling results remain contingent upon it. Although templating always occurs in a modeling exercise, albeit tacitly, formalizing these procedures as part of modeling protocols improves the rigor in coastal-change modeling overall: not least through recognition that templating procedures must always be documented in reporting modeling results.

ACKNOWLEDGMENTS

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LITERATURE CITED

BEETS, D.J.; VAN DER VALK, L., and STIVE, M.J.F., 1992. Holocene evolution of the coast of Holland. *Marine Geology,* 103, 423–443.

BROWN, L.F. and FISHER, W.L., 1977. Seismic stratigraphic interpretation of depositional systems: examples from Brazil rift and pull-apart basins. *American Association of Petroleum Geologists Special Publication,* 26, 213–248.

CAPOBIANCO, M.; STIVE, M.J.F.; JIMENEZ, J.A., and SANCHEZ-ARCILLA, A., 1998. Towards the definition of budget models for the evolution of deltas. *Journal of Coastal Conservation,* 4, 7–16.

CARTER, R.W.G., 1988. *Coastal Environments: An Introduction to the Physical, Ecological and Cultural Systems of Coastlines.* London: Academic, 617p.

CHORLEY, R.J.; SCHUMM, S.A., and SUGDEN, D.E., 1984. *Geomorphology.* New York: Methuen, 498p.

COWELL, P.J. and THOM, B.G., 1994. Morphodynamics of coastal evolution. *In:* CARTER, R.W.G. and WOODROFFE, C.D. (eds), *Coastal Evolution: Late Quaternary Shoreline Morphodynamics.* Cambridge: Cambridge University Press, pp. 33–86.

COWELL, P.J.; HANSLOW, D.J., and MELEO, J.F., 1999a. The shoreface. *In:* SHORT, A.D. (ed.), *Handbook of Beach and Shoreface Morphodynamics.* Chichester: Wiley, pp. 37–71.

COWELL, P.J.; ROY, P.S.; CLEVERINGA, J., and DE BOER, P.L., 1999b. Simulating coastal systems tracts using the shoreface translation model. *Society of Economic Palaeontologists and Mineralogists Special Publication,* 41. 62, 165–175.

COWELL, P.J.; ROY, P.S., and JONES, R.A., 1995. Simulation of LSCB using a Morphological Behaviour Model. *Marine Geology,* 126, 45–61.

COWELL, P.J.; STIVE, M.J.F.; NIEDORODA, A.W.; SWIFT, D.J.P.; DE VRIEND, D.J.; BUIJSMAN, M.C.; NICHOLLS, R.J.; ROY, P.S.; KAMINSKY, G.M.; CLEVERINGA, J.; REED, C.W., and DE BOER, P.L., 2003. The Coastal-Tract (Part 2): Applications of aggregated modeling to lower-order coastal change. *Journal of Coastal Research* (this volume).

COWELL, P.J.; STIVE, M.J.F.; ROY, P.S.; KAMINSKY, G.M.; BUIJSMAN, M.C.; THOM, B.G., and WRIGHT, L.D., 2001. Shoreface Sand Supply to Beaches. *Proceedings 27th International Conference on Coastal Engineering* (American Society of Civil Engineers), pp. 2495–2508.

CURRAY, J.R., 1964. Transgressions and regressions. *In:* MILLER, R.C. (ed.), *Papers in Marine Geology.* New York: McMillan, pp. 175–203.

DEAN, R.G. and MAURMEYER, E.M., 1983. Models of beach profile response. *In:* KOMAR, P. and MOORE, J. (eds), *CRC Handbook of Coastal Processes and Erosion.* Boca Raton: CRC Press, pp. 151–165.

DE VRIEND, H.J., 1998. On the predictability of coastal morphology. *In:* BARTHEL, K.-G. *et al.* (eds.), *Third European Marine Science and Technology Conference—Conference Proceedings.* Luxemburg: European Communities, pp. 289–300.

DE VRIEND, H.J., 2003. On the prediction of aggregated-scale coastal evolution. *Journal of Coastal Research,* (this volume).

DE VRIEND, H.J.; CAPOBIANCO, M.; CHESHER, T.; DE SWART, H.E.; LATTEUX, B., and STIVE, M.J.F., 1993. Approaches to long-term modelling of coastal morphology: a review. *Coastal Engineering,* 21, 225–269.

FISHER, W.L. and MCGOWAN, J.H., 1967. Depositional systems in the Wilcox Group of Texas and their relationship to the occurrence of oil and gas. *Gulf Coast Association of Geological Societies Transactions,* 18, 105–125.

GELFENBAUM, G.; SHERWOOD, C.R.; PETERSON, C.D.; KAMINSKY, G.M.; BUIJSMAN, M.; TWICHELL, D.C.; RUGGIERO, P.; GIBBS, A.E., and REED, C., 1999. The Columbia River littoral cell: A sediment budget overview, *Proceedings of Coastal Sediments '99* (Long Island, New York, American Society of Civil Engineers), pp. 1660–1675.

HAIGH, M.J., 1987. The holon: hierarchy theory and landscape research. *In:* AHNERT, F. (ed.), *Geomorphological Models: Theoretical and Empirical Models.* Catena Supplement 10. Catena Verlag, pp. 181–192.

HALLERMEIER, R.J., 1981. A profile zonation for seasonal sand beaches from wave climate. *Coastal Engineering,* 4, 253–277.

KOMAR, P.D., 1973. Computer models for delta growth due to sediment input from rivers and longshore transport. *Geological Society of America Bulletin,* 84, 2217–2226.

LEATHERMAN, S.P., 1983. Barrier dynamics and landward migration with Holocene sea-level rise. *Nature,* 301, 415–18.

NICHOLLS, R.J.; BIRKEMEIER, W.A., and LEE, G-H., 1998. Evaluation of depth of closure using data from Duck, NC, USA. *Marine Geology,* 148, 179–201.

NIEDORODA, A.W; REED, C.W.; SWIFT, D.J.P.; ARATO, H., and HOYANAGI, K., 1995. Modeling shore-normal large-scale coastal evolution. *Marine Geology,* 126, 1/4, 181–200.

NIELSEN, P., 1992. *Coastal Bottom Boundary Layers and Sediment Transport.* Advanced Series on Ocean Engineering Vol. 4. Singapore: World Scientific, 324p.

NITTROUER, C.A. and STERNBERG, R.W., 1981. The Formation of Sedimentary Strata in an Allochthonous Shelf Environment: The Washington Continental Shelf. *Marine Geology,* 42, 201–232.

PARSONS, B.S.; SWIFT, D.J.P., and WILLIAMS, K., in press. Quaternary Facies Assemblages and their bounding surfaces, Chesapeake Bay mouth: Implications for stratigraphic analysis. *Journal of Sedimentary Research.*

PILKEY, O.H.; YOUNG, R.S.; RIGGS, S.R.; SMITH, A.W.; WU, H., and PILKEY, W.D., 1993. The concept of shoreface profile equilibrium: A critical review. *Journal of Coastal Research,* 9, 255–278.

ROELVINK, J.A. and STIVE, M.J.F., 1991. Sand transport on the shoreface of the Holland coast. *Proceedings 22nd International Conference on Coastal Engineering,* New York: American Society of Civil Engineers, pp. 1909–21.

RHOADS, B.L. and THORN, C.E., 1996. Toward a Philosophy of Geomorphology. *In:* RHOADS, B.L. and THORN, C.E. (eds), *The Scientific Nature of Geomorphology.* Chichester: Wiley, pp. 115–143.

ROY, P.S.; COWELL, P.J.; FERLAND, M.A., and THOM, B.G., 1994. Wave dominated coasts. *In:* CARTER, R.W.G. and WOODROFFE, C.D. (eds), *Coastal Evolution: Late Quaternary Shoreline Morphodynamics.* Cambridge: Cambridge University Press, pp. 121–86.

SLOSS, L.L., 1962. Stratigraphical models in exploration. *Journal of Sedimentary Petrology,* 32, 415–422.

STERNBERG, R.W., 1986. Transport and accumulation of river-derived sedimentation the Washington continental shelf, USA. *Journal of the Geological Society* (London), 143, 945–956.

STIVE, M.J.F. and DE VRIEND, H. J., 1995. Modelling shoreface profile evolution. *Marine Geology,* 126, 235–248.

STIVE, J.F.; CAPOBIANCO, M.; WANG, Z.B.; RUOL, P., and BUIJSMAN, M.C., 1998. Morphodynamics of a tidal lagoon and the adjacent coasts. *Proceedings of the 8th International Biennial Conference on Physics of Estuaries and Coastal Seas,* DRONKERS, J. and SCHEFFERS, M., (eds). The Hague, 1996, pp. 397–407.

STIVE M.J.F.; DE VRIEND, H.J.; COWELL, P.J., and NIEDORODA, A.W., 1995. Behavior-oriented models of shoreface evolution. *Proceedings Coastal Dynamics '95* (American Society of Civil Engineers), pp. 998–1005.

STIVE, M.J.F.; ROELVINK, D.J.A., and DE VRIEND, H.J., 1991. Large-scale coastal evolution concept. *Proceedings 22nd International Conference on Coastal Engineering* (New York, American Society of Civil Engineers), pp. 1962–74.

SWIFT, D.J.P., 1976. Continental shelf sedimentation. *In:* STANLEY, D.J. and SWIFT, D.J.P., (eds.) *Marine Sediment Transport and Environmental Management.* New York: Wiley, pp. 311–350.

THIELER, E.R; PILKEY, O.R.; YOUNG, R.S.; BUSH, D.M., and CHAI, F., 2000. The use of mathematical models to predict beach behavoir for U.S. coastal engineering: A critical review. *Journal of Coastal Research,* 16, 48–70.

THORNE, J.A. and SWIFT, D.J.P., 1991. Sedimentation on continental margins, II: application of the regime concept. *In:* SWIFT, D.J.P.; OERTEL, G.F.; TILLMAN, R.W., and THORNE, J.A. (eds), *Shelf Sand and Sandstone Bodies—Geometry, Facies and Sequence Stratigraphy.* Special Publication No. 14 of the International Association of Sedimentologists. Oxford: Blackwell Scientific Publications, pp. 33–58.

WANG, Z.B.; KARSSEN, B.; FOKKINK, R.J., and LANGERAK, A., 1996. A dynamic/empirical model for estuarine morphology. *Proceedings of the 8th International Biennial Conference on Physics of Estuaries and Coastal Seas,* 1996. DRONKERS, J. and SCHEFFERS, M. (eds). The Hague, pp. 279–286.

WOODROFFE, C.D.; CHAPPELL, J.; THOM, B.G., and WALLENSKY, E.,

1989. Depositional model of a macrotidal estuary and floodplain, South Alligator River, Northern Australia. *Sedimentology*, 36, 737–56.

WRIGHT, L.D., 1995. *Morphodynamics of Inner Continental Shelves.* Boca Raton: CRC Press, 241p.

ZHANG, Y.; SWIFT, D.J.P.; NIEDORODA, A.W.; REID, C.W., and THORNE, J.A., 1997. Simulation of sedimentary facies on the Northern California Shelf: implications for an analytical theory of facies differentiation. *Geology*, 27, 635–638.

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The Coastal-Tract (Part 2): Applications of Aggregated Modeling of Lower-order Coastal Change

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ABSTRACT

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The *coastal-tract* approach to coastal morphodynamics, described in the companion paper (The Coastal-Tract Part 1), provides a framework for aggregation of process and spatial dimensions in modeling low-order coastal change (*i.e.*, evolution of the shoreline, continental shelf and coastal plain on time scales of 10^2 to 10^3 years). Behavior-oriented, coastal-change models encapsulate aggregate dynamics of the coastal tract. We apply these models in a coastal-tract framework to illustrate the use of the concept, and to explore low-order morphological coupling under different environmental settings. These settings are characterized by data-models that we have constructed from four contrasting continental margins (NW Europe, US Pacific, US Atlantic, and SE Australia).

The gross kinematics of the coastal tract are constrained and steered by sediment-mass continuity. The rate of coastal advance or retreat is determined quantitatively by the balance between the change in sediment accommodation-space, caused by sea-level movements, and sediment availability. If the lower shoreface is shallower than required for equilibrium (negative accommodation), then sand is transferred to the upper shoreface (NW Europe, US Pacific, and SE Australian cases modelled) so that the shoreline tends to advance seaward. This tendency also occurs when relative sea level is falling (coastal emergence). Coastal retreat occurs when the lower shoreface is too deep for equilibrium (positive shoreface accommodation). This sediment-sharing between the upper and lower shoreface is an internal coupling that governs first-order coastal change. The upper shoreface and backbarrier (lagoon, estuary or mainland) also are coupled in first-order coastal change. Sediment accommodation-space is generated in the backbarrier by sea-level rise (and reduced by sea-level fall), but the amount of space is also moderated by influx of fine sediments from the coast, or sand and mud from fluvial sources. Remaining space can then be occupied by sand transferred from the upper shoreface causing a retreat of the latter (transgressive phases modelled for NW Europe, US Atlantic, and SE Australian cases).

ADDITIONAL INDEX WORDS: *Aggregate change, shoreface, backbarrier, lagoon, scale, coastal tract, coastal-tract cascade, templating, data-model, behavior-oriented models, morphological coupling, sea level, sediment supply, coastal evolution, coastal management, sea-level rise, transgression, regression, barrier, continental-shelf, sediments, accommodation-space, numerical model.*

INTRODUCTION

In the companion paper, *Part 1* (COWELL *et al.*, 2003, this volume), we have described concepts of the *coastal tract*, *coastal-tract cascade* and *coastal-tract templating*. They provide a framework for modelling coastal morphodynamics on a wide range of scales (from seconds to millennia). We use these principles here (in Part 2), by reference to large-scale models and their application, to elucidate the aggregate-scale processes governing coastal-tract evolution. Further details can be obtained from earlier works describing these models (Table 1).

The paper is aimed also at demonstrating the common el-

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ements of low-order coastal change. As outline in Part-1, low-order behavior entails coupling of the upper shoreface to both the backbarrier and the lower-shoreface (*i.e.*, continental shelf). Illustrations presented below draw from our comparative modeling of coasts from a wide range of settings in Australia, North America and Europe. In each case we have assumed alongshore homogeneity, so the illustrations do not deal with the more complicated cases involving heterogeneity referred to in Part 1. These were beyond the scope of our modeling so far and are the subject of ongoing research.

Scale and Forcing

Beyond the annual time scale in the coastal ocean, with rare exception more than enough power exists to entrain and transport sediment. Most transport involves to-and-fro displacements of sediments with very small net deposition or erosion. Measurable changes in coastal morphology results from these small net amounts accumulating over many years. Thus, prediction of processes acting over time scales of decades and longer entails the difficult task of resolving small net changes in a system characterized by large fluctuations (DE VRIEND, 1998, and 2003, this volume).

Because limited opportunity exists to measure environmental forcing in detail over such long periods, we must also rely on generalized representations of waves and currents, and assume applicability over long distances: *i.e.*, long-term change involves large length scales (Part 1). Figure 1a shows forcing scales ranging from individual storm events to changes in worldwide climate or sea level. The progression from event time scales ($< 10^0$ yrs) to millennia demands increasing generalization of theory and data, but generalization entails decreased resolution achievable from measurements and predictions (Figure 1a). Thus, *coastal-tract templating* procedures require an appreciation of coastal-change dependence on time-, and length-averaged quantities.

