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Large Scale Coastal Behavior '93



U.S. Geological Survey
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**U.S. Department of the Interior
U.S. Geological Survey**

Large-Scale Coastal Behavior '93

edited by
Jeffrey H. List¹

Open-File Report 93-381

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Foreward

Large-Scale Coastal Behavior '93 was held in St. Petersburg, Florida in November, 1993. This international conference was conceived as a means of bringing together a multi-disciplinary group of researchers addressing the physical evolution of coastal areas on a temporal scale of decades and a spatial scale of the littoral cell. A primary goal of the conference was to stimulate coastal research at these scales, furthering understanding of the impact of major anthropogenic modifications, sea-level rise, and altered hydrodynamic forcings on the coastal system.

This volume contains extended abstracts from the conference's 66 oral and poster presentations. With 60% of the abstracts contributed from 15 countries other than the United States, the conference truly represented an international gathering of scientists actively involved in large scale coastal behavior research.

The Conference Organizing and Scientific Committee gratefully acknowledges many individuals who assisted with the organization of the conference and the production of this volume. These include Sabina Goldgof, Gail Dunlop, Karen Monroe, Lance Thornton, Sandy Coffman, and Ife Davis.

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COASTAL DEVELOPMENT IN SURINAME AT DIFFERENT TEMPORAL AND SPATIAL SCALES

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Over the last 6000 years, the tropical chenier coast of Suriname (Figure 1) is characterized by relatively large and rapid changes. This implies different methods of approach related to temporal and spacial scales. Four scales are distinguished:

- Small scale coastal behaviour: not treated in this paper.
- Meso scale coastal behaviour: the actual development of coastal sections over a period of months to a few decades.
- Large scale coastal behaviour: the development of the entire Suriname coast on a time scale of decades to a few centuries (?).
- Meta scale coastal behaviour: the geological evolution of the coastal plain on a temporal scale of centuries to millenia.

Meso scale coastal behaviour.

Mudbanks

In Suriname the meso scale coastal development is determined by a system of extensive, shoreface-attached, mudbanks, which migrate with an average velocity of 1,5 km/yr in a westward direction. They are separated by interbank troughs. The source of the sediment is the Amazone River which supplies yearly some $11-13 \times 10^8$ tons of sediment into the Atlantic Ocean. Approximately 20% of this sediment is transported westward along the North coast of South America. Roughly $1,5 \times 10^8$ tons of this transport is in suspension and $1,0 \times 10^8$ tons in the form of the above mentioned mudbanks (Eisma et al., 1991). Their migration is due to accretion at the westside of the mudbanks and erosion of the eastside.

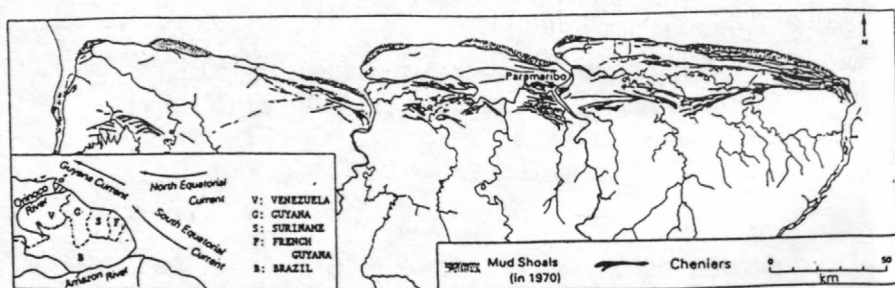


Fig. 1. The coastal area of Suriname.

The Suriname coastal waters can be classified as a micro-tidal, low to moderate (wave) energy environment. The velocity of the westward directed sea current at 50 cm above the bed, is between 10 and 50 cm/s.

The concentration of the Amazone-borne suspended material is high. Regularly, fluid mud is formed, especially in periods of strong winds when water turbulence is high. An important property of fluid mud is that it causes wave attenuation. Under the dominating hydrodynamic regime, fluid muds are deposited at the west side of the mudbanks. Therefore, mudbanks are advancing westward (Augustinus et al., 1989).

Over a period of time, due to compaction and maturation, the fluid mud changes into a more or less firm clay. Obviously, this clay does not have the wave damping property of the fluid mud. For that reason, waves are able to attack the eastside of the mudbanks and cause erosion.

Cheniers

In the interbank areas, the waves can reach the shore. This produces an erosive environment in which cheniers are formed.

In certain cases, e.g. due to a large local sand supply, cheniers are accumulating in a seaward direction, and form chenier bundles. All the cheniers are subject to the westward migrating mudbanks.

Coastal dynamics

In general, the Suriname coast can be characterized as a system of westward migrating erosion and sedimentation areas, driven by the mudbanks (Augustinus et al., 1989). The average 'wavelength' of the system is 45 kilometer. Together with the migration velocity of 1,5 kilometer per year this results in a periodicity of about 30 years.

Large scale coastal behaviour.

Balances of net accumulation and net erosion have been made for the period 1947-1981, using 5 series of aerial photographs (Table 1). It appeared that the Suriname coast has a net erosion from 1947 to 1966 and a net accretion from 1966 to 1981 (Augustinus et al., 1989). In East Suriname, this large scale trend commenced earlier than in West Suriname. The time scale of the large scale behaviour is unknown yet.

TABLE 1

Average net displacement (km²) of the Suriname coastline, deduced from a comparison of five series of aerial photographs.

	1947-1957	1957-1966	1966-1970	1970-1981
Suriname coast	-30.5	-15.2	+14.1	+57.4

The most successful hypothesis to explain this large scale coastal behaviour is based on changes in the direction and force of the Northeastern trade winds, which generate the waves in this area. The angle between the direction of wave propagation and the coast determines the ratio between the longshore wave energy flux and the cross-shore wave energy flux (Augustinus, 1987). A more easterly direction of the trade winds favours the longshore wave energy flux. This results in a growing length of the mudbanks (= a growing area of coastal accretion). A more northerly component will result in extensive winnowing processes and a more pronounced chenier formation. Stronger winds will cause a higher turbulence of the water and therefore favour fluid mud formation.

A statistical analysis has indicated that since the sixties the trade winds in the windy season (January-April) showed a shift towards the east combined with an increase in wind force. This coincides with a steady growth of the mudbanks and the above mentioned net coastal accretion, in other words confirms the hypothesis (Eisma et al., 1991).

Meta scale coastal development.

The evolution of the Suriname chenier plain, which started about 6000 years BP, is characterized by a net lateral accretion of clayey sediments. The period of lateral coastal extension can be divided into three sedimentation phases: Wanica, Moleson and Comowine, interrupted by periods in which extensive chenier bundles have been formed. The latter coincide with periods of lowering sea level.

A relation between a lowering sea level and the formation of cheniers is not very obvious. However, superimposed on the large scale coastal behaviour, sea level oscillations might offer an explanation. The chenier bundles which separate the sedimentation phases could possibly mark the coincidence of a period with more northerly trade winds and a period with lowering sea level.

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Mathematically shaping of tidal inlets.

by Willem T. Bakker¹ and Huib J. de Vriend²

1 Introduction.

De Vriend et al. [1] introduced a conceptual model of the dynamic behaviour of tidal inlets, making use of physical principles as well as empirical relationships.

Especially, the behaviour of the outer delta is investigated: in a two-line model (a diffusion-type model with the action of breaking waves as agent) a set of sources and sinks is introduced, simulating supply c.q. subtraction of sand from the outer delta by ebb- c.q. floodchannels.

The amount of material, supplied by the ebb-channel to the outer delta is derived from the shape of the outer delta itself and has (in the model) no relationship with the morphological state of the tidal basin.

A first step to a link between inner basin and outer delta is Bakker's analytical computation [2] of the tide-averaged morphological evolution of a prismatic tidal channel closed on one side.

The present paper is a continuation of [2]. The prismatic tidal channel has been replaced by a linearized tidal network of a multiple inlet system; some results concerning the most Western part of the Wadden Sea are shown.

Although the model is not yet operational, the outline is given and some preliminary results are shown.

2 The Lorentz method.

With the Lorentz method, a linearized method of tidal computation, the discharge q_{here} at a boundary of a tidal channel is calculated as:

$$q_{here} = \beta z_{here} - \alpha z_{there} \quad (1)$$

where z denotes the vertical tide (being an abbreviation of $Re(z \exp(i\omega t))$; i.e. α and β are complex numbers); *here* denotes the considered boundary of the channel and *there* the other end of the channel.

In a network of channels, the dynamic equation (1) for each channel and the continuity requirement at each of the N nodal points lead to a system of equations:

$$\overline{A}z = \overline{Z} \quad (2)$$

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where the $N \times N$ matrix \bar{A} contains the α - and β - coefficients for the various channels and the N -dimensional vectors \bar{z} and \bar{Z} contain the unknowns and the boundary conditions respectively. Thus the vertical tide in a node n of the network equals:

$$z_n = \sum_{k=1}^K b_{kn} Z_k \quad (3)$$

where b_{kn} is an element of the inverse matrix \bar{A}^{-1} ; K =number of boundary points (all other $N-K$ Z_k -values are zero). As an example, fig.1 and 2 show the Lorentz [1] schematization of the Western Wadden Sea, and fig.3 to 6 the modulus of the coefficients b_{kn} for the 4 inlets.

3 Tide-averaged flux of water over the shoals.

Tidal asymmetry is an important morphological agent in the basin. Since a linear model is unable to describe this, shoals and channels are treated separately.

Assuming the shoals to be at mean water level, the net transport of water \bar{Q}_{sh} over a shoal with width B_{sh} can be shown to be:

$$\bar{Q}_{sh} = \frac{1}{4} \hat{z} B_{sh} \hat{v}_{sh} \cos \phi_{sh} \quad (4)$$

Here \hat{z} and \hat{v} denote the amplitude of local vertical and horizontal tide and ϕ is the phase difference between z and v .

4 Tide-averaged flux of water through channels.

Also through channels a tide-averaged flux of water takes place above the level of LW. Here holds a similar formula for the net transport as (4), however with a factor 1/2 instead of 1/4.

The gradient of the tide-averaged water fluxes above LW-level (as well on the shoal as in the channels) acts as rain in the drain of the sewage system, consisting of the part of the channels below the LW-level; non-linear steady-flow calculations have been carried out for finding the distribution of this flow through the network. Friction of the channels has been enhanced by taking oscillatory tidal motion into account.

From these computations result the steady velocity \bar{v} , used for sediment transport computations.

Extension of this model is intended, by implementing the radiation stress, caused by the tidal motion into the dynamical equations. Also wind effects easily can be included.

5 Tide-averaged sand flux through the inlets.

For a combination of a harmonic and a steady flow (characterized by δ , the ratio between steady flow velocity \bar{v} and tidal velocity amplitude \hat{v}) a third-power sediment transport formula yields a half-tide averaged sediment transport (S) per unit of width:

$$(S)^* = S_1 v^3 \left(\frac{3}{2} \delta + \frac{4}{3\pi} \right) + O(\delta^2) \quad (5)$$

Here S_1 is the transport, when the (instantaneous) velocity is 1 m/sec.

Averaging over the total tide, the term " $4/3\pi$ " in (5) cancels out, strictly. However, as only the top of the horizontal tide causes big sand transport, considerable gain of accuracy can be obtained by taking existing asymmetry in horizontal tide into account and taking different values for \hat{v} during flood tide and ebb tide.

6 Effect of resonance on the stability of the inlet system.

An advantage of the linearization of the tide is the possibility of unraveling (on one hand) and clustering (on the other hand) of tidal features in the tidal basin. In this way one can simulate in a very condensed way the response of the tidal basin to changes of the inlet system, without loosing essential physics like resonance. This may be elucidated in the following way.

In the paper, the discharge q in a tidal channel in the k^{th} inlet is expressed as:

q - (outer boundary condition) * (influence tidal basin) * (influence friction gorge)

or, in formula:

$$q = \frac{A_g g}{\omega L_g} Z_k \frac{-\zeta^*}{1 - \zeta^*} \frac{1 + e^{4i\theta}}{2i} \quad (6)$$

where A_g and L_g are cross section and length of the tidal gorge. The value of θ depends on the bottom friction in the gorge [3] and is zero if friction is absent and $\pi/4$ for infinite large friction; thus the last factor in (6) varies between 1 (no friction) and 0 (infinitely large friction). In the following, this factor will be called " f ".

In the complex number ζ^* the resonance characteristics of the tidal basin (as function of the seaward boundary condition in the k^{th} inlet of the multiple inlet system) are comprised. The value $1 - \zeta^*$ is the ratio between the vertical tide at the outside of the gorge, compared with the one at the inside of the gorge. When ζ^* is 1, infinitely large tidal discharges in the gorge result (eqn. (6)). The variable denotes:

$$\zeta^* = \frac{\omega^2 L_g}{f, g} \frac{\sum_{n=1}^n [O\zeta]_n}{A_g} \quad (7)$$

In (7) O is the area, representative for the n^{th} node of the system in the tidal basin (fig. 1) and ζ_n is the complex influence coefficient b_{kn} , of which the module is depicted in fig. 3 to 6.

7 Future extensions

- Flood channels through the outer delta can be modelled as separate channels in the network.
- Ebb-and flood channels in the outer delta can be allowed to migrate, according to an empirical channel migration model. The channel dimensions depend on the orientation and the seaward boundary condition. If the orientation becomes hydraulically too unfavourable, new channels may form and take over.
- Using a cross-shore profile description [4], the spatial resolution may be increased. Thus a quasi-3D description of the edge of the outer delta may be obtained.

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- [3] Verslag van de Staatscommissie Zuiderzee 1918-1926 (1926), Algemeene Staatsdrukkerij (in dutch)
- [4] Bakker, J. Xu & O. Koshizawa, (1991), Coastal computations on the effect of offshore breakwaters at Niigata. Rijkswaterstaat, Tidal Waters Division and University of Hawaii, Oc. Engng. Dept.



Figure 1. Schematization tidal basins, Western Wadden Sea

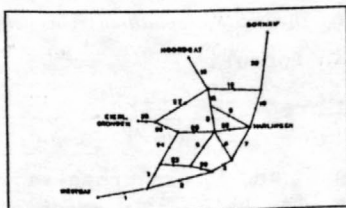


Figure 2. Schematization channel system, Western Wadden Sea

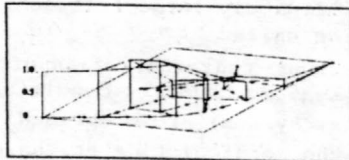


Figure 3. Influence Coefficients Westgat

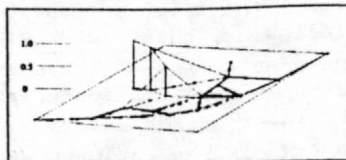


Figure 4. Influence Coefficients Eierl. Gronden

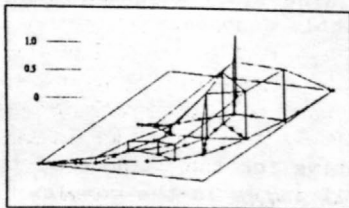


Figure 5. Influence Coefficients Noordgat

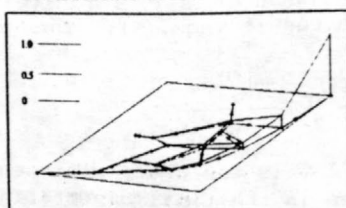


Figure 6. Influence coefficients Bornrif

Bristow, C.S.¹, Horn, D.P.² and Raper, J.F.².

In this paper a variety of sources are used to reconstruct the long-term accretion history of Scolt Head Island and determine the effect of long-term low magnitude events (sea level and sediment supply), medium term climatic cycles and short-term high magnitude (storm) events. The objective of this research is to predict how short term small scale nearshore processes and landform responses on beaches and spits are aggregated to define medium scale coastal system behaviour and coastline configurational change. Knowledge of such behaviour defined in terms of energy and sediment budgets will make it possible to study the thresholds and feedbacks in coastline evolution, and indicate the possible pathways between alternative landform configurations. Such models may also be coupled to predictions of North Sea sea level changes climatic change and storminess to predict the response of the coast to sea level rise.

Scolt Head Island is a barrier island on the North Norfolk coast in eastern England. The island is approximately 6.5 km long, with a sand and shingle beach along the north coast and numerous shingle bars and recurved spits to the south which enclose salt marshes. The island is a nature reserve and part of a heritage coast which has had very little development or anthropomorphic interference. An extensive drilling survey has been carried out to determine the structure of the recurved spits, and this has been combined with monitoring the active gravel bars to produce a comprehensive model of spit development and barrier island evolution on this macrotidal coast.

At least 20 recurved spits have been identified, and these are the dominant geomorphic element of the island. Consequently the process of spit formation appears to be the key to understanding the development of Scolt Head Island. The recurved spits record the progressive growth of the island from east to west. These were previously believed to be vertical and horizontal extensions of a solid gravel platform forming the skeleton of the island (Steers 1960). Recent drilling of over 50 shallow boreholes suggests that this is not always the case. Monitoring of recent movement of gravel bars on Scolt Head Island indicates that spit formation and migration is controlled by

wave action rather than longshore drift (Davis 1991). 3D modelling of sedimentary architecture based on mean particle size measured from boreholes samples (Raper et al. 1992) indicates that spits at Scolt Head Island have a shoreface structure with overwashing on the lee side. This is consistent with an origin as a wave dominated feature driven up from the tidal delta, rather than a detached shoreface recurved around the end of the island. Work by Oost et al. (1992) and Bristow, et al. (1992) has shown that spit development can be associated with a similar set of developmental stages from detached shoreface, to wave dominated intertidal bar, to supratidal spit. These stages are reversible in the same way as beach stages except when the spit bar develops a supratidal component in storm overwashing or through widespread deflation and dune building when it may become stabilised. The essential controls on 'spit' stages appear to be the balance between tidal flows around the spits and the wave-induced flows and sediment transport over it. This work is a development from the standard view of spits as detached shorefaces fed by longshore drift (Schwartz 1972).

A detailed historical record of the evolution of Scolt Head Island has been compiled by Allison (1985). Historical evidence of accretion on Scolt Head Island dates back to maps constructed from 1797. There are older maps and charts but their accuracy is questionable. The area was remapped in 1822 and 1825 at the time a dyke was constructed on salt marshes south of the island. Subsequent maps include the first edition of the Ordnance Survey Six Inch Series in 1886 and the second edition in 1904 with subsequent maps in 1949, 1958, 1969-70 with revisions in 1977. In addition there are a series of twelve plane table maps of Far Point constructed between 1927 and 1958 by Steers (1960). More recently airphotos have been used to observe changes in this area. Historical data provides us with a series of planform views of the accretion. It appears from the maps that the island elongated westwards by almost 1km between 1886 and 1904. A very rapid progradation of almost 5.5m per annum. During this time there was one major storm reported in 1897 with very destructive tides. It is possible that the elongation was triggered by this storm. However it is more likely that this progradational event represents the cartographers view of an area which became stabilised as a subaerial feature around this time. The elongation of the island is believed to have taken somewhat longer and may have been initiated in the 1820's with the enclosure of saltmarshes. The saltmarsh enclosure included blocking several tidal channels and a consequent reduction in the tidal prism which would have shifted the balance between tidal flux and wave driven processes at the end of the island. It is interesting to note that the enclosure of saltmarshes for improved agriculture has probably helped to improve the harbour at Brancaster through the construction of a natural

breakwater while at the same time leading to the gradual decline of the harbour due to increased sedimentation in some of the tidal channels. Analysis of aerial photographs shows that westward accretion of the island is continuing today with gravel spits developing on the tidal delta and being driven onto the end of the island by waves. The movement of sediment on the end of the island has been monitored using pebble tracers and detailed point surveys of the gravel spits. The topographic surveys can be used to calculate the volumes of sediment being moved on the end of the island and hopefully this will enable a sediment budget to be constructed so that future developments can be predicted.

The evolution of Scolt Head Island provides a suitable case study for research into the way in which small scale process and landform behaviour patterns can be aggregated into medium scale coastal system behaviour, with specific reference to landforms such as beaches, spits and barrier island bars which are composed of non-cohesive sediments. Accretion is episodic rather than continual and represents a state of punctuated equilibrium in the geomorphic system. Observations of accretion on the western end of the island, Far Point, indicate that wave refraction is the dominant spit forming process. A combination of high tide and critical wave heights are required for refraction to be effective. Pulses of accretion have been observed on diurnal and annual time scales, but it is clear that longer period fluxes in sediment supply and sea level are responsible for larger scale accretion events. The recurved spits on Scolt Head Island are arranged in clusters which may be linked to episodic increases in the rate of sea-level rise (Pethick 1980) or possibly changes in tidal regime. Bristow et al. (in prep) suggest that recent extension of the island may be a response to a reduction in the tidal prism following dyke construction in the 1820's.

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THE HISTORICAL EVOLUTION OF THE COASTS OF VENICE, ITALY

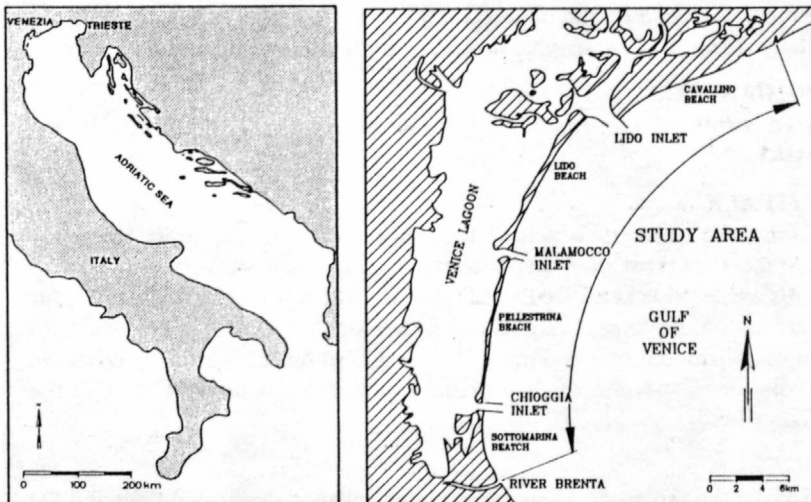
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INTRODUCTION

The coasts off the Venice Lagoon have for a long time been affected by man-made interventions.

The natural conditions at the location are a lagoon separated from the Adriatic Sea by low barrier islands. These islands were kept stable by a constant supply of sand from both south and north. The openings between the islands were wide and shallow and constantly changing. A location map and the area of interest are shown below.



Location map of Venice Lagoon and the adjacent coasts.

The hydrographic conditions in the north-western corner of the Adriatic are dominated by two wind systems: the Scirocco from south-east and the Bora from north-east. The tide is semi-diurnal with ranges up to 1.1m. The bed material in the area is fine sand.

HISTORICAL EVOLUTION

The natural quasi-stable barrier islands caused problems to habitation and navigation already back in the 1500's. The wide, shallow and constantly changing inlets caused difficulties for the navigation, and the low islands were occasionally breached during storms. Further, siltation took place within the lagoon due to the discharging of the rivers.

As early as from the 14th to the 17th century the rivers were diverted around the lagoon. From the middle to the turn of the 19th century, very large inlet jetties were constructed in the three inlets, navigation channels were dredged and sea walls at the islands were reinforced or established. All these initiatives caused changes in the evolution of the coastal zone. Due to a continually increasing problem of flooding of Venice and changing ecological conditions within the lagoon, the Consorzio Venezia Nuova initiated the planning of mitigating measures, including construction of gates across the three inlets.

The evaluations of effects of the new works on the evolution of the coasts have been based on the understanding of the historical evolution and the study of sediment transport processes by numerical models.

NUMERICAL MODELLING

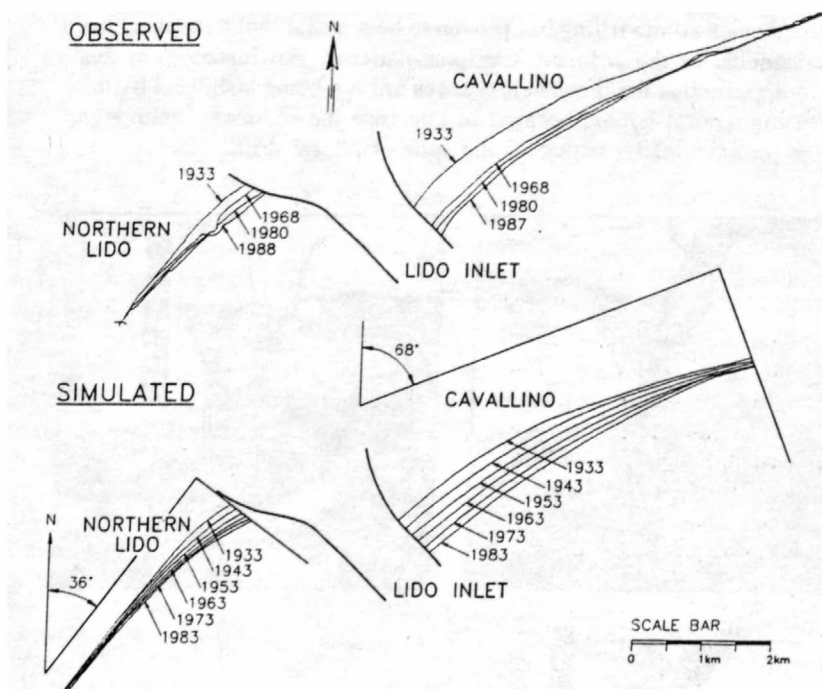
Two suites of numerical models have been applied, LITPACK and MIKE.

LITPACK

Along the adjacent beaches, the littoral drift is the most important transport mechanism. These beaches can be considered as quasi-uniform. This assumption is utilised in LITPACK, which is a suite of modules for calculation of waves, currents and littoral drift. The results are used both in establishment of a sediment budget and in modelling of the coastline evolution. Examples of the verification of the model are shown in the figure below.

MIKE

Around the three inlets, the interactions between the littoral drift and the strong tidal currents are important. Here, the sediment transport patterns are studied by the following types of models: Nearshore waves by a stationary directionally decoupled parametric model MIKE 21 NSW, waves within the tidal inlets where diffraction is important by a time-dependant vertically integrated Boussinesq model MIKE 21 BW, hydrodynamics by a depth-integrated, fully dynamic model, MIKE 21 HD and sediment transport and initial bed level changes by a physically based intra-wave period model, MIKE 21 ST.