Changes that occur as the result of a single storm event can be studied in terms of the sediment movement brought about by a time series of waves, currents and water levels (sub-event scales $\leq 10^{-1}$ in Figure 1a). However, net morphological changes related to a sequence of storm and calm intervals tend to be less than those of each individual event so that analyses depending on summing individual event responses is prone to error. Thus, the sum of forcing-parameter variations may be characterized better as a statistical climatology. This allows the system response to be assessed in relation to the whole range of forcing conditions. Although this approach sidesteps the residual-error problem, the gain in reliability has a cost of reduced precision. That is, predictions based on such an approach can approximate average future conditions only, rather than individual realizations for a given sequence of events.

Representation of Aggregate Forcing

The large time scales relevant to the coastal-tract favor adoption of the highly aggregated approach to forcing embodied in *the sedimentation-regime* concept formalized by THORNE and SWIFT (1991) and THORNE *et al.* (1991) based on SLOSS (1962). The theory states that, over long time periods, the controls on coastal deposition are summarized by the so-called *Sloss variables* that include a) accommodation space available for deposition, b) sediment supply (rate and composition), and c) intensity of sediment transport. Variation (loss or gain) in accommodation space occurs due to changes in relative sea level. Sediment supply rates and composition (principally range of grain sizes) depend on deltaic, shelf and littoral sources and sinks outside the tract. The sediment-transport intensity governs sediment-dispersal mechanisms within a tract, but this third Sloss variable obviously also affects sediment supply.

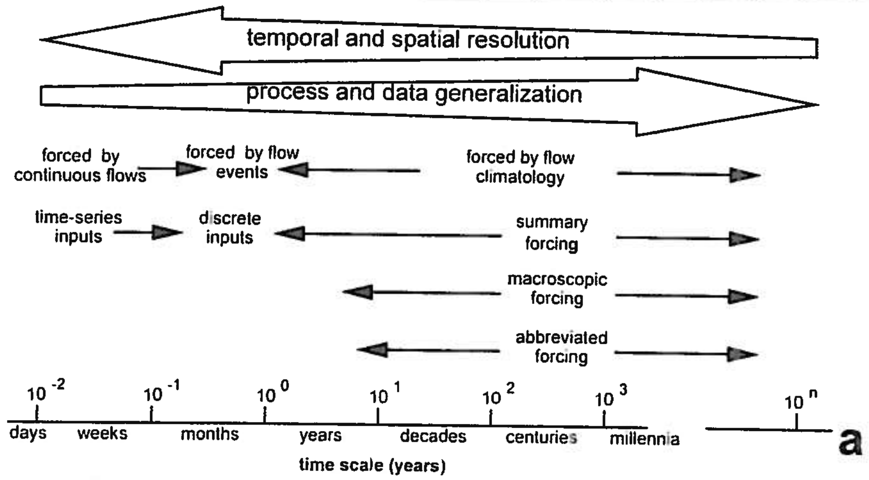
Table 1. *Existing Large-Scale Behavior Models encapsulating principles of elements within the coastal tract. (CP = Cross-shore Profile. IDV = 1 Dimensional Vertical representation of depth-integrated sediment dynamics.)*

Acronym	Model Name	Type	Reference
ADM	Advection-Diffusion Model	CP model with 1DV macroscopic sediment dynamics	NIEDORODA *et al.* (1995)
HPM	Hinged Panel Model	CP model with 1DV input-filtered sediment dynamics	STIVE and DE VRIEND (1995)
STM	Shoreface Translation Model	CP model of morphokinematics	COWELL *et al.* (1995)
ASMITA	Aggregated Scale Morphological Interaction between a Tidal-inlet system and the Adjacent coast	Box model with spatially lumped, macroscopic sediment dynamics	BUIJSMAN (1997) STIVE *et al.* (1998)

The regime concept postulates that equations of state can be derived in terms of the Sloss variables. Thus, Sloss variables are macroscopic parameters that provide a highly aggregated (bulk) representation of conditions forcing the coastal tract. Various approaches related to this concept have been proposed (Figure 1a) such as *macroscopic forcing* (NIEDORODA *et al.*, 1995), *summary forcing* (STIVE and DEVRIEND 1995), and *abbreviated forcing* (COWELL *et al.*, 1995). In the first approach the forcing is reduced to time-, and space-averaged conditions which include a sediment-dispersal regime that reflects the general flow climatology and its variation across the continental shelf. In the second approach, the forcing is summarized according to key, representative conditions such as the power-equivalent wave and a long time-averaged current (*input filtering* of DE VRIEND *et al.*, 1993). In the last approach, the forcing is abbreviated to a reduced set of Sloss variables: process parameters aggregated in terms of sediment *accommodation* space, sediment supply and morphological regime.

MODELS AND METHODS

The concept of the *coastal tract* emerged through our development of numerical models for long-term coastal processes (Table 1). These models compete and support each other by adopting different methods for aggregating coastal features and processes into sub-systems (*i.e.*, comprising the coastal tract). These models have one thing in common how-



Model	Spatial Representation	Time Scale (years)	Forcing	Variables
Event	1-, 2-, or 3-D	10^{-2} - 10^{-1}	Direct/ continuous	Meteorological & Oceanographic
Event Sequence		10^{0} - 10^{1}	Direct/ Statistical	
ADM	Depth- & Width- Averaged	10^{2} - 10^{3}	Macroscopic	Sloss Variables
HPM			Summary	
STM			Abbreviated	Reduced Sloss Variables
ASMITA	System Averaged (non-spatial)		Summary	Sloss Variables
Hybrids	generally space averaged		as required	as required

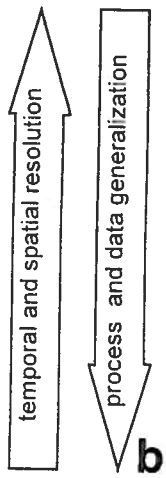


Figure 1. Scale and generalization in a) representation of *coastal-tract* forcing, and b) modeling framework for *coastal-tract* analyses.

ever: their sub systems share sediments and thereby interact dynamically (*morphological coupling*) on time scales of change related to size of sub systems. As part of the PACE project, the models were tested and compared, with the conclusion that each approach has contrasting strengths and weaknesses that collectively yield insights into coastal-tract behavior.

Scale and Process Representation: Comparative Modeling

Figure 1b summarizes the modeling approaches that can be applied to the coastal tract. Conceptually this summary extends Figure 1a by specifying the models with respect to their representation of forcing, governing variables and spatial dimensions.

Event-scale Models

Event-scale models are the smallest scale in the range, of which two types have been defined: a) *initial sedimentation / erosion* (ISE) models and b) *medium-term morphodynamic* (MTM) models (DE VRIEND *et al.*, 1993). Both include algorithms that provide computation of bed changes from sediment continuity based on transport fluxes driven by waves and currents. The ISE approach does not feed back the computed bed changes into another cycle of computation. Although useful as a diagnostic tool, the ISE approach cannot deal with long-term predictions involving successive morphology changes (*i.e.*, coastal evolution).

Models of the MTM variety are designed to cope with this problem. These models however can suffer from inadequately proscribed boundary conditions, numerical instabilities that promote growth of spurious morphological features, and un-

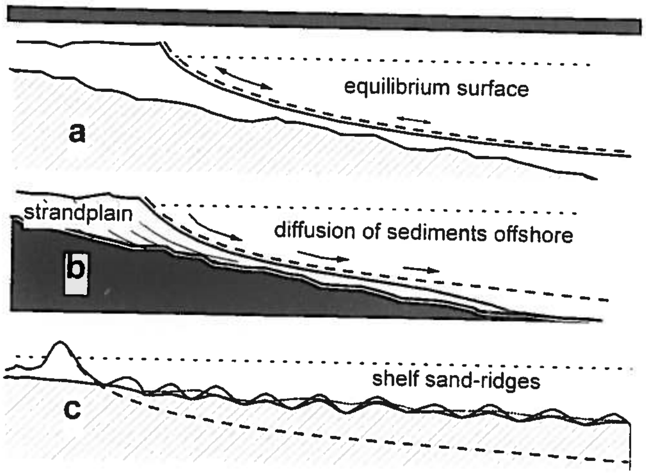


Figure 2. Inherited shelf slope compared to equilibrium shoreface after a period of sea-level stillstand and profile evolution: a) fully adjusted shoreface, b) shelf steeper than equilibrium shoreface providing sediment accommodation space offshore and deltaic sediment supply at the coast, and c) shelf much shallower than equilibrium shoreface causing negative accommodation space offshore.

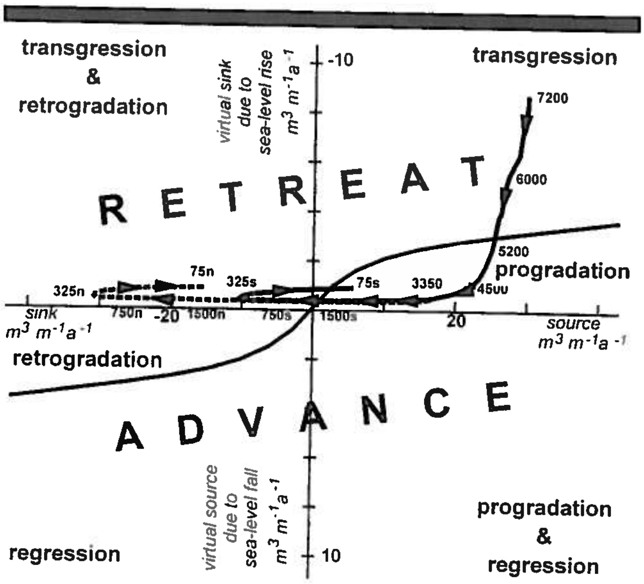


Figure 3. Evolution of the central Netherlands coast (Hoek of Holland to Den Helder) as a time trajectory in sediment-supply/accommodation phase space (abscissa and ordinate respectively, scaled in cubic meters per year per meter of shoreline). Numbers along the trajectory indicate time (years BP): suffixes n and s denote north and south of Haarlem respectively.

expected interactions between system components that cannot be anticipated from knowledge of how the separate parts behave in isolation (DE VRIEND *et al.*, 1993). Furthermore, there are difficulties in representing inputs (forcing and boundary conditions) over long periods of time. The problem has been addressed by reducing these inputs to statistical summaries (*e.g.*, SHERWOOD, 1995). Because most work is done by large storms however, it is presently not clear how to weight the statistical representation of inputs appropriately: the classical *magnitude-frequency* problem in geomorphology (WOLMAN and MILLER, 1960). Even if these problems can be suppressed, the non-linearity of the large-scale processes must lead to progressive departures of predictions from real behavior: *i.e.*, because the latter is a time series of unique realizations (ensuing from a climatological history), whereas the former are merely estimates of central tendency for a system sensitive to initial conditions.

Nevertheless, event-scale modeling is useful as a diagnostic tool in coastal-tract modeling. It can be applied with large-scale models that require internal calibration and estimates for boundary conditions. That is, synthetic data computed from event-scale models can be used to supplement incomplete field-data sets if adequate representation of the physics exist for the small-scale processes; and these may well include subtle but important effects such as bed armouring, wave-ripple effects, and 'lidding' of the wave boundary layer (REED *et al.*, 1999).

Aggregate-scale Models

For long-term predictions, we can by-pass the problem of having to predict and time-integrate very small differences in highly variable systems. This is achieved by deriving macroscopic variables (*e.g.*, Sloss variables) that define the physics directly on the scale of interest. For these systems, ISE processes simply become noise, although estimates of them inform us about the intensity of sedimentation regime.

Our physical concepts of the *coastal tract* derive significantly from ideas underlying the Sloss variables. Accommodation space can be either positive (under-filled) or negative (over-filled) and commonly varies spatially over the profile (Figure 3 in Part 1). The concept applies equally to the volume created on a shelf (Figure 2) or within an estuary by sea-level rise that exceeds the sedimentation rate. Available hydrodynamic power controls the rate of morphological change. Characterization of the power changes according to the system, but generally takes some form of aggregated wave- or flow-climatology. The grain size and supply rate of sediment are usually externally controlled. The detail at which these are quantified depends on the nature of the coastal-tract system under investigation.

Large-scale models shown on Figure 1b utilize the Sloss-variables, but in different ways (see Table 1 for acronyms, details and documentation). The HPM incorporates all the Sloss-variables: translation of the form-invariant upper shoreface occurs in response to interaction between sediment supply and changes in accommodation space due to sea-level variations. Depth-averaged variables representing the power-equivalent wave and current environment govern sediment transfer between hinged panels that characterize the lower, middle and upper shoreface (STIVE and DE VRIEND, 1995).

The ADM also contains the Sloss variables although it does not employ representative hydrodynamic power in transporting sediments. Rather, the model aggregates transport events into macroscopic processes (advection and diffusion) that are calibrated against measures of hydrodynamic power for specific sites (CAREY *et al.*, 1999). Again there is an interaction between sediment supply (positive or negative) and sea-level related change in accommodation space, with responses governed by the advection-diffusion processes.

The STM operates with a reduced set of Sloss variables since the hydrodynamic forcing is expressed through morphological regime: *i.e.,* sediment-transport intensity represented implicitly via coastal-profile geometry that varies through time with forcing parameters. Geometries for the upper shoreface, lower shoreface, and backbarrier are calibrated against field data, including radiometrically measured morphological history and hindcastings of associated wave and tide regimes, using empirical relations and other event-scale models (COWELL *et al.*, 1999b, 2003). The balance between changes in process-related profile geometry, accommodation-space (due to changes in sea level), and sediment supply governs profile kinematics.

ASMITA bears a similarity to the panel model and also the ESTMORF model for tidal basins (WANG *et al.*, 1996). Accommodation-space is represented by a series of control volumes (or panel elements) for the shoreface, ebb- and flood-tide deltas, tidal-inlet channel, and tidal-basin channel-network. Morphodynamic change in each element is specified by time-averaged physics that are related to the Sloss variables. An equilibrium system is developed and then perturbed to study the responses of the system elements and the interaction between these elements.

Comparative tests revealed different strengths and weaknesses among these models. Work in progress thus aims to combine the individual modeling approaches and the models themselves. An example is the nested ASMITA-ADM hybrid (see Figure 12 further on) that simulates lagoon-shoreface coupling over space and time scales not possible using either model alone.

Data-Model Templating

Coastal-tract delineation for a site specific case requires development of a *data-model* that aggregates process and morphological properties as they exist in nature (Part 1). Application of numerical models and empirical analyses to predict site-specific, low-order, coastal change both require a data-model before prediction can proceed. The data-model links the process model to that part of nature represented in the process model. Such a linkage is necessary to calibrate the process model and to assess its results.

Generally, data-model development occurs intuitively without explicit recognition as a formal step in coastal research. The risk here is that ambiguities occur during analysis, modeling and interpretation of results, introducing additional uncertainty into coastal-change prediction, as well as tensions between practitioners of data-driven and model-driven approaches to large-scale coastal behavior (*cf.,* PILKEY, 1993). The coastal-tract concept therefore seeks to formalize procedures for linking coasts in nature to models that have much lower dimensionality: *i.e., templating.*

First-order change in the coastal tract is constrained by the environmental setting: *i.e.,* zero-order in the tract-cascade (Part 1). The environmental setting depends mainly upon sediment supply and steepness of the continental shelf and the hinterland (Figure 2). The shelf steepness determines how far the shoreface was from equilibrium at the onset of sea-level stillstand (or near stillstand) at the end of the post-glacial marine transgression (5 to 6 ka BP). Steeper shelves and hinterlands are more likely to have shorefaces that are too deep to be in equilibrium for a given coastal-ocean climatology and endemic sediment sizes (Figure 2b). Backshores under these conditions are more likely to be a mainland beach than a barrier lagoon (see Figure 1b in Part 1). Parts of the southeast Australian coast (ROY *et al.*, 1994) and much of the Pacific US coast (KAMINSKY *et al.*, 1997) have this type of setting.

The converse is true for low gradient shelves and hinterlands where a) the shelf surface is more likely to be shallower than the equilibrium shoreface, and b) the backshore is more likely to comprise a lagoon, in-filled to some degree depending on sediment supply (Figure 2c). The coasts of eastern USA (SWIFT, 1976) and northwest Europe (BEETS *et al.*, 1992) are examples. The stage of coastal evolution therefore depends on a) how far the shoreface was from equilibrium at the end of the post-glacial marine transgression, b) the size of the lagoon at that time, and c) the sediment supply since then.

COASTAL-TRACT MODELING

Description of our comparative modeling provides a vehicle to elucidate the general principles outlined in Part 1 regarding coastal-tract behavior. This objective is enhanced because the modeling exploited data sets from a wide range of environments, allowing us to study low-order change under contrasting conditions.

The coastal-tract principles are not entirely new: they were implicit in earlier kinematic models for first approximation of long-term coastal change, and in related ideas. In a narrow sense, the Bruun Rule is an upper-shoreface response to a reduced set of Sloss variables, that include sea-level rise and sediment supply, with the response involving sand transfers between the shoreface and the backshore or backbarrier (DEAN and MAURMEYER, 1983). In a broader sense, SWIFT (1976) extended CURRAY's (1964) ideas into a general framework for long-term coastal change entailing *transgression* (landward retreat) and *regression* (seaward advance) of the shoreline due to sea-level rise and fall, with corresponding tendencies toward *retrogradation* and *progradation* due to net sediment losses or inputs alongshore.

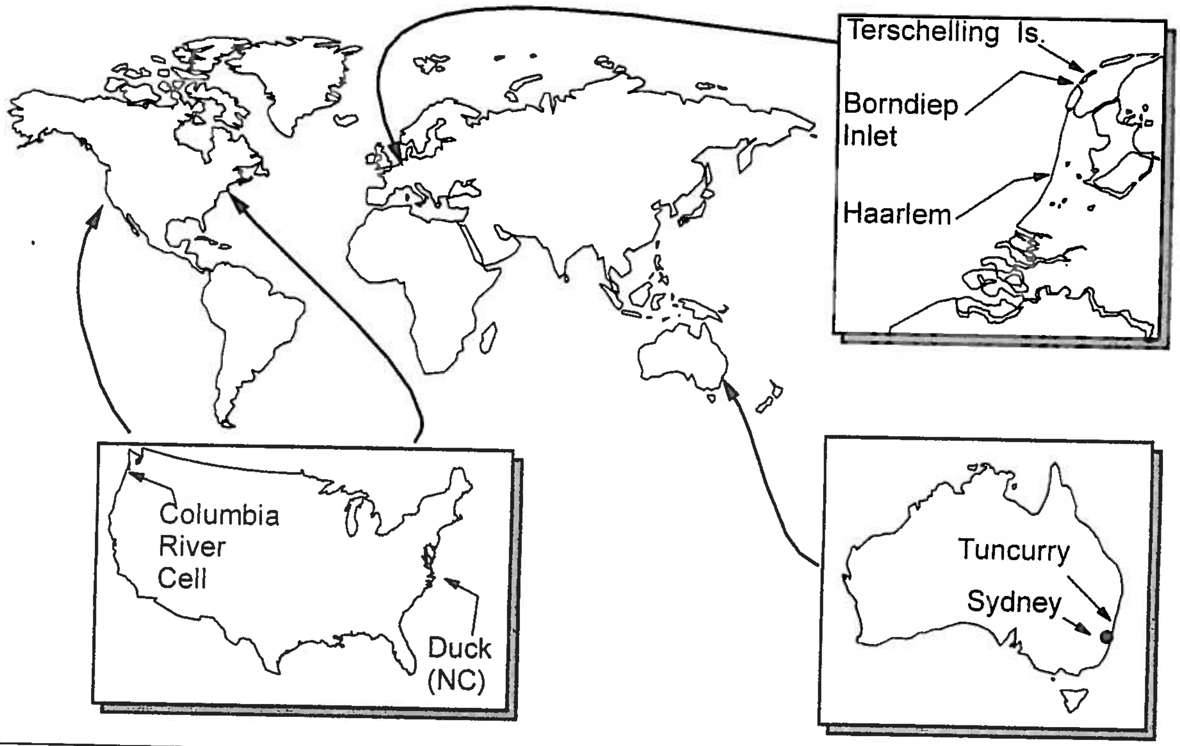
Overall therefore, the kinematics of the coastal tract schematized by Figure 3 in Part 1 are constrained and steered by sediment-mass continuity in response to the Sloss variables. We consider the general kinematics formally in the following section, after which we examine more specific modes of tract behavior. The examination proceeds within the coastal-tract framework (Part 1) through application of computer models to morphostratigraphic data sets (identified in Tables 1 and 2 respectively).

Gross Coastal Tract Kinematics

SWIFT's (1976) concepts can be quantified in the context of the coastal tract, and related back to the familiar *Bruun Rule*

Table 2. *Morphological and stratigraphic data sets used in comparative modeling and illustration of aggregated processes within the coastal tract, and their locations.*

Data Set	Comparative Properties	Summary Reference
Netherlands: Haarlem, Central Holland	Closed barrier (inlet-free) coast with low-gradient shelf and low (−ve) sediment input	BEETS, *et al.*, 1992; STIVE and DE VRIEND, 1975
Netherlands: Terschelling, Wadden Friesland	Barrier-inlet tidal basin coast with lo-gradient shelf and low (+ve) sediment input	SHA, 1992; VAN DER SPEK, 1994
SE Australia: Tuncurry Bay	Closed barrier (inlet-free) coast with steep shelf and low (+ve) sediment input	ROY *et al.*, 1994, 1997
NW USA: Columbia River coastal cell (Tillamook Head to Point Grenville)	Mixed closed and barrier-inlet coast with steep shelf and moderate (+ve) sediment input	STERNBERG, 1986; PETERSON and PHIPPS, 1992; WOLF *et al.*, 1997; KAMINSKY *et al.*, 1997
E. USA: Duck NC	Mixed closed and barrier-inlet coast with steep shelf and sediment input during post glacial transgression	FIELD *et al.*, 1979; MEISBURGER *et al.*, 1989



(BRUUN, 1962), if we consider the sediment balance of the upper shoreface. We adopt the assumption that the upper shoreface to a first approximation is form invariant relative to mean sea-level over time periods ($\gg$1 yr) for which profile closure occurs (NICHOLLS *et al.*, 1998). We can represent the upper shoreface by an arbitrary, but usually concave-up, profile $h(x)$ to a depth h_* and a length L_*, in which x is the distance from the shore (DEAN, 1991). Sediment-volume conservation for profile kinematics requires that

$$\frac{\partial h}{\partial t} + c_p \frac{\partial h}{\partial x} = 0 \qquad (1)$$

or via $h = MSL - z_b$

$$\frac{\partial z_b}{\partial t} + c_p \frac{\partial z_b}{\partial x} = \frac{\partial MSL}{\partial t} \qquad (2)$$

where c_p is the horizontal translation rate of the shoreline position. The sediment-transport balance equation for a fixed spatial control volume is

$$\frac{\partial z_b}{\partial t} + \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} + s = 0 \qquad (3)$$

where $q_{x,y}$ are the cross-shore and alongshore sediment transports, and s is a local source or sink. These equations may be combined to yield

$$c_p = \frac{\partial MSL}{\partial t}\left(\frac{\partial h}{\partial x}\right)^{-1} - \frac{\partial q_x}{\partial h} - \frac{\partial q_y}{\partial y}\left(\frac{\partial h}{\partial x}\right)^{-1} - s\left(\frac{\partial h}{\partial x}\right)^{-1} \qquad (4)$$

or, after cross-shore integration over L_*,

$$c_p h_* = \frac{\partial MSL}{\partial t} L_* - (q_{x,sea} - q_{x,dune}) - \frac{\partial Q_y}{\partial y} - s \qquad (5)$$

in which Q_y is the alongshore transport integrated over L_*.

In the absence of littoral transport gradients and other sources or sinks (including sand exchanges with the lower shoreface and backbarrier) the above reduces to the standard *Bruun Rule* (BRUUN, 1962):

$$c_p = \frac{\partial MSL}{\partial t}\left(\frac{L_*}{h_*}\right) \tag{6}$$

Equation 5 is similar to the DEAN and MAURMEYER (1983) version of the Bruun Rule, an analytical precursor of the coastal-tract concept. The shoreline-change rate is determined quantitatively by the balance between the 'sink' term, for accommodation-space generated due to sea-level rise (first term on the right-hand side), and sediment availability (being the sum of sinks and sources, the last three terms on the right-hand side of equation 5). The relative sea-level change is a virtual sink/source term since there is no absolute loss, although the response is comparable to the impact of a real source/sink regarding horizontal movements of the upper shoreface.

The source and sink terms in equation 5 allow the qualitative Curray-Swift model of coastal evolution to be quantified as a time trajectory in sediment source/sink phase space: *e.g.*, evolution of the well-documented central Netherlands coast between Hoek of Holland and Den Helder in Figure 3. The trajectory is based on a) estimates derived from radiometric data by BEETS *et al.* (1992), listed in the Annex, for the period 5000–0 years BP; and b) the results of STM simulations for 7200–5000 BP. The line separating advance and retreat of the coast is fitted for the trajectory in the top-right quadrant, with its mirror image assumed for the bottom-left quadrant in the absence of other data. The trajectory bifurcates after 2000 BP because differences develop in rates of shoreline change averaged alongshore north and south of Haarlem (see Table 2 for location). The shape of the advance/retreat-threshold curve demonstrates that coastal evolution is governed mainly by a) sediment supply (±) under near-stillstand sea-level conditions (such as those predominating in the late Holocene), and b) change in accommodation space when sea-level changes rapidly (such as during global glaciation and deglaciation).

Low-order Shoreface Processes

Dynamics

We depart from conventional definitions of the shoreface which, as a component of the coastal tract, we regard ostensibly as comprising the entire sub-aqueous continental-shelf surface (Part 1). We assume that the main flows responsible for shore-normal sediment fluxes on the shoreface comprise wind waves and swell, wind-driven flows (upwelling and downwelling), tidal currents (especially in the vicinity of estuaries), and surf-zone flows (especially undertow) on the upper shoreface (NIEDORODA and SWIFT, 1991; STIVE *et al.*, 1991; WRIGHT, 1995; COWELL *et al.*, 1999a). On time scales relevant to coastal-tract behavior, aggregate sediment dynamics respond to a coastal-flow climatology that can be characterised only as a magnitude-frequency distribution.

The effects of wind-wave and swell asymmetry in this climatology can be thought of as advecting sediments onshore, if we assume that residual wave transports act only in the shoreward direction. Other components of the flow climatology drive sediments onshore and offshore through irregular cycles of varying amplitude and periodicity, such that net fluxes behave diffusively. In aggregate, these sediment dynamics are analogous to large-scale turbulence in which horizontal eddy diffusion characterizes across-shore sediment dispersal, with the length scale of the eddies increasing with distance from shore. This concept of macroscopic sediment dynamics was introduced by NIEDORODA *et al.* (1995) who represented the time-averaged across-shore transport-volume flux as

$$Q_x = q_{x,ad} + q_{x,dif} + q_{x,g} + s_x \tag{7}$$

for each point along the shoreface profile at a positive distance x from the shore, where subscripts *ad*, *dif*, and *g* denote advection, diffusion, and gravity respectively, while s is a source or sink effect due to net along-shelf transports at any point.

The first term on the right in equation 7 captures the advective effect of wave asymmetry on sand transport. This term is a lumped representation of the depth-dependent sediment flux, due to combined wave- and current-driven transports and effects related to bottom slope. Onshore directed wave-asymmetry effects dominate on shallower parts of the shoreface. The second term represents the long-term depth-averaged total-load transport, the behavior of which is controlled in aggregate by an average-annual horizontal diffusion coefficient (that is also depth dependent) and the cross-shore gradient of the representative sediment concentration (a depth dependent variable).

The concept underlies the Advection-Diffusion Model (ADM in Table 1) which expresses the long-term sediment-transport regime associated with a site-specific coastal-ocean climatology. The ADM assumes that coastal-tract morphology is shaped by space- and time-averaged processes. The ADM *data-model* thus requires elimination of local relief through spatial-averaging over several kilometers along the cross-shore profile. Usually a sediment input (or loss) at the shoreward end of the profile is set to represent the average rate of terrestrial input per unit coast length, but sediment can also be directed into the model offshore, as would occur from alongshelf-transport gradients. The hydrodynamic and diffusion parameters are fitted by successive trials until a uniformly bypassing profile of the correct shape is obtained. An actual sea level history is then used to drive the evolution of the profile.

The aggregate processes represented in equation 7 are illustrated in the modeling of the Columbia River coastal tract (Table 2) using the ADM (Table 1). Essentials of the data-model (Figure 4a) include an aggraded shoreface overlying an erosion surface dating from 7000 years BP (WOLF *et al.*, 1997), and a prograded strandplain, interrupted by inlets to barrier lagoons (KAMINSKY *et al.*, 1997). In reality the ocean floor beyond the continental shelf lies at depths larger than 1200 m but, for practical reasons, the data-model ocean floor lies at about 300 m (which has only minor influence on the ADM calculations). The external forcing includes sea-level rise, and a sediment supply from the Columbia River estimated to total 17.5 million m^3a^{-1}, with sediment losses be-

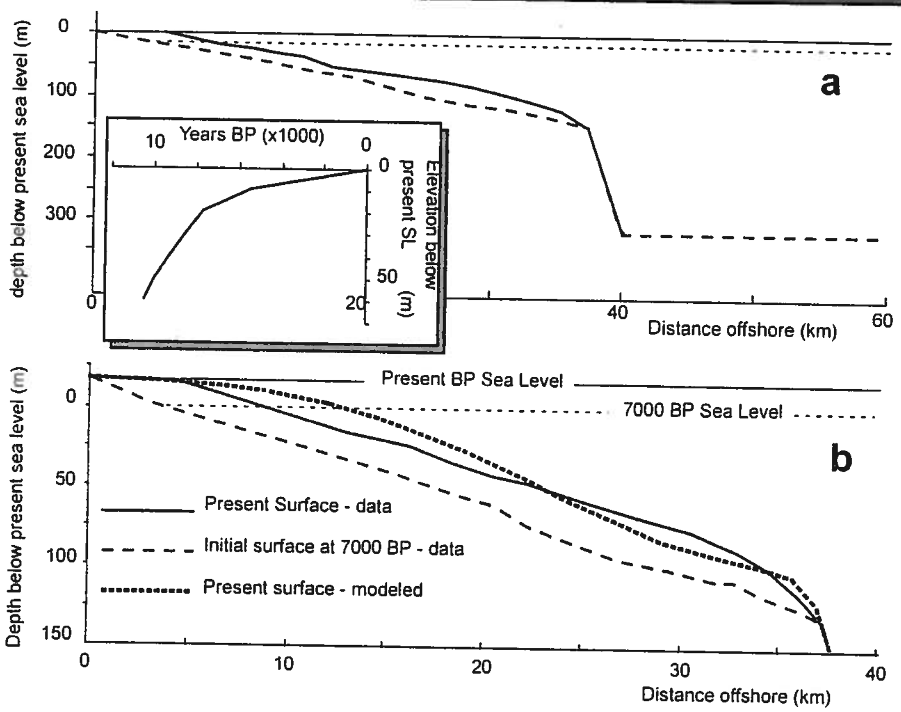


Figure 4. Columbia River coastal-tract: a) Data-model for the including initial surface (dashed) dating from 7000 years BP beneath the present surface (solid line) with sea-level curve (inset); and b) ADM simulation of Columbia River coastal-tract evolution from 7000 BP to present.

yond the continental-shelf edge (STERNBERG, 1986) and to the backbarrier tidal basins estimated at 0.7 and 2.4 m^3a^{-1} respectively. The result is a net sediment input of 14.4 million m^3a^{-1}, averaged over the approximately 150 km alongshore distance characterized by the Columbia-River tract. Since the shoreface of the Columbia tract has experienced net aggradation during the past 7000 years, the initial shoreface was necessarily deeper than equilibrium, providing positive sediment-accommodation space on the shelf (*cf.* Figure 2b).