Comparison of measured coastline evolution data and model results.

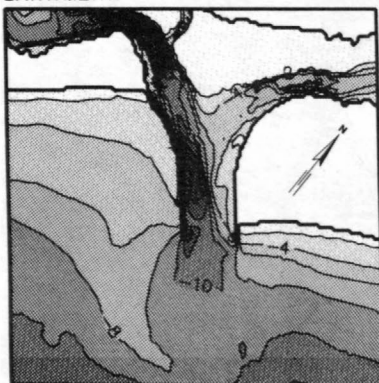
Due to the complexity of the latter set of models it is important to carefully select the conditions to be studied in detail. This has been done based on calculation of 1) the contributions to the yearly transport from all wave and current components in the statistics in 3 offshore positions along the study area, and 2) the littoral drift along the Cavallino and Sottomarina beaches, 3) the transport capacity within the inlets as function of the tidal range. 5 design conditions and their relevant frequencies of occurrence have been defined. In the following figures are examples of calculated wave-current sediment transport fields around Lido inlet at the time of maximum ebb flow.

COMMENTS

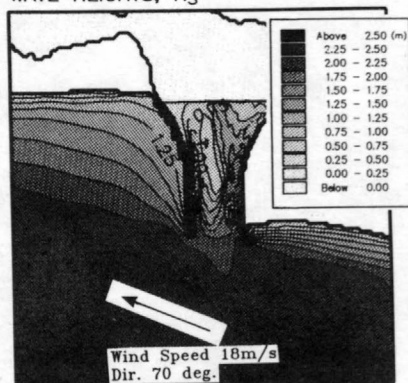
At the Venice tidal inlets the navigation problems were for many years solved by the establishment of large jetties combined with initial dredging. The long-term evolution of the downdrift beaches has been severe leeside erosion. During the latest decades bypass of the littoral drift has started causing the need for an average maintenance dredging in the order of magnitude of the littoral drift.

The mathematical modelling has proven to be a useful tool to support the understanding of the sediment transport patterns. For instance, in this case it appears that the downdrift beaches are not being stabilised by the increasing natural bypass because in this case the sediment - although able to pass the inlet - settles off the zone of littoral drift.

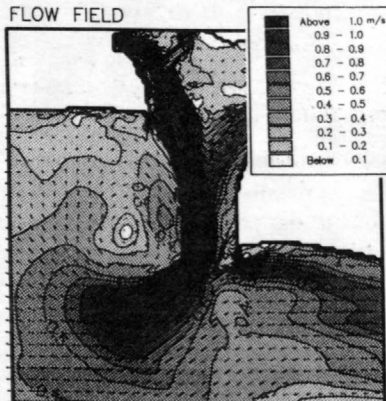
BATHYMETRY



WAVE HEIGHTS, H_s

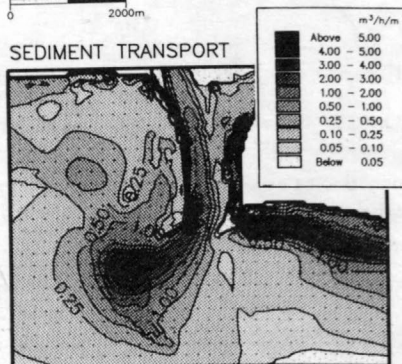


FLOW FIELD



0 2000m

SEDIMENT TRANSPORT



ACKNOWLEDGEMENT

The studies have been undertaken by DHI for Consorzio Venezia Nuova.

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REGIONAL SEAFLOOR CHANGE NEAR ST. MARYS ENTRANCE, GEORGIA/FLORIDA AND ITS INFLUENCE ON SHORELINE RESPONSE

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Hydrographic surveys of regional nearshore morphology provide a direct source of data for quantifying seafloor changes. Historically, these data have been collected in conjunction with regional shoreline position surveys by the U.S. Coast and Geodetic Survey (currently National Ocean Service of the National Oceanographic and Atmospheric Administration). However, bathymetric data are often overlooked as a primary source of information for assessing small-scale (large areas) coastal evolution or site-specific (large-scale) response to natural and human-induced processes. This neglect may be related to the amount of analysis necessary to attain an accurate result. Comparison of digital bathymetric data for the same region but different time periods provides a method for calculating net movements of sediment into (accretion) and out of (erosion) an area of study. Several manual and automated techniques have been used for making quantitative estimates of change. Moody (1964) superimposed contour data from charts of different time periods to determine change. Pierce (1969) used data point comparisons for exact geographic positions on charts of different time periods to calculate volumetric changes. Until the 1980s, these two procedures were standard practice (primarily contour overlay) for evaluating historical changes in nearshore bathymetry (e.g., Stauble and Warnke 1974), particularly those related to inlet systems (Dean and Walton 1973, Olsen 1977). However, in recent years, bathymetric data have been processed using commercially available surface modeling software (Hansen and Knowles 1988; List, Jaffe, and Sallenger 1991).

Unlike bathymetric surveys, analysis of historical shoreline position data has been applied routinely for quantifying rates of shoreline retreat and advance (e.g., Caldwell 1966; Morton 1979; Leatherman 1984; Byrnes et al. 1991; McBride et al. 1991). Coastal scientists, engineers, and planners often use this information for estimating the magnitude and direction of sediment transport, monitoring engineering modifications to a beach, examining geomorphic variations in the coastal zone, establishing coastal erosion setback lines, and verifying shoreline change numerical models. The purpose of this study was to quantify changes in shoreline position and nearshore bathymetry to identify trends in small-scale (large area) coastal evolution, and to evaluate the impact of natural processes and human influences on shoreline response near Cumberland Island, Georgia and Amelia Island, Florida. Digital data for the area between St. Andrew Sound, Georgia, and Nassau Sound, Florida, and seaward from the high-water shoreline to the 12-m depth (NGVD) contour or the limit of data were used to assess change from the 1850s to 1992.

Shoreline Change

The magnitude and direction of shoreline position change were evaluated for Cumberland and Amelia Islands using six topographic surveys (1857/71, 1924, 1933, 1957, 1973/74,

1991). In most cases, change measurements exceeded potential errors; however, in some cases, areas showing little change over short time intervals were considered insignificant. Four primary results summarize the findings of the historical shoreline change study. First, average long-term shoreline position change is net progradational for the Cumberland-Amelia barrier island system (Figure 1). In other words, the magnitude of shoreline advance exceeded retreat for the study area. Between 1857/70 and 1991, Cumberland Island prograded at a net rate of 1.5 m/year; Amelia Island prograded 0.4 m/year between 1857/71 and 1991. Second, five areas of substantial shoreline movement were identified for the study area. The northern margin of the Cumberland Embayment, southern Cumberland Island at the jetty, and the northern end of Amelia Island at the jetty show large shoreline advances since 1857/71. The historical protuberance near Fernandina Beach and the southern 6 km of Amelia Island show net shoreline retreat. Third, only one area of long-term chronic erosion is identified. This includes the southern 4 km of Amelia Island where the rate of shoreline retreat has steadily increased since 1871. Finally, for the five areas of significant change, most movement took place between 1857/71 and 1924. This time period encompasses jetty construction at St. Marys Entrance that is related to deposition on adjacent shorelines. However, direct association between retreat along southern Amelia Island and jetty placement is not as straightforward because inlet processes at Nassau Sound exert significant influence on adjacent shoreline response.

Bathymetric Change

Bathymetry data from three time periods were analyzed to assess regional changes in coastal response (1855/75, 1924, and 1953/79). Prior to jetty construction in the early 1900s, the morphology of the ebb-tidal delta at St. Marys Entrance appeared similar to those associated with other natural inlet systems separating offset barrier islands (Hayes and Kana 1976, Hayes 1991). Regional characteristics of nearshore morphology for the period 1855/75 show several important features that depict the framework of this system. First, the orientation of contours defining shoals is skewed to the south, suggesting a net southerly directed transport of sediment for the entire region. The tidal channel at all three inlet

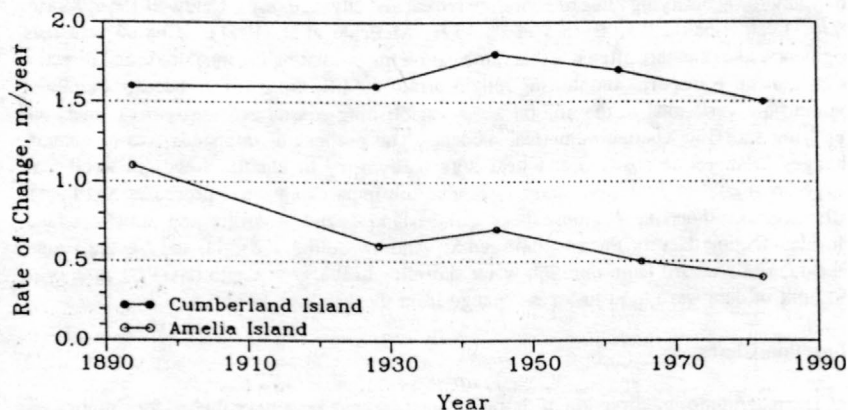


Figure 1. Cumulative change in shoreline position relative to the 1857/71 shoreline

systems (St. Andrew Sound to the north, St. Marys Entrance, and Nassau Sound to the south) exits the coast with a southeast orientation. Shoals associated with St. Andrew Sound are oriented to the southeast as well. In addition, the ebb-tidal delta identified with St. Marys Entrance is strongly skewed to the south. Second, the shoal complex associated with St. Andrew Sound is by far the most extensive subaqueous feature in the study area, and the channel is greater than 15 m deep in some places. Sand deposits defined by the 6-m (NGVD) depth contour extend approximately 8 km offshore and at least 10 km south along Cumberland Island. Third, Stafford Shoal, a sand deposit located midway along Cumberland Island, is approximately the same size as the ebb-tidal delta at St. Marys Entrance and extends 4 to 5 km offshore. Fourth, the ebb shield associated with St. Marys ebb-tidal delta is well defined by the 6-m (NGVD) depth contour. The ebb-tidal delta extends about 4.5 km seaward from the coast adjacent to the entrance and connects with the shoreline approximately 5 km south of the main channel where a headland existed. The deposit appears to provide an efficient conduit for sediment bypassing from Cumberland to Amelia Islands. Fifth, Nassau Sound has the smallest shoal deposits, primarily confined to within 2 km of the coast. The general decrease in sand storage in ebb-tidal deltas from north to south correlates well with a decrease in tidal prism in backbarrier environments. Sixth, the inner shelf area north of St. Marys Entrance is broader and more gently sloping than that associated with Amelia Island to the south. Finally, the 6-m (NGVD) depth contour away from tidal inlets and Stafford Shoal is located within 1 km of the shoreline and within 0.5 km of the coast along central Amelia Island.

Spatial and temporal change in bathymetry between 1855/75 and 1953/79 were quantified using Intergraph's MGE Terrain Modeler, a surface modeling routine. The most pronounced change is that associated with the ebb-tidal delta at St. Marys Entrance. Approximately 90 million cu m of sediment are associated with the development of this feature. Conversely, the channel entrance between the jetties shows a net loss of approximately 40 million cu m, suggesting that the delta has been a net sink of littoral and nearshore shelf sediment since jetty construction. However, most deposition associated with a primary portion of the ebb-tidal delta (documented by comparing incremental changes between 1870/75 and 1992, and defined by boundaries of the 1992 bathymetric survey; see Byrnes and Hiland 1993) occurred by the 1950s, and the amount of change since has decreased (Figure 2). As in previous discussions, net southerly drift of sediment is illustrated for the St. Andrew Sound ebb-tidal delta and Stafford Shoal. Regions of deposition are situated downdrift (south) of erosion zones. In addition, the orientation of deposition related to the modern ebb-tidal delta at St. Marys Entrance is to the south. Erosion on the southern half of the historical ebb-tidal delta and adjacent coastline of Amelia Island is associated with deposition along northern Amelia Island at the fillet and along the central portion of the island. Deposition along the northern 4 km of Amelia Island is related to a reversal in sand transport associated with localized wave refraction across the historical ebb-tidal delta. The nearshore region along the southern 6 km of Amelia Island indicates erosion; however, nearshore deposition along the southern 3 km of the island, particularly at the northern margin of the inlet, also is shown. For changes that have occurred throughout the period of record, the only direct connection that can be made between delta evolution and shoreline change is the large adjustments that were shown everywhere in the system between the mid-1800s and 1924. Since then most system changes associated with the channel and ebb-tidal delta appear decoupled from shoreline response.

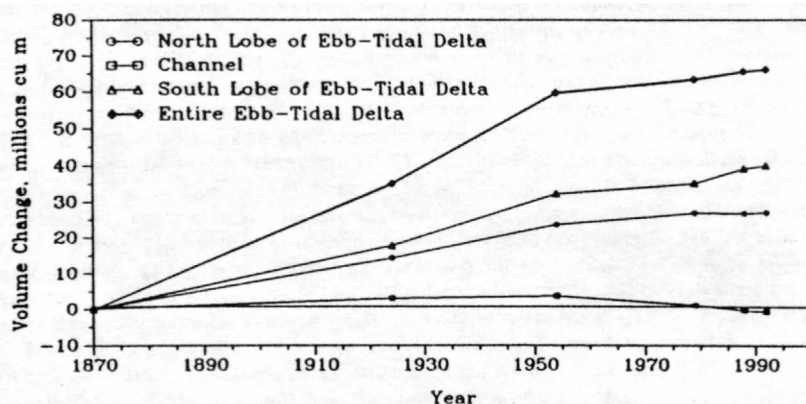


Figure 2. Cumulative changes in sediment volume at St. Marys ebb-tidal delta as determined by the limits of the 1992 bathymetric survey

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Behaviour-Oriented Models Applied to Long Term Profile Evolution

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SUMMARY A possible approach for the development of simple predictive methods of shoreface profile evolution is described. We hereby introduce the model concept and describe its applicability on the base of *diffusion-type* formulations over time scales from seasons, to years and decades.

INTRODUCTION

We consider uniform sandy beaches of several km's length where we assume that the along shore variation of the average coastal profile and of basic hydrodynamic processes can be neglected. On the contrary, the processes on the cross-shore direction are important in determining the profile evolution. In this situation, *large scale coastal behaviour*, and specifically the response of the coastal system to uniform environmental forcings, can be computed on the base of the principle of sand-mass conservation and of geometric rules that allow the evaluation of the displacement of sandy beaches and the change in position of the coastline. This *static* approach does not require a rigorous definition of the involved processes but rather their parametric representation. Nevertheless it allows to evaluate possible scenarios of beach morphology evolution over a variety of time scales. For coastal management applications however the *dynamic* evolution is also of interest.

Our main interest lays on the upper to middle shoreface, for processes on time scale from years to decades and on the total shoreface, for processes on time scale of decades. This implies that we are dealing with what we would define as *long-term modelling*: modelling on a time scale longer than can be handled by existing validated process-based (mathematical-physical) models (De Vriend et al., 1993a). The profile development on the scale of years to a decade is most probably not well represented by process based models, both because of possible deficiencies and because in reality there will be more than just cross-shore processes responsible for the profile development.

BEHAVIOUR ORIENTED MODELLING

Our objective is to obtain modelling tools that may be applied in a context where little experimental information is available, with the aid of validated short term process-based models. They should be able to reproduce static conditions under uniform environmental forcings, but also to give an assessment of the dynamic transitions between different static conditions following the application of varying forcings. Possible formal approaches to extend the use of process based models may be based on the manipulation of the available mathematical formulations to simplify the hydrodynamic input conditions (input filtering) in order to obtain a representative set of inputs, or to simplify the physical processes (process filtering) to reformulate the model on the scales of interest.

These approaches however, even if formally correct, present a number of practical difficulties that do not allow to solve the problem completely on the larger scales when either computational resources or available data are lacking. This implies that we have to fall back on inductive concepts like in (Stive et al., 1990). In practice this means that we might have to adopt some assumptions based on our physical intuition and our expectations of the process behaviour. This approach may be termed *behaviour oriented modelling*; it is intended to overcome the practical difficulties by directly reproducing the qualitative behaviour of some characters of the profile evolution while maintaining a

parametric representation. In practice this approach tries to "implicitly filter" both inputs and processes.

To identify interesting behaviour (cf. summary of Table 1), on the smaller scales we have evidences and data from the real life to rely on. On the contrary, on the longer scales, where there is a scarcity of data, we may rely on a *qualitative observation* of the results of simulations made with short-term process based models.

Observed Process	Qualitative Behaviour
• Profile Displacement in the Long Term	Inclining Profile Looks like a "diffusion" of sediment in the upper part of the profile
• Fate of Sand Supplies and Perturbations	The Spreading Looks like a "diffusion"
• Accreting Profile	Inclining Profile Looks like a "diffusion"
• Berm Formation	of sediment in the upper part of the profile
• Eroding Profile	Declining Profile Looks like a "diffusion"
• Bar Formation	of sediment around the bar
• Bar Movement in the Long Term	Propagating "wave" according to the wave climate

Table 1 - Qualitative Behaviour to be reproduced

DIFFUSION-TYPE FORMULATIONS

Diffusion is considered as the tendency to "smooth gradients"; thus declining may be seen as a local diffusion, if considered as a function of the cross-shore position, inclining as a local diffusion, if considered as a function of depth. In favor of the hypothesis of diffusion there is also the way the profile responds in time (fast initial response and slow settlement). Whenever diffusion is concerned, these observations allow us to apply *diffusion-type formulations* for behaviour oriented models, thus identifying the fundamental parameters as the space-varying coefficients of a rather simple dynamic equation. It is important to note that the choice of the class of diffusion-type model equations is not derived rigorously from any basic process-based model equations, it is selected only because its solution exhibits the proper behaviour for our application. Other formulations could be applied to reproduce other aspects of the behaviour (e.g. the "wave like" motion).

With appropriate initial and boundary conditions the *cross-shore position* can be described as a function of *profile depth* $x(z)$: $\frac{\partial x}{\partial t} = \frac{\partial}{\partial z} \left(D(z) \frac{\partial x}{\partial z} \right) + S(t, x, z)$. The same basic equation may be applied to describe the evolution of the *profile depth* $z(x)$ as a function of the *cross-shore position*: $\frac{\partial z}{\partial t} = \frac{\partial}{\partial x} \left(D(x) \frac{\partial z}{\partial x} \right) + S(t, x, z)$. In particular, the first formulation is an extension of the n-line model with an infinite number of contour lines; a variation of the second one has been used for the evaluation of dune erosion by storms (Kobayashi, 1987). $S(t, x, z)$ is an external source function which depends on time, on the cross-shore distance (x) and on the profile depth (z). $D(z)$ and $D(x)$ are respectively a depth dependent and a cross-shore distance dependent diffusion coefficient. The spatial variation of the diffusion coefficient allows us to represent the variation of morphological timescale with the position, and an asymmetry in the long-term residual sand displacement across the profile. The idea is to have all the information about the typical site climate, the sand characteristics and the degree of activity of the various profile zones summarized into $D(z)$ or $D(x)$.

The calibration of this parameter is the key element of the model definition: all information, on hydraulic and sediment characteristics as well as on shorter-term dynamics is stored in it. Human induced inputs as well as other "natural corrective" terms which would eventually be required, are summarized into $S(t, x, z)$. The final objective of the experience that can be gained in this way is to be able to directly express the parameters that give shape and value of $D(\cdot)$ as functions of mean environmental parameters (wave input and water level variations) and of geometrical characteristics.

For the first formulation we may assume the diffusion $D(z)$ to be at maximum at the top and almost negligible at the lower part of the profile; a peak value is expected in correspondence of the typical breaking depth. This is well in agreement with physical considerations. It can be demonstrated that

the time dependence of the response is of exponential type (in agreement with Kriebel and Dean, 1993, or Kriebel and Dean, 1985). There is even more empiricism in the second formulation as far as $D(x)$ has to be shaped in relation to "where we want" the sediment to diffuse. In this case it is useful to relate the peak of the diffusion with the position where the waves break and distribute the diffusion in the surf zone. Reference shapes, and possible reference formulations, may be defined using the concept of *equilibrium profile* or identifying the *average profile* over the time scale of interest.

The same basic equation may be applied to describe the evolution of the actual cross-shore position referred to an "equilibrium profile position" $x(z)$ - $x_e(z)$, or the actual cross-shore position referred to an "autonomous profile position" $x(z)$ - $x_a(z)$ that well represent the situation that arise when introducing a disturbance along the profile. The latter is the formulation we are currently using to examine the cross-shore spreading of nourishment (Stive et al., 1992; De Vriend et al. 1993b).

With the direct inclusion of transport terms it is of course possible to better reproduce some peculiar aspects of modification of profile shape. Both formulations have in principle the potential to reproduce the various situations of Table 1 at the extent they are simple enough.

A clear practical limitation is represented by a stationary diffusion term (and transport) in the formulation. What we may expect in such a case is a sort of "mean evolution" in the modelled period, governed by the way we conduct the comparison of profiles. The stationarity has also strong implication on the behaviour of the solution and, on ultimate analysis, on the character of the profile evolution that has to be reproduced. This means that all the listed model equation are restricted to reproduce profiles which are evolving in a constant way (e.g. constantly eroding or accreting).

We have to define a beach boundary condition and a seaward boundary condition. Boundary conditions must be defined in relation to the specific formulation chosen and to the specific application. In order to specify the boundary condition we have to rely on a representation at least at conceptual level of what is over the boundaries and what the boundaries represent. Table 2 gives a summary of applicability of the various formulations.

Formulation	Applicability	Beach Boundary	Sea Boundary
$x(z)$	Response to Sea-Level Rise	$x_B' = x_B'(0)$	$x_S = x_S(0)$
$x(z)$ - $x_a(z)$	Nourishment or Sand Extraction (with the support of a Process Based Model)	$x_B' = 0$	$x_S = 0$
$x(z)$ - $x_e(z)$	Nourishment or Sand Extraction (Displacement from equilibrium)	$x_B' = 0$	$x_S = x_S(0)$
$z(x)$	Eroding Profile on the Short scale; Bar Formation	$z_B = z_B(t)$	$z_S' = z_S'(0)$
$x(z)$	Accreting Profile on the Short Scale; Berm Formation	$x_B = x_B(t)$	$x_S = x_S(0)$

Table 2 - Applicability of the various formulations

Onshore we should specify at least a simplified dune erosion scheme to be coupled with the dynamic model in order to reproduce the role of dune as reservoir or supplier of sand; we see here also the possibility to establish a link with coastline models. Seaward the fundamental problem is to define where the boundary should be. For the processes we are considering this boundary is placed on the lower shoreface. A valuable and useful approach to determine the seaward extent (closure depth) was developed by Hallermeier (1991), who defined it as the annual shoreward boundary of his shoal zone. In general a time scale dependence of the seaward extent of the active zone (the closure depth) can be expected. Insight into the time scale dependence of the closure depth and the relative activity across the active zone is of importance for more accurate reproduction and prediction of the behaviour.

In principle it is easy to see that $z(x)$ applies more for the description of the erosion process rather than accretion process. Analogously $x(z)$ applies more in case of berm formation rather than bar formation. These observations suggest the possibility to switch from one formulation to another on the base of a criterion to discriminate between accretion and erosion conditions or on the base of a simple condition for bar formation.

CONCLUSIONS

By application of a detailed process-based, cross-shore morphodynamic model and some inductive assumptions, the characters of shoreface profile evolution are being studied in relation to time scales. The results give qualitative and quantitative indications on how to reproduce such characters in a simple way on the base of the so-called *behaviour-oriented modelling* approach. The behaviour to be reproduced may be based on field evidence and on specific characters inferred from process models. Periods from storm to seasons are interesting as far as they allow us to check the concept against well validated short term process models. However the real longer term objective is to obtain a predictive method to establish the behaviour of shoreface profile on the longer scales.

One objective of our study is to assess whether and to what extent the behaviour oriented approach, and specifically the diffusion model concept, stands in practice, and to find simple and manageable parameterized expressions for the diffusion coefficient as a function of boundary conditions, geometrical features and environmental parameters (in particular wave climate data and water level variations). For situations where few data are available and quick evaluations are required, the behaviour oriented concept will play its role in summarizing and give the maximum added value to the available information.

ACKNOWLEDGMENTS

This work has been carried out as part of the European Community MaST G6 and MaST G8 Coastal Morphodynamics Projects (Directorate General for Science, Research and Development, MaST Contract no. 0035-C and no. MAS2-CT92-0027) and with the support of the Coastal Genesis Research Programme of The Netherlands (Rijkswaterstaat, Tidal Waters Division) and of the Researchers of Excellence Programme for Catalan Universities of the Generalitat de Catalunya.

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LONG-TERM MORPHODYNAMIC EVOLUTION
OF BEACHES AND BARRIERS:
EXAMPLES FROM PARAGLACIAL COASTS

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Beaches and barriers on many high-latitude coasts comprise mixtures of fine and coarse clastic materials forming a distinctive morphodynamic environment (Carter and Orford, 1993). Over relatively long time scales (decades to millennia) these coasts show evidence of self-organization through textural maturation, sand-gravel decoupling, and morphological feedback within an essentially invariant process domain. Factors such as sea-level change and sediment availability act as fundamental controls on the rate of self-organization. Such coastal behaviour suggests the presence of low-dimensional non-linear systems characterized by specific phase configurations (Middleton, 1990). These concepts, which have their origins in the early work of Johnson (1925) and coworkers, have been developed further by Orford et al. (1991) among others.

Simple heuristic models of shoreline behaviour from examples in Europe and North America are used to explore the formation, stabilization and breakdown of such coasts, particularly under the widely pervasive influence of sea-level rise. Long periods of slow evolution are punctuated by short episodes of rapid reorganization (Fig. 1). The slow phases are marked by trends of increasing energy absorption and the gradual reworking, partitioning, and textural sorting of material

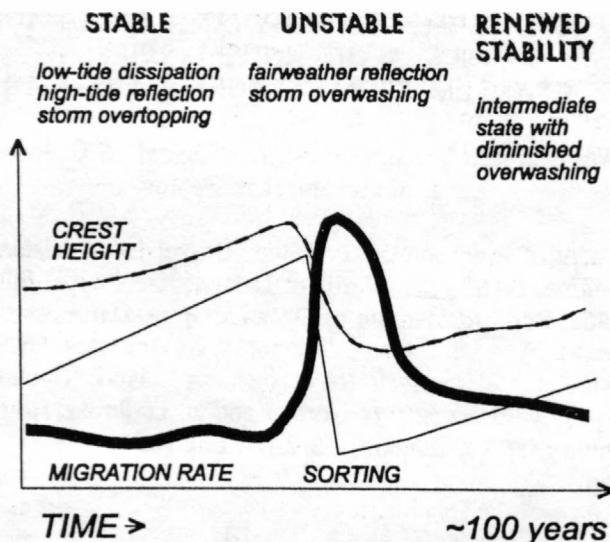


Figure 1: Schematic representation of catastrophic transition from stable to unstable morphodynamic states (attractors I and II of Carter and Orford, 1993) for swash-aligned single-ridge gravel structures.

towards transport minima. Such information-creating (entropy-defying) behaviour results in the effective near-field dissipation of energy and the development of stable beach configurations within which secular input fluctuations are largely absorbed. During intervening episodes of rapid change, the stable coastal structure is broken down, sediment is remixed, and transport is reestablished. These events may be associated with variation in any system input, the dominant factors being sediment supply, wave climate, and rate of sea-level rise. The trigger may be a local sediment deficit or weather event or a wider regional change leading to threshold exceedance in the morphodynamics of the shore system. At the local level, adjacent beaches and barriers may, at any one time, show quite different behaviour depending on the antecedent states of individual coastal cells.

Some of the best examples are provided by drift-aligned systems responding to diminishing (or non-accelerating) sediment supply. Through mechanisms of proximal reworking, alongshore propagation of shoreline perturbations and littoral cell growth, this type of barrier illustrates many of the feedback characteristics inherent in the long-term evolution model. A particularly fine example is found along the north shore of Chedabucto Bay, Nova Scotia, where, in response to both a rapidly rising sea level and a steep alongshore energy gradient, the barrier system has transgressed landward and reorganized texturally over a distance of 40 km. In this case, there is evidence on the inner shelf of a relict barrier system that has been outflanked and oversteepened by rising sea level. Other variations to the basic model exist when and where shoreline bathymetry is deep, so that most of the sediment is stored subaqueously in a barrier platform, as in St. George's Bay, southwest Newfoundland (Shaw and Forbes, 1992); alternatively, where shoreface bathymetry is shallow, rapid progradation can occur with relatively little sediment supply, as at Dungeness on the south coast of England, Derrymore spit in the west of Ireland (Taylor et al., 1986), and a number of sites in southeast Newfoundland (Shaw and Forbes, 1987).

Swash-aligned systems, exemplified by numerous examples along the Atlantic coast of southeastern Canada (Carter et al., 1990; Forbes et al., 1990; Forbes and Syvitski, 1994), respond to diminishing sediment inputs by marginal reworking and landward migration, consuming previously deposited progradational facies. Some systems intermediate between swash- and drift-alignment display complex offset and washover morphologies. Subtle changes in the input parameters affecting sediment supply and/or morphodynamic response can result in large-scale beachface remodelling (Orford and Carter, 1985), renewed progradation, or other major morphological adjustments such as those defining the limits of beach-ridge bundles in prograding systems.

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Abstract

The morphodynamic evolution of a sandy coastal environment encompasses variations on the temporal scale spanning from high frequency turbulent bursts of order seconds to glacial thaws and associated sea-level rise of order millenia. The problems faced by researchers in the field of large scale coastal behaviour (LSCB) involve an assessment of all the appropriate forcings on a morphodynamic system. To ensure that the most important forcings are included, at the most important timescales, filtering techniques are employed. Of particular importance are the possible effects of a short-term change, be it 'man'-made (e.g. dredging of an approach channel) or naturally occurring (e.g. extreme storm), on the long-term response of the system.

Recent advances in computational power have enabled the development of dynamically linked morphological area models which include the major process modules of wave propagation, current distribution and associated sediment transport, to predict the resulting bed evolution (see for example [1]). Such process models already incorporate a certain degree of process filtering, e.g. depth-averaging, to enable practical applications to predictive analyses. The study of LSCB over the time period of years to decades using this type of model requires further input filtering and/or process filtering. This paper describes work carried out at HR Wallingford using the coastal area morphodynamic model, PISCES [2] to investigate the feasibility of tidal schematisation in order to carry out longer term predictions.

Introduction

PISCES is a process-based model designed to investigate the response of a sandy seabed to relatively short-term events, up to a spring-neap cycle. This limitation in timescale is primarily due to the computational effort required to run the model in this integrated mode, in order to simulate the various hydrological and sedimentological processes. To run PISCES for longer simulation periods than this design scale will therefore require some (further) schematisation of either the inputs or the processes, or both.

The architecture of PISCES allows the calculation of the tidal residual sediment flux field to drive the morphodynamic module, effectively removing the intra-tidal bed updating procedure. This approach gives rise to a significant reduction in computational overheads, thus allowing longer simulation periods. However, the applicability of such a technique is limited to specific morphological scenarios.

Consider a coastal morphodynamic system where the dominant sediment transport processes are due to tidal currents. If the timescale for bed level changes is long compared with the tidal period it is sufficient to model the bed evolution by calculating tidal residual fluxes over a static bed, and updating the model bathymetry at the end of the tide. Furthermore, if the timescale for bed level changes is short compared to a typical spring-neap cycle estimates of residual sediment fluxes for spring and neap tides can be used to drive the morphological module, linearly interpolating between these two fields. This method has another advantage of allowing the determination of the bed evolution through a spring-neap cycle without having to model the intermediate tides. The use of sediment flux residuals rather than fluid flux residuals ensures accurate representation of the processes associated with entrainment from the bed.

Methodology

The long-term evolution method described is defined as follows:

1. Time $t=0$ is defined arbitrarily as HW springs
2. For the initial bed configuration spring (S) and neap (N) tidal residual sediment fluxes are determined
3. Predictor stage. Linear interpolation between the S and N fields yields the estimated new bed after 14 tides
4. Based on this new bathymetry the N tide residuals are recalculated
5. Corrector stage. Re-starting from the initial bed interpolation between the original S and new N tide yields the corrected bed after 14 tides
6. Recalculate the new fields for this bed (in this case the neap tide and the predictor spring tide) and continue.

By this means the bed evolution over 14 tides is represented by only three hydrodynamic tidal calculations, giving rise to a significant saving in computational effort.

To demonstrate the approach, PISCES was applied to a specific application of a spoil dispersal test, using model geometry based on a hypothetical test case of a river mouth [1], amended for tidal currents only. The initial bathymetry is presented in Figure 1, showing the large spoil heap with a peak height of 5m, approximately 600m offshore in the area of stronger tidal streams that run parallel to the coast. Nominal spring and neap tides of range 2m and 1m respectively were specified, giving rise to peak currents of 80cm/s and 50cm/s.

Figure 2 shows the updated bathymetry after 28 days, which required 12 full-tide calculations. Cross-sections through the spoil heap along the main flow streamline are presented in Figure 3.

Conclusions

The method of tidal schematisation described provides a relatively accurate means of applying PISCES to longer term simulations, *for specific scenarios*. Due to the modular architecture of PISCES the approach is automatic, and relatively quick, requiring 2 days elapsed time for the 1 month simulation presented. As a result longer simulations of up to a year are possible.

The method assumes a continuous adjustment in the flow field between spring and neap tides and appreciable errors would be introduced if the current distribution for a specific application varies greatly between the spring and neap tides (eg gyre formation). De Vriend et al [3] describes a further schematisation resulting in the definition of the optimum morphological tide for scenarios where the current distribution is similar between the springs and neaps. Other errors are introduced if, for example, at mean tide and lower range tides the tidal currents are insufficient to entrain sediment from the bed. However this latter error is less significant if the effects of waves are included.

Regarding the effects of waves, further work is being carried out to investigate the efficacy of combining this tidal schematisation with a wave filtering technique, such as the single representative wave method [3], amended for monthly filtering rather than annual.

Acknowledgements

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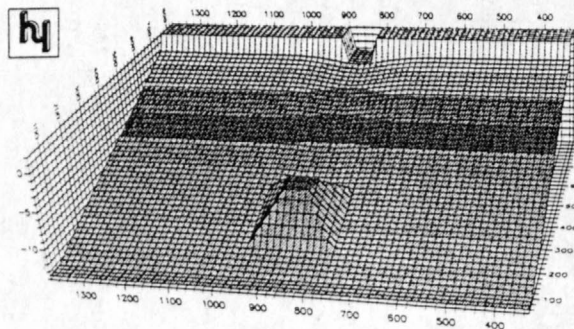


Figure 1 Initial bathymetry

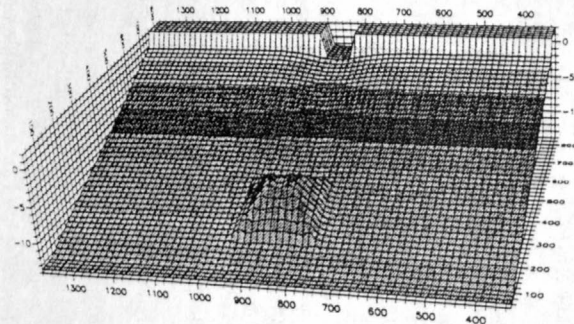


Figure 2 Updated bathymetry after 28 days

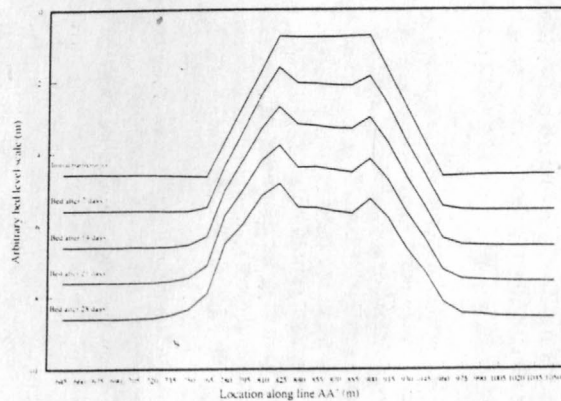
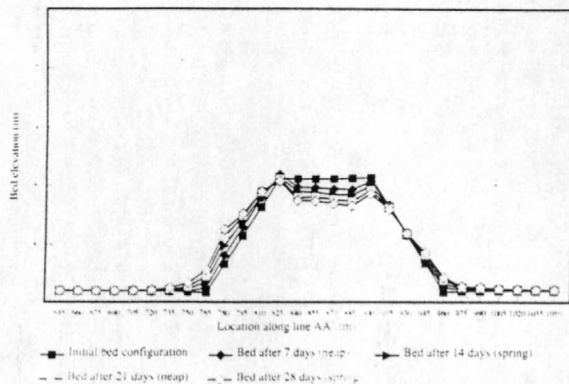


Figure 3 Bed level cross-sections over simulation period

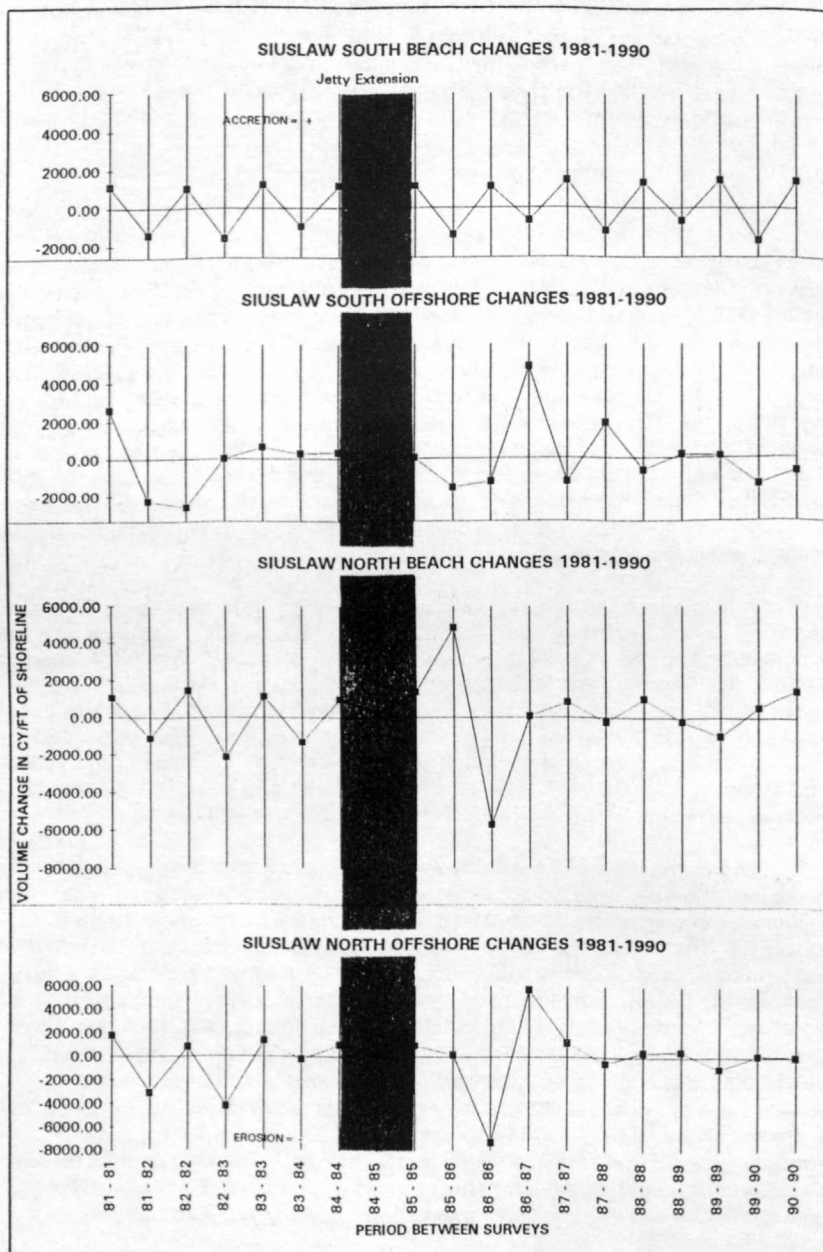
SEASONAL EROSION/ACCRETION CYCLES IN A LITTORAL CELL

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From 1981 through 1990 semiannual beach and nearshore surveys were taken along 18 kilometers of shoreline near the entrance to the Siuslaw River, Oregon. Changes in 82 profiles from the dune crest out to a depth of at least -5.5 meters MLLW were analyzed. The study area is at the northern end of the largest littoral cell along the Oregon coast, stretching almost 80 kilometers. For the entire length of this littoral cell the shoreline is backed by extensive dune fields. Two major rivers, the Umpqua and Siuslaw, and a large estuary, Coos Bay, occur within this littoral cell. The jetties at the Siuslaw River are within 13 kilometers of the northern cell boundary at Heceta Head. It can be argued that the area north of the jetties is a subcell of the larger littoral cell. During the study, jetties at the mouth of the Siuslaw River were extended over one thousand meters seaward. This had a pronounced effect on the seasonal erosion/accretion cycle to the north, but did not seem to affect the area to the south similarly.

A permanent survey baseline was established with profiles every 76 meters apart for the first 760 meters on either side of the Siuslaw River entrance and 305 meters apart for the remaining distance. The baseline extended 8.2 kilometers north of the Siuslaw River entrance and 9.7 kilometers to the south. The area above MLLW was surveyed by standard land survey techniques. The remainder of each profile was surveyed using a helicopter survey technique developed by Portland District. Horizontal control was on the order of 0.3 meters, and vertical control varied from 0.03-0.1 meters. Surveys were done within a two week time frame in March and September of each year with a few exceptions.

The main purpose of the study was to account for beach changes after jetty extension. Surveys were scheduled as early in spring as feasible in order to document winter geomorphic conditions and in early fall to document summer conditions. The study began in May 1981. Jetty extension began in summer 1984 and was completed by September 1985. The final survey was in September 1990. The following figures compare profile changes between surveys north and south of the jetties. The survey periods are indicated as 1981-1981, 1981-1982, etc, which correspond to Summer 1981, Winter 1982, etc. Beach volume changes include the cumulative change for all measured profiles above MLLW between the indicated survey periods. Offshore volume changes are the cumulative change for all profiles above MLLW. Positive values occur fairly regularly for the summer reflecting accretion between the spring and fall surveys. The change from fall to the following spring is commonly negative indicating erosion. This pattern is less regular offshore and there appears to be a large perturbation following jetty completion in 1985.



The south beach appears relatively uninfluenced by jetty construction, illustrating the classic seasonal erosion/accretion cycle throughout the study period. The north beach begins classically, but by the winter of 1984, after jetty construction begins, instead of erosion there is accretion. This increases through the winter following jetty completion in 1985. By the summer of 1986 there is substantial erosion in both the beach and offshore areas north of the jetties. The winter of 1986 is again anomalous in that the entire offshore area shows a large net accretion. These effects appear to be dampening out by the end of the study period in 1990, although the classic seasonal pattern has not been reestablished.

These and other analyses suggest differences between the areas north and south of the jetties and between the beach and offshore areas. Other analyses show a striking difference in alongshore volume changes north and south of the jetties. The jetties may act like an artificial headland providing more sheltering to the north from large winter waves predominantly from the southwest. During summer wave conditions which tend to move littoral sediment southward, more sediment would be trapped north of the jetties. This is substantiated by the finding that there was net accretion throughout the area north of the jetties between 1981 and 1990.

If the beach south of the jetties represents the natural seasonal erosion/accretion cycle, and the beach north of the jetties shows how this cycle was interrupted by jetty construction, why are the offshore areas so irregular? Perhaps the simplest explanation is that the measured profiles do not extend seaward far enough to contain the offshore boundary of cross shore transport. Examination of the offshore bathymetry showed that the offshore bar was not always present (Carstarphen, 1991). The assumption is that material was carried offshore beyond the extent of the measured profiles and did not always move back onshore by the next measurement. There is also simply a matter of scale and accuracy. The offshore portion of the profile is commonly two to three times as long as the beach portion and the offshore depth accuracy is approximately 1/3.

Some conclusions from the study include:

- 1) The littoral cell south of the jetties represents a relatively natural system, unaffected by jetty extension.
- 2) The portion of the littoral cell north of the jetties acts independently of the larger littoral cell.
- 3) Extension of the jetties temporarily affected the northern portion of the littoral cell.
- 4) The seasonal erosion/accretion cycle is less well defined offshore and the linkage with the beach is complex.

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NATURAL AND ANTHROPOGENIC INLET-INDUCED SHORELINE CHANGE IN SOUTHEASTERN NORTH CAROLINA

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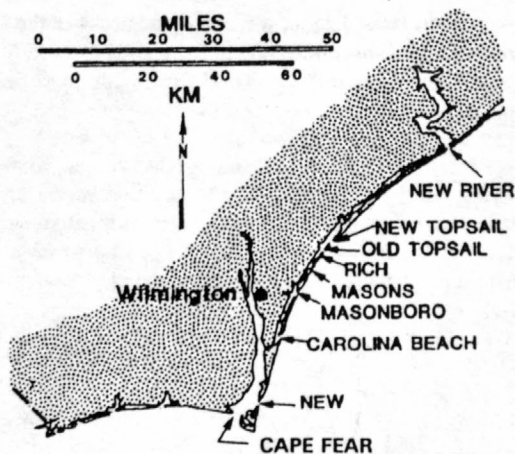
The low mesotidal setting in southwest Onslow Bay from New River Inlet to Cape Fear and the extreme northeast sector of Long Bay in the vicinity of the Cape Fear River is characterized by eight transgressive barriers ranging from 2 to 40 km long (Fig. 1). Five of these barrier islands have shorefront development. Inlets which separate these barriers are a diverse group of wave-dominated features. Five migrating and four stable inlet systems have been identified. Three of the stable inlets have been dredged extensively or stabilized by structures. With the exception of one inlet, Rich Inlet, all migrating inlets have been modified to some extent for navigation purposes. In this 120 km coastal sector, nearly all natural and human-induced shoreline changes identified as critical erosion zones are associated with inlets.

Figure 1. Inlet study area showing the location of the major inlets located along the North Carolina coast in southwestern Onslow Bay and the northeast end of Long Bay.

The data base for this study consisted of approximately 50 sets of aerial photographs dating from 1938 to 1990, historic maps and current and historic bathymetric charts. Analysis of these data sets indicate the following generalizations apply to the inlets in this coastal cell.

1) Erosion and accretion cycles along stable inlet shorelines were related to cyclical changes in the symmetry of ebb deltas. The cycles were associated with repositioning of the main ebb channel and corresponding position changes in the flood channels and where bars attached to the adjacent barrier islands. Cycles ranged in duration from five to 20 years and cycle length correlated with inlet size and storm history. Cycles were shortened by storms; cycles were typically longer in larger inlets.

Southeastern North Carolina Inlets



The process of channel extension and abandonment provides a mechanism where sand packets of varying size (range: 5×10^3 to 4.0×10^6 m³) are bypassed downdrift to the adjacent barrier island.

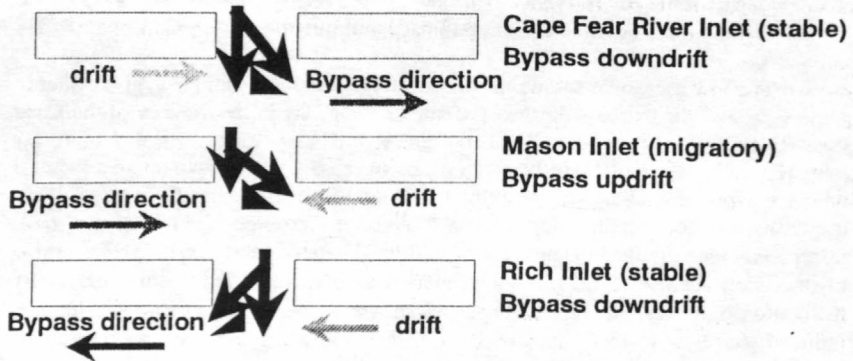


Figure 2. Relationship of inlet location and stability, local drift direction, and direction of shoreward movement by bypassed bar packets following ebb delta breaching.

Bald Head Island, the barrier adjacent to the Cape Fear River Inlet accreted southwestward more than 800 m from 1885 to 1915 following an ebb delta breaching event. This scenario pre-dated the major modifications of the harbor entrance channel.

2) Unstable inlets migrated to the southwest at average annual rates varying from 5.0 to 40.0 m. In addition to the erosion of the downdrift barrier island, migration resulted in truncation and realignment of the trailing shoreline (Fig. 3). Rates of shoreline recession ranged up to 12 m/year for as long as a decade. Erosion rates decreased as the updrift barrier planform adjusted to the position of the inlet.

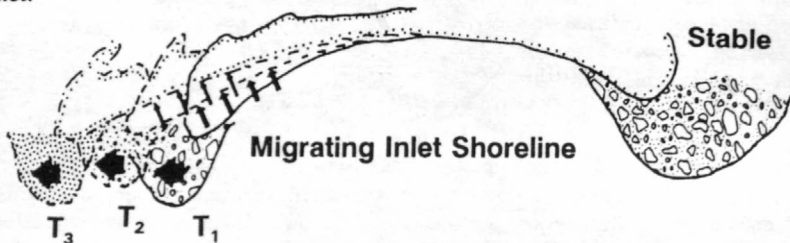
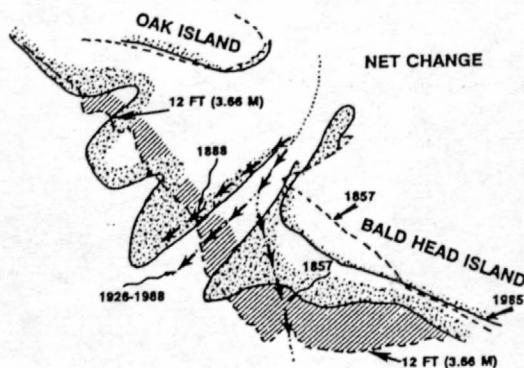


Figure 3. Trailing barrier realignment. Figure illustrates updrift erosion is a consequence of inlet migration and planform adjustments.

Concurrent with the process of migration, packets of sand were bypassed to the updrift shoulder of the inlets when ebb deltas were breached. The cycle of ebb delta breaching varied in length from 2 to 20 years. Deflection of the updrift channel and simultaneous changes in the symmetry of the ebb delta caused migration and attachment of small bar complexes (100 m x 50 m) within one km updrift of the main channel. Frequently, these temporary shoreline convolutions promoted rapid erosion leeward of where the sand packets attached. The sand packets frequently prompt periods of relatively rapid inlet migration as the bars move laterally along the barrier spit and into the adjacent estuary.

3. Modification of inlets by dredging resulted in an increase in tidal prism, a correspondingly larger retention capacity of the ebb delta and disruption of the natural ebb delta breaching cycle. Dredging changes are reflected in an extension and deepening of the ebb channel across the ebb platform. Bypassing to the downdrift shoals ceases when the ebb channel is bifurcated during channel maintenance. Shoreline erosion occurs when the sand supply to the shoreline is reduced by inlet dredging. Typically, the maximum rate of erosion will lag the breaching of the shoals by several decades, especially in the larger inlets such as the Cape Fear River Inlet. Inlet stabilization by jetties mimicked the impacts of dredging. The modification of inlets by hard structures has reduced sand bypassing and increased erosion on one or both sides of the inlet. Significant long term coastwise erosion associated with inlet modifications is exemplified by Masonboro Island, a 12 km long barrier island. Located between a jettied system and a continually dredged artificial inlet, Masonboro Island has experienced a major reduction in sand supply and washover topography has increased along 80% of the island's length.

Figure. 4. Shoreline and shoal changes, Cape Fear River Inlet, 1857-1985. Ebb channel deflection resulted in erosion of Bald Head Island. A fixed shore-normal channel orientation and shoal segmentation has drastically reduced the downdrift sand supply.



Despite increased knowledge concerning the natural process of change associated with inlets, management of these systems is difficult. The magnitude of change caused by either natural or man-induced processes is impossible to predict with sufficient accuracy to assist coastal managers in making sound decisions; each

inlet must be studied in detail to ascertain the specific magnitude and direction of change and its effect on human occupation and commerce. Planning is enhanced by knowledge of the basic characteristics of inlets and the likelihood of change. We conclude that the decade-to-century scale coastwise sand budget and ultimately the shoreline retreat rate are negatively affected by the increased retention capacity of modified inlets. Because of the large numbers of inlets and the storm frequency in southeastern North Carolina, sand loss to the adjacent shoreface is likely to be much higher than in regions where natural systems occur.