Modeled offshore diffusion of sediments supplied at the coast by the Columbia River give a net accreted volume close to the measured volume of post 7000 BP deposition on the shoreface totalled out to the shelf edge (Figure 4b). Although the results indicate that diffusion rates in nature are higher than those used in the model, the discrepancy may be due to limited viability of the alongshelf-homogeneity assumption for the Columbia shoreface: *i.e.*, fine-sediment deposits occur in field data as a distinct band trending diagonally across the shelf (STERNBERG, 1986). Nevertheless, the ADM results overall are consistent with the data-model. Both indicate that diffusion is incapable of displacing all sediments to the lower shoreface. Thus, the rate of sediment supply at the coast by the Columbia together with the onshore advection due to waves resulted in simulated progradation of the upper shoreface in addition to overall aggradation of the lower shoreface (Figure 4b).

The advection-diffusion processes are continuous across the entire shoreface, but the terms in equation 7 and the balance between them varies systematically with water depth and distance from the coast. Thus, the advective effects offset the diffusive effects to the greatest degree in shallow water. The variation in relative importance of these effects across the shoreface (*i.e.*, first two terms in equation 7) forms the basis of the separate but coupled behavior of the upper and lower shoreface.

Upper-lower Shoreface Coupling

In shallower depths on the shoreface, the enhanced effect of shoreward sediment advection (first term in equation 7) can overwhelm the seaward-acting diffusion effects. The Hinged-Panel Model (HPM in Table 1) represents advective processes involving BOWEN's (1980) version of Bagnold's *energetics* transport model, with the seaward-directed diffusive effects simply represented by downslope gravity. The HPM was developed to analyze a shore-normal profile extending from the surf zone to the lower shoreface on an open ocean coast with uniform alongshore bathymetry. The profile is portrayed as discrete sections (panels) representing the upper, middle and lower shoreface. The profile of the upper shoreface is assumed to keep its shape while translating on-, or offshore. The sediment transfer between panels is computed from width-, and depth-averaged variables representing the power-equivalent wave and current environment. Profile changes are constrained by sediment mass conservation.

Central Netherlands: Haarlem. Application of the HPM to the Holland coastal tract demonstrated the combined effect of an external littoral supply of sediments to the tract and the transfer of sand from the lower to the upper shoreface within the tract (STIVE and DE VRIEND, 1995). The data-

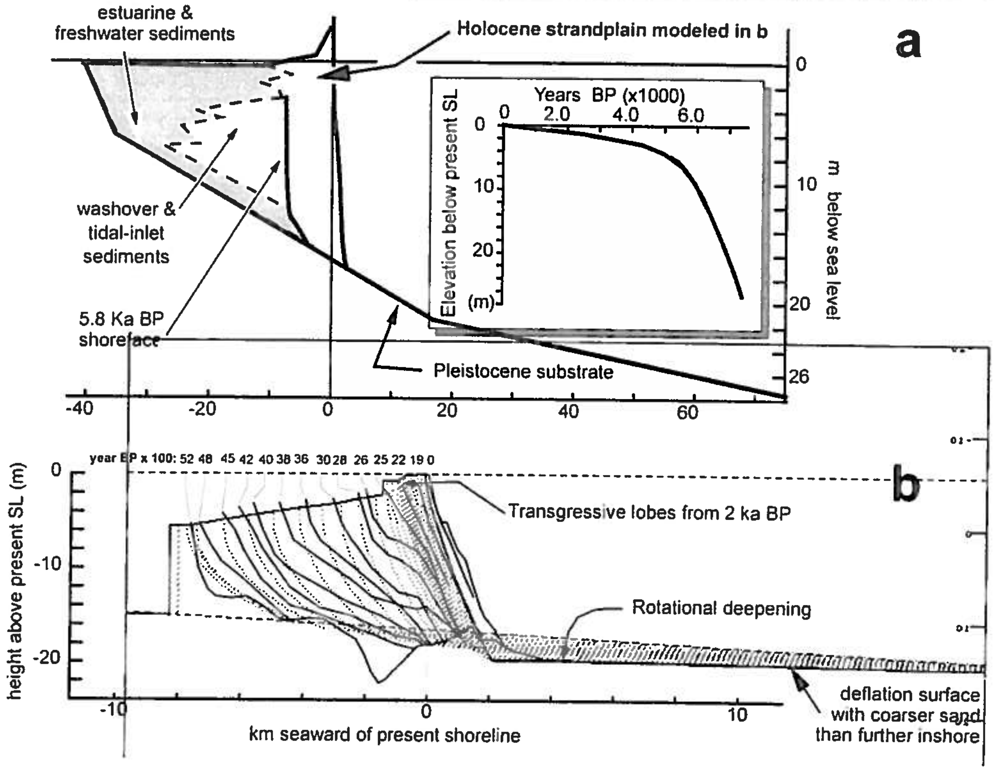


Figure 5. Seaward translation of the upper shoreface in the Holland coastal tract driven by littoral sediment feed and a deepening of the landward 10 km of the lower shoreface, with the deepening decreasing progressively in the offshore direction: a) data-model; and b) STM simulation overlaid on data-model isochrons (solid lines). The steeply dipping dotted lines within the strandplain depict positions of the STM shoreface during previous time steps: i.e., synthetic isochrons. The dotted lines on the lower shoreface are earlier erosion surfaces.

model was based on morphostratigraphic surveys calibrated by radiocarbon dates from a strandplain that formed along the full length (100 km) of the cell after about 5000 BP, when inlets to the tidal basin closed (BEETS *et al.*, 1992; VAN DER VALK, 1992). The data-model comprises a gently sloping lower shoreface (Figure 5a) indicative of a negative accommodation capacity offshore (Figure 2c), a strandplain volume estimated at some 6×10^6 m^3 supplied between 5000 and 2000 BP, with less than half the amount thought to be supplied from adjacent coastal tracts (BEETS *et al.*, 1992). The remainder is thought to have been reworked from the shoreface, primarily from the subaqueous tidal deltas but also from the landward segment of the lower shoreface.

STIVE and DE VRIEND (1995) used the HPM to simulate these processes. The kinematics are reproduced in Figure 5b using the Shoreface Translation Model (STM in Table 1) applied to the same data-model, beginning with a littoral feed of 30 m^3 a^{-1} per meter of coastline from 5200 to 3600 BP, then declining by 5 m^3 a^{-1} m^{-1} until 2000 BP when it was stabilized at -0.5 m^3 a^{-1} m^{-1}. The cross-shore transfer of sand to the upper shoreface was modeled by adjusting profile geometry through time. More specifically, a deepening of the upper-lower shoreface boundary in the STM (L_* in equation 5) was imposed at a rate of 0.17 m per 100 years consistent with, although less than, rates of measured bathymetric change (STIVE *et al.*, 1991; HINTON *et al.*, 1999). The lower shoreface deepening extended 10 km seawards at rates that decreased linearly with distance. The simulated process (Figure 5b) provided 49 percent of the strandplain volume.

These results suggest that the coupling between the upper and lower shoreface on the Holland coast is an important factor offsetting tendencies toward shoreline recession due to littoral sand losses from the tract. This stabilizing form of low-order coastal behavior has been overlooked in coastal-management studies in general (COWELL *et al.*, 2003). Nevertheless, each of our data sets (Table 2) contains evidence for this type of behavior. For example a coarse sediment lag exists on the lower shoreface at each of the field sites, and such deposits are widespread elsewhere throughout the world (COWELL *et al.*, 1999a, 1999b, 2003), indicative of surface lowering.

SE Australia: Tuncurry. Amongst the first descriptions of coarse sediments on the lower shoreface came from the Holland coast (VAN STRAATEN, 1965). They have been analyzed in detail for the Tuncurry site in SE Australia (Table 2) regarding their relationship to the underlying parent material from which they were shown to derive (ROY *et al.*, 1997). Typical of SE Australia, the lag deposits occur in water depths greater than roughly 20 m and extend as a shore-parallel band up to 10 km in width (ROY *et al.*, 1994).

Simulation of the Tuncurry-strandplain formation using the STM (Figure 6) illustrates genesis of the lag deposits

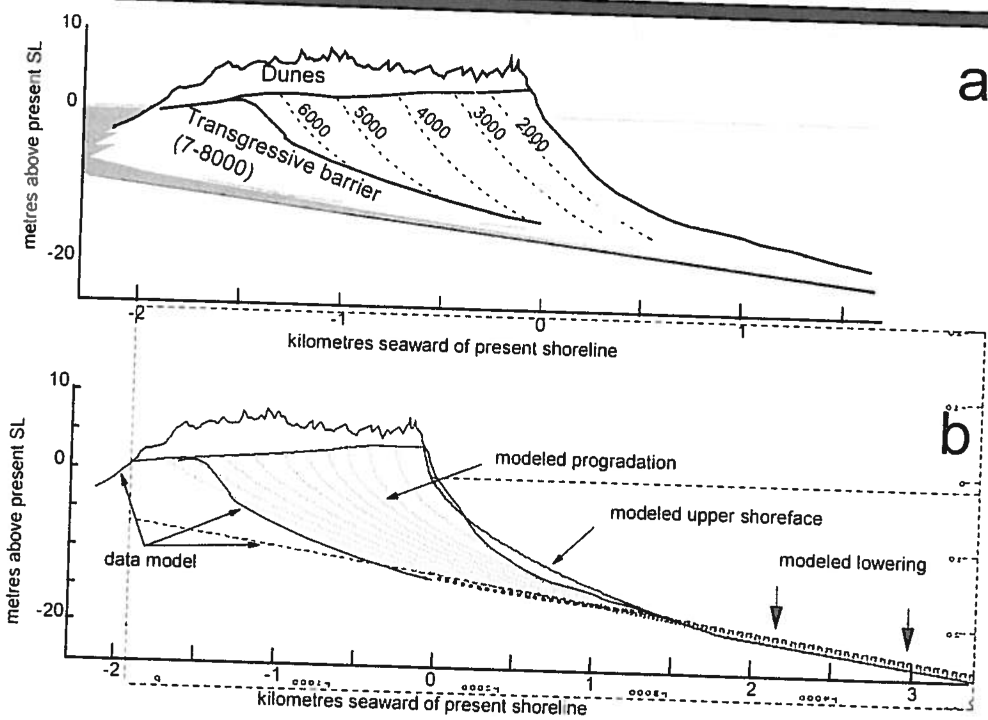


Figure 6. Tuncurry coastal tract, SE Australia: a) data-model based on radiometric dating of cores samples and ground-penetrating radar, with the isochrons labelled in years BP (ROY *et al.*, 1997); and b) STM simulation output overlaid on data-model.

through low-order behavior similar to that evident on the Holland coast. The Tuncurry coastal tract characterizes a cell spanning 11.5 km of coastline in which progradation has averaged 0.3 m a^{-1} throughout the Holocene (Fig 6a). The simulation involved a constant sea level for 6000 years, a progressive increase in runup elevation from 1.5 to 4.0 m above MSL due to increasing wave exposure, a steady increase in the depth of the upper-lower shoreface boundary at a rate of 0.002 m per year, decreasing linearly with distance for 10 km across the lower shoreface, and a constant littoral sand feed of 0.25 m^3 per year per meter of shoreline (Figure 6b). Sand supply from the lower to the upper shoreface due to shoreface deepening averaged 1.1 $m^3 a^{-1} m^{-1}$ over the 6000 year period, thus supplying 80 percent of the sand comprising the prograding strandplain. The sand transfer is well below detection and prediction limits on annual time scales. Nevertheless, such small net residuals aggregate to produce all of the mean trend behavior within the Tuncurry coastal tract.

NW USA: Columbia River Coast. The Columbia River coastal tract (Table 2) provides a particularly valuable opportunity to substantiate Haarlem-Tuncurry type coupling between the upper and lower shoreface. Cycles of earthquake-induced subsidence and subsequent rebound occur at roughly 500 year intervals in the NW Pacific coastal region (KAMINSKY *et al.*, 1997). These tectonic processes cause episodes of sudden rise in relative sea levels, estimated at roughly 1–2 m, followed by gradual isostatic re-emergence. Responses to these events are evident in heavy mineral seams and shore-parallel lineaments in dunes within the strandplains (MEYERS *et al.*, 1996; KAMINSKY *et al.*, 1997). The data-model (Figure 7a) used with simulations was described in connection with Figure 7a, but for the shoreface coupling experiments the data-model also incorporates surface morphology generalized from the Long Beach strandplain, and a sea-level curve in which the magnitude of subsidence events is proportional to the duration between them based on marsh data (ATWATER *et al.*, 1995).

The results of the Columbia simulation show that it is impossible to reproduce present shoreface morphology without progressive deepening of the lower shoreface. Under conditions of earthquake induced sea-level fluctuations simulated with a time-invariant shoreface in the STM, the prograding sediment wedge produced a pronounced bulge in the shoreface (Figure 7b) that is absent from the data-model (Figure 7a). The sand contained within the submarine bulge reduces the sub-aerial volume so that progradation of the simulated strandplain is significantly less than in the data-model.