**Simulation Modelling of LSCB:
Parametric Scaling, Markovian Evolution and Site Idiosyncrasies.**

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Large-scale coastal behaviour (LSCB) entails geomorphological change that is state dependent and subject to stochastic boundary conditions. The uncertainty this causes in coastal evolution is compounded by an incomplete understanding of the morphodynamic processes involved, especially under highly idiosyncratic conditions. The most common and reliable approach to LSCB modeling is based upon sediment-mass budget considerations. The process uncertainties can be sidestepped to some extent by adopting a parametric representation of processes in terms of sediment-mass geometries. A parametric formulation allows LSCB to be modelled by incorporating relationships based upon fully deterministic theory as well as heuristic knowledge drawn from the distinct qualitative patterns that exist in nature. It also allows introduction of a greater level of complexity, such as inclusion of fine-grained estuarine sedimentation, whilst maintaining a fairly simple model. Such a model can simulate the effects of storm erosion and recovery, sea-level change, and variations in coastal sediment budgets.

Investigation of coastal evolution at geological time scales (order 10^3 years) illustrates how underlying trends in LSCB are likely to be masked by fluctuating modes of LSCB. For example, computer experiments simulating coastal change, at an open coast site in southeastern Australia during the post-glacial marine transgression, show that a deficit of $75 \text{ m}^3 \text{ yr}^{-1}$ in the alongshore transport budget was required to reduce the elevation of interstadial barriers by the amount indicated from seismic and C^{14} data (Fig. 1, Inset C). This deficit, which is only 0.04% of the estimated total net alongshore flux, is unlikely to be detectable through either field measurements or based upon existing transport relationships. As sea level neared its present elevation, the landward-migrating barrier in Figure 1 encountered a more embayed coastline. The sediment-budget was transformed to one of surplus during the stillstand, resulting in 6,000 years of shoreline progradation (Fig. 1, Inset B). The progradation occurred at an average rate of 0.3 m yr^{-1} , which again is undetectable, given that interannual and interdecadal fluctuations of the shoreline position are typically in the order of 10^2 m . It seems likely therefore, that underlying trends in LSCB are only resolvable through modeling. Such

trends are related primarily to sea-level changes and sediment budgets that persist over geological time scales. Under these conditions, processes involve constant and smoothly-changing sand-body dimensions that are associated with systematic variations in boundary conditions.

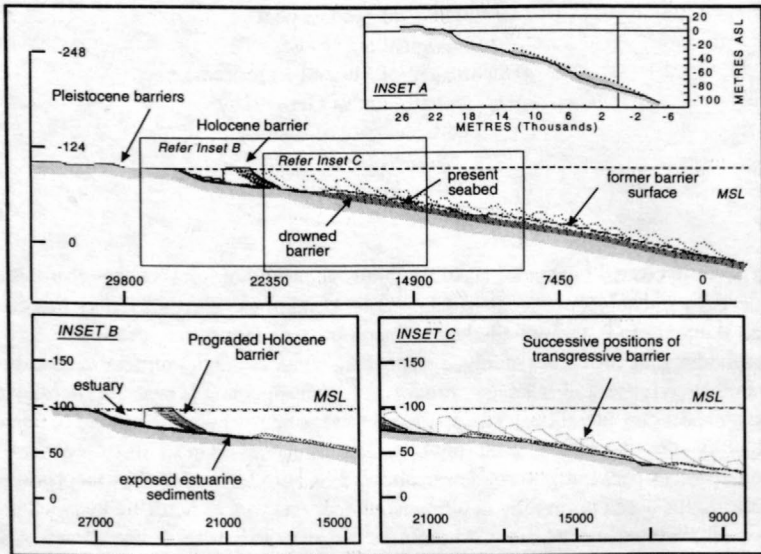


Figure 1. Computer simulation of coastal evolution during and following the post-glacial marine transgression on the Tuncurry shelf, 220 km north of Sydney, Australia. The initial substrate (Inset A) features two interstadial barriers, the surfaces of which were lowered by 3 m, based upon chronostratigraphic data. Sea level was rising at 15 mm yr^{-1} (modelled in 1.0 m increments) when this lowering took place (Inset C). Barrier progradation ensued following the onset of the stillstand (Inset B).

However, at historical and engineering time scales (order 10^1 to 10^2 years), stochastic fluctuations in boundary conditions, involving storms and short-term sea level changes, can be expected to cause irregular variations in dimensions of sand bodies. Morphodynamic evolution at these time scales is strongly Markovian and seems amenable to quantitative analysis only on a probabilistic basis. Unfortunately, at these time scales, little is understood at present about environmental fluctuations. In southeastern Australia, these fluctuations are related to such phenomena as the El Nino Southern Oscillation (ENSO), which has interannual cycles, and other climatic processes, as yet unidentified, that have strong interdecadal signals to which coastal sand bodies are especially responsive. Consequently, it is difficult to predict the effect that anthropogenic modifications have upon LSCB since the background fluctuations

at these time scales are so strong, and the likelihood of sensitive dependence upon initial conditions, due to Markovian inheritance, is so great. A compounding effect of state dependence is the long adjustment time (order 10^3 years) of macroscopic geomorphological behaviour, evident in the simulation modeling (Fig. 2). The long adjustment time exists even when instantaneous morphodynamic response is assumed. For example, the initial shoreline position is the same for both cases in Figure 2. The macroscopic disequilibrium in Figure 2B results in a markedly greater shoreline recession.

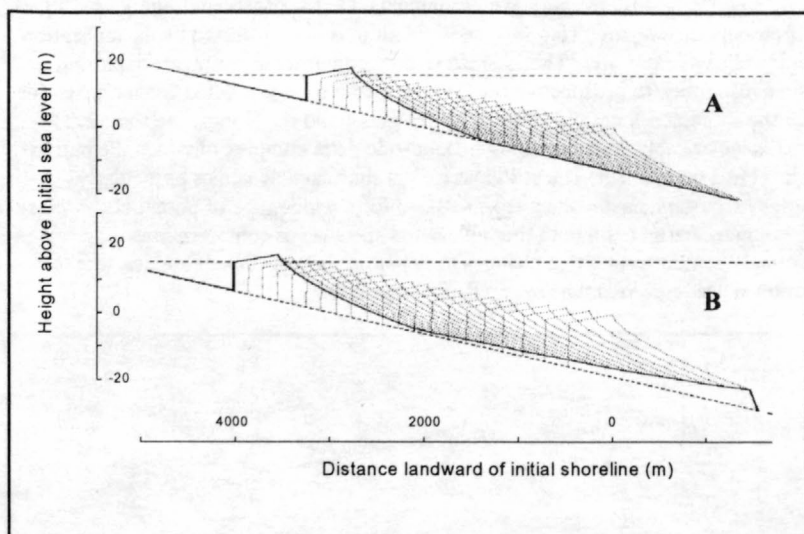


Figure 2. Macroscopic adjustment by a barrier to rising sea level (1 metre in 20 steps - the results are independent of the sea level increment). A) Constant-form migration of the barrier occurs when initial conditions correspond to those which give an equilibrium translation rate. B) When initial conditions differ from those of macroscopic equilibrium, the translation rate gradually approaches the uniform rate in A.

State sensitivity in LSCB (illustrated in Figure 2) derives from the dynamically-related dimensions of coastal morphologies. These dimensions also vary strongly through time as boundary conditions fluctuate. Variations in morphological parameters exert their importance through changes in temporary sediment storage within the coastal sand body itself. Scale relationships between these dimensions are explored through simulation to simplify development of probabilistic methods for predicting LSCB. Important dimensions in this respect are the width of the barrier, and depth and shape of the shoreface. Unfortunately, with possible exception of the shoreface (Hallermeier)

depth, little is known about these fundamentally-important dimensions and the processes controlling them. Furthermore, the prospects for deriving dimensional relationships from process modelling do not seem encouraging in the face of the complexity in morphodynamic processes underlying LSCB. The problem is compounded by real-world complications, such as biotic effects, which usually are excluded from morphodynamic models.

Process-based, forward modelling therefore seems of limited viability in prediction of LSCB, even for fairly conventional environments. On the other hand, the versatility of parametric modelling, involving inverse simulation, is demonstrated by its application to highly idiosyncratic sites. This is illustrated by computer experiments exploring the possible responses, to prospective sea-level rise, of a semi-protected barrier beach for which the shoreface is heavily affected by sea grass, and the alongshore sediment budget is inextricably linked to a larger flood-tide delta complex of which the barrier beach is but a component. The sea grass makes shoreface dynamics particularly uncertain. Coastal adjustments were simulated for a wide range of potential shoreface morphologies, based upon both theoretical and speculative considerations. Figure 3 is an example demonstrating that changes to the alongshore sediment budget, due to variation in sea level, overwhelm any Bruun-type effects.

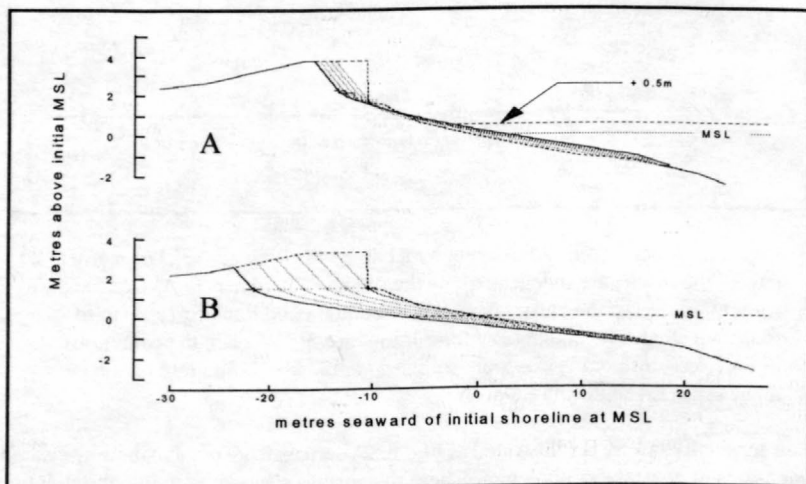


Figure 3 Modelled shoreline response to: a) a 0.5 m rise in sea level (decade increments - 50 year scenario); and b) a deficit in the alongshore transport budget at 0.5% of total transport over a five-year period (yearly increments). Site: Mackerel Beach, Broken Bay, located 35 km north of Sydney, Australia.

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THE ROLE OF
MATHEMATICAL MODELLING IN
LARGE-SCALE COASTAL MORPHOLOGY

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Since the KNAW-Colloquium on LSCB in Amsterdam, 1989 (see De Vriend, 1991), much work has been done on mathematical modelling in this field. Existing concepts have been elaborated and new ones have been developed and tested in real-life applications (for instance, see De Vriend et al., 1993). An updated state-of-the-art review is therefore expedient.

To that end, the various model concepts, their potentials and their limitations are considered from an LSCB point of view. Pilot applications are discussed for a range of models, such as process-based diagnostic models, medium-term simulation models, linear and non-linear Fourier-mode evolution models, formally averaged models and semi-empirical behaviour models. The role of analysis, reduction and assimilation of data is considered, and possibilities to handle uncertainty are indicated. Finally, needs for further research and data are identified and research strategies are discussed.

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LARGE SCALE RESPONSE OF THE NILE RIVER DELTA SYSTEM

by

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INTRODUCTION

The Nile River Delta system extends from the west near Alexandria to Rafah on the east, a distance of some 450 km. From approximately 500 BC to 1100 AD, the Nile is reported to have had up to ten outlets distributed such that the sediment supply caused a large shoreline advancement near the center of the Nile Delta system. In approximately 1100 AD, the number of outlets was reduced to two, Rosetta on the west and Damietta on the east. In its natural state, the Nile is estimated to have delivered approximately 70 million m³/year of sediment to the coastal system. Although most of this was in the fine size range, it is believed that approximately 20 million m³/year of this was between 0.02 mm and 0.2 mm in size. The coarser size fraction resulted in significant promontories at the mouths of the two outlets. The earliest dams were constructed on the Nile in 1902; these undoubtedly caused some reduction in the quantities of the coarser sediment reaching the coastline. However, the sediment supply was essentially eliminated with construction of the High Aswan Dam in 1964. The dams have captured much of the coarser sediment and have reduced the peak flood flows required to transport the sediment. This paper examines, through numerical modelling, the Delta system characteristics and provides a basis for extrapolating its evolution into the future.

NUMERICAL MODEL

A one-line numerical model is applied which incorporates the equations of continuity and transport. Required parameters in the modelling are the effective wave height and direction, the effective depth of sand motion and the volumes of beach sized sediment originally transported to the coast by the various outlets.

The Pelnard Considered equation is the linearized and combined form of the equations of transport and continuity

$$\frac{\partial y}{\partial t} = G \frac{\partial^2 y}{\partial x^2}$$

in which the nominal orientations of x and y are parallel and perpendicular to the shoreline, respectively. The term G is the so-

called "longshore diffusivity" and represents the tendency for a shoreline perturbation to spread out and embodies the effective wave height, H , the depth of effective sand motion, $h_s + B$, and the longshore sediment transport coefficient, K . Since the planform geometry of the Delta and the shoreline change rates, $\frac{\partial y}{\partial t}$ are known it is possible to develop an estimate of the longshore diffusivity value of $0.0032 \text{ m}^2/\text{s}$, as shown in Figure 1. A depth of effective motion of 10 m and a transport coefficient, K , of 0.2 (based on sediment size) yield an effective breaking wave height of 0.4 m, in reasonable accord with measurements.

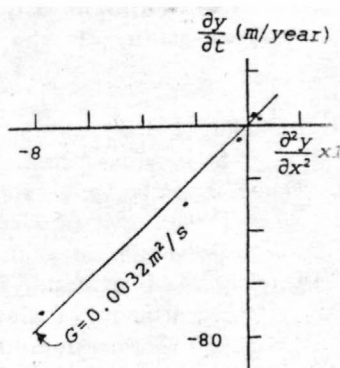


Figure 1. Determination of the Longshore Diffusivity, G .

The estimated supply rates of beach-sized sediment by each of the two outlets were estimated from the following, which is derivable from the linearized form of the transport equation

$$Q_R = 2 \frac{G(h_s + B)}{\tan \frac{\delta}{2}}$$

in which $\frac{\delta}{2}$ is the half interior angle of the promontory. Based on the above equation, the estimated supply rates of the two branches prior to dam construction are:

$$(Q_R)_{\text{Rosetta}} = 5.3 \text{ million m}^3/\text{year}$$

$$(Q_R)_{\text{Damiatta}} = 3.2 \text{ million m}^3/\text{year}$$

With the information representing the system determined as described above, a numerical model was developed and exercised using the full nonlinear transport equation. The model extended from Alexandria to Rafah and was "exercised" for the period from 500 BC to 2000 AD. The changes in the Nile tributaries as described in the Introduction and the reduction in sediment supply due to dam construction were represented. The model predicts, as is found in nature, that the erosion is very extreme in the vicinity of the Rosetta and Damiatta promontories and rapid, but less extreme in the vicinity of the convex outward portion of the Delta which was formed by the ancient tributary system. Zones of moderate accretion are located adjacent to the zones of rapid promontory erosion. The model provides a basis for extrapolation and planning for future shoreline evolution.

A Sediment Budget for Saco Bay, Maine, and an Evaluation of the Long- and Short-term Geologic and Oceanographic Processes

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Saco Bay is a 12 km wide rock-framed, arcuate sandy embayment in southwestern coastal Maine. The areal extent of the bay is 75 km² with a relatively open geometry, a barrier beach system, mainland sandy beaches, and numerous rocky headlands and islands. Its barrier beach system is the longest in Maine and extends from the Saco River (100 m³/sec discharge) in the south to the Scarborough River (3 m³/sec discharge) in the north, which protects the most extensive salt marsh in the region.

The barrier system supports dense coastal development and has provided the setting for environmental disputes about the siting of high-rise buildings, marinas, seawalls, ocean outfalls, and dredging projects. To assist coastal planners, we have evaluated the sediment budget and geological response of the barrier system to long- and short-term sea-level rise, mapped the shoreface, estuary, and dunes, measured sand volumes in depositional reservoirs, and begun to evaluate the rates of exchange between several of the reservoirs. We have integrated numerous data sets in this study including side-scan sonar images, seismic reflection profiles, ground penetrating radar profiles, vibracores, grab samples, current meter measurements, hydrographic profiles, beach profiles and historical aerial photographs. The primary goal of this study is to identify both the sand volume of depositional reservoirs as well as the exchange rate between them to construct a sediment budget.

Saco Bay was deglaciated about 14.0 ka. The sea, in contact with retreating ice, invaded inland beyond the present coast and glaciomarine mud was deposited in a blanket across much of the region. Tens of meters of glaciomarine mud overlie till and bedrock in most of Saco Bay. Between 14.0 and 10.5 ka relative

sea level fell to below present as isostatic rebound outpaced eustatic sea-level rise. As sea level fell, fluvial sand was deposited over glaciomarine mud on land, and possibly offshore. We believe that the Saco River, which deeply incises sandy, glaciated terrain, delivered a large volume of sand to the sea at least until 10.5 ka, the time of lowstand. A sandy shoreline complex at 50 - 60 m present water depth marks the lowstand position of the sea. During the early Holocene, sea-level rise was rapid (up to 30 mm/yr) and only patchy sand deposits mark this transgression between 50 and 25 m water depth. Sea-level rise slowed gradually since 5 ka to less than 1 mm/yr. In the past 2 ka sea level has risen only 1 meter. Tide gauges, however, record a more rapid, 2 mm/yr rise within the past 60 years, suggesting a recent acceleration in the rate of sea-level rise.

During the past few millenia of slow sea-level change, the present dune system reached an average width of 200 m and the back-barrier salt marsh accumulated up to 6 m of peat over tidal flat deposits. At the Saco River mouth and other small inlets, beach ridges underlie salt marshes and connect to the core of the present barrier. These relationships provide evidence of seaward progradation from simple spits and an overall widening of the barrier, probably related to the relative stability in sea level in the last few thousand years. In addition, the geomorphology demonstrates net northward littoral sand transport.

Beach and dune sand volumes, investigated with ground-penetrating radar and augering, contain an estimated 26 million m³, although most dunes are covered by development. The barrier ranges from 150 - 500 m wide and 2 - 13 m thick. For most of its length, the shoreline position has varied less than 10 m during several decades of profiling and aerial photography.

The shorelines showing the greatest variability in time are those areas adjacent to tidal inlets. Net accretion occurs adjacent to the Scarborough River inlet and varies from 0.4 to 3.8 m/yr (1914 - 1993). Net shoreline erosion has occurred at Camp Ellis beach adjacent to the 1.3 km Saco River jetty. At this location erosion rates have averaged 0.4 to 0.6 m/yr since 1953. Recent Camp Ellis beach erosion is notable because the Saco River, which has 30 times the flow of the Scarborough River, must have been the ultimate source of sand to the bay's beaches, dunes and marsh in the past.

The Saco River carries sand through base-draining turbine races in its dams. Just below the final dam the channel contains

poorly sorted, coarse sand and granules. The lower Saco River is choked with fine sand that extends, in the form of bars, offshore of the river-mouth jetties. Side scan sonar images record ebb-oriented bedforms throughout the estuary.

Although for much of the year the lower Saco is a vertically stratified to partially mixed estuary, ebb dominance has also been documented by hydrographic surveys. The upper estuary is dominated by seaward flow with stronger average and bottom ebb current velocities (3 - 63 cm/sec). The lower estuary is also ebb dominant but stronger flood currents (4 - 16 cm/s) reflect the influence of the saltwater wedge under normal flow.

Significant downstream sand transport may occur during spring freshets. In the April freshet of 1993 (6 times the strength of that in 1992) the estuary experienced seaward discharge during the entire tidal cycle, peaking late in the ebb cycle, and replacing the saltwater tidal prism.

On the basis of the new hydrodynamic measurements, inference from dune and spit geomorphology, and the absence of sandy glacial sources offshore, we believe the Saco River has been the ultimate source of most sand to the bay's beaches. Dredging of the Saco River mouth for harbor maintenance has transferred an estimated 1 million m^3 to the beach as nourishment in this century. If this dredged sand represents the river's bedload discharge, the river could have transported greater than half of the sand in the barrier system in the past 5 ka.

A seafloor sediment map constructed from side scan sonar images and grab samples, reveals that significant amounts of well-sorted sand occur in water depths less than 30 m. This shoreface sand apron is also found in numerous seismic reflection profiles as a landward-thickening wedge contiguous with the coastal barrier. Several offshore vibracores demonstrate that less than a meter of sand overlies glaciomarine sediments seaward of the toe of this wedge. A strong seismic reflector is correlated with the unconformity above the glaciomarine sediment. This unconformity is overlain by up to 8 m of backbarrier, barrier and shoreface sediments near the modern shoreline, but less than 1 m of total Holocene sediment in 15 m water depth. The unconformity is relatively smooth, but locally, channels up to 5 m deep and 100 m broad are interpreted as inlet scars. Larger, filled paleovalleys of the ancestral Scarborough and Saco Rivers are incised greater than 10 m into glaciomarine sediments offshore. On the basis of an offshore isopach map of sediments above the unconformity, approximately 65 million m^3 of sand

occurs in the shoreface. This portion of the barrier complex is the largest reservoir of sand in Saco Bay.

Hydrodynamic measurements during fair weather and storms were used to estimate short-term sand transport potential. Currents were sampled and averaged to produce 30 minute vectors and burst-mode 1 second vectors to analyze contributions of wind-driven flow, tidal currents, and wave-orbital motion. These data were compared to wave and wind data from the Portland Large Navigation Buoy. Combined flows during "northeasters" and Hurricane Bob (August 1991) reached 30 - 40 cm/sec. Flow directions are governed by wind stress and tidal stage. Both onshore and offshore flows, capable of moving sand in water deeper than 30 m, have been measured. Wave orbital measurements are consistent with 1 - 2 m wavelength ripples in coarse sand and gravel imaged by sonar on the seabed. The shoreface sand wedge, including its seaward edge, is active many times each year. For a given storm event, the balance between offshore-directed downwelling flow and onshore-directed upwelling flow determines the net sand transport. The cumulative effect over a year (and longer periods) of transport events leads to volume changes in the shoreface sand wedge and, presumably, a shoreline response. Beach accretion observed following Hurricane Bob is consistent with net onshore flow measured seaward of the beach. The balance between downwelling and upwelling over longer time scales may determine how the coastal barrier system responds to sea level.

We have quantified volumes of sediment in various compartments of the Saco Bay barrier system and hypothesize that sand sources are primarily the Saco River, and secondarily, offshore glacial deposits, early Holocene fluvial sands, and lowstand beach deposits. We have begun to understand oceanographic and estuarine processes that contribute to the sediment budget, but need to establish rates of transport among coastal environments. While analysis of the short-term rate of sand transport between the river, beach and shoreface is just beginning, there appears to be a significant involvement in the sediment budget from human actions, particularly in dredging and beach nourishment. How recent human activity and sea-level rise will affect the long-term behavior of the coastal barrier system can only be determined with a complete understanding of rates of geologic and oceanographic processes.

THE SEDIMENT BUDGET ON RUGGED COASTS: SPECIAL CONSIDERATIONS

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INTRODUCTION. The west edge of the United States is a collision coast characterized by a narrow continental shelf, irregular shoreline, and narrow beaches. Many of the beaches are backed by seacliffs. Most are slowly retreating. Along the more rugged reaches of this coast, the prism-shaped littoral sediment lens with the beach as the upper, landwardmost part, typically rests atop a resistant shore platform (Fig. 1). The platform typically joins the seacliff at a sharp angle near mean sea level. The volume of the littoral sediment lens is stable under equilibrium conditions, which requires a balanced sediment budget. A change in the volume of sand and coarser material in the lens results when the sum of all entering and exiting material fluxes is not zero. Reversible movements of sediment, such as result from seasonal climatic changes in the Pacific Basin, affect the distribution of sediment within the lens, but do not affect the volume of the lens. Reversible onshore and offshore movements of material, however, define the dynamic equilibrium surface profile and to some extent the basal profile of the lens as well as its landward (backbeach line) and seaward (shorebase) limits. A change in the budget of sediment in the lens produces a change in the distance between the shoreline and backbeach line, i.e., mean beach width.

The ability to predict and manage the mean width of the beach is of special concern to public officials and coastal property owners. Width is a direct measure of both the recreational potential and the protective function of a beach. A sediment budget analysis consisting of an historic calibration phase, then a predictive phase in which selected components of the budget are changed, is a useful tool to estimate future mean beach widths.

Barriers that act to modify the longshore movements of sediment play a significant role in the sediment budget on a rugged coast. The quantitative impact of littoral

barriers on the net longshore sediment transport rate, Q_n , and the control that rate imposes on mean beach width, \bar{w} (Q_n), are the subject of this investigation.

APPROACH. To estimate \bar{w} (Q_n), the alongshore component of energy flux, p_b , and the availability of sediment for transport, V_e , must be addressed. V_e is the balanced sum of all fluxes entering and leaving a designated coastal reach of length, x_1 , except Q_n , which is the net longshore sediment transport rate out of that reach. Topographic and bathymetric perturbations affect beach widths along rugged coasts. They reduce the actual net longshore sediment transport rate, Q_n , from the potential net rate, Q_{np} . Q_{np} , for the purposes of this work, is the net rate as it would be if the shoreline and all isobaths were straight, aligned with the mean orientation of the shoreline in x_1 , and the volume of the littoral sediment lens, V_l , was such that the underlying substrate (Fig. 1) was never scoured. Q_{np} is the maximum net transport flux for a given net p_b .

Refraction perturbations, controlled by isobaths that deviate from straight and shore-parallel, affect the angle at which waves break and the amount of energy reaching the inner littoral zone. A flat-lying rocky stream delta is a good example of a refraction perturbation. Wave-blocking and diffraction perturbations, typified by a reef or detached breakwater, also affect the angle at which waves break and the amount of energy reaching the inner littoral zone. Sediment-blocking perturbations are parts of the substrate that pierce the surface of the littoral sediment lens. Examples are groins, jetties and headlands (Fig. 2) that act as barriers that locally align the shoreline. The volume of sediment in the littoral sediment lens effects the capacity of all these perturbations, natural and artificial, to affect p_b . In turn, these impediments to longshore transport act like capacitors or gates that regulate the downcoast movement of sediment proportional to the amount that reaches them from upcoast. Hence they partially control the mean width of the beach.

The regional orientation of a rugged coast plays an especially important role in the evolution of sediment-blocking barriers and the plan configuration of the coast. Where Q_{np} is large and $Q_n \approx Q_g$ (Q_g = gross transport rate), hook-shaped bays dominate the coastal planform (Fig. 2, left). Where $Q_n \ll Q_g$, and Q_{np} is smaller, bays tend to be more symmetric and the budget of sediment, and mean beach width, is often more affected by the rate at which seacliffs retreat, shore platforms are downworn, and streams discharge coarse sediment, than the incoming longshore

flux. Deeper symmetric bays may be closed systems to longshore throughput. Sea level rise plays a slightly different role along rugged shorelines with large than with small calculated Q_{np} 's.

FIELD EXAMPLE. Sediment budget analyses were done for two littoral cells with different regional orientations in southern California. These nearby cells serve to illustrate the contrasts between coasts where $Q_n \approx Q_g$ and Q_{np} is large, and $Q_n \ll Q_{np}$ and Q_{np} is smaller.

The Malibu Littoral Cell is 75-km long and oriented in an east-west direction. $Q_{np} \approx Q_n \approx 8.4 \times 10^5 \text{ m}^3/\text{yr}$ -east (measured) at its western boundary. Here, on the Oxnard Plain, the beach is long, straight, and wide. Q_n in segments of the high relief downcoast part of the cell varies within a 25% range of $1.5 \times 10^5 \text{ m}^3/\text{yr}$ -east. Mean beach widths for the downcoast segments are controlled by the orientation of the coast between littoral barriers, the entering longshore flux, stream contributions from high gradient near-coast watersheds, and, importantly, the geometry of the blocking downcoast littoral barrier. Thanks to these barriers of all shapes and sizes, a beach exists along this coast even though the actual net longshore sediment transport rate is only about 20% of the potential rate. The headland shown in Figure 2 (right), with a submerged reef extension, is an example of a medium-sized sediment-blocking barrier in the Malibu Cell.

Q_{np} is much lower along the 21-km long Laguna Beach Littoral Cell (or series of mini-cells). This reach is oriented northwest-southeast. Measured Q_n at its upcoast boundary is zero; Q_n leaving the cell at Dana Point is about $6000 \text{ m}^3/\text{yr}$ -southeast. Pocket beaches are contained within deep symmetric compartments between headlands and reefs that extend like continuous groins beyond the headlands to depths of up to 15m. Most beaches in this cell have been stable (net shoreline change averaging $\pm 0.2 \text{ m/yr}$) in the past 50 years. Seacliffs provide a relatively large portion of the small amount of coarse sediment that annually reaches the deepest of these bays. As the seacliffs are progressively armored, preventing beach retreat, a relative rise in sea level will narrow the beaches unless they are artificially nourished.

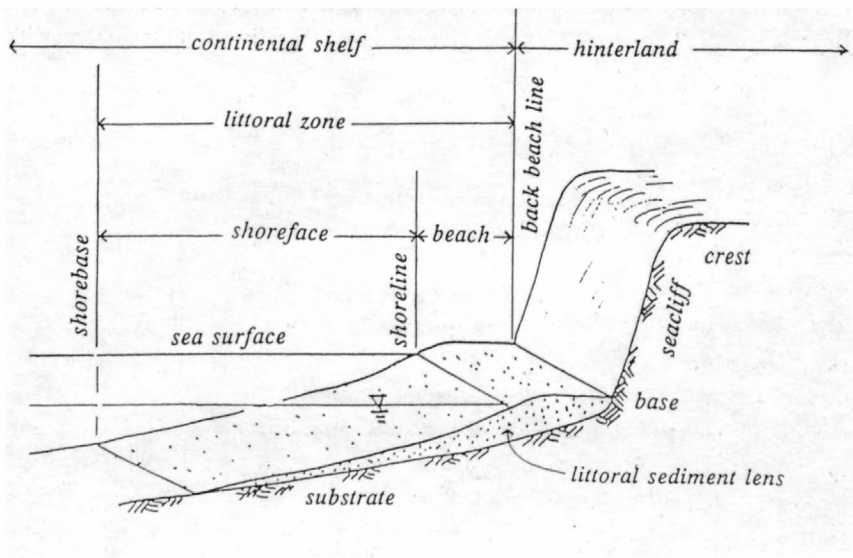


Figure 1. Definition sketch, littoral sediment lens.



Figure 2. Typical sediment-blocking structures in the Malibu Littoral Cell.

**A New Method of Investigating Temporal and
Large-scale Spatial Shoreline Trends**

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Two fundamental questions relative to shoreline studies are: (1) What is the best approach to quantify historical shoreline movement and subsequently, to predict future shoreline locations? and (2) Given that the majority of shoreline studies utilize linear methods to quantify shoreline movement (e.g., an end point rate), what are the errors associated with applying these linear methods to non-linear systems?

We have developed a quantitative method to model the linearity versus non-linearity of temporal historical shoreline movement. Among other applications, we utilize this tool as an aid in predicting future shoreline locations in time and in space and to block the coast into similar regional longshore (spatial) segments based on common historical, temporal shoreline trends.

We employ a complexity penalty method, known as the Minimum Description Length (MDL) criterion, to select a polynomial model most representative of historical shoreline movement at a location. The MDL method scores a model according to both its complexity and its accuracy by assessing relationships among: (1) the number of parameters in the model; (2) the training model error; and (3) the underlying level of data noise. Constant models signify stable shorelines, linear models indicate consistency in the long-term trend, quadratics denote the occurrence of one change in the long-term trend, cubic models suggest two changes in the long-term trend, etc.

Using a data base consisting of T-sheet and aerial photographic shoreline position/time data for transects spaced at 50 m intervals from southern New Jersey to Florida (U.S.A.), we have tabulated the following results: 12% of the shorelines are stable, 23% linear, 63% quadratic, and 3% cubic. Thus, two-thirds of the shorelines sampled along the eastern seaboard display nonlinear historical shoreline movement. Data that span 1943 - 1982 (=40 years) along Cape Canaveral indicate that this reach is somewhat anomalous to the eastern seaboard data set in that 36% of the shoreline is stable, 51% linear, 13% quadratic, and no shorelines are cubic. Last, we discuss scientific and policy implications of using linear methods to quantify non-linear shoreline systems.

OFFSHORE BATHYMETRY AND TIME STEP IN SHORELINE RESPONSE NUMERICAL MODELS

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INTRODUCTION

Shoreline response (numerical) models (SRM) are successful and robust in predicting long-term evolution of the shoreline for a wide variety of wave, beach, and boundary conditions. Basic input for this elementary class of models is a time series of wave height, period, and direction specified at some offshore boundary. For long-term (order of years to decades or centuries) and regional-scale SRM simulations, there are ambiguities and problems with predictive capability in specifying the boundary condition which governs wave propagation and subsequent shoreline configuration and in deciding on a time interval for stepping through a long simulation period.

Wave transformation from deep to shallow water may be performed using either a simplified model (Snells law) or a sophisticated model applicable to irregular bathymetry. Complex wave transformation models are, in principal, more rigorous, but require substantial resources. Also, the time evolution of the offshore morphology is unknown, and uncertainties and inaccuracies introduced with the assumed wave input and time discretization interval cast doubt on the relevance of sophisticated models. One limitation of simpler wave models has been the inability to reproduce realistic (not straight) shoreline planforms over long coastal reaches. This paper provides guidance for specifying both the offshore boundary condition and the time discretization interval for long-term, large-extent SRM applications.

PROCEDURE

In order to refract incident waves, offshore depth contour orientations must be specified. This is normally done in one of three ways (Figure 1): (a) offshore contours are taken to be parallel to the shoreline, (b) offshore contours are taken to be straight lines parallel to the longshore (x) axis, and (c) offshore contours are taken to follow the main trend of the coast. Assumption (a) is most consistent with the original formulation of SRM theory for which the beach profile shape is the same along the shoreline and constant over time.

Irrespective of initial shoreline planform and waves, all three standard assumptions result in a gradual smoothing of the shoreline through time to tend toward a straight line if no local perturbations are present. In contrast, broadly curved and gently undulating shorelines are commonly observed, even on sandy beaches without headlands, most likely resulting from along-shore variations in offshore bathymetry. Figure 2 is a schematized illustration of the interrelations that control the shoreline planform. The dashed line from nearshore to offshore bathymetry represents the incorrect link that is inherent in SRM theory and that causes the shoreline irregularities to be smoothed out in a long-term perspective.

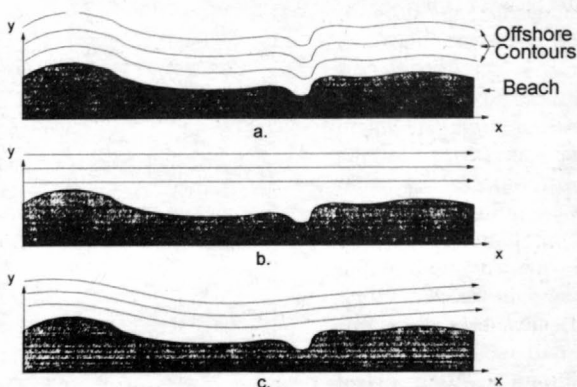


Figure 1. Definitions of offshore contours: a) parallel to shoreline, b) parallel to x-axis, c) along main trend of coast.

The dashed line from nearshore to offshore bathymetry represents the incorrect link that is inherent in SRM theory and that causes the shoreline irregularities to be smoothed out in a long-term perspective.

According to SRM theory, waves cause the profile to move between the limits of a certain berm elevation above datum and a depth of closure $\approx 2H$, where H is a statistical wave height. Therefore, it is reasonable and consistent to assume that the shoreline change produced by a wave affects the bottom contours to the depth of closure, with deeper areas unaffected by the change in shoreline planform. To investigate this effect, the algorithm that calculates offshore contours in the shoreline response system GENESIS (Hanson and Kraus 1989) was altered to allow an external specification of the offshore contour.

The basic philosophy is that it is possible to define offshore contours to represent an *effective* offshore bathymetry forming the breaking wave conditions necessary to produce persistent large-scale features of long-term shoreline planform. Within this framework, large-scale shoreline orientation, engineered and natural perturbations in bathymetry, and internal boundary conditions can be inserted to produce smaller-scale undulations on the shoreline.

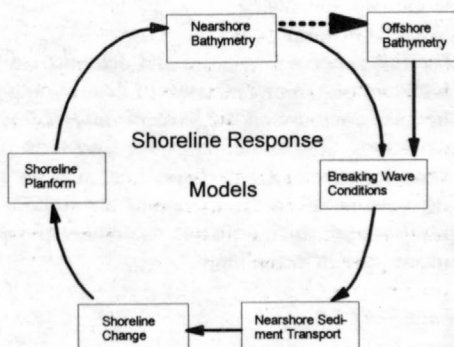


Figure 2. Schematized illustration of the interrelations that control shoreline planform change.

Figure 3 shows an example calculation indicating the approach of the shoreline planform for a somewhat exaggerated smoothed offshore contour, where the smoothing in the planform results from wave refraction. Calculated shoreline position is plotted at 3-month intervals for two years. An advantage of the proposed method for equilibrium planform control in comparison to other methods, such as introducing so-called background erosion, is that after the large-scale trends in bathymetry are accounted for by the offshore contours, actual or proposed nearshore structures can be cleanly introduced.

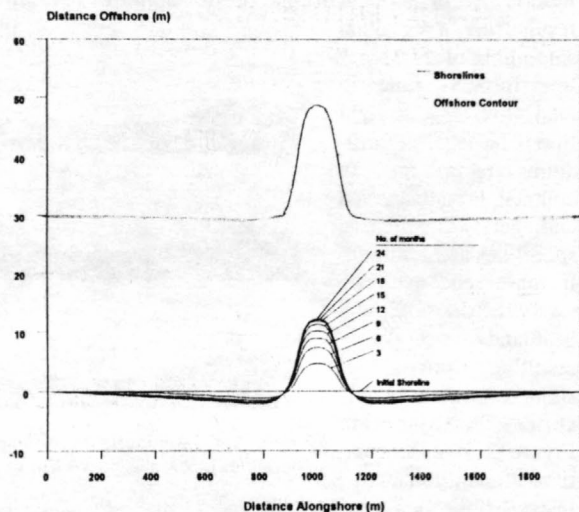


Figure 3. Sample calculation of an initially straight shoreline affected by waves transformed over an irregular bottom.

The full paper will analyze and demonstrate in detail the importance of the offshore boundary condition and ways of determining the shape of the representative effective offshore contours on the basis of observed or expected trends in long-term shoreline evolution. It will then consider long-term predictions (order of several decades or centuries) with SRMs. Guidelines will be given on wave input time discretization intervals based on experience of the authors with regional-scale shoreline evolution projects and on simulations examining physical accuracy, numerical accuracy, and uncertainty in wave input.

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SORTING DIFFERENTIATION PROCESSES ALONG THE ISRAELI
MEDITERRANEAN SHORE AS INDICATED BY GRAIN-SIZE POPULATIONS

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Populations of surface sediments are the only information actually available that integrates the net effects of all long-term sedimentary processes. They usually reflect large scale geological resolution while single samples, laminae or trapped sediments reflect the small scale, instantaneous dynamic conditions of a very short time span. Following the terms of Otto, we study the "sedimentation units" which are deposited under essentially long-term constant physical conditions, as opposed to "laminations" which are deposited due to momentary fluctuations in wave and current velocity.

Surface sediment samples taken along the southern and central Israeli Mediterranean coast, part of the Nile littoral cell, indicate the interplay between grain-size sorting differentiation processes acting cross-shore and along the shore. The outcomes show that longshore sediment transport, deposition, and erosion are dictated by the processes acting normal to the shore.

The littoral sediments were sampled in a systematic stratified sampling survey from the active layer (sedimentation unit). The samples were taken in profiles normal to the shoreline in eight fixed subenvironments of the beach. A few hundred profiles were sampled along 120 km along the shore (from Rafah to Ashqelon, at Nizanim, Palmahim, Bat-Yam, and Herzlia). Eight empirical

populations (super samples) correspond to eight environments normal to the shore were constructed: deep inshore, shallow inshore, step, mid swash-zone, top swash-zone, mid backshore, far backshore, and coastal aeolian sands.

The relative high sample number in this study (a few thousand) allowed abandonment of the single sample approach, and a population concept was used. Super samples, which are grain size probability models, are constructed to describe the sediments with emphasis on the size-sorting patterns between the environments. The supersamples are combinations of all the single samples in a given geographical setting and morphodynamic environment.

By studying the configuration of the various super samples it is possible to use the grain-size populations as natural tracers. One of the outcomes was that the artificially constructed supersamples emerge as distinct grain-size populations, appearing as unimodal, bimodal or polymodal distributions. Thus, an attempt to understand the beach as a sedimentary depositional environment requires the probabilistic approach.

The longshore sediment transport has five components which are related to a large scale model:

- a) Constant winnowing of fine sediment (<0.040 mm to 0.060 mm) from the inshore to the deep offshore.
- b) Storing of sand in the backshore of the southern part of the studied area with the typical grain-size 0.250 mm. Under favorable conditions (lower sea level, climate), the sediments are moved and winnowed further inland to the coastal dunes which cover large areas of the coastal plain, as their size range is within the competence of the wind transport capacity. This long-term mechanism causes a deficit of the grains down stream in the entire beach environments.

c) The part of the grain-size populations which is mostly exposed to the longshore transport is the inshore population. This bimodal grain-size distribution is subjected to two modes of cross-shore winnowing: offshore and inland. It moves relative rapidly by the longshore currents, following the rhythmic patterns of the nearshore dynamics.

d) The grains moving in the swash-zone follow a zigzag motion along the beachface. The movement is relatively slower than the movement in the inshore and suffers much from natural (beach rocks, aeolianites) and artificial (groins, harbors) hindrances.

e) Time effect. The velocity of the various size populations is significantly different, causing a long term, large scale sorting differentiation effect.

The results demonstrate that in the southern part (Rafah - Ashkelon) a stage of equilibrium has been achieved. The slow moving size populations as well as the fast moving are resident. From Nizanim northwards, there is a deficiency of the wind winnowed grains and the slow moving coarse grains. Further down stream, in Rishon - LeZion, Bat-Yam area, the bimodal sediment populations strengthen this appearance. Further north, passing the large natural hindrance of the Jaffa cape, the sediments at Tel Aviv - Herzlia area are the end products of these sorting differentiation processes and form a homogeneous, very well sorted unimodal population of very fine sand in all the various littoral environments. Further to the north only this grain size population can be found in all the littoral environments.

The longterm transport of the detrital nilotic sands along the coast is accompanied by dynamic sorting-differentiation processes. They are responsible for the large scale appearance of the Israeli littoral sediments in four sedimentary cells from south to north.

PREDICTING MORPHODYNAMIC CHANGE
FROM TIDAL RESIDUAL VECTORS
AT A LARGE TIDAL INLET, TAURANGA HARBOUR,
NEW ZEALAND

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Extended Abstract

In order to achieve port expansion, the Port of Tauranga Ltd undertook a major capital dredging programme [6 million cubic metres] in 1992 through and adjacent to a large tidal inlet and delta system located between a barrier island and barrier tombolo system. Sediment transport pathways through the inlet were previously researched by Davies-Colley and Healy (1978a,b) and Black (1984), based upon considerable verifying field data (Healy et al., 1987). Such inlets are regarded as dynamic equilibrium systems (Boothroyd 1985) when comprised of predominantly Holocene stratigraphic sequences (Davis and Healy 1992), and thus the dredging would likely impact upon the tidal inlet morphodynamics, possibly leading to interaction with the coastal littoral drift system and adjacent sandy shorelines (Healy 1980).

Accordingly, as part of the Environmental Impact Assessment investigations, the DHI System 21 hydrodynamic model was run (Bell, 1991) with the aim of attempting to assess the impact of dredging-induced changes on the harbour hydrodynamics. As demonstrated by Black (1984) and applied by Healy (1985) from modelling during the major Tauranga Harbour study of 1983-85 (Barnett, 1985) the full tidal cycle residual vectors for current speeds in excess of sediment threshold - here taken as 0.3 m/s for the medium sands on the floor of Tauranga Harbour - can be interpreted in terms of the major sediment transport pathways.

In particular the "sediment transport residual vectors" (i.e. full tidal cycle vector residuals exceeding 0.3 m/s, the currents capable of moving sandy sediments) should indicate the transport pathways as well as sediment scour and deposition zones. Interpretation is based on the pattern of residual vectors in relation to the current regime, from which it is theoretically possible to predict morphodynamic change, assuming that the tidal currents are the prime sediment transport driver (de Lange, 1988).

For the application at Tauranga, we were primarily interested in the difference between the sediment transport residuals of the bathymetric configuration after the major dredging program, compared with the then existing bathymetric configuration.

The EIA study carefully reviewed the full tidal cycle currents, changes in channel discharges, maximum flood and ebb flow vector differences, and mean and spring tide residual differences, and 'sediment transport' residual patterns as presented from the numerical modelling.

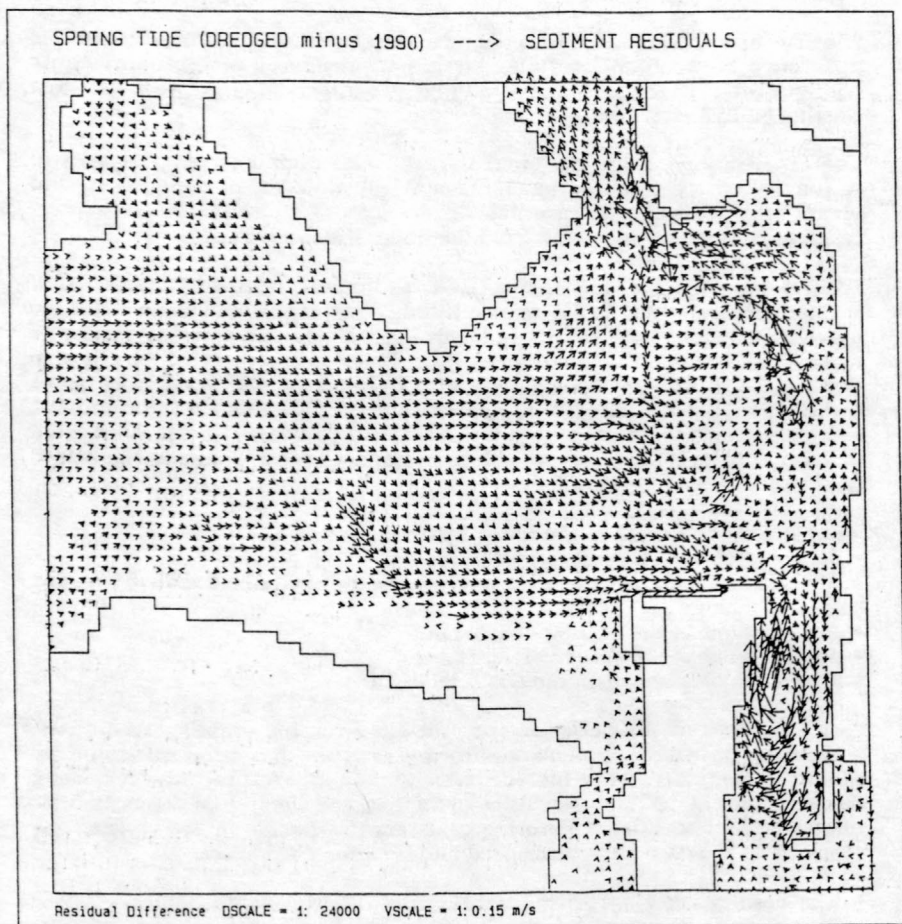
From assessment of the spring tide 'sediment transport' residuals, it seemed (Healy et al., 1991) as if a number of potential changes could be identified. Essentially the dredging had deepened the shipping channels to such an extent that they acted as a major 'ditch' into which the ebb flows were attracted, drawing the ebb flow across the flood tidal delta rather than flowing directly to the entrance. This influenced the implied sediment transport patterns, especially on and adjacent to the flood tidal delta, and in particular on the ebb shield and adjacent channels. The major implications were (Fig.1):

- enhanced ebb flow in the lower Western Channel
- deposition on the western sector of the ebb shield
- deposition in the shipping channel where the ebb shield spilled into the shipping channel
- stronger flow in the Outumoetai channel
- deposition near Matakana barrier island
- no discernible change on the ebb tidal delta.

As a condition of the development consents from the granting authorities, a detailed post-development monitoring program has been instituted by the port company. This includes repeated hydrographic surveys along transect lines at 200 m intervals over the ebb and flood tidal deltas, and air photography and RCM monitoring to assess the change in the system. The monitoring program has commenced and is to run for 5 years.

Initial results show surprising agreement with the model interpretations in that a new scour zone has developed on the flood ramp of the flood tidal delta, and parts of the ebb shield are becoming intertidal when previously all of the ebb shield remained sub-tidal.

Fig.1 The differences in "sediment transport" spring tidal cycle residual velocity vectors between the then existing 1990 and the post major dredging program bathymetry. Predictions of morphodynamic change are based on the pattern of vectors in relation to the current regime.



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**QUANTIFICATION OF MORPHOLOGICAL CHANGES
OF SOME SUPRATIDAL SANDS IN THE GERMAN BIGHT (GERMANY)**

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The North-Frisian supratidal sands Japsand, Norderoogsand and Süderoogsand are located at the seaward border of the Wadden Sea area of the Federal State of Schleswig-Holstein, Germany (Fig. 1). They do not possess a vegetation cover and occupy a total area of about 25 km². Earlier investigations (TAUBERT 1982) show that they are migrating coastward since at least 1947 with a mean velocity of 23 m/a, obviously as a result of the local yearly mean high water (MHW) rise of 0.49 cm.

According to BRUUN rule most of the sediment, released by the drifting of the sands, should have been transported in a seaward direction to be deposited under the wave-base. To test this hypothesis mass balances of the three sands have been carried out for four to five periods between 1947 and 1991. The calculations were carried out with the software package "MORAN", which was written in the cause of a scientific project of the German Coastal Engineering Board (SIEFERT 1989). This simple package not only enables it's user to calculate the sediment budget, but turnover rates as well, thereby allowing the scientist to make statements about the actual intensity of sediment movements in an area (HOFSTEDE 1991).

From 1947 untill 1991 the sands drifted in a coastward direction over a distance ranging between 250 m in the south and 1600 m in the north. Allthough the yearly MHW-rise did not change significantly between 1947 and 1991, the drifting rate of the two southern sands was about twice as high during the period 1967-1991 as before. This phenomena could be the result of the observed increase in storm surge intensity and frequency in the German Bight since about 1960 (SIEFERT 1984).

As a result of the migration about 23 to 43 Mio m³ of sediment was released from the western side of the sands (Fig. 1).