Simulation of time-varying shoreface responses to sudden sea-level rise using the STM involved reducing water depth (h_*) at the upper-lower shoreface boundary from 20 m to 15 m upon the occurrence of a subsidence event. Gradual shoreface deepening during the subsequent period of gradual rebound was simulated by increasing h_* at a rate of 0.01 m per year, reducing progressively to zero 2000 m further seaward (Figure 7c). The effect of shoreface deepening was to subdue the submarine bulge in the prograding sand wedge. In time steps immediately following an earthquake, the effect involved transfer of sand seaward. The aggregate effect however over a full subsidence-rebound cycle was net displacement of sediment to the sub-aerial portion of the prograding

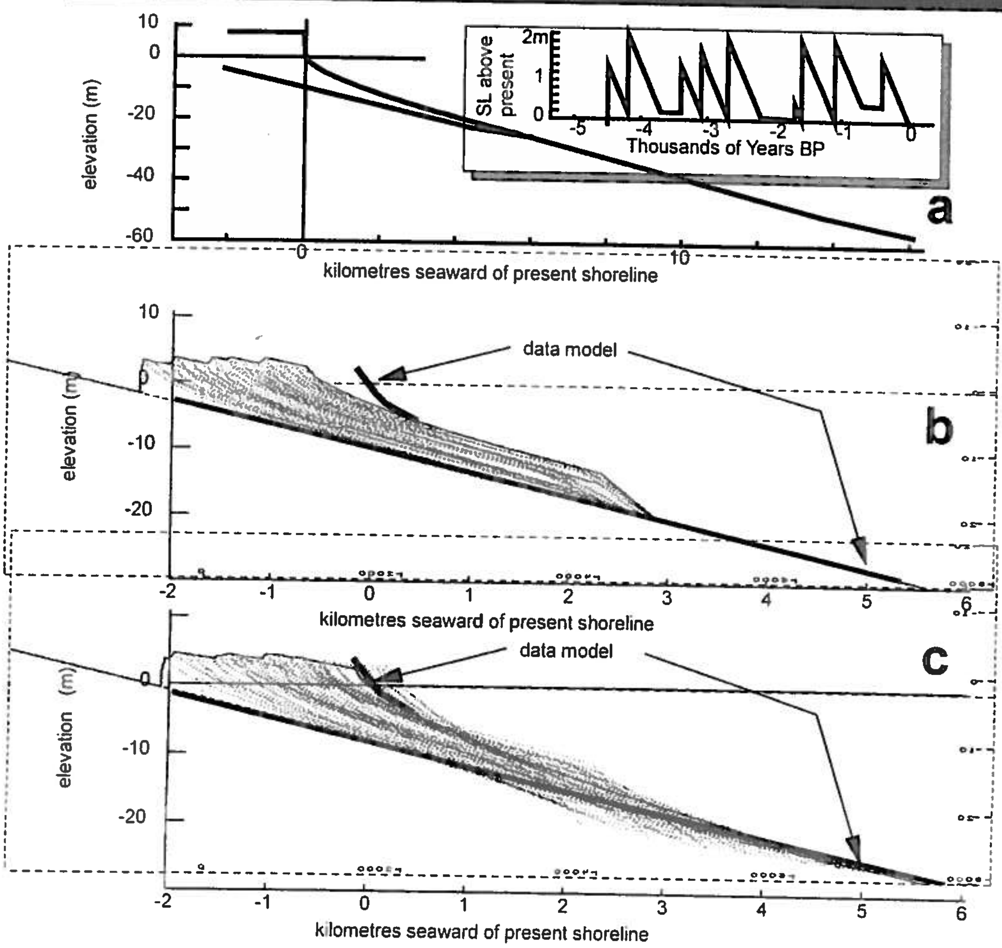


Figure 7. Columbia coastal tract, SW Washington, USA subject to seismically induced sea-level variations and littoral sediment feed: a) data-model, b) STM simulation without shoreface deepening, and c) STM simulation with shoreface deepening between earthquake events.

sand wedge. The simulated wedge is geometrically consistent with the data-model.

Falling Sea Level. A similar requirement for coupling between the upper and lower shoreface exists during falling sea levels in general. Since we assume that the upper shoreface can attain an equilibrium on the decadal time scale (ie., it has invariant time-averaged form extending seaward to a time-scale dependent closure depth), then the toe of the upper shoreface descends at the same rate as the falling sea level. The lower shoreface however must also degrade so that its elevation remains less than that of the upper shoreface. Otherwise a shear-stress shadow zone would form if the water depths increased further inshore, as would occur in the absence of coupling. Thus, as in the previous examples (Figures 4 to 6), a coupling exists involving a deepening of the landward portion of the lower shoreface with a corresponding transfer of sediment from the lower to the upper shoreface constrained by mass-conservation. The consequence is a prograding strandplain (COWELL *et al.*, 1999b).

Our Tuncurry data set (Figure 8a) contains stratigraphic evidence of shoreface behavior as sea level fell through tens of meters during the onset of the last glacial period 50–30 thousand years ago. Sea-level estimates were derived from uplifted coral terraces in New Guinea (CHAPPELL and POLACH, 1991). Strandplains developed in response to falling sea levels and are now located in 30–80 meters of water off the present coast. Seismic data and thermoluminescence dates from cores through the strandplains show that pre-existing sediments comprising the substrate were eroded to depths greater than 15 m below the present seabed, and subsequently backfilled with upper-shoreface sediment to produce the strandplains (ROY *et al.*, 1997).

At least two phases of strandplain development occurred, as indicated by the formation of backbarrier lagoonal deposits midway along the sequence (about 11 km offshore from the present coast in Figure 8a). These distinct phases correlate with the fluctuations in the overall trend of falling sea level (Figure 8a inset). The fluctuations were ignored however in STM simulations: only sea-level trends apply. Figures 8b and 8c show the STM simulation of strandplain formation after sea level had fallen continuously through 80 m. During this process, deepening of the lower shoreface was imposed in the simulation to a distance of 10 km seaward of the upper shoreface (*i.e.*, the active width of the lower shoreface specified in

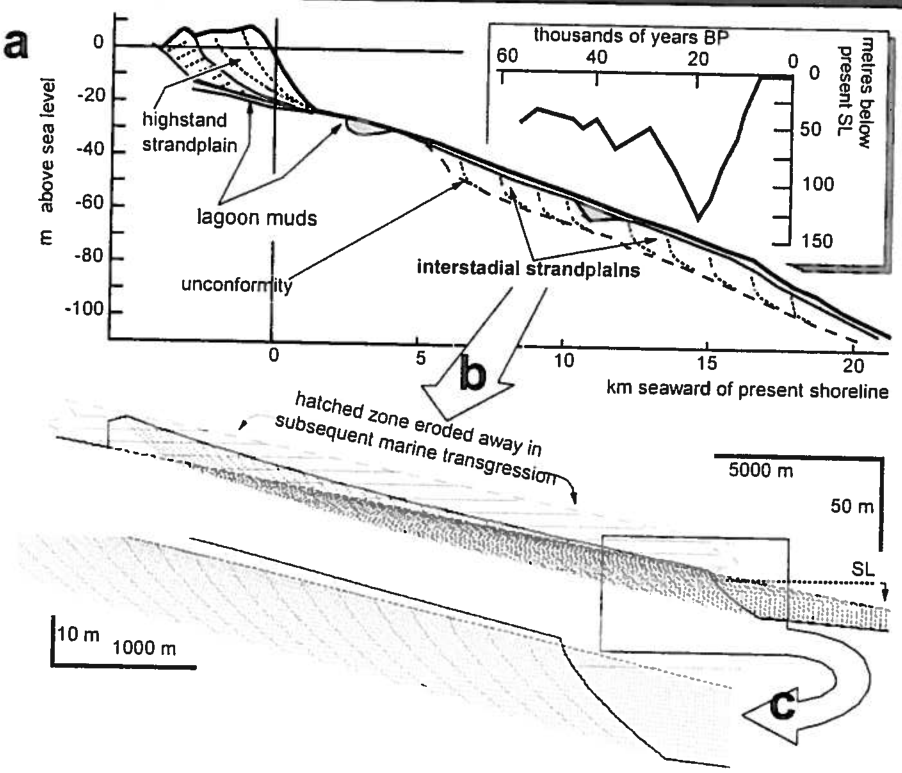


Figure 8. Shoreface response to falling sea level at Tuncurry, SE Australia (Table 3): a) data-model, including strandplains and the sea-level curve (inset); b) STM simulation of strandplain formation; and c) detail of prograding upper shoreface, in which the dotted sub-horizontal lines represent previous erosion surfaces of the lower shoreface when falling sea level was higher, and the dotted steeply dipping curved lines are the buried depositional surfaces of the prograding shoreface.

the STM). This distance was estimated from the extent of the corresponding zone identified under present conditions (simulated in Figure 6). Lowering was achieved by adjusting h_*. Stratigraphic evidence suggests the upper part of the strandplain has been removed in the subsequent Holocene marine transgression and current period of sea-level stillstand (COWELL *et al.*, 1995; ROY *et al.*, 1997). Overall, the process caused a massive reworking of the pre-existing substrate to a depth roughly equivalent to upper-shoreface closure depth (as defined for the decade to century time scale).

Low-Order Backbarrier Processes

Dynamics

The backbarrier component of the coastal tract includes the sub-aerial beach, dunes and lagoon/estuary or coastal lowland equivalent, if present (*cf.* zones A,B,C in Figure 1, Part 1). The backshore can be ignored for the very long times scales (10^4–10^6 years) relevant to the basin-fill problem in exploration geology. Then, transgressive and regressive surfaces in seismic records signify the gross behavior of the continental shelf in response to large (tectono- and glacio-eustatic) sea-level changes and variations in sediment supply (CAREY *et al.*, 1999). Ignoring the backbarrier even on these geological time scales however, requires the simplifying assumption that deposits in lagoons and estuaries can be aggregated into a generalized backstepping coastal sediment wedge during periods of rising sea level. Similarly, but at the other extreme, analysis of short-term coastal change (*e.g.,* dune-erosion models) also can ignore explicit linkages with the backbarrier. These effects are usually represented implicitly in models on intermediate scales (HANSON *et al.*, 2003, this volume), as boundary conditions (Part 1).

A more complete account of coastal behavior however, must include shoreface-backbarrier coupling (Figure 2 in Part 1). For example, in environments with continental-shelf slopes typical of the Columbia and Tuncurry sites, if the backbarrier accommodation space generated by sea-level rise is occupied fully by deposition of fine sediments, then the rates of coastal recession (transgression) can be almost 50 percent less than in the absence of fine sediments. On the other hand, a doubling in the rate of sand bypassing from the shoreface to the backbarrier can increase recession rates by up to 40 percent (Figure 9). Generally, rates of fine-sediment deposition in the lagoon depend on the availability of such sediments from fluvial sources, or from muds mobilized through shoreface erosion. Biogenic sediment production may add to backbarrier sedimentation.

Transfer of sand from the shoreface to the backbarrier driven by sea-level rise causes *barrier rollover* (LEATHERMAN, 1983), a process represented analytically by the Generalized Bruun Rule (DEAN and MAURMEYER, 1983). Barrier rollover

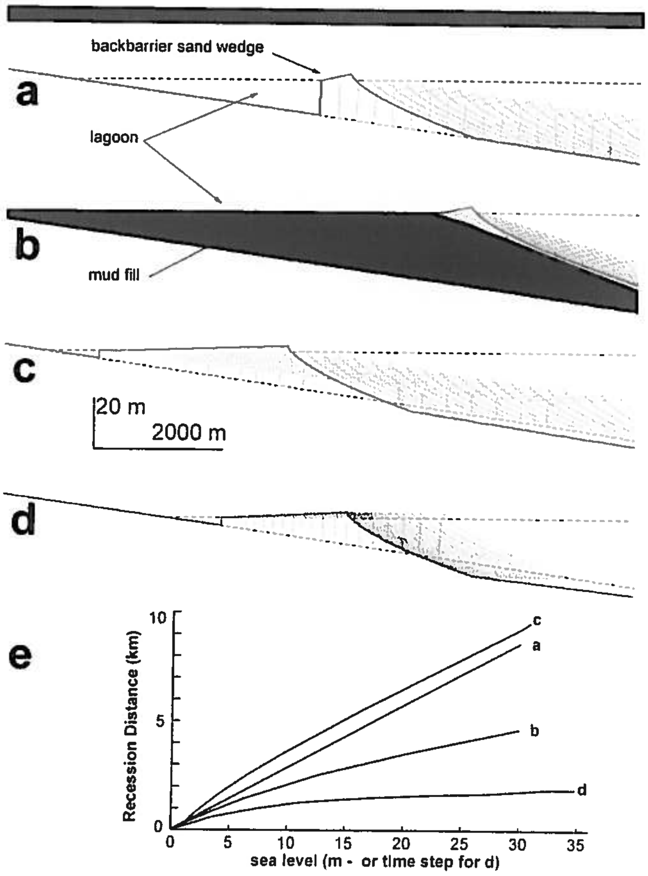


Figure 9. Dependence of shoreface-recession on backbarrier behavior for idealised conditions on a 0.2 degree shelf slope: a) simple barrier rollover with time-invariant backbarrier sand wedge; b) mud deposited at the same rate as sea-level rise; c) width of backbarrier sand wedge increasing with time; d) increasing width of backbarrier sand wedge with constant sea level; and f) comparison of shoreface recession distances in a to d. A sea level rise of 1 m per time step applies to a, b, and c. Sea level is constant in d. All other conditions not specified are the same in each case.

involves bypassing of sediments from the shoreface to the backbarrier (Figure 9a) through washover, tidal-inlet, and transgressive-dune processes (third-order processes in the coastal-tract cascade). Conceptually, the Generalized Bruun Rule aggregates all these processes, simply by assuming that they collectively cause a given rate of sediment bypassing. In the special case analyzed by Dean and Maurmeyer, this is the rate that permits the barrier to maintain a constant form during transgression (in an alongshore spatially averaged sense).

Generally, tidal-inlet processes dominate sediment exchanges between the shoreface and the backbarrier, even in areas where these processes are episodic: *e.g.*, for microtidal barrier islands (LEATHERMAN, 1983). These processes include an interaction between the upper shoreface, the inlet channels, the channels and shoals on the ebb-tide delta and the tidal flats in the basin (ostensibly the flood-tide delta). These morphological elements operate collectively as a third-order sediment-sharing system within the coastal-tract cascade (Table 1 in Part 1). The interaction is characterized by ASMITA (Table 1) in which each component of the system seeks to maintain a dynamic equilibrium with its hydraulic forcing. Equilibrium is expressed by equations of state that relate gross morphologic parameters (specifically the aggregate volume of the positive or negative relief within each component) to relevant forcing parameters (*e.g.*, tidal range or tidal prism). The equations of state are empirically derived relations: *e.g.*, between tidal deltas and the tidal prism (EYSINK, 1991).

Variations in external forcing (*e.g.*, average tidal range) shift the equilibrium state for any or all of the units. Disturbance to the equilibrium of one unit demands changes in each of the others, since all units share a common total sand volume (at third order). The mutual changes result in an exchange of sediment between the various units until the equilibrium state is re-established. Diffusion processes govern the sediment transport between the units, each of which maintains a time-averaged concentration of mobile (suspended) sediments related to the hydraulic forcing. Disturbances in the overall equilibrium cause adjustments involving net sand transfers between the shoreface and backbarrier. Net transfer to or from the backbarrier causes landward and seaward displacement of the upper shoreface respectively.

In application of ASMITA, an equilibrium system is developed and then perturbed to study the responses of the system components and the interaction between these components. An interesting result emerges from the different response times associated with the various components in the system. Once the system is perturbed a series of interactions develop between the components (inlet, tidal basin, and shoreface) that cannot be predicted based on understanding of the time-scale of separate adjustment of individual components to their equilibrium state.

In principle, this third-order behavior is affected also by the interaction between the dynamics of the backbarrier sand wedge (primarily the gross flood-tide delta unit) and the fine-sediment fill (muds). Higher rates of mud deposition decrease the tidal prism; the construction of polders in the lagoon can have the same effect. Similarly, the hydrodynamics of the inlet itself regulates discharge, influx and retention times (flushing rates) of water containing fine suspended sediments.

Shoreface-Backbarrier Coupling

Overall, the rate of sand bypassing from the shoreface to the backbarrier controls the rate of upper-shoreface translation landward, illustrated in Figure 9 for idealised conditions with a shelf slope comparable to Columbia and Tuncurry coastal tracts. This bypassing is controlled by the available sediment accommodation space in the backbarrier, and the rate at which this space is regenerated by any sea-level rise. Thus, if the rate of sand bypassing from the shoreface to the backbarrier remains constant, coastal recession occurs at a uniform rate (Figure 9a). If accommodation space available is reduced through mud deposition, then sand transfer from the shoreface to the backbarrier is reduced, which in turn reduces the rate at which the upper shoreface recedes (Figure 9b). If backbarrier space is available to accommodate growth

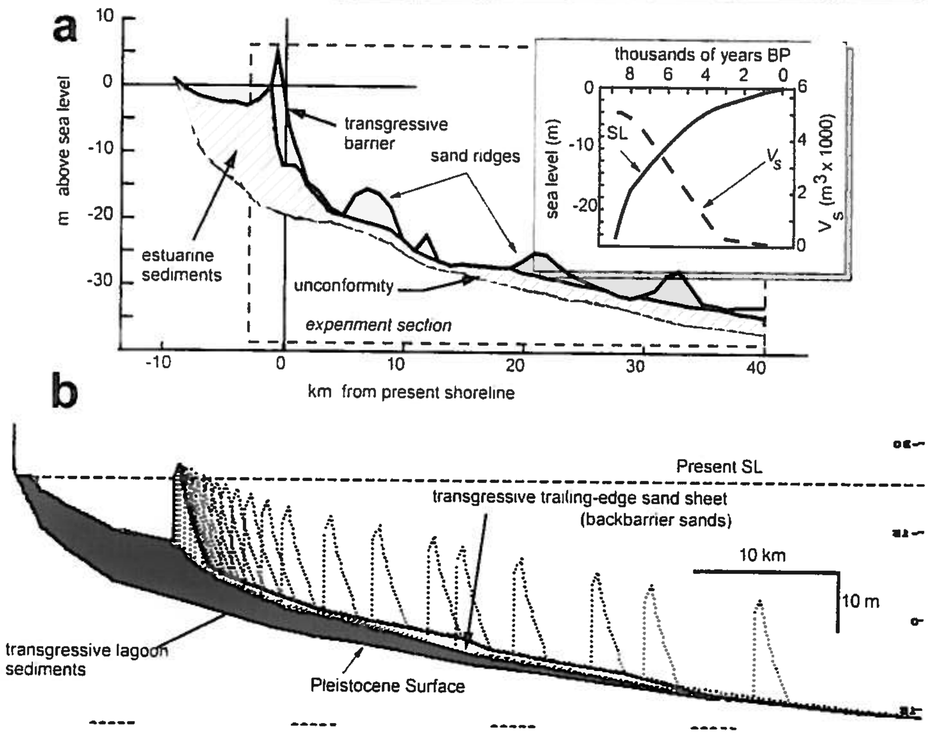


Figure 10. Duck coastal tract, NC USA: a) data-model, and b) STM simulation covering the last 20 m of post-glacial sea-level rise, but before the onset of progradation. Mud deposition was at 30 percent of sea-level increments.

in the backbarrier sand wedge (ostensibly the flood-tide delta), then such growth elevates the rate of shoreface retreat (*cf.*. Figure 9c and 9a) since the upper shoreface and backbarrier sand wedge is a sediment-sharing system. Therefore, even in the absence of sea-level rise, growth of the backbarrier sand wedge occurs at the expense of the shoreface (Figure 9d).

The linear recession of the shoreface for simple barrier rollover across the 0.2° substrate (Figure 9e) occurs at the same rate as inundation would occur over a non-erodable substrate of the same slope. Mud deposition in the lagoon or variations in sand bypassing to the backbarrier cause departures from the inundation rate (b and c in Figure 9e). Despite mud deposition and growth of the tidal delta both occurring at a constant rates (Figure 9b and c), the recession rates in these cases were not constant (Figure 9e). This behavior reflects fundamental non-linearity due to morphological state dependence in coastal evolution (*i.e.*, 'sensitive dependence upon initial conditions'). Although in principle the results of simulations such as those in Figure 9 could be 'interpolated' to decadal time scales for coastal-management purposes, state-dependent effects would need somehow to be taken into account.

E USA: Duck NC. The importance of backbarrier mud deposition in mitigating shoreface retreat is evident for the Duck coastal tract (Figure 10). In this case, the transgressive barrier is so diminutive that the aggregate behavior depends more on the backbarrier than the shoreface itself. The Duck simulation also shows the additional effect of an external littoral supply of sand (V_s in Figure 10a inset). This supply results in the formation of a transgressive sand sheet on the lower shoreface (Figure 10b). This sand sheet blankets the lagoonal muds as they emerge seaward of the receding barrier driven landward by rising sea levels. We assume that this sand sheet was subsequently reworked to form, at least in part, the sand ridges located at present on the lower shoreface (Figure 10a). In reiterative simulations, each with the same sea-level curve, mud deposition was the main controlling variable. Simulated recession (Figure 10b) that gave a final location of the transgressive barrier corresponding to the data-model (Figure 10b) was achieved with mud accretion rates set at 30 percent of sea-level increments.

Central Netherlands: Haarlem. Simulations suggest that the early Holocene evolution of the Holland coastal tract (7000–5800 BP) was entirely regulated by coupling between the upper shoreface and the backbarrier (Figure 11). The coupling involved regulation of backbarrier accommodation space by variations in dimensions of the upper shoreface. Our concepts on this coupling are illustrated further through the following assumptions used to develop the data-model.

1. A progressive increase occurred in depth and width of the shoreface during the post-glacial sea-level rise (Figure 11 inset).
2. The transgressive backbarrier and shoreface formed an amorphous sediment wedge of low relief before about 6000 BP when flows and sediment transfers from the shoreface to the backbarrier were unimpeded because the barrier was poorly developed.

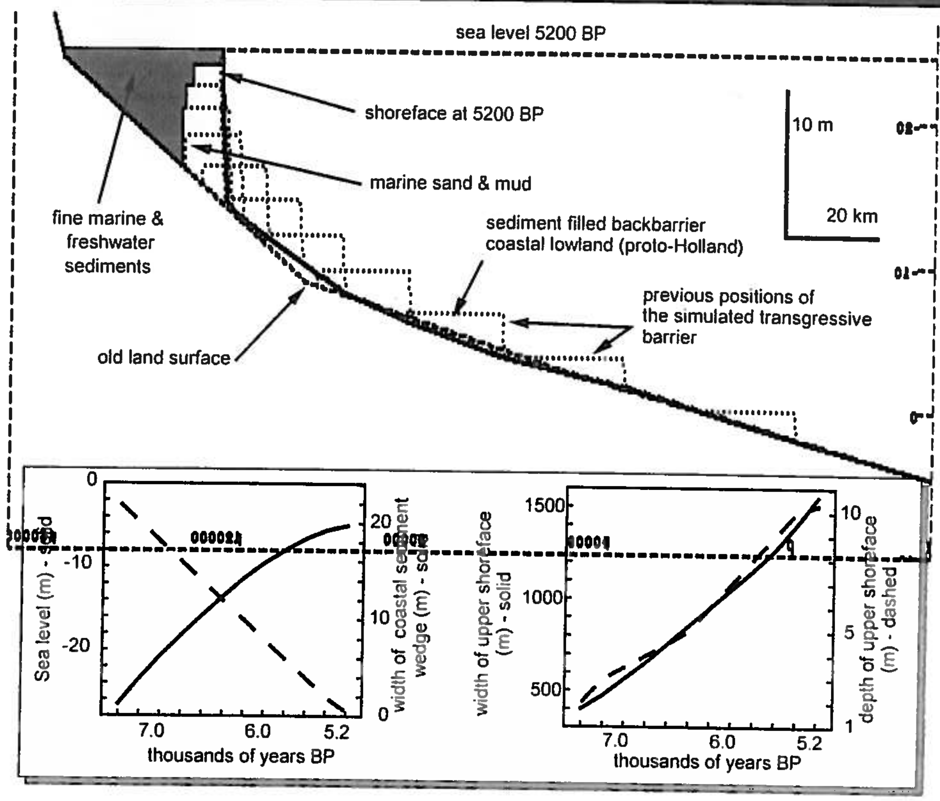


Figure 11. STM simulation of early Holocene evolution of the Holland *coastal tract* driven by post-glacial sea level rise. The insets show time-dependent changes in variables estimated in the *data-model*; Figure 5a gives the *data-model* morphology.

3. Both fine and coarse sediment fractions were conserved when re-exhumed by shoreface retreat: sediments were simply re-incorporated into the backbarrier further landward.
4. The backbarrier filled completely with sediment as fast as new accommodation space was created by the rising sea level.
5. The time-averaged littoral sediment supply was bypassed to the backbarrier at V_s = 66 m^{3}a^{-1} during the transgressive phase of Holocene evolution (8.0–5.8 ka BP).
6. Sediment transfers between backbarrier and shoreface became increasingly impeded and subject to normal inlet and overwash processes after about 6000 BP.
7. Shoreface-connected ridges (VAN DE MEENE, 1994) can be ignored in first-order tract behavior because they are higher order features.

Assumption 1 reflects the increasing fetch, water depth and hence wave energy, as the North Sea gradually flooded after the last glacial maximum. We adopted this assumption based on fetch- and depth-limited hindcasting of waves in a North Sea of reduced area: *i.e.*, when the basin first began to flood some 8–7000 years BP. Greater frictional attenuation than at present also can be inferred given that bed gradients were less (*c.* 0.01°) before hydro-isostatic subsidence ensued due to flooding of the basin (LAMBECK, 1995). We justify the second assumption through a) field evidence from exposures of late transgressive deposits (CLEVERINGA, 2000), and b) our interpretation of early conditions when the wave climate (Assumption 1) was probably subordinate to meteorological and astronomical tides.

Assumptions 3 and 4 stem from Assumptions 1 and 2; but they are based also on field data that indicate that the provenance of both coarse and fine sediments was largely the North Sea, with recycling through the backbarrier during rapid marine transgression between 8.0 and 5.8 ka BP (BEETS, *et al.*, 1992, 1995). In reality, sandier sediments occur in the seaward end of the transgressive wedge, while the fine fraction was trapped in the landward extremities of the tidal basin, with freshwater sediments and peat occurring along the mainland shoreline (VAN DER SPEK, 1994).

The sediment supply rate (Assumption 5) was estimated from the total Holocene sediment volume contained within the coastal lowlands of the Holland tract at present. This estimate (V_s = 66 m^{3}a^{-1}) is more than double the rate (V_s = 30 m^{3}a^{-1}) estimated for the period 5.8 to 2.0 ka BP when the coast was prograding (BEETS *et al.*, 1992): we applied the latter rate to the simulation shown in Figure 5. The higher rate of supply in the early Holocene assumes that 36 m^{3}a^{-1} of fine sediments were included in the total sediment feed from littoral sources sequestered by the backbarrier. Obviously, the fine fraction was not incorporated into the depositional volume during the later shoreface-progradation phase because the flows on the shoreface were too energetic to permit deposition of fine sediments. The pre-existing floor of the North Sea (including lowstand floodplain and deltaic deposits of the Rhine), and erosional-retreat of promontories to the north and especially the south of the tract provided the source of coarse and fine sediments for the Holland tract during the transgression (BEETS *et al.*, 1992, 1995).

Assumption 6 reflects a decreasing capacity for bypassing of shoreface sediments to the backbarrier as inlets became choked and dune development restricted the occurrence of washover. Such changes in conditions are evident in geological data and are attributed to reduced accommodation space (decreasing rate of sea-level rise) relative to the sediment supply volume (BEETS, *et al.*, 1992, 1995).

Geometric parameters were manipulated in the STM in accordance with these assumptions. Fine and coarse fractions were treated as an undifferentiated barrier-backbarrier sediment wedge in the simulation of early Holocene transgressive phase: *i.e.*, transgressive deposits landward of the shoreface (washover, flood-tide delta, estuarine basin, fluvial-delta and marsh sediments) were lumped together. We simulated this coastal-lowland sediment wedge by maintaining backbarrier width in the STM at values much greater than the lagoon width (Figure 11). During later phases, the effects of Assumption 6 were introduced by applying a progressive reduction in barrier width (Figure 11 inset).

Application of a progressive increase in shoreface dimensions (Assumption 1) caused the simulated elevation and width of the modeled backbarrier to grow through time, and thus also the volume of sediment sequestered by it (Figure 11). Although sediments probably were also mined from the lower shoreface during the transgression (BEETS *et al.*, 1992), in the manner illustrated by Figure 9c, rates of shoreface growth and external sediment feed used in simulations were sufficient to prevent the occurrence of much substrate erosion (Figure 11). Finding evidence for such erosion on the present lower shoreface is problematic because of the amount of modern reworking. Overall, the simulated interplay between shoreface behavior and the sequestering of sediments into the backbarrier resulted in translation of the upper shoreface to within 1200 m of the 5800 BP shoreface location in the data-model (or less than 1.5% of a potential translation distance exceeding 80 000 m).

The simulated reduction in rate of sand bypassing to the backbarrier, with continued littoral sand supply from outside the tract, caused the modeled evolution to flip spontaneously from transgression to regression at 5200 BP (Figure 11). This fundamental change from coastal recession to progradation, despite continued sea-level rise (albeit at a declining rate), was induced entirely through reduction in this sand bypassing.

N Netherlands: Wadden Islands. The first-order coupling of the shoreface and backbarrier is regulated by second-order processes that govern the rate of sand bypassing between the two complexes. This cascade relationship can be represented, for example, by nesting ASMITA within the ADM, STM or HPM (Table 1). Thus, ASMITA computations for the shoreface-backbarrier sediment exchanges were inserted for the sediment-budget in the landward-most grid cell of the ADM.

The sink term on the inner shoreface (s_x in equation 7), defined as the region interacting directly with the backbarrier (computational grid cell, $x = X_1$), becomes

$$s_{X_1} = s_b/L_y \tag{8}$$

where s_b is the source/sink term for shoreface-backbarrier sediment exchanges and L_y is the length of coastline represented by the coastal tract. Figure 12 illustrates results of the ADM-ASMITA hybrid model applied to shoreface-backbarrier coupling for the Terschelling barrier island and its Borndiep inlet (The Netherlands Wadden Sea coast, Table 2). The Terschelling-Borndiep data-model assumes that the sediment demand of the Borndiep only affects 20 km (L_y) of the coast up-drift from Terschelling (SHA, 1992). The dimensions of the inlet morphologies are given in Table 4 and the shoreface data-model is shown in Figure 12b.

The simulation involved a sea-level rise of 0.002 m a^{-1} for 200 years, then increasing to 0.004 m a^{-1} for a further 300 years. We assume that, during this period, the inlet morphologies maintain constant surface areas and that there is no net littoral sediment feed. Under these conditions, the rising sea level creates a sediment demand in the Borndiep tidal basin which causes the evolution of the tidal inlet morphologies shown in Figure 12a. In the first 200 years all units are close to their equilibrium states. After the increase in sea-level rise, all units evolve toward new equilibrium volumes, and the sediment sink (s_b) doubles from 0.6 million m^3 a^{-1} to a peak of 1.1 m^3 a^{-1} after 500 years. Simulations were undertaken with and without the inlet (Figure 12b). The comparison shows that inlet behavior predicted by ASMITA has significant consequences for shoreface behavior as the effects feed through the ADM. With the inlet, onshore directed advective transport in the ADM causes erosion of the shoreface beyond water depths of 7 m. Without the inlet, sediment eroded from the upper shoreface is transported further offshore.