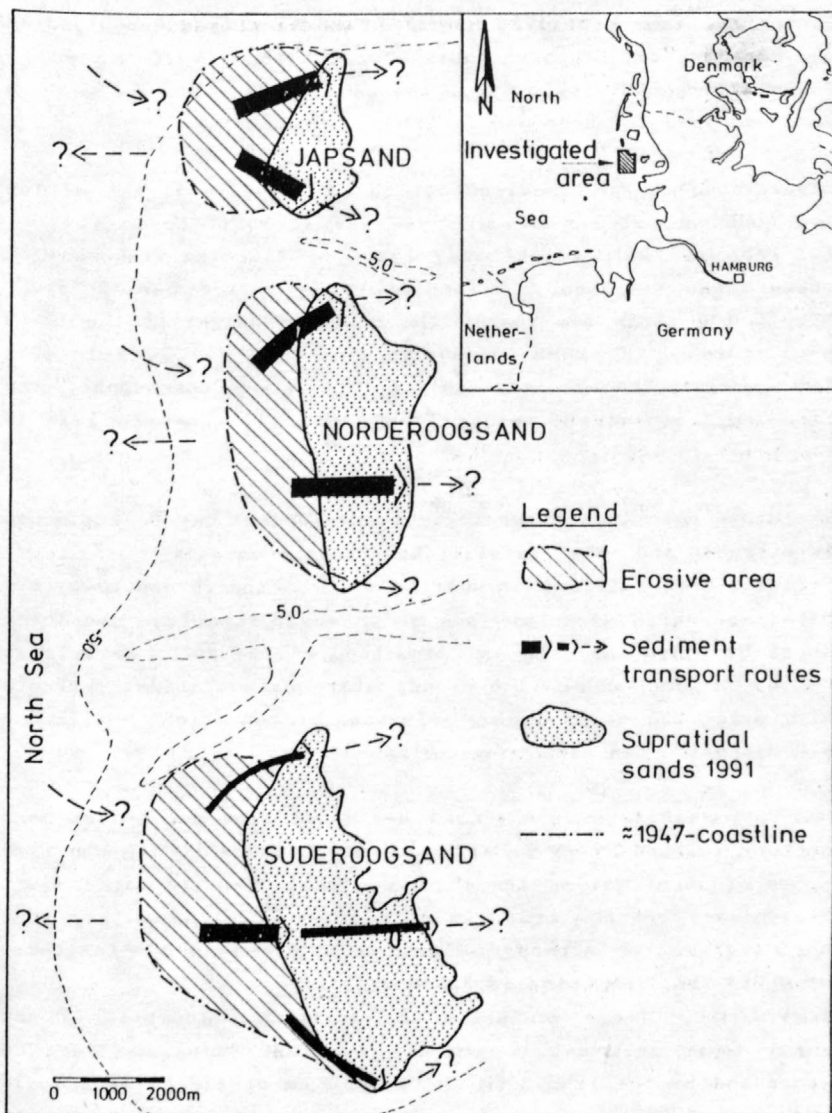


Fig.1: Investigated area with sediment transport routes (generalized)

During the same period 32 Mio m³ of sediment was deposited at the northern and southern ends as well as on top of the sands. So in all probability most of the sand eroded from the western side was used to raise and lengthen the sands.

Resulting from this sedimentation the relative height of the two southernmost sands above MHW was in 1991 the same as it had been in 1947, which means that the flooding frequency of these sands remained the same over the entire period (about 120 to 160 times per year). The relative height of the third sand increased from MHW +6 cm in 1947 to MHW +25 cm in 1991 (an absolute height-increase of 40 cm!). Consequently the flooding frequency decreased from about 300 times in 1947 to 130 times in 1991.

The above described morphological development can be explained by overwash and coast parallel transport processes.

During storms material is whirled up from the foreshore by orbital and surf currents. This sand is exported from the foreshore by storm currents and deposited at the spits as well as on top of the sands. If a steady state exists, tidal currents will erode the spits during fair-weather conditions untill the old situation has been re-established.

Modern predictions upon global sea level rise during the next century (OERLEMANS 1989, WIGLEY & RAPER 1992) lie in the same order of magnitude as the observed local MHW-rise since 1947. Preliminary results from global climate models on the other hand suggest an increase of mean wind speed in mid-latitudes of up to 25%, resulting in 55% higher waves!

According to these scenarios the drifting velocities of the sands would at least remain the same as during the last 50 years and by the year 2050 up to 4000 ha of tidal flats would have been lost.

According to the coastal configuration model of HAYES (1979) the northfrisian supratidal sands nowadays persist in a transitional phase to barrier islands under the prevailing wave

and tidal regime. Thus a relatively stronger increase in mean wave height compared to the tidal range increase could perhaps trigger the formation of one or two new barrier islands in the position of the supratidal sands. If this happens the drifting velocity would certainly decrease dramatically and the whole area would become more or less stabilized.

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The Energy Spectrum of Shoreline and Sand Bar Morphology

R.A. Holman¹, T.C. Lippmann² and W.A. Birkemeier³

Background

Field studies of nearshore morphology tend to focus on variability at a particular temporal scale. For instance, recent experiments at Duck, North Carolina have focussed on sand bar response in the storm band, with typical periods of one or two weeks. On longer scales, the annual cycle of beach change has been the subject of numerous studies, while inter-annual variability is now being discussed based on several long time series. Changes over longer time scales has been the province of geologists, with only a very weak connection to the shorter-scale processes.

Of course, processes of morphological response actually exist as a continuum whose relative importance can be expressed in an energy spectrum. This is useful for two reasons. First, the spectrum shows the strength of variability at all frequency bands and reminds us of the potential interlinking of different bands. Second, there are aspects of spectra that can be indicative of the overall nature of the process. For instance, a red spectrum is usually associated with a process that is self similar (or fractal) across a broad range of scales. Since we believe that feedback between the bathymetry and the wave field is probably quite important, this sort of nonlinear dynamical behavior is, in fact, expected.

Our objective, then, is to compute the energy spectrum of several nearshore morphological features over as wide a frequency range as possible. Comparisons will also be made with the spectrum of significant wave height over the same, approximately decade long, time span.

Data Sources and Analysis

The region of study for the morphological time series will be Duck, North Carolina, the site of the US Army Corps of Engineers Field Research Facility (FRF). Two data sources will be used. For over a decade, the staff of the FRF have collected monthly bathymetry data from the dune to seaward of the outer bar using the CRAB surveying system. Additionally, since 1986, daily video time exposures have been analyzed to show the shoreline and inner bar location and morphology. While the time exposures will provide the primary data for the time series analysis, the CRAB data will be folded in to the analysis to provide better low frequency statistics.

From the two-dimensional time exposure and bathymetry data, three time series will be extracted: the location of the shoreline, the location of the inner bar crest, and the location of the outer bar crest, when present. Results will then be calculated using traditional spectral techniques for the super-annual frequencies and higher order spectra for inter-annual signals.

Wave height data have been collected from a waverider located in approximately 13 m depth for over a decade. Cross-spectral comparisons will be made with bathymetric signals.

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Results

Previous work has already indicated the potential importance of inter-annual signals (Lippmann and Holman, in press). It is expected that these will contribute to redness in the spectrum. However, spectral analysis has not yet been carried out, so any comments on results would be purely speculative. Of particular interest will be spectral shape between wave height and bathymetric response as well as the band-by-band coherence and phase estimates.

Interannual Changes in Bed Elevation: Outer Surf Zone, Duck, NC 1981-1991

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A unique data set of highly accurate nearshore surveys is used to examine the temporal and spatial scales of nearshore profile response. The surveys were conducted between January 1981 and December 1991 on four profile lines along the mid-Atlantic coast of the United States in Duck, North Carolina at the Waterways Experiment Station's Field Research Facility (FRF). The ongoing FRF survey program includes twice-monthly surveys of four profile lines covering an alongshore distance of approximately 1 km (two profiles are located at each end of the FRF property). The profiles extend from the dune crest seaward to approximately the -8 m depth contour. In addition, wave, tide and wind measurements have been made on a much more extensive (>daily) schedule.

This work will concentrate on the portion of the profile deeper than ~4.0 m, seaward of the highly mobile inner bar. This coverage includes the portion of the profile generally considered to contain the 'closure depth' for the majority of the storms impacting this section of coast. The inner bar's development and rapid movement were examined by Sallenger *et al.* [1985], Howd and Birkemeier [1987], and more recently by Lippmann and Holman [1990] and Lippmann *et al.* [1993]. This rapid variation, both in space (Order 10-100m) and time (hours to days) cannot be resolved by the twice-monthly sampling scheme in the present data set.

The outer profile, in marked contrast with the inner profile, is dominated by responses at a time scale on the order of years (Figure 1). The large scale, coherent (over

the 1 km extent of monitored beach) depositional events are seen to decrease in frequency and magnitude in an offshore direction. The periods between depositional events are characterized by slow erosion, during which we assume sand moves back to shallower water.

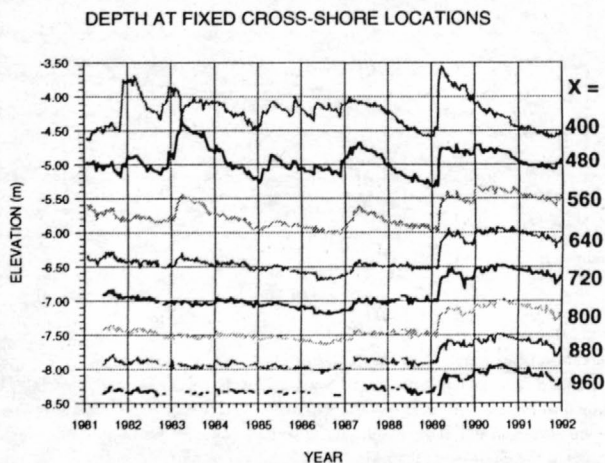


Figure 1. Time series of bed elevation at fixed cross-shore locations on FRF profile line 62, approximately 500m north of the pier. Distances offshore from the FRF base-line are shown along the right side of the figure. The average MSL shoreline position occurs at approximately 100m.

This paper will characterize the nature of these responses and attempt to correlate the elevation changes with process measurements. Initial results seem to indicate that at least some of these depositional events span individual storm events (Figures 2 and 3), and that profile responses can be very different to storms with similar maximum wave heights. Both observations suggest that storm sequencing is extremely important in understanding the magnitude of profile change. Sequence analysis includes not only the relative and maximum wave heights of the storms, but their duration and the length of the 'recovery' period between successive storms.

These are to some extent included in a new depth-dependent form of Dean's Parameter, $S = H^2/(whT)$, where H is the incident wave height, T the incident wave period, h the mean depth and w is the settling velocity of the sediment. Correlations of this parameter with Δh , the change in depth between surveys, while not stunning, are significantly improved over the traditional Dean Parameter ($|r| \sim 0.4$ compared to ~ 0.10 for the Dean parameter). It is also found that different temporal averaging schemes (Wright, *et al.*, 1985) optimize the correlation. The optimal averaging period is short (days-weeks) for the inner profile, and generally increases (to the order of months) at the offshore limits of

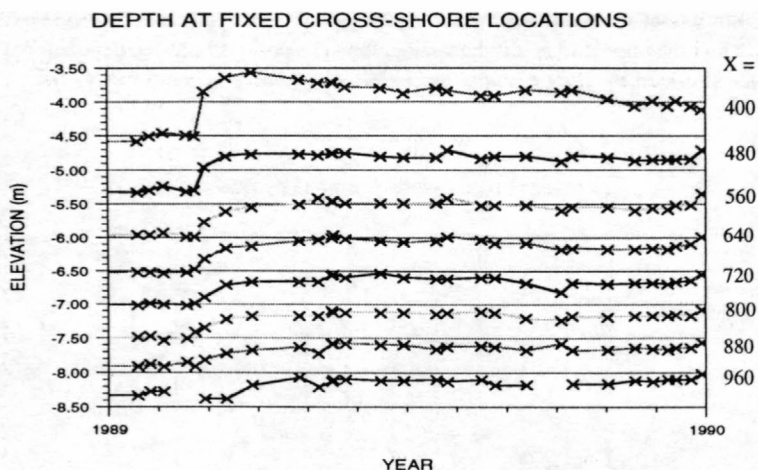


Figure 2. Elevation time series from Profile 62 in 1989 for the same locations as Figure 1, giving an expanded view of the large depositional event. Symbols mark the times of the surveys also shown in Figure 3 in relation to storm events. Note that the deposition occurs over a number of surveys, and that it continues at the further offshore positions after erosional recovery begins at the shallowest location ($x = 400\text{m}$).

this study.

Spectral and cross-spectral techniques are also used to investigate the response of the bed elevations to wave forcing. The results, while preliminary, tend to support the above findings that annual cycles do not dominate the bed elevation response in the surf zone despite being prominent in the forcing.

ACKNOWLEDGEMENTS

This analysis has been supported by Cooperative Agreement 1434-93-A-1097 between the U.S. Geological Survey, Coastal Geology Program, and Duke University (P.A.H. and H.F.S.). The U.S. Army Corps of Engineers supported the data collection and W.A.B. through the Field Research Facility Measurements and Analysis Work Units as part of the Coastal Flooding Research Program at the Coastal Engineering Research Center.

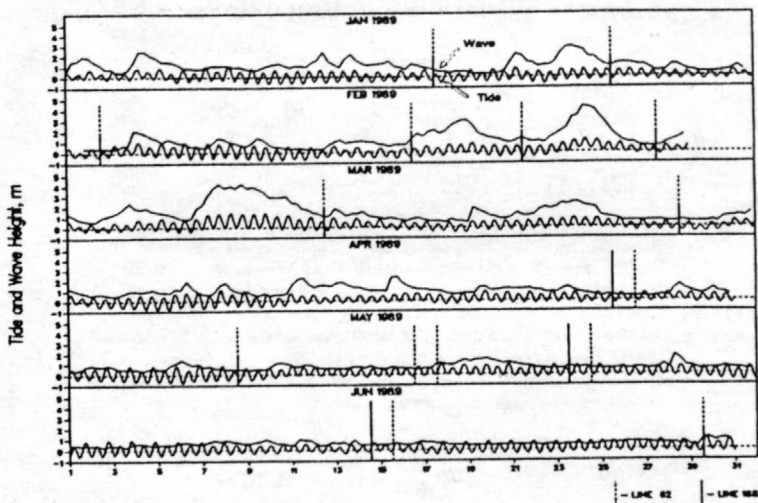


Figure 3. Offshore significant wave heights and tide elevations for the first half of 1989. Times of surveys of line 62 are shown by the vertical dotted lines. Note that they bracket the storms through the February and March period quite well. Despite the occurrence of at least three discrete storms, the elevation data in Figure 2 appear to show a single large response. After Lee and Birkemeier, 1993.

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A dynamical model for tidal offshore sand banks and sand waves

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Abstract

A simple depth-averaged morphological model is described which describes the interaction between a tidal flow and an erodible bed in a shallow sea. In this model the wave-length and the orientation of the fastest growing bed mode agree well with those of tidal sand banks. However, this model only predicts the growth of large-scale sand ridges. We describe a 3D-extension of this model which gives besides a local fastest growing mode corresponding to the sand banks, also a local fastest growing bed mode with the characteristics of sand waves.

1 Introduction

In the seabed several rhythmic patterns can be observed. These range from very little ripples ($\lambda \sim 1$ cm), which can be seen at the beach at low water level, to large tidal sand waves and tidal sand banks. These tidal waves and banks are situated in shallow seas all over the world, where large tidal velocities occur, see Off (1963).

Experimental research gives some characteristic measures of tidal sand banks, see for example Pattiaratchi et al. (1987) and for sand waves see Lanckneus et al (1992) and Huntley et al (1993). The wavelength (distance from crest to crest) of sand waves is approximately 500 m and the wavelength of sand banks varies between 2 and 10 km. The height (horizontal distance between crest and trough) of the banks can reach 30 m, in general the height of both waves and banks varies between 4 to 10 m. The shape of the waves and banks can be very non linear (steep fronts); the mild slope of the banks is often covered with sand waves. The crests of the banks have an anti-clockwise orientation with respect to the major tidal axis, the crests of the sand waves is perpendicular to the major tidal axis. As far as we know tidal sand banks are stationary phenomena, the waves are not.

1.1 Aim

We are interested in the conditions which cause perturbations of the sea bed to grow due to positive feedback with the tidal current. Thus we try to find a dynamical explanation for the existence of the tidal waves and banks. In order to achieve this we develop a model depending on several input parameters which can be adapted to a particular sea. Subsequently we examine if this model yields information on wavelength, height, shape and evolution time of a possible tidal wave or bank. Then we can compare the results with experimental knowledge

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about the banks, and with other models as in Huthnance (1982a,b) and de Vriend (1990).

2 Models

2.1 Depth-averaged model

The model consists of the depth-integrated shallow water equations and conservation of sediment. As the velocities are not very large, we assume that the sand particles will not go into suspension such that the transport may be described as bedload. This process is parameterized by an empirical relationship. This expression describes that transport increases when velocities increase (in a non-linear way) and that sand is easier transported downhill than uphill. Now we arrived at a set of four differential equations for four unknowns (u , v , ζ and h). In order to make this model as simple as possible we choose the boundaries infinitely far away. This means that we study the growing process of offshore sand banks.

2.2 3D model

An extension of the model presented in the previous subsection is a description of the flow in the vertical direction, and assuming that the sediment transport depends on the bottom shear stress instead of the depth averaged velocities. Here we start with a flow model consisting of the 3D shallow water equations. In order to make it possible to compare the results with the 2D-model, we have to model the bottom boundary condition in the 2D model in a more complicated way. Firstly we have to allow a rotation of the bottom velocity compared to the depth averaged velocity (Ekman rotation due to the Coriolis effect), secondly we have to allow a time lag between the bottom velocity compared to the depth averaged velocity (turbulence generated at the bottom needs time to travel upwards the water column). This bottom boundary condition needs four parameters which can be computed from the 3D shallow water model.

3 Linear Stability Theory

The quantities in the model are made dimensionless with relevant scales. It then follows that if we assume the relevant morphological length scale to be much smaller than the tidal wave length, we find a basic solution in which the velocities depend on time only. This describes a tidal ellipse over a flat bottom. We wish to investigate the stability of a particular basic state with a fixed ellipticity of the tidal ellipse. Subsequently we admit a general disturbance (having the whole range of wavevectors) to the flat bottom and investigate if a disturbance initially tends to decay (basic state is stable) or tends to grow (basic state is unstable). We get the averaged initial growth rate by averaging over a tidal cycle. This is allowed because the morphological time-scale appears to be in the order of hundreds of years and is subsequently much larger than a tidal period. The analysis for the depth-averaged model is described in detail in Hulscher et al. (1993).

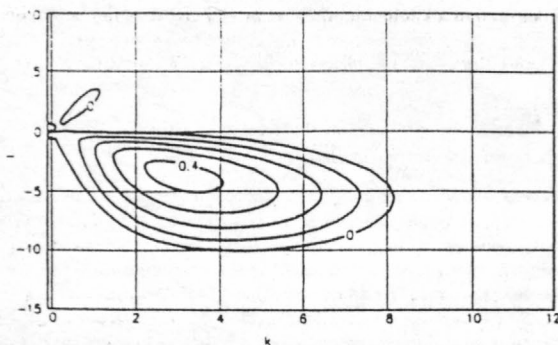


Figure 1: Contour plot of the growth rates of bed form perturbations, in the (k, l) wavenumber plane for a unidirectional tide.

3.1 Sand banks in the depth-averaged model

In the depth-averaged model we can determine the stability of the bed form disturbances in a straightforward way. The unidirectional tide yields stable and unstable disturbances depending on the wave vector \vec{k} of the disturbance. In figure 1 the lines correspond to wave vectors with the same positive growth rate. The initial fastest growing mode is $|\vec{k}| \sim 5$ which in this case corresponds to a wavelength $\lambda \sim 8$ km and the direction $\theta \sim 60^\circ$, where θ is the angle between the wavevector of the disturbance and the principal tidal current direction. We note that the very long wavelengths, though they have smaller growth rates, are unstable too. The stability picture for a circular tide gives the same fastest growing wavelength, but as the circular tide gives no preferred direction, this wavelength can have any direction. The difference between the unidirectional and circular tide is the growth rate of the long waves. The circular tide gives always stable long bed waves, the unidirectional tide gives unstable long bed forms. We conclude that this linear stability model yields initial fastest growing wavelengths and directions which agree with the ones we observe in nature. The results of this linear stability model also agree with the models of Huthnance (1982) and de Vriend (1990).

3.2 Sand waves in a 3D model

To determine the growth rates in the 3D-case is more complicated. We investigated an extension of the depth-averaged model in the following way. First we relate the bedload transport to the local bed shear stress instead of the depth-averaged velocity. The bed shear stress belonging to the basic state can be computed from the 3D model (including veering of the velocity field and a phase difference between bottom shear stress and depth-averaged currents). The bed shear stress induced by the bottom perturbations is approximated by the assumption that there is a similar relationship between the perturbed bed shear stress and depth averaged perturbed velocities as there were in the basic state. This enables us to calculate new growth rates. Now we find two local maximum growth rates, one which coincides with the fastest growing mode in the depth averaged model corresponding to sand banks, and another one with a

wavelength of about half a kilometer and orientated perpendicular to the depth-averaged major flow axis. These characteristics of this second peak agree well with the characteristics of sand waves as described in e.g. Huntley et al (1993).

4 Conclusions and discussion

The simple morphological model predicts local fastest growing modes which corresponds well with the large scale rhythmic patterns as observed in shallow seas. As the sand waves were not present in the depth-averaged model but they do are in the local 3D extension we can conclude they are connected with the vertical structure of the tidal flow.

The evolution of sand banks and sand waves is a nonlinear process. The linear stability theory only yields information on the initial fastest growing bed forms. This approach however is no longer correct when the flat bottom has developed to a wavy pattern with small but finite amplitude. The growth will be affected by nonlinear interactions and possibly other wavelengths will be selected. The initial fastest growing modes can differ from the dominant modes in the fully developed sand bank and sand wave pattern. This means that we must be very careful by comparing the results of a linear theory with observations.

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Numerical Modeling of Shoreline Change from Longshore Transport in the Platte Bay Region, Lake Michigan

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INTRODUCTION

Valuable wetlands behind beaches of the Great Lakes are in danger of being destroyed by coastal erosion. The Platte Bay region of northeastern Lake Michigan was studied to determine the processes responsible for coastal erosion in a wetland rich area (Fig. 1). The study area is a 40-km stretch of coast in a broad embayment with complex offshore bathymetry developed by glaciation (Fig. 2). The Platte Bay region coast is varied and includes erosional areas backed by high cliffs (over 100 m high) and accretional areas with active dune fields.

METHODS

Our predictions of shoreline changes and the fate of wetlands behind the beaches of the Platte Bay region incorporate information from numerical modeling of waves and sediment transport, historical shoreline comparisons, and a beach and nearshore monitoring program. The scheme for modeling the gradients in longshore sediment transport has three components: 1) deepwater wave input, 2) propagation of deepwater waves to shore, and 3) calculation of longshore sediment transport from shallow water wave parameters. The deepwater wave climate is derived from 32 years of wave hindcasts, once every three hours, at a station offshore of the middle of the study area (Hubertz et al., 1991). Waves were propagated to shore using a directional, parametrically spectral numerical model, HISWA (Holthuijsen et al., 1989). This model includes wave generation by wind and wave dissipation through bottom friction, white capping, and shallow water breaking. The convergences and divergences of longshore sediment transport are determined from calculations using standard formula (Komar and Inman, 1970; Shore Protection Manual, 1984) and HISWA wave parameters at the seaward edge of the surf zone.

To evaluate the modeling effort, predicted shoreline changes are compared to long-term historical trends (from 1860 to present) and results from a current beach monitoring program (9 stations in the study area, Fig. 2).

RESULTS

Fair-weather conditions dominate the wave climate of northeastern Lake Michigan (Fig. 3). Over 90 percent of the time waves are less than 1.5 meters high, have short (3 to 7 second) periods, and arrive from any offshore direction. In contrast, the larger waves have longer periods, are associated with the passage of large weather systems, and tend to arrive from either the southwest (pre-frontal) or north (post-frontal)(Fig. 4). The largest waves occur about once every 5 years and have heights greater than 5 meters and periods as long as 10 seconds. A goal of this study is to quantify and compare the contributions of the infrequent large storms and the more frequent small storms to shoreline change.

A HISWA model run for a large storm shows the modification of pre-frontal deepwater waves (Fig. 2). The input wave height was 5 m, wave period was 10s, waves were from the southwest with a 17 degree directional spread, and winds were also from the southwest at 20 m/s. The model was run with default model coefficients and dissipation by white capping, shallow water breaking, and bottom friction enabled. The energy transport vectors show a convergence in the center of the study area where a shoal focuses wave energy and divergences in the southern and northern study area where deep bays spread wave energy. This pattern is supported by recent surveys which show the central study area is depositional. Longshore sediment transport gradients are then calculated for this and other scenarios to determine the long-term net changes to the shoreline of this 40-mile stretch of coast.

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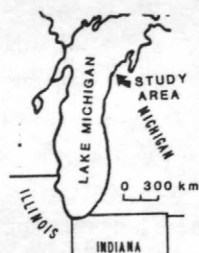
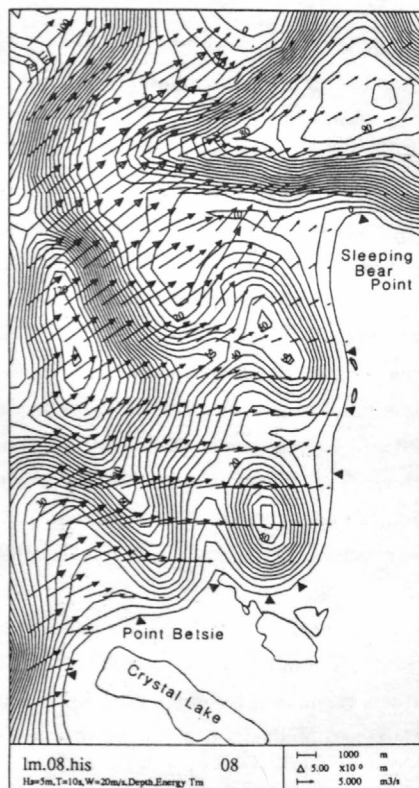


Figure 1- Location Map

Figure 2- HISWA model output showing bathymetry and energy transport vectors for a large storm (5 m, 10 s, waves from the southwest, 20 m/s wind blowing from the southwest). Filled triangles indicate beach monitoring station locations. Depth contours interval is 5 m. A 1000 m scale bar is located in the lower right of the figure.

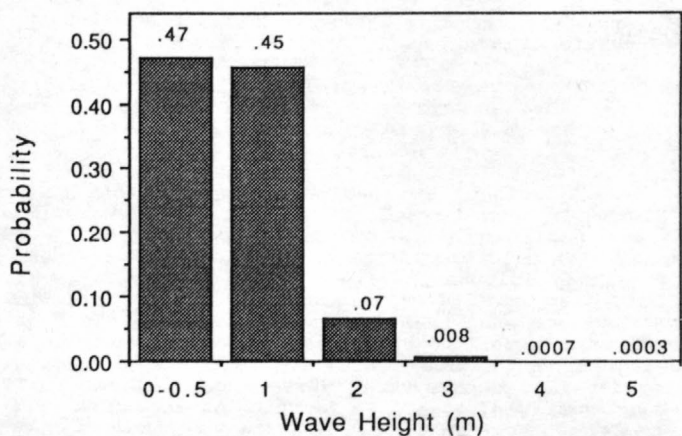


Figure 3- Deepwater wave height distribution for hindcast at a station offshore of the middle of the study area.