The deepening of the shoreface is greatest furthest inshore (Figure 12b). This occurs because, for the hybrid-nesting scheme applied here, the sediment demand of the tidal basin directly affects only the first wet element (X_1) of the ADM. The ADM grid size in this example is about 1000 m. In reality however, significant direct influence of the ebb-tidal delta extends further from the coast: probably up to several kilometers (~5 km). A larger grid size would better capture the direct inlet effects further offshore, but this would detract from the resolution. Alternatively, the model could be refined by dividing the immediate sediment demand over more grid-cells according to a specified distribution (*e.g.*, linear decrease of sediment demand in offshore direction through several grid cells). Nevertheless, Figure 12 adequately demonstrates the principles of nested processes within the cascade hierarchy.

DISCUSSION: SCALE CONSIDERATIONS IN COASTAL MANAGEMENT

Prediction of Mean-trend Coastal Change

The illustrations of low-order coastal behavior in the previous sections may seem to involve scales of little interest in routine coastal management. However, observations on these

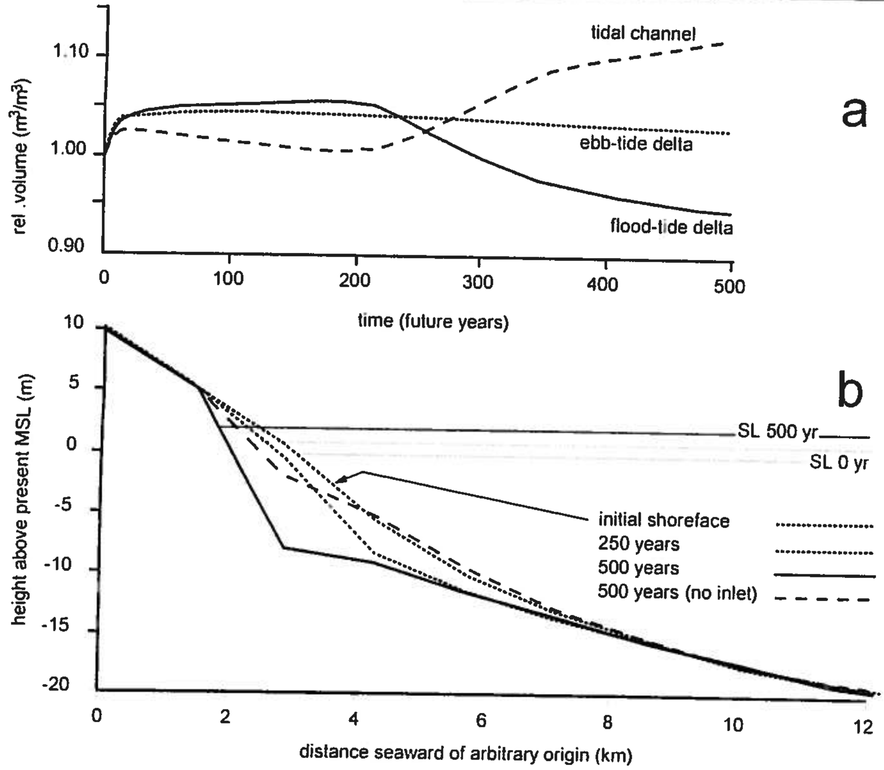


Figure 12. Simulation of coupled shoreface-backbarrier for the barrier island Terschelling and its Borndiep tidal inlet, on The Netherlands Wadden Sea coast, using a hybrid ADM-ASMITA model (Table 1): a) ASMITA predictions of tract-unit volumes (relative to initial values); b) ADM simulation of shoreface evolution with ASMITA inlet effects embedded (plus one case without inlet for comparison).

large scales are required for the occurrence of sufficient change and spatial coverage to resolve the modes, trends and rates in low-order behavior. Once resolved, these trends can be scaled down: from millennia to years if necessary. Thus, provided that the simulation has been wound up numerically to its operational state, the quantitative output over small time steps (years to decades) remain of immediate relevance to core decision making in coastal management, especially where discrimination between chronic and acute coastal erosion is concerned.

From what we know of the problems of up-scaling, coastal-management based on down-scaled predictions can hardly be less valid, even if currently they are less often contemplated. Moreover, downscaling probably provides the more reliable predictions past the decadal time scale. Whereas upscaled predictions can make little use of field data of direct relevance to these time scales, the modeling described above can exploit geological data. Assimilation of these data into behavior models constrain predictions on the shorter time scales. Data that are especially useful in this regard include stratigraphy derived from acoustic seismic, ground-penetrating radar and coring, especially where time-calibrated through radiometric dating. Predictions from aggregated models however only address probable future states. Aggregation precludes exact predictions for particular future instances (i.e, individual realizations).

Table 3. *Surface areas and equilibrium volumes for elements of Borndiep inlet, The Netherlands* (BIEGEL, 1993).

	Area (m²)	Equilibrium Volume (m³)
Flood-tide delta	1.77E + 08	1.24E + 08
Inlet channel	9.87E + 07	3.00E + 08
Ebb-tide delta	2.00E + 07	1.33E + 08

Scale Considerations in Delineation of Upper and Lower Shoreface

The lower shoreface is by far the most extensive region under most circumstances. It covers most of the continental shelf, out to the shelf break. Its character varies geographically, reflecting the evolutionary modes of coastal tracts in contrasting environments (Figure 2). Although the classical definition of the shoreface emphasizes the dominance of surface gravity waves in determining the morphology, the importance of other classes of flow in governing sediment dispersal has been recognized increasingly in recent years (NIEDORODA and SWIFT, 1991; WRIGHT, 1995). Recognition of the wide range of sedimentary processes operating on shorefaces has served to blur the offshore demarcation of the shoreface so much that generally it is possible only to trace the process continuum out to the shelf break (*e.g.*, NIEDORODA *et al.*, 1995). The continental-shelf slope is included in the coastal-tract concept since it is in principle also part of this deterministic process continuum. The slope is a region where the effects of gravity become the dominant transport agent. In reality, the process continuum manifested by active sedimentation probably extends onto the shelf slope only under conditions involving high levels of fine sediment input at the coast, where the shelf is narrow, or during sea-level low stands.

The upper shoreface contains the surf zone and extends some distance seaward of it. This distance is time-scale dependent since the upper shoreface is defined as the zone in which detectable changes in bed elevation can occur during a specified period. That is, adjustment of the upper shoreface in response to varying input forcing can be assumed to occur instantaneously for practical prediction of low-order change: *e.g.*, over a single time-step used in large-scale models of the shoreface (STIVE and DE VRIEND, 1995). The morphological time scale, shown schematically in Figure 3 of Part 1 can be estimated directly from field measurements of profile closure, the most commonly applied version of which is the annual closure depth (HALLERMEIER, 1981). Closure depths generally increase with the time period over which observations are made (NICHOLLS *et al.*, 1998; HINTON *et al.*, 1999), so the extent of the upper shoreface increases with time scale (and thus the time step used in a model).

Short-term (*e.g.*, annual) closure depths discriminate the seaward limit to morphological change occurring at the sub-time step level in models of low-order coastal behavior. This seaward limit (*i.e.*, h_* and L_* in eq. 5) and the shoreline circumscribe the *active zone* defined by STIVE and DE VRIEND (1995). Since the active-zone closure depth increases with time scale, longer time steps can be used in modeling longer evolutionary sequences, but only to a limited degree. Low-order coastal change is characterized by significant morphologic change on the lower shoreface (Figure 2 and 3 in Part 1), as shown by the illustrations in the previous section. For low-order coastal evolution therefore, closure provides a measurable index for the time scale of morphological coupling between the upper and lower shoreface. Extrapolation of this index for time spans beyond several decades is the only empirical approach available at present, given the limited temporal coverage in existing data sets for the lower shoreface.

On medium scales, closure is typically observed near the seaward limit of the active zone influenced by breaking waves. The repeated onshore and offshore migration of bars produce large gross fluxes of sediment and a distinct associated closure. At longer time scales, closure can decouple from the active zone and move onto the middle/lower shoreface. Preliminary analysis with the ADM suggests that closure steadily moves offshore with increasing time scale and the primary control shifts from breaking waves to shoaling waves and more general shoreface processes. Repetitive profile observations (since 1965) on the Holland coast to 16 m depths show similar behavior, together with a shoreward closure related to breaking waves (HINTON and NICHOLLS, 1998; HINTON *et al.*, 1999). On half the profiles at a 25-year time scale, the shoreward closure is followed by reopening of the profile on the middle shoreface and a final seaward closure near the limit of the data (about 16 m water depth). Reopening was associated with slow, near-continuous profile erosion: the eroded sand is inferred to move onshore to the active zone (STIVE *et al.*, 1991). The temporal pattern of reopening is nearly linear: extrapolation suggests that in about 50 years, all the profiles on the Holland coast will show reopening, and in about 100 years, all the profiles will be morphodynamically active across the entire surveyed profile.

Process measurements along the US East Coast (NIEDORODA *et al.*, 1985; WRIGHT *et al.*, 1991; 1995; BEAVERS *et al.*, 1999) and in Spain (GARCIA *et al.*, 1998) show that the shoreface is so active that, as time scale increases, closure is likely to move further seaward than the depths indicated from bathymetric data available at present (WRIGHT, 1995; NICHOLLS *et al.*, 1998). Furthermore, closure depth concepts has been applied to sandy systems. In using closure as a time-scale index for low-order change further out on the lower shoreface of the more generic coastal tract (*e.g.*, Figure 4), the influence of different sediment grades (*i.e.* sand, silt and clay) must be taken into account.

SUMMARY AND CONCLUSIONS

Low-order coastal evolution involves systematic coastal change upon which all other morphological changes are superimposed. This type of change manifests itself in coastal management as chronic problems such as shoreline migration that may persist for decades to centuries. For coastal management, low-order coastal change is of first order importance because it controls systematic trends in shoreline movement and morphology, and thus also the shift through time in locations impacted by higher-order changes (*e.g.*, responses to individual storms). At any given time (*i.e.*, on the sub-decadal time scale), the rates and magnitudes of higher-order changes dominate community perceptions, because these changes dominate the morphological-response signal in the short term.

Although important (non-negligible) on the decadal time scale, study of lower-order change requires a longer view to gain resolution (*i.e.*, to discriminate it from the high-order variance). Thus, model calibration is best achieved by hindcast comparison with geological signals of coastal change recorded in deposits laid down over centuries to millenia. The results of these studies can be extrapolated down to the decadal time scale to provide low-order trends upon which higher changes (*e.g.*, measured in monitoring programs) are superimposed. Predictions from aggregated models however only address probable future states. Aggregation precludes exact predictions for particular future instances (i.e, realizations).

Quantitative prediction of rates of change and future shoreline positions must contend with the climatology of forcing and resulting sediment-transport regime prevailing at any given site. That is, predictions must capture the residual effects of a very large number of fluctuations in the direction and intensity of sediment transport summed over time spans (decades to millenia) that are typically four to six orders of magnitude longer than individual transport events (hours to days). Generally, transport residuals are smaller than or sim-

Annex. *Estimates are based on: a) for the period 5000 BP to present, analyses of geological data by* Beets *et al. (1992); and b) for 7400 to 5000 BP on STM simulations using a sea-level history derived from modeling by* Lambeck *(1995) involving eustatic and hydro-isostatic effects. Data after 2000 BP are split into sub cells 'south' and 'north' of Haarlem (i.e., Hoek of Holland-Haarlem, and Haarlem-Den Helder respectively).*

Time Calendar Years BP	Rate of Sea-level Rise mm a^{-1}	Shoreline Displacement Advance (+) or Retreat (−) m a^{-1}		Terms in Coastal-evolution Trajectory (eq. 5): C_ph^* $m^2 a^{-1}$		Accommodation (+/−) Due to Sea-level Change $m^3 a^{-1} m^{-1}$	Combined Sources: Supply (+) & Loss (−) $m^3 a^{-1} m^{-1}$	
7400–7200	17.5	−185.3		−1853.2		−8.75	30	
7200–7000	15.5	−132.5		−1324.8		−7.75	30	
7000–6800	14.0	−102.5		−1024.7		−7.00	29	
6800–6600	12.5	−72.8		−728.5		−6.25	29	
6600–6400	12.0	−47.2		−472.4		−6.00	28	
6400–6200	11.5	−37.6		−376.4		−5.75	28	
6200–6000	11.0	−28.2		−282.3		−5.50	27	
6000–5800	9.0	−13.7		−136.8		−4.50	27	
5800–5600	8.0	−0.62		−62.0		−4.00	26	
5600–5400	7.0	−2.4		−24.0		−3.50	26	
5400–5200	5.0	2.6		25.8		−2.50	25	
5150–4500	2.0	2.1		21.0		−1.00	22	
4700–2000	1.0	1.6		16.0		−0.50	17	
2000–1000	0.75 ± 0.25	0.3	−1.7	−17	−17	−0.38	3	−17
1125–375	0.75 ± 0.25	−1.1	−3.9	−31	−39	−0.38	−11	−39
500–150	0.75 ± 0.25	−0.6	−2.7	−27	−27	−0.38	−6	−27
150–0	1.75 ± 0.25	0.45	−1.65	−16	−17	−0.38	5	−16.5
Sub-cell:		south	north	south	north		south	north

ilar to the predictive error for gross transport based on standard methods. Thus, scaling up estimates of the residuals to quantify the transport regime associated with the forcing climatology is unlikely to provide reliable predictions (De Vriend, 2003, this volume).

We attempt to sidestep this problem through a behavior-oriented approach in modeling low-order coastal change. This approach deals with the forcing climatology directly through a highly aggregated representation of the processes. We have developed three types of behavior models that capture forcing climatology in different ways: *i.e.,* as macroscopic forcing, summary forcing and abbreviated forcing. We applied these models in comparative experiments on long-term coastal change using data sets from the contrasting environments of SE Australia, Pacific and Atlantic USA, and European North Sea Coasts (Netherlands).

Data-models were compiled for each data set to define coastal tracts with spatial and process dimensions consistent with the behavior models applied, and to discriminate between boundary conditions and internal components. The internal components of the tract interact by sharing a common pool of sediments. The data-templating procedures for developing the data-models consume about 90 percent of the modeling effort: the numerical experiments are a minor component of the work overall. The level of expertise required for data-model templating makes these types of models unsuitable as coastal-management tools for use by novices.

We assumed alongshore homogeneity in morphology and processes for each data set when compiling data-models. Thus, each coastal cell was represented (through alongshore averaging) as a cross-shore profile of unit width. The experiments then entailed comparison of simulated coastal-change against coastal evolutions evident in the data-models. Synthesis of the modeling results demonstrates the following coastal-tract principles:

- The gross kinematics of the coastal tract are constrained and steered by sediment-mass continuity. The rate of coastal advance or retreat is determined quantitatively by the balance between the sediment accommodation-space generated due to sea-level rise (or lost due to sea-level fall), and sediment availability (being the sum of external sinks and sources). The relative sea-level change is a virtual sink/source term since there is no absolute loss, although the response is comparable to the impact of a real source/sink regarding horizontal movements of the upper shoreface.
- The upper and lower shoreface are coupled in first-order coastal change. Sediments (particularly the fine fraction) diffuse from the upper out across the lower shoreface (*e.g.,* the US Columbia River coast), but if the lower shoreface is shallower than required for equilibrium, then sand is transferred to the upper shoreface from the landward portion of the lower shoreface. This transfer offsets coastal-recession tendencies caused by other factors (*e.g.,* the central Netherlands coast) or, in the absence of these, the transfer produces a seaward advance of the coast (*e.g.,* the Tuncurry coast in SE Australia), and during periods of falling sea level.
- The upper shoreface and backbarrier (lagoon, estuary or mainland) are coupled in first-order coastal change. Sediment accommodation-space is generated in the backbarrier by sea-level rise (and reduced by sea-level fall), but the amount of space is moderated also by influx of fine sediments from the coast, or sand and mud from fluvial and biogenic sources. Remaining space can then be occupied by

sand transferred from the upper shoreface causing a retreat of the latter: *e.g.*, transgressive phases of the central Netherlands coast and the US-Atlantic (Duck) coast during the mid Holocene, and the Netherlands Wadden Sea coast under present conditions. Conversely, an increase in the tidal prism of active tidal basins (due to changes in tidal forcing) causes a transfer of sand from the backbarrier to the shoreface causing seaward advance of the latter.