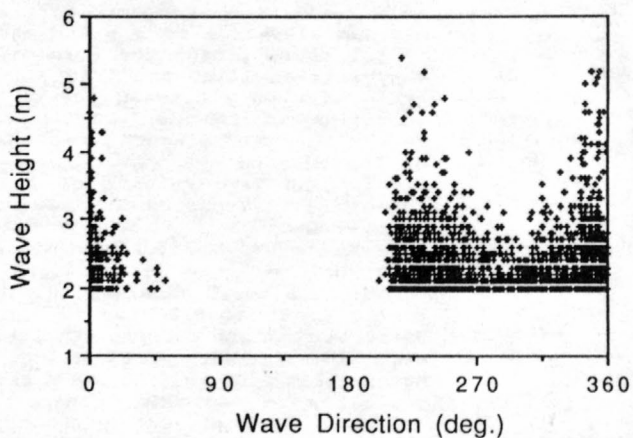


Figure 4- Deepwater wave height as a function of wave direction for large waves (heights greater than 2 m). There were no large waves from between about 50 and 200 degrees because the fetch was too short to generate waves larger than 2 m. A similar plot showing all waves fills in this area of the graph.

SEDIMENT BUDGET FOR THE
WEST COAST OF TAIWAN
A comparative study between results from
numerical models and bathymetric maps.

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A study of wave, current, and sediment transport on the west coast of Taiwan has been carried out to study the future impact on an offshore barrier island of the construction of a huge landfill area which will block some of the supply of sediment to a number of barrier islands forming a large sandspit. The study has included set up and calibration of two dimensional hydrodynamic and sediment transport models and one dimensional coastal sediment transport models. The sediment budget calculated by the models has been compared with sediment budget calculated by comparing detailed bathymetric surveys carried out in 1990 and in 1991 and comparing coastline location on maps from 1962 to 1991. Below is shown some results from a study of the stability of the island Wai-San-Ting-Chou.

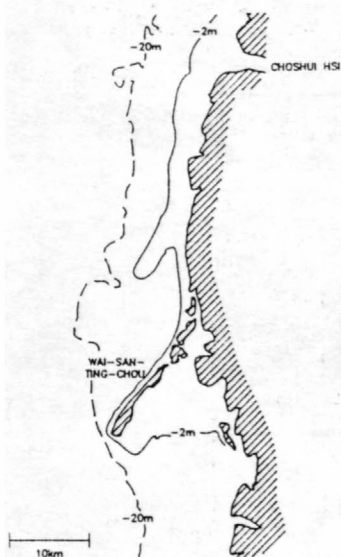


Fig. 1

In fig. 2 is shown the calculated sediment transport rates across 5 sections. Sediment budget shown in the ovals has been calculated for the areas between the profiles by comparing the sediment transport rates in and out of each section. Sediment

The barrier island is located on the west coast of Taiwan, see fig. 1. The island form a barrier between a large shallow area with tidal flats toward south and the Formosa Strait. The maximum elevation is 2 m and the slope of the shore facing the Formosa Strait vary between 1:120 and 1:190. The mean grain size vary between 300 μm on the shoreline to 200 μm on 10 m water depth. The predominant waves generated by the monsoon are coming from N with significant wave heights of 1 - 2 m. Three to four typhoons cross Taiwan each year producing 5 - 6 m waves coming from NE to NW. The dominant current is generated by the semi-diurnal tide producing a coast parallel current with a velocity up to 0.8 m/s.

Based on wave and current statistic, the sediment characterization and bathymetric from 1991, a one dimensional coastal zone sediment transport model (LITPACK) has been set up to calculate the yearly net sediment transport along the north coast of the southernmost barrier island. The model calculate a scheme of sediment transport rates which is weighted corresponding to their occurrence frequency.

budget shown in the boxes has been calculated by calculating the difference in volume inside the actual area between a bathymetric map from 1990 and a bathymetric map from 1991.

The net change in volume in the whole area is $-4.3 \text{ mill m}^3/\text{y}$ (model result) and $4.1 \text{ mill m}^3/\text{y}$ (difference between bathymetric maps from 1990 and 1991) corresponding to an average retreat rate of 75 m/y . The location of the shoreline shown on maps from 1962, 1979, 1982 and 1991 has been used to calculate the change in sediment volume inside the actual area. The average retreat of the coastline is 50 m/y corresponding to a net loss of sediment about $3 \text{ mill m}^3/\text{y}$.

The difference in sediment budget calculated by the model and calculated from the 1990 to 1991 bathymetric maps is small compared to the absolute volumes. The difference between the 30 year average change based on the maps and the result from the model is somewhat larger but still small compared to the absolute volumes. The difference can be explained by change in the slope of the profiles, changes in sediment characteristics etc. during the last 30 years. The overall conclusion is that a one dimensional coastal zone sediment transport model like LITPACK make a fairly accurate picture of the yearly net change in volume and coastal retreat of a shoreline. Once such a model has been set up to calculate a present situation and the results has been verified, the model can be run using different conditions e.g. sheltering from new structures and partly blocking of supply of sediment due to reclamations. These topics will be treated in the paper.

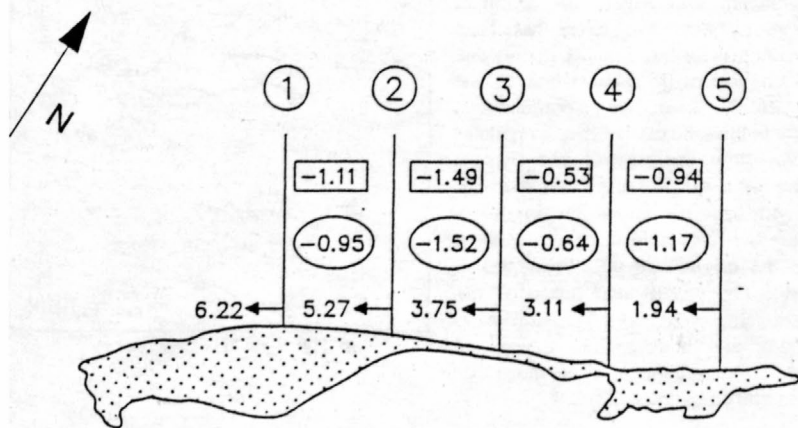


Fig. 2 Sediment budget. Values with arrows are transport rates. Values in ovals are sediment budget for each section calculated by the model. Values in boxes are sediment budget calculated based on repeatedly bathymetric surveys.

EROSION AND ACCRETION OF THE EBRO DELTA COAST: A LARGE SCALE RESHAPING PROCESS

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The Ebro Delta coast is a 50 km long sandy shoreline which has developed during the last five centuries from the sediment supplied by the Ebro River (Fig. 1). Main morphological features are the two spits enclosing two big lagoons (to the north and to the south of the river mouth) which indicate a net longshore transport directed towards the south in the southern hemidelta and towards the north in the northern part. After several centuries of growth, the deltaic trend of evolution changed a few decades ago mainly due to the nearly total reduction of solid river discharge. In fact, since then the evolution of the delta may be considered to be primarily wave dominated.

The aim of this study is to analyze the observed coastline changes to obtain a conceptual large-scale evolution model for the Ebro Delta coast. The observed changes will be associated to different processes acting on specific time and spatial scales.

DRIVING AGENTS

i) River discharge

A dramatic decrease in the sediment discharge by the Ebro River has been detected during the last decades, mainly due to: (i) the construction of dams in the course of the river, which produces a decrease in the solid discharge and regulates the river flow downstream and (ii) the decrease in rainfall at the Ebro drainage basin and flow deviations for irrigation. Presently the river solid discharge is about 5% of the original supply (Palanques *et al.*, 1990). The useful sand discharge has been estimated as 32,000 m³/yr (Jiménez *et al.* 1990), and it is mainly transported towards the northern part of the Delta (Guillén and Jiménez, 1993).

ii) Wave conditions

Wave characteristics in the Ebro Delta coast can be roughly described by an average offshore H_s of 0.72 m, and an average T_m of 3.9 s. Three main components can be distinguished: eastern - E and NE-, southern and northwestern waves. Eastern components, characterized by higher and more energetic waves, are predominant in morphological terms.

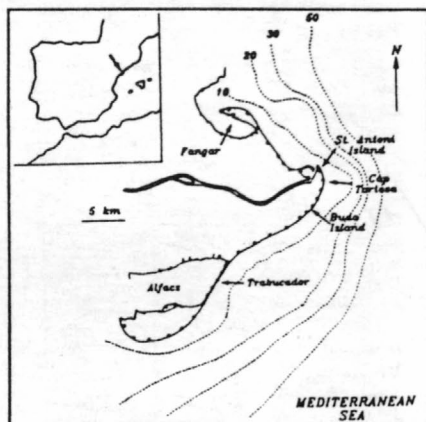


Fig. 1. The Ebro Delta coast.

iii) Sea level changes

Short-term

Like most of the Mediterranean coast, the Ebro Delta is a microtidal environment, with a maximum astronomical tidal range of about 0.25 m. Meteorological tides associated to the offshore passage of low-pressure systems can be observed several times per year. Under these conditions, surges up to 1.0 m have been recorded.

Long-term

There is a lack of information on sea-level changes in the Ebro Delta coast due to the absence of long-term tidal records in the area. Emery *et al.* (1988) analysed neighbouring sea-level records, estimating a sea-level rise of 0.8 mm/yr for the Alicante station (SE Spain) and of 1.4 mm/yr for the Marseille station (SE France). However, a larger relative sea-level rise (RSLR) is to be expected in the area due to local subsidence and compaction. Sánchez-Arcilla *et al.* (1993) have estimated a RSLR about 3 mm/yr applying a modified Bruun rule.

SHORELINE AND AREA CHANGES

Shoreline changes for the period after dam construction have been calculated using aerial photographs. Three campaigns corresponding to the years 1957, 1973 and 1990 were used. Moreover, several beach surveys taken from 1988 to 1992 were available to estimate beach profile changes at a shorter scale.

The comparison between shoreline positions from 1957 to 1990 can be seen in Fig.2. Erosive areas are the central lobe (with a maximum erosion retreat of 1500 m), the northern exposed coast and the Trabucador Bar. Moreover, the inner part of the Trabucador Bar has experienced some displacement towards the mainland. The combination of this displacement and the outer coast erosion has resulted in a rotation of 2° clockwise around its north end-point since 1957.

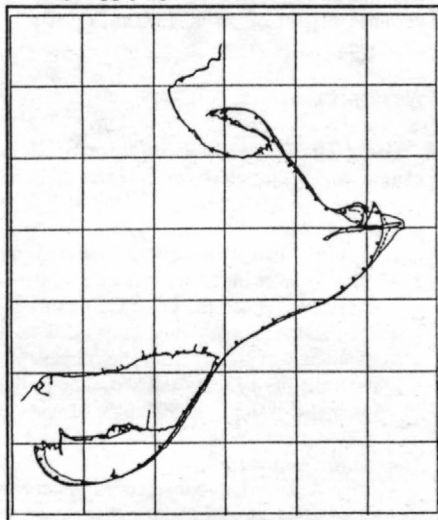


Figure 2. Shoreline evolution between 1957 and 1990.

Accretive areas are the two spits (the northern one has prograded 1000 m towards the coast and the southern one advanced 700 m towards the SW during the last 30 years) and the northern part of the Trabucador Bar.

If these shoreline displacements are converted into area changes, results indicate that the total subaerial deltaic area has increased from 1957. Fig. 3 shows the rates of area changes for two periods (1957/73 and 1973/90). It can be observed that during the first period (1957/73) gross and net rates of area change were higher than during the second period. This can be associated with the fact that during the first analysed period, took place most of the deltaic coastline response to the change in natural conditions (decrease in solid river discharge, etc...). Thus, during this stage, the St. Antoni Island and mouth area were reshaped to a configuration closer to the present one. The rates of change during the second period (1973/90) are smaller because the deltaic coastline is continuously evolving towards a equilibrium configuration, and the corresponding coastline rate of change is therefore reduced.

The origin of this net area increment must be associated with the low solid river discharge and with the sediment redistribution along the coast. In this respect, the siltation of the north lagoon is produced by the deposition of sand in a shallow zone, which implies that the same amount of sediment will correspond to a wider subaerial area than that in the outer coast.

If the net increment of the deltaic area is converted to volumetric change -under the adoption of a realistic closure depth- a small volumetric gain of about $15,000 \text{ m}^3/\text{yr}$ is obtained.

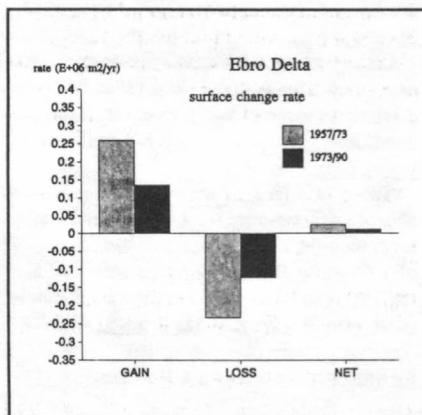


Figure 3. Rates of deltaic area change.

PROCESSES

Using all the available information shoreline changes have been split into several components acting on different scales and associated to different processes.

Short-term changes

Short-term changes have been determined from beach surveys taken during the period 1988-92. These changes are reflected in the seasonal movements of the shoreline and bar/trough systems. Since the coastline of the Ebro Delta shows segments with very different orientations, the profiles must be studied taking into account the angle of wave incidence. This explains why, under the same offshore wave climate and under similar nearshore wave heights, inner bar systems and shoreline position evolve in a different way depending on its location and therefore, on the angle of wave incidence.

Medium-term changes

Medium-term behaviour refers to the coastal response over several years, in such a way that seasonal/cyclic and shorter changes are considered as "noise". This response has been derived from aerial photographs and beach surveys following the methods proposed by Dolan *et al.* (1991). The medium-term evolution trend of the deltaic coast has been associated with longshore transport gradients acting along the coast (Jiménez and Sánchez-Arcilla, 1993). The results show that at this scale, the sediment budget in the coastal area can be considered as closed, and the main response is a redistribution of the sand along the coast. At this scale a closure depth of 7 m has been calculated from beach profile changes (Jiménez and Sánchez-Arcilla, 1993).

Episodic changes

Episodic coastal events are coastal morphodynamic changes due to the action of a set of driving forces with a long return period. These highly energetic forces produce important erosive processes in a very short time span during heavy storms. These changes are noticeable specially at sensitive stretches such as the Trabucador Bar (a barrier beach). An example of these changes occurred in early October 1990, when an eastern storm caused a breach of 800 m long and with a maximum depth of 0.4 m below the MSL. During this

process approximately 70,000 m³ of sediment were removed from the subaerial part of the beach and transported to the inner bay (Sánchez-Arcilla and Jiménez, 1993). This amount is about 70% of the total volume eroded from the barrier during a normal year due to longshore transport gradients. This process occurs by the action of very high waves and in a situation of mean water level surged by meteorological conditions.

Long-term changes

Long-term coastal behaviour must be considered as the integral result of all agents acting on the coast at smaller scales plus the action of agents whose consequences are only visible at this scale. The temporal scale ranges from decades to centuries.

These changes will need a relatively long time to be detected although they are continuously acting. An example of this is the landwards movement of the inner side of the Trabucador Bar. This movement can be associated with overwash processes during a normal year (several overwash events per year are usual) and inlet formation during episodic events, as previously mentioned. These processes may be considered as a time-averaged phenomenon, due to storm events over a certain time-period (see e.g. Leatherman, 1983).

Other factors acting at this scale are cross-shore feeding, aeolian transport and changes due to RSLR. Using sediment budget considerations, Sánchez-Arcilla *et al.* (1983) proposed that presently the Ebro Delta coast is nearly in equilibrium, in such a way that under the present scenario of RSLR and low river solid discharge, the net sand balance in the coastal area of the Ebro Delta is close to zero.

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NEARSHORE BARS AND LARGE SCALE COASTAL BEHAVIOUR

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Most of the existing literature about large scale coastal behaviour stresses the relations between the external forcing mechanisms (relevant processes) and the observed tendencies in the morphology. These types of relations may to some extent point at the observed tendencies in morphology, but the spatial variability in coastal behaviour is often hard to relate to the spatial variability of the external forcings. At the same time, the internal variations in boundary conditions in association with spatially non-varying external conditions may often explain the observed behaviour.

The present literature on the dynamics of nearshore bars mainly focusses on the migrational tendencies of these bars during a single event, like a storm period. In multiple-barred systems almost all emphasis is given to the behaviour of the inner bar systems which are largely influenced by swash-backwash processes as well as breaking waves and mean current processes. The observed migrational tendencies are described by morphodynamical models which are based on for example the break-point theories, the undertow processes and the long period waves.

The net responses of these multiple-barred systems on the time-scale of years are hardly studied at all. Larson and Kraus (1992) started with the analysis of the DUCK-data set, but most of the databases are too short in time and/or lack the morphological observations in the outer nearshore zone.

The spatial differences in coastal behaviour along the central part of the Dutch coast are discussed by Wijnberg and Terwindt (1992). In the present study the behaviour of nearshore bars is studied at three specific locations along the central part of the Dutch coast, near Egmond in the North, Zandvoort some 20 km to the South and Katwijk another 20 km to the South.

The differences in processes and morphological behaviour of bars over a cross-shore profile justify a division in the areas:

- the "swash bar area" between the fore dune and the inner nearshore bar;
- the "nearshore bar area", which is the part of the profile where the inner and outer nearshore bars are positioned;
- the "offshore area" seaward of the outer nearshore bars.

The quantitative description of the cross-shore profile development and the nearshore bar behaviour in morphometric terms requires a mean profile to

which the individual measured profiles are compared. Herein, the mean profile of the nearshore bar area is used.

The mathematical expression used to describe this mean profile in the nearshore bar area is:

$$h = [1 - x_r^b] \Delta h + h_r$$

in which

h = computed height [m];

x_r = normalized offshore distance (normalized to the width of the nearshore bar area) [-];

b = exponent;

Δh = difference in bed height between the landward and seaward boundaries [m];

h_r = reference height at the seaward boundary [m].

The residual profile, defined as the individual measured profile minus the fitted profile is used for the description of the morphology. The dimensions of the bar features include the position of the bars, the bar spacings, the amplitude of bars, the width of bars, the slope of bars etc.

An overview of the characteristics of the morphology at Egmond, Zandvoort and Katwijk is given in table 1. The temporal scale of interest is in the order of decades. All the specific sites are multiple barred and not influenced by man-made structures like harbour moles or seadikes.

	Katwijk	Zandvoort	Egmond
mean slope	1:130	1:150	1:115
amount of bars	2	3	2
longshore rhythmicity (km)	-	-	1-5
cross-shore migration rate (m/y)	60	60-70	30
return period of a bar (y)	4-5	3-4	~ 15
lifetime of a bar (y)	6-8	10	15-20

Table 1 Characteristics of the nearshore zones at the study sites

The behaviour of nearshore bars is to some extent a reflection of large scale coastal behaviour. The actual relation between bar behaviour and large scale coastal behaviour, however, is far from understood. The residual volumes of sediment in the nearshore bar area over the years are showing irregular alternations (Fig. 1). The integration of these yearly sediment volumes may give a net sediment volume over the decades (see Terwindt and Kroon, 1993). However, an underlying trend is not visible and the alternations differ with those of morphological development of the bars. This is partly due to the

limited amount of 27 years in the database.

All the nearshore bars at the specific sites along the central part of the Dutch coast are migrating in a net seaward direction. This migration is not continuous, but shows a certain return period over years (table 1). The net migrations in seaward direction are supposed to occur after successive storms, the net stand still phases are related to calm weather periods. Assuming a major influence of long period waves on the modulation of the nearshore barred topography (see Kroon, 1990) the sudden seaward migration of the outer nearshore bar may be induced by morphological changes in both the inner and outer nearshore zone. In the inner nearshore zone the flattening and landward shift of the waterline position after a couple of storms result in change in landward location of the reflection area for the long waves. Hereby, the former structure of the long period waves and the barred topography are out of resonance and quickly striving to attain a new quasi-equilibrium. In the outer nearshore zone the outer bar may come below a threshold level, where it will stop resonating with the long waves. At that moment, the inner nearshore bars are free to move offshore.

In addition, it is also expected that the observed spatial differences in bar appearance and bar migration rates at the three specific sites along the Dutch coast may be induced by internal differences in the boundary conditions, such as the mean slope. Field data of hydrodynamics are generally in favour of this view, since there is for example no significant longshore variability in the mean annual wave climate along the central part of the Dutch coast. There is also no reason to assume that the supply of incoming infragravity energy will vary substantially along this coast.

It is obvious though that the time-scales involved in meso-scale coastal behaviour (= bar dynamics) may vary from a decade near Zandvoort to more than two decades near Egmond. At the same time the spatial dimensions of the morphological features are also different. It is important to realize that this regional variability in meso-scale coastal behaviour is probably a response to varying internal boundary conditions.

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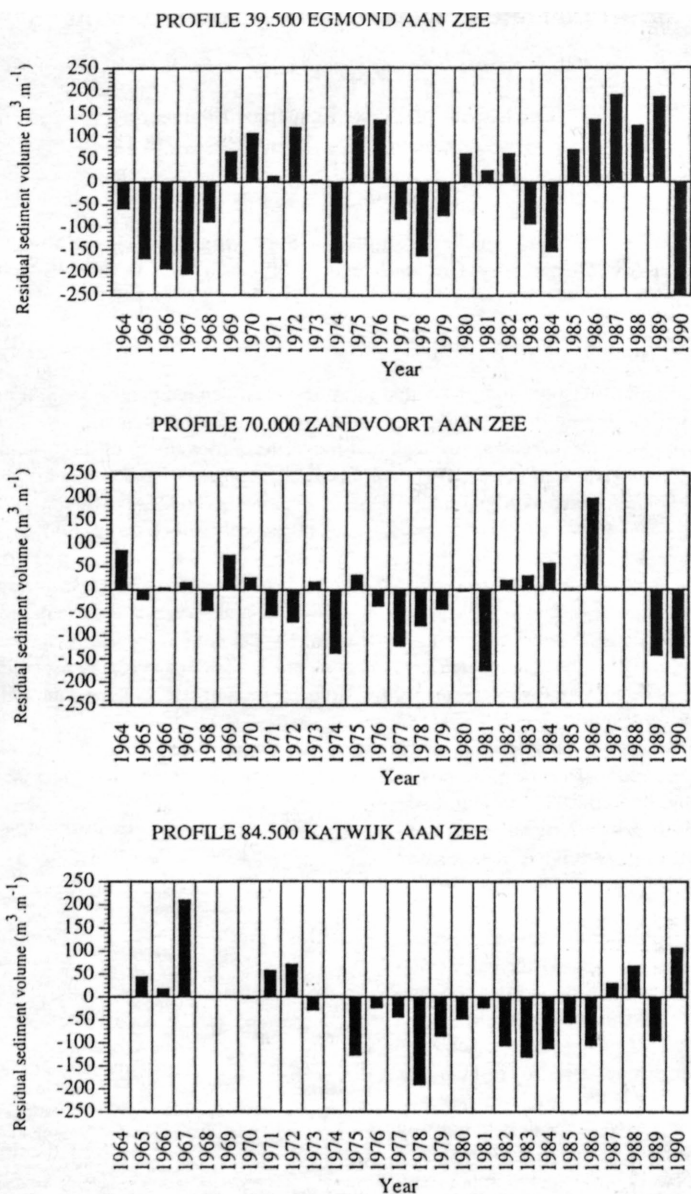


Figure 1. Residual sediment volumes for three different locations over the period 1964-1990

PREDICTION OF CROSS-SHORE SEDIMENT TRANSPORT AT DIFFERENT SPATIAL AND TEMPORAL SCALES

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INTRODUCTION

Coastal sediment transport and associated morphological change can be viewed on spatial and temporal scales ranging from instantaneous motion of single grains on a time scale comparable to that of the local turbulence to seasonal and longer term movement of large sand bodies such as longshore bars and tidal deltas. Sediment transport, its forcing, and engineering activities are conveniently classified according to representative space-time scales for the interaction between morphology and flow. *Microscale* refers to changes from sub-wave period to several wave periods (time scale $t = \text{sec} - \text{min}$; space scale $\ell = \text{mm} - \text{cm}$). At *mesoscale*, net transport rates over many wave periods are evaluated ($t = \text{min} - \text{hr}$, $\ell = \text{cm} - \text{m}$). *Macroscale* involves seasonal changes ($t = \text{days} - \text{months}$, $\ell = \text{m} - 100 \text{ m}$), and *meegascale* corresponds to changes over coastal reaches on the order of one or more littoral cells ($t = \text{year} - \text{decades}$, $\ell = \text{km} - 100 \text{ km}$). In Fig. 1, the shaded area denotes compatible time-space scales for calculations, and white spaces denote time-space scales considered by the authors to be incompatible.

A central concern is how to unite or reconcile calculation approaches of different scales, for example, the microscale of basic physical processes and the meso- and macroscales of engineering applications. The present study considers this problem by simplifying to one dimension involving cross-shore sand transport and beach profile change.

Successful engineering cross-shore models have been developed for calculating transport rates at mesoscale (Kriebel and Dean 1985, Larson and Kraus 1989), and only limited success has been achieved with models that aim to resolve the transport rate on the microscale. Microscale-based models employ smoothing and filtering in the calculation procedures to ensure stable model behavior in carrying over to the mesoscale,

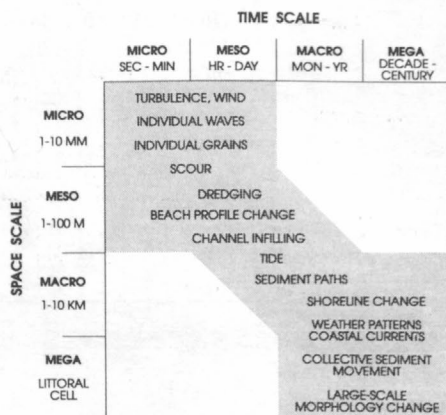


Fig. 1. Space-time scales in sediment transport and morphology-fluid interactions.

especially for the case of monochromatic waves (Dally and Dean 1984, Hedegaard et al. 1991). In a unified approach, formulations of transport relationships at different scales should allow for a smooth transition between different scale ranges based on physical principles. The objectives of this study are to examine calculation procedures for the cross-shore transport rate at different spatial and temporal scales and to investigate how transport relationships developed for the mesoscale may be used for verifying microscale transport formulas and in beach profile evolution simulations on a macro- and megascale.