- Rates of first-order coastal evolution due to upper-shoreface couplings with the backbarrier and lower shoreface are governed by a) second-order processes of the latter two sub systems, and b) the zero-order forcings of relative sea-level change and rates of external sediment supply or loss to the coastal tract.

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LITERATURE CITED

ATWATER, B.F.; NELSON, A.R.; CLAGUE, J.J.; CARVER, G.A.; YAMAGUCHI, D.K.; BOBROWSY, P.T.; BOURGEOIS, J.; DARIENZO, M.E.; GRANT, W.C.; HEMPHILL-HALEY, E.; KELSEY, M.H.; JACOBY, G.C.; NICHENKO, S.P.; PALMER, S.P.; PETERSON, C.D. and REINHART, M.A., 1995. Summary of coastal geological evidence for past great earthquakes at the Cascadia subduction zone. *Earthquake Spectra*, 11, 1–18.

BEAVERS, R.; HOWD, P.; BIRKEMEIER, W. and HATHAWAY, K., 1999. Evaluating profile data and depth of closure with solar altimetry. *Proceedings of Coastal Sediments '99*, American Society of Civil Engineers, 479–490.

BEETS, D.J.; VAN DER VALK, L., and STIVE, M.J.F., 1992. Holocene evolution of the coast of Holland. *Marine Geology*, 103, 423–443.

BEETS, D.J.; CLEVERINGA, P.; LABAN, C. and BATTEGAZORE, P., 1995. Evolution of the lower shoreface of the coast of Holland between Monster and Noordwijk. *Netherlands Royal Geological Service Bulletin*, 52, 235–247.

BIEGEL, E.J., 1993. *Morphological changes due to sea-level rise in tidal basins in the Dutch Wadden Sea versus concepts morphological response model MORRES.* Report IMAU-93.14. Utrecht University: Institute for Marine and Atmospheric Research Utrecht, 57 p.

BOWEN, A.J., 1980. Simple models of nearshore sedimentation, beach profiles and longshore bars. *In*: MCCANN, S.B. (ed.), *The Coastline of Canada*, Geological Survey of Canada Paper, 80–10, 1–11.

BRUUN, P., 1962. Sea-level rise as a cause of shore erosion. *Journal of Waterways Harbors Division*, American Society of Civil Engineers, 88, 117–130.

BUIJSMAN, M.C., 1997. *The impact of gas extraction and sea-level rise on the morphology of the Wadden Sea.* Report H3099.30, Delft Hydraulics, Delft, 78p.

CAREY, J. S.; SWIFT, D. J. P.; STECKLER, M.; REED, C. W. and NIEDORODA, A. W., 1999. High resolution sequence Stratigraphy modeling: effects of sedimentation processes. *Society of Economic Palaeontologists and Mineralogists Special Publication* 62, 151–164.

CHAPPELL, J. and POLACH, H., 1991. Post-glacial sea-level rise from a coral record at Huon Peninsula, Papua New Guinea. *Nature*, 349, 147–49.

CLEVERINGA, J., 2000. *Reconstruction and modelling of Holocene coastal evolution of the western Netherlands.* University of Utrecht: Geologica Ultraiectina, No. 200, 197p.

COWELL, P.J.; HANSLOW, D.J. and MELEO, J.F., 1999a. The Shoreface. *In*: SHORT, A.D. (ed.), *Handbook of Beach and Shoreface Morphodynamics.* Chichester: Wiley, 37–71.

COWELL, P.J.; ROY, P.S.; CLEVERINGA, J., and DE BOER, P.L., 1999b. Simulating coastal systems tracts using the shoreface translation model. *Society of Economic Palaeontologists and Mineralogists Special Publication*, 62, 165–175.

COWELL, P.J.; ROY, P.S. and JONES, R.A., 1995. Simulation of LSCB using a Morphological Behaviour Model. *Marine Geology*, 126, 45–61.

COWELL, P.J.; STIVE, M.J.F.; NIEDORODA, A.W.; DE VRIEND, H.J.; SWIFT, D.J.P.; KAMINSKY, G.M., and CAPOBIANCO, M., 2003. The Coastal-Tract (Part 1): A conceptual approach to aggregated modelling of low-order coastal change. *Journal of Coastal Research* (this volume)

COWELL, P.J.; STIVE, M.J.F.; ROY, P.S.; KAMINSKY, G.M.; BUIJSMAN, M. C.; THOM, B.G. and WRIGHT, L.D, 2001. Shoreface Sand Supply to Beaches. *Proceedings of the 27th International Conference on Coastal Engineering*, American Society of Civil Engineers, 2495–2508.

CURRAY, J.R., 1964. Transgressions and regressions. *In*: MILLER, R.C. (ed.), *Papers in Marine Geology*, New York: McMillan, 175–203.

DEAN, R.G., 1991. Equilibrium beach profiles: characteristics and applications. *Journal of Coastal Research*, 7, 53–84.

DEAN, R.G. and MAURMEYER, E.M., 1983. Models of beach profile response. *In*: Komar, P. and Moore, J. (eds), *CRC Handbook of Coastal Processes and Erosion*, Boca Raton: CRC Press, 151–165.

DE VRIEND, H.J.; CAPOBIANCO, M.; CHESTER, T.; DE SWART, H.E.; LATTEUX, B. and STIVE, M.J.F., 1993. Approaches to long-term modelling of coastal morphology: a review. *Coastal Engineering*, 21, 225–269.

EYSINK, W.D., 1991. Morphologic response of tidal basins to changes. *Proceedings of the 21st International Conference on Coastal Engineering*, American Society of Civil Engineers, 1948–1961.

FIELD, M.E.; MEISBURGER, E.P.; STANLEY E.A. and WILLIAMS, S.J., 1979. Upper Quaternary peat deposits on the Atlantic inner shelf of the United States. *Geological Society of America Bulletin*, New York, 90, 618–628.

GARCIA, V.; JIMENEZ, J.A.; SANCHEZ-ARCILLA, A.; GUILLEN, J. and PALANQUES, A., 1998. Short-term relatively deep sedimentation on the Ebro Delta coast: Opening the closure depth. *Proceedings of the 26th International Conference on Coastal Engineering*, American Society of Civil Engineers, 2902–2912.

HALLERMEIER, R.J., 1981. A profile zonation for seasonal sand beaches from wave climate. *Coastal Engineering*, 4, 253–277.

HINTON, C. and NICHOLLS, R.J. 1998. Depth of closure along the Holland coast. *Proceedings 26th International Conference on Coastal Engineering*, American Society of Civil Engineers, 2913–2925.

HINTON, C.; NICHOLLS, R.J. AND DUNSBERGEN, D. 1999. Profile Re-Opening on the Shoreface of the Holland Coast. *Proceedings Coastal Sediments '99*, American Society of Civil Engineers, 535–550.

KAMINSKY, G.M.; RUGGIERO, P.; GELFENBAUM, G. and PETERSON, C., 1997. Long term coastal evolution and regional dynamics of a US Pacific Northwest littoral cell. *Proceedings of Coastal Dynamics '97*, American Society of Civil Engineers, 614–623.

LAMBECK, K., 1995. Late Devensian and Holocene shorelines of the British Isles and North Sea from models of glacio-hydro-isostatic rebound. *Journal of the Geological Society*, 152, 437–448.

LEATHERMAN, S.P., 1983. Barrier dynamics and landward migration with Holocene sea-level rise. *Nature*, London, 301, 415–418.

MEISBURGER, E.P.; JUDGE, C., and WILLIAMS, S.J., 1989. Physiographic and geological setting of the Coastal Engineering Research Centre's field research facility. Washington, *US Army Corp of Engineers, Miscellaneous Paper CERC*, 89-9, 28 p.

MEYERS, R.A.; SMITH, D.G.; JOL, H.M. and PETERSON, C.D. 1996. Evidence for eight great earthquake-subsidence events detected with ground penetrating radar, Willapa barrier, Washington. *Geology,* 24, 99–102.

NICHOLLS, R.J.; BIRKEMEIER, W.A. and LEE, G-H., 1998. Evaluation of depth of closure using data from Duck, NC, USA. *Marine Geology,* 148, 179–201.

NIEDORODA, A.W. and SWIFT, D.J.P., 1991. Shoreface processes. *In:* HERBICH, J.B. (ed.), *Handbook of Coastal and Ocean Engineering,* Vol. 2, Houston: Gulf Publishing Co., 736–769.

NIEDORODA, A.W; REED, C.W.; SWIFT, D.J.P., ARATO, H. and HOYANAGI, K., 1995. Modeling shore-normal large-scale coastal evolution. *Marine Geology,* 126, 1/4, 181–200.

PETERSON C. and PHIPPS J. B., 1992. Holocene sedimentary framework of Grays Harbor basin, Washington, USA. *SEPM Special Publication No.48,* Society for Sedimentary Geology, 237–285.

PILKEY, O.H.; YOUNG, R.S.; RIGGS, S.R.; SMITH, A.W.; WU, H. AND PILKEY, W.D., 1993. The concept of shoreface profile equilibrium: A critical review. *Journal of Coastal Research,* 9, 255–278.

REED, C.W.; NIEDORODA, A. and SWIFT, D.J.P., 1999. Modeling sediment entrainment and transport processes limited by bed armoring. *Marine Geology,* 154, 143–154.

ROY, P.S.; COWELL, P.J.; FERLAND, M.A. and THOM, B.G., 1994. Wave dominated coasts. *In:* CARTER, R.W.G. and WOODROFFE, C.D. (eds), *Coastal Evolution: Late Quaternary shoreline morphodynamics,* Cambridge: Cambridge University Press, 121–186.

ROY, P.S.; ZHUANG, W-Y; BIRCH, G.F.; COWELL, P.J. and COGXIZN LI, 1997, *Quaternary Geology of the Forster-Tuncurry coast and shelf, southeast Australia.* Geological Survey Report: GS 1992/201, Sydney: NSW Dept. of Mineral Resources, 405p.

SHA, L. P., 1992. *Geological research in the Ebb-tidal delta of the 'Het Friesche Zeegat', Wadden Sea, The Netherlands.* Haarlem: Rijks Geologische Dienst, 20p.

SHERWOOD, C.R., 1995. *Measurements and modeling of suspended-sediment transport on the Northern California continental shelf.* PhD dissertation, Seattle: University of Washington, 175p.

SLOSS, L.L., 1962. Stratigraphical models in exploration. *Journal of Sedimentary Petrology,* 32, 415–422.

STERNBERG R.W., 1986. Transport and accumulation of river-derived sedimentation the Washington continental shelf, USA. *Journal of the Geological Society,* London, 143, 945–956.

STIVE, M.J.F. and DE VRIEND, H. J., 1995. Modelling shoreface profile evolution. *Marine Geology,* 126, 235–248.

STIVE, J.F.; CAPOBIANCO, M.; WANG, Z.B.; RUOL, P. and BUIJSMAN, M.C., 1998. Morphodynamics of a tidal lagoon and the adjacent coasts. *In:* DRONKERS, J. and SCHEFFERS, M. (eds), *Physics of estuaries and coastal seas: Proceedings of the 8th International Biennial Conference on Physics of Estuaries and Coastal Seas.* Rotterdam: Balkema, 397–407.

STIVE, M.J.F.; ROELVINK, D.J.A. and DE VRIEND, H.J., 1991. Large-scale coastal evolution concept. *Proceedings 22nd International Conference on Coastal Engineering,* American Society of Civil Engineers, 1962–74.

SWIFT, D.J.P., 1976. Continental shelf sedimentation. *In:* STANLEY, D.J. and SWIFT, D.J.P., (eds.) *Marine Sediment Transport and Environmental Management,* New York: Wiley, 311–350.

THORNE, J.A. and SWIFT, D.J.P., 1991. Sedimentation on continental margins, II: application of the regime concept. *In:* SWIFT, D.J.P.; OERTEL, G.F.; TILLMAN, R.W. and THORNE, J.A. (eds), *Shelf sand and sandstone bodies—Geometry, facies and sequence stratigraphy,* Special Publication No. 14 of the International Association of Sedimentologists, Oxford: Blackwell Scientific Publications, 33–58.

THORNE, J.A.; GRACE, E; SWIFT, D.J.P. and NIEDORODA, A.N., 1991. Sedimentation on continental margins, III: the depositional fabric—an analytical approach to stratification and facies identification. *In:* SWIFT, D.J.P.; OERTEL, G.F.; TILLMAN, R.W. and THORNE, J.A. (eds). *Shelf sand and sandstone bodies—Geometry, facies and sequence stratigraphy,* Special Publication No. 14 of the International Association of Sedimentologists, Oxford: Blackwell Scientific Publications, 59–87.

VAN DE MEENE, J.W.H., 1994. *The Shoreface-Connected Ridges Along the Central Dutch Coast.* Utrecht: Netherlands Geographical Studies 174, 222p.

VAN DER SPEK, A., 1994. *Large-scale Evolution of Holocene Tidal Basins in The Netherlands.* Ph.D. Thesis, Utrecht: University of Utrecht, 191p.

VAN DER VALK, L., 1992. *Mid- and Late-Holocene coastal evolution in the beach-barrier area of the western Netherlands.* Ph.D. thesis, Amsterdam: Vrije Universiteit, 235p.

VAN STRAATEN, L.M.J., 1965. Coastal barrier deposits in south and north Holland—in particular in the area around Scheveningen and Ijmuiden. *Meded. Geol. Sticht.,* NS, 17, 41–75.

WANG, Z.B.; KARSSEN, B.; FOKKINK, R.J. and LANGERAK, A., 1996. A dynamic/empirical model for estuarine morphology. *In:* DRONKERS, J. and SCHEFFERS, M. (eds), *Physics of estuaries and coastal seas: Proceedings of the 8th International Biennial Conference on Physics of Estuaries and Coastal Seas.* Rotterdam: Balkema, 279–286.

WOLF, S. C.; HAMER, M. R. and MCCROY P. A., 1997. *Quaternary Geologic Investigations of the continental shelf offshore southern Washington and northern Oregon.* Open-file report 97–677. U.S. Department of the Interior, U.S. Geological Survey. 4 plates.

WOLMAN, M.G. and MILLER, J.P., 1960. Magnitude and frequency of forces in geomorphic processes. *Journal of Geology,* 68, 54–74.

WRIGHT, L.D., 1995. *Morphodynamics of Inner Continental Shelves.* Boca Raton: CRC Press, 241p.

WRIGHT, L.D.; BOON, J.D.; KIM, S.C. and LIST, J.H., 1991. Modes of cross-shore sediment transport on the shoreface of the Middle Atlantic Bight. *Marine Geology,* 96, 19–51.