CROSS-SHORE TRANSPORT RATE MODELS

Larson and Kraus (1989) developed the numerical model SBEACH to predict profile evolution produced by storms, employing a mesoscale description with calculation time steps of 5-30 min and resolving morphologic features of 1-10 m. The cross-shore transport rate in the surf zone was expressed in terms of wave energy dissipation per unit water volume and an equilibrium energy dissipation (Kriebel and Dean 1985). Seaward of the surf zone, the net transport rate decreased exponentially with distance offshore depending on breaking wave height and grain size. In the swash zone, the transport rate decreased linearly from the seaward end of the swash zone to the runup limit. Mesoscale transport formulas were developed with data from large wave tank (LWT) experiments.

The cross-shore transport formulas in the swash and the offshore zone employed in SBEACH were largely empirically based, being derived from simultaneous analysis of wave properties and calculated cross-shore transport rate from measured profile change. In the following, the connection between these empirical, mesoscale formulas and descriptions of the cross-shore transport rate at the microscale is discussed.

Cross-Shore Transport Rate in the Offshore Zone

If sediment is transported offshore primarily through advection from the surf zone, where the sediment is suspended because of strong turbulence from wave breaking, a mass balance per unit width for an infinitesimal element in the cross-shore plane yields,

$$-\frac{\partial}{\partial x}(Vc) + \frac{\partial}{\partial z}(wc) = \frac{\partial c}{\partial t} \quad (1)$$

where V is the local cross-shore velocity, w the sediment fall speed, c the sediment concentration, x the cross-shore coordinate pointing offshore, z the vertical coordinate pointing upwards, and t the time. Introducing self-similar V - and c -profiles in Eq. 1 expressed as $V=V_B(x)\Psi_1(\xi)$ and $c=c_B(x)\Psi_2(\xi)$, where $\xi=z/d$, Ψ_1 and Ψ_2 are shape functions, and the subscript B refers to conditions at the sea bottom, integration from the sea bottom, located at $z=-d$, to the water surface gives for steady conditions,

$$\frac{d}{dx}(c_B V_B d) + \frac{w c_B}{A} = 0 \quad (2)$$

$$A = \int_{-1}^0 \Psi_1 \Psi_2 d\xi \quad (3)$$

A closed-form solution for the concentration at the sea bottom c_B may be derived to Eq. 2 for a V_B that is a known, arbitrary function of x ,

$$\frac{c_B}{c_o} = \frac{V_o d_o}{V_B d} e^{-\frac{1}{A} \int_0^x \frac{w}{V_B d} dx} \quad (4)$$

where the subscript o refers to a point where c_B is specified, such as the break point, also taken to be the origin of the x -axis. Volume transport per unit width q at a location x is,

$$\frac{q}{q_o} = e^{-\frac{1}{A} \int_0^x \frac{w}{V_B d} dx} \quad (5)$$

where q_o is the transport rate at x_o . Eq. 5 represents a mesoscale transport rate that decreases exponentially with distance offshore, in agreement with empirical results from analysis of LWT data.

Cross-Shore Transport Rate in the Swash Zone

A general expression for the sediment transport by waves on a sloping beach is,

$$q = B \tau_b^{3/2} \tan \beta \quad (6)$$

where τ_b is time-averaged bottom shear stress, β local beach slope, and B a coefficient encompassing sediment properties such as grain size and density (Watanabe 1982, Madsen 1991). Eq. 6 may be used to derive the approximate shape of the transport rate distribution in the swash zone, under the assumption that the magnitude of the transport is determined by the conditions at its seaward boundary. The shear stress is assumed to be related to the velocity according to $\tau_b = 0.5 f \rho u_b^2$, where f is a friction coefficient, ρ water density, and u_b time-averaged velocity in the swash taken to be uniform through the water column. In order to employ Eq. 6 to calculate transport rates, an estimate is needed of the local swash velocity representing the net of the uprush and backwash. A simple approximation of u_b is obtained from the speed of the bore front u , estimated as,

$$u^2 \sim u_s^2 - 2g\Delta h \quad (7)$$

where u is the bore speed when a swash cycle begins at the location x , g is acceleration of gravity and Δh is the elevation difference between the center of the bore at x_s and a point on the foreshore.

Although bottom friction may be the main agent for transporting the sediment, it is assumed here that the decrease in velocity on the foreshore is mainly determined by the transformation of kinetic energy to potential energy and not by energy loss. Also, the representative local

velocity of sediment transport at a point on the foreshore is taken to be proportional to the speed of the bore front as it passes that point. Substitution and defining the transport rate at x_r as q_r yields,

$$\frac{q}{q_s} = \left(1 - \frac{2g\Delta h}{u_s^2}\right)^{3/2} \frac{\tan\beta}{\tan\beta_s} \quad (8)$$

where β_s is the local slope at x_r . Eq. 8 is valid for elevations for which $2g\Delta h < u_s^2$; in the limit $2g\Delta h = u_s^2$, the runup height R is attained, and Eq. 8 can be written:

$$\frac{q}{q_s} = \left(1 - \frac{\Delta h}{R}\right)^{3/2} \frac{\tan\beta}{\tan\beta_s} \quad (9)$$

For a plane-sloping foreshore described by $\Delta h = \tan\beta(x_r - x)$, where $\tan\beta$ is the constant foreshore slope, Eq. 9 reduces to,

$$\frac{q}{q_s} = \left(\frac{x - x_r}{x_s - x_r}\right)^{3/2} \quad (10)$$

where x_r is the location of the runup limit. For the lower foreshore (x close to x_r), Eq. 10 is well approximated by a straight line, as found in analysis of the LWT data.

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Techniques for Long-Term Morphological Simulation under Tidal Current Action

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Introduction

Although hydrodynamics and morphodynamical changes are strongly coupled, resulting in mutual adjustment within an evolving overall system, time scales of both components are quite different : water motion can strongly change every second under wave conditions (letting aside turbulence time scales), or every hour under tidal conditions, while significant bed changes will usually occur after many weeks, months, or even years.

For morphodynamic computation, this difference in time scales gives rise to at least 2 problems :

- * the cost of the simulation of the coupled system where small time step is required from hydrodynamics variability, while long duration of the simulation is linked to slow bed evolution ;
- * the ability of the modelling of short term fluctuating sediment transport and bed changes to be extended to long term trends, for which these fluctuations can appear as a kind of noise which scrambles these trends ; moreover, the amplification of errors all along the path from instantaneous velocity to long-term evolution - going through instantaneous sediment transport, net sediment transport (over the tide, e.g.), short-term bed evolution - can make it quite large the effect on long-term behaviour of a small error in hydrodynamics.

This paper presents both aspects of input filtering and of tidal process filtering, describing the results of the work carried out by the LNH within the european research programme G6 Coastal Morphodynamics (Latteux, 1992) as well as experimentations previously performed. This presentation will focus on the case of non cohesive sediment, although the behaviour of cohesive sediment has also been investigated within the same european programme (Villaret et al., 1992).

Input filtering

The problem is to find out a limited set of tides which can represent the whole tide variability. The questions are :

- how many "representative" tidal conditions are required to simulate properly the effect on the bed of this whole variety of the tide climate, according to the case to be dealt with ?
- and how to determine them ?

These questions have been examined using various approaches :

- investigation on the case of a schematic velocity distribution of tidal current,

- utilization of measured currents, for all classes of tides, in several regions characterized by typically different flow pattern (Latteux, 1990a),
- numerical experimentation in the case of a schematic estuary opening in a coastal region dominated by strong currents (de Vriend et al., 1993). This is a rather severe case because the non-linearities of hydrodynamics in such a configuration lead current patterns to depend significantly on tidal conditions.

For the two first cases, the main tentative conclusions were as follows :

- when bed and coast topography are rather simple, a unique tide condition can be found, using a sediment transport formula and the measured currents for each class of tide range. This condition is in between mean tide and mean spring tide, due to the non linearity of the transport formula ; the more the used formula is non linear, the farther this tide condition is from mean tide.
- on the other hand, when topography leads to irregular variation of current, tidal cycle has to be discretised using several classes of tide conditions ; these conditions and their related weighting factors have then to be determined from computation of sediment transport at all the measurement points.

In the last case, the long-term (19 years) cycle of the tide has been discretised in about 20 classes of tide range with their associated occurrence frequency. For each class, hydrodynamic and morphodynamic computations over 1 tide have been performed ; bed changes for each class have then been combined, with weighting coefficients corresponding to the occurrence frequency of the class, to determine the reference bed evolution on one year (fig.1a).

Yearly tidal cycle has then been schematised using 1, 2 or 3 representative tides. The best "unique tide" (or so-called "morphological tide") was found to be in between mean tide and mean spring tide, leading to currents about 12 % higher than currents induced by mean tide ; this tide condition raises a relative error on the bed evolution - computed as the ratio of mean quadratic deviation between reference and filtered computations of bed changes to mean quadratic bed changes - of about 6 % (fig.1b). Use of 2 representative tides (neap and spring tides) has reduced significantly this error, to about 1.5 % ; on the other hand, use of 3 tides (neap, mean and spring tides) does not improve significantly the results : the error falls to about 1.2 %.

But numerical experimentation has also pointed out that another single tide, used together with a proper scaling factor, could lead to better results : as a matter of fact, 5 months of spring tide led to an error on annual bed changes close to 3 %, far less than for the "morphological tide" : it means clearly that the "morphological tide", which leads to the most accurate average transport rate, does not lead to the most accurate bed evolution patterns.

The above-mentioned experiments have shown that there are no automatic procedures for such input filtering, but at best some advices and some recipes. As a matter of fact, it must be the art of the engineer to determine, with regards to his specific problem, how to schematise the variability of the input, as it is done for reduced scale modelling. This work will be greatly helped by the existence and the in-depth use of a sufficient amount of current field measurements.

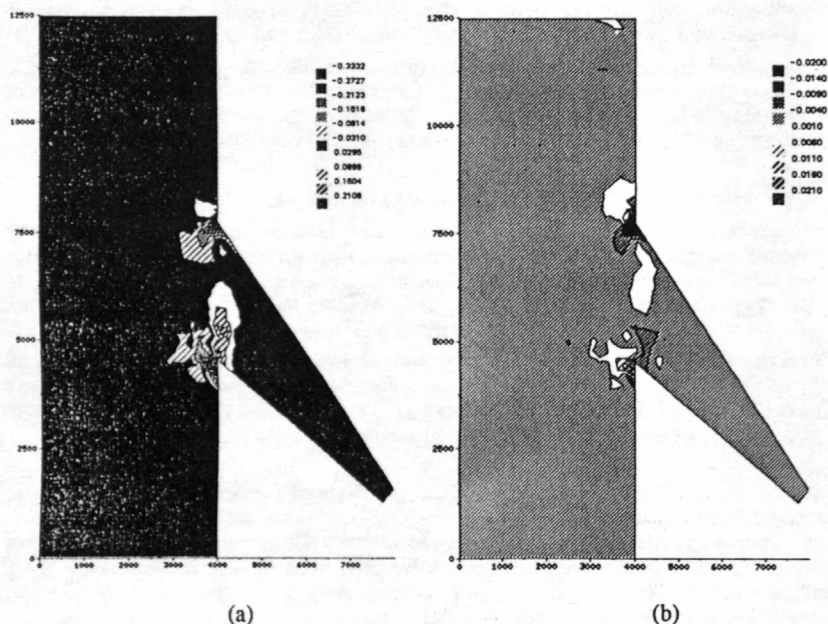


Figure 1 : Tidal input filtering on the case of the schematic estuary : (a) bed topography changes in one year, according to the reference case - (b) errors in bed topography changes when computed with a single representative tide.

Tidal process filtering

Independantly of input consideration, 2 problems appear in long term morphodynamic computations :

- flow pattern is modified as bed is changing, and should then be re-computed continually, leading to very heavy computation costs ;
- number of computational time steps should be huge if each involved tide had to be simulated.

For the first problem, schematic adjustment of the flow (no current deflection nor free surface changes ; "new" velocity is computed from initial flow discharge and from "new" water depth, given the bed evolution) can be used as long as bed changes are small enough compared to water depth.

As regards the second point, 4 methods have been developed at LNH to simulate the effect of repetitive identical tides by reproducing only a few of them (Latteux, 1990b) :

- extrapolation over N tides of the results computed on the first tide,

- lengthening of the tide : N successive tides are simulated by a single one, extended N times (and in which the time step can similarly be extended),
- centering in time, over the N tides, of the first method by a kind of Runge Kutta method,
- and first order expansion of the sediment discharge as a function of the bed evolution, assuming moreover no current deflection nor free surface level variation due to bed changes (schematic flow adjustment).

These methods have been investigated and compared on the 1-D case of a progressive wave over a complex topography, and, for some of them, on the 1-D case of a bar under a rigid lid and then on the previously described case of the schematic estuary.

The main results are as follows :

- such filtering methods can save a lot of CPU time (at least a factor 10 in the computation costs), without too significant an error.
- suitability of the various filtering methods depends on the phenomenon to be reproduced. As an example, the extrapolation method is not fitted for propagative cases, as the migration of bedforms, but more for evolution at more or less fixed place (erosion/sedimentation due to works ...).

Acknowledgements

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J. V. Letter, Jr.¹, B. P. Donnell², and N. J. Powell³

Atchafalaya Bay in Louisiana contains one of the continental U.S.'s largest and most dynamic deltas. Subaerial delta growth rates in the bay during the 1970's and 80's ranged from 8 sq km per year of land loss to 11 sq km per year of growth, with a net average growth exceeding 2 sq km per year. These changes have significant implications for aquatic and wetlands habitat, coastal flooding, water quality, and maintenance of navigation in the region, and obtaining realistic predictions of future delta growth has been a priority task for the U.S. Army Corps of Engineers.

One component of the Corps' effort to predict Atchafalaya Bay delta growth and its impacts has been unprecedented 2-dimensional numerical modeling of sedimentation for a 50 year period. The Corps' TABS-MD system of numerical models was used to describe hydrodynamics (currents and water levels caused by river discharges, tides, and waves), circulation and transport (salinity), and sedimentation (transport, deposition, erosion, bed modification, and subaerial growth). The large bathymetric changes occurring over the 50 year period required multiple updates of predicted hydrodynamic responses to delta development.

The models were verified to satisfactorily reproduce past observed conditions in Atchafalaya Bay, then predictions were made of delta growth and its impacts through the year 2030 with and without project features such as flood damage reduction levees. The models predicted that the delta will encompass over 300 sq km within 50 years and have a pronounced impact on the physical environment of Atchafalaya Bay and adjoining marshes. Human efforts can influence both the evolution of the delta and its impacts on the environment.

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LARGE SCALE SEDIMENT MOVEMENT IN A MACROTIDAL HIGH-ENERGY COASTAL AREA: THE FRENCH COAST OF THE ENGLISH CHANNEL

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A good knowledge of sediment movement is particularly important in order to estimate the current state of the coastline and to predict its future evolution in relation to the probable rise of sea level during the next decades. In a high-energy environment such as the French coast of the English Channel, the combination of strong tidal currents and wave action provokes two main effects: intensive coastal erosion with correlative coastal flooding, and strong infilling of estuaries. The first aspect has been extensively studied in the Channel Islands Bay, the second mainly in the Bay of Seine (fig.1). This paper briefly describes the main aims of these two complementary projects and gives some global results, with special attention to the large scale methodology used.

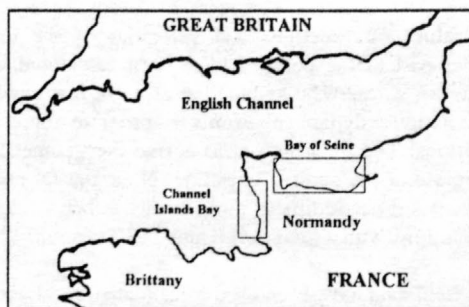


Fig.1: project sites

QUANTIFICATION OF SEDIMENT MOVEMENT ON THE UPPER FORESHORE OF BEACHES IN A MACROTIDAL ENVIRONMENT

The western coast of the Cotentin Peninsula is a linear low-lying sandy coast threatened by strong marine erosion. In order to understand and predict coastal changes, and to define coastal protections with limited negative impacts, large scale field investigations have been initiated to obtain the most accurate data for sediment transport models.

In the Channel Islands Bay, the propagation of swells coming from the North Atlantic is strongly modified by a complex bathymetry due to numerous shoals and islands. Consequently, along the beaches, longshore currents and sediment transport are alternately oriented southward or northward, closely depending on the direction of wave propagation. The morphological characteristics of coastal sedimentary structures, with many spits and longshore bars, provide a good pluriannual residual approach to estimate the directions of sediment transports.

Data on directional waves appear to be fundamental parameters for the analysis of sediment movement on beaches and for the understanding of coastal erosion. Field measurements of offshore directional waves during a one year period, at three locations on the foreshore area, show:

- the decreasing of wave heights from North to South along the western coast of the Cotentin Peninsula, with a maximum annual significant height of 4.36 m;
- the importance of swells coming from a west direction in a very narrow window;
- the bimodal distribution of wave energies with two main peak periods: 3-4 s and 9-10s.

The correct values of directions and velocities of the longshore currents occurring on the upper part of the beaches have been calculated using a model of shoreward propagation of waves. The knowledge of wave features has permitted the simulation of about 50 hydrodynamical events in order to represent mean annual hydrodynamic conditions. The main aim is to define the geometry of the different coastal cells occurring along the studied coastline. Nine coastal compartments have been identified. The related net sediment movement has been calculated to explain long term coastal evolutions with a modified Kraus L.S.T. formula.

Two kinds of field data have been used to validate our numerical model:

- firstly, large scale morphological observations obtained by comparison of vertical aerial pictures over the past 50 years,
- secondly, large scale field experiments using radioactive tracer techniques, which permitted to estimate the directions of net sand transport and to calculate the drift rates during a six month period.

Relating to the global sea level rise, it appears that the studied coastline consists of areas of about ten kilometers length, either in erosion or in sedimentation, separated by stable zones in which long term sediment dynamics is moderate. On a large scale, each of these coastal cells is characterizing by its own sediment movement, the main consequences being the construction of ebb and flood deltas, the infilling of small estuaries, the growth of spits and local spectacular coastline recessions.

ACCUMULATION OF SEDIMENTS WITHIN ESTUARIES

Estuaries generally constitute sediment traps for the storage of sediments supplied from both the land and the sea. During the holocene period, in the example of Bay of Seine, reworked marine sediments, originally derived mainly from rivers during Pleistocene period, have selectively filled the river valley. If the general distribution of bottom sediments on the shelf is largely the result of reworking of Quaternary deposits, the combined action of tidal currents and waves results in the mobility of offshore sands, which contribute with river inputs to create high rates of sedimentation within the different estuaries of the bay.

A large scale study has permitted to understand the present patterns of sand transport on the shelf, extensively analysed using different techniques:

- the general distribution of bottom sediments reflects the tidal current velocities, responsible for the development of sedimentary bedforms which have been used to postulate regional patterns of sediment movement. Different bedforms described following a side-scan sonar survey suggest a net sand transport directed toward the estuary mouths of the eastern bay.

- estimates of direction of net sand transport and drift rates have been made at different locations using radioactive and fluorescent tracer techniques. The observed tracer dispersion patterns are elongated according to the rectilinear nature of tidal currents. The net transport vector directions are strongly correlated with the asymmetry in peak tidal flows, i.e. in the direction of the dominant flood currents. The influence of waves also provoke locally a shoreward transport and more generally a longshore drift.

All data tend to show that the dominant bedload transport is oriented from offshore to offshore, contributing to the infilling of estuaries. At the present time, the estuaries of the Bay of Seine are almost entirely filled in, and the estuary mouth shoals form an exterior arc. When sand from the shelf reaches the estuary mouths, it is redistributed in different circulation cells by waves and tidal currents.

The comparison of hydrodynamic surveys available for the Bay of Seine since 1834 has shown an important increase of depth on the shelf correlated to the increase in sedimentation rate in estuaries, particularly in the Seine estuary. If the present sediment dynamic within estuaries depends mainly on local hydrodynamic conditions, the knowledge of long term sediment budgets closely depends on large scale data registrated as long as possible during periods of sea level rise.

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LARGE SCALE HIGH RESOLUTION COASTAL SURVEYING WITH THE SHOALS SYSTEM

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Large scale coastal behavior is important in the design of US Army Corps of Engineers (CE) beach erosion control and hurricane protection projects. Project limits are specified by the local sponsor usually represented by a city or county government. However, coastal processes and littoral cells are not delineated by political boundaries and modifications to an existing shore invariably effects conditions elsewhere in the cell. To better understand the effect a beachfill may have on a region, advanced numerical models are being developed to forecast regional scale changes using high speed supercomputers like the CRAY YMP which makes large scale modeling feasible. But these models are only as good as the data used to calibrate and verify them and regional scale field data are costly and time consuming to obtain. The CE is currently involved in a joint project with the Canadian government to design and construct a hydrographic survey system capable of rapidly collecting accurate hydrographic survey data.

The SHOALS (Scanning Hydrographic Operational Airborne Lidar Survey) system is being developed to augment the US Army Corps of Engineers existing fathometer-based hydrographic survey capabilities. The SHOALS system uses laser technology to perform accurate, high speed, hydrographic surveying of the coastal environment. The system is mounted on a Bell 212 helicopter and has an operational speed ranging from hovering to 100 knots at altitudes ranging from 100 to 1,000 meters. The laser fires at 200 Hz as it is scanned in an arc across the forward direction of the helicopter. Nominally, the aircraft will fly at approximately 40 knots and 200 meters altitude producing a laser scan width of 110 meters wide with individual depth soundings spaced 3 meters apart in both the longitudinal and lateral directions of the line of flight. With these performance specifications, 720,000 soundings can be collected each hour over an 8 km² survey area. A vessel equipped with a fathometer traveling at 6 knots would require approximately 240 hours to survey the same area at equal sounding density.

The SHOALS system will be field tested in May and June 1993 off Sarasota, FL. Sarasota was selected because it has a variety of environmental conditions in the bay including sand, mud, and vegetated bottom type, and varying water quality conditions. The offshore gulf region offers greater depths and waves to evaluate system performance on and within the region the Corps has constructed several beach erosion control projects and maintains several shallow and deep draft navigation projects that can be used in the field tests. This paper will present results of the field tests.

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FIELD MEASUREMENTS AND LARGE SCALE MORPHOLOGICAL CHANGES IN A TIDAL INLET SYSTEM IN THE WADDEN SEA.

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1. INTRODUCTION

Based on the sand sharing principle (Dean, 1988), it is postulated that changes in the coastal area, including beaches and nearshore zone, adjacent to inlets are correlated with changes in the inlet basin system. Of particular interest in this study are changes at the intermediate scale i.e time scales of decades and length scales that are typical for morphological units as outer delta, inlet and basin. The changes at these time and length scales can be thought of as the integrated results of smaller scale processes. However, the success of explaining changes at the intermediate scales using this approach is limited, as often the result of the integration depends on the time-sequence of the smaller scale events. Therefore a more promising avenue is to focus on the large scale coastal behaviour itself and try to establish empirical relations with the gross parameters characterising the driving forces i.e tidal currents and waves. The partial closure of the basin of the Friesche Zeegat (Figure 1), one of the inlet basin systems of the Dutch Wadden Sea, in 1969 offers a unique opportunity to study large scale morphological changes in response to a change in the hydrodynamics.

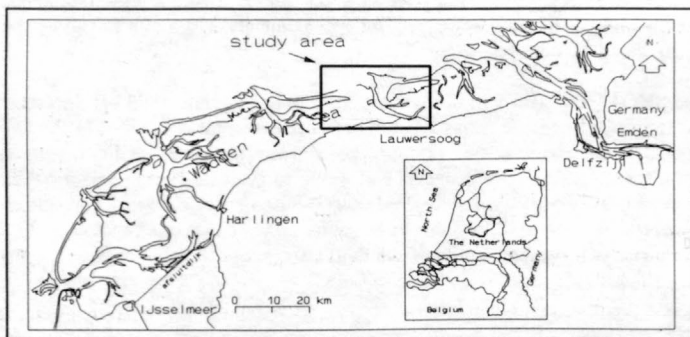


Figure 1. Study area Friesche Zeegat inlet.

2. OBJECTIVES

The objectives of this study are:

- (-) to determine how the various morphological units of the Friesche Zeegat have adapted to the partial closure of the basin and how the observed changes relate to the hydrodynamics;
- (-) to delineate the sediment transport mechanisms and;
- (-) to suggest ways to incorporate these findings in a Large Scale Coastal Model (LSCM).

3. RESEARCH METHODS AND ANALYSIS

Morphological changes in response to a change in hydrodynamics due to partial closure of the Friesche zeegat inlet have been studied based upon analysis of bathymetric and hydraulic data.

3.1 DATA SETS

Tide characteristics of the basin have been studied from data collected over the last thirty years at the station Lauwersoog - situated in the southern part of the tidal basin - and at a station lying in the vicinity of the inlet.

Bathymetric data are the basis of the analysis of morphological adaptation of the inlet system due to partial closure of the basin in 1969. Bathymetric surveys have been carried out nearly every 5-years since 1967.

In 1991 and 1992 field campaigns were carried out at several locations in the tidal inlet, tidal basin and outer delta. During 13-hours measurements on different vessels in a transect, sediment concentration as well as tidal velocities have been registered. During autumn of 1992, measurements of tidal velocities, waves and sediment concentration extending over a period of one month have been carried out at stations in the vicinity of the inlet.

3.2 ANALYSIS

Using a harmonic analysis the tidal constituents of waterlevel and current speed were determined. Especial attention was given to the Eulerian mean current and the asymmetry in the tidal current speed. Current speed and concentration were used to calculate sediment transport.

The bathymetric data was used to study changes in cross-sectional area and sediment volume at intervals of 5 years since the partial closure of the basin of the Friesche Zeegat. Analysis include sediment budgets for the system as a whole and the morphological sub-units, outer delta, inlet, basin and adjacent coast.

4. HYDRAULIC SETTING OF THE FRIESCHE ZEEGAT INLET

Wave records over the period 1979-1986 indicate that for prevailing W-N wind directions, the average significant wave height decreases in a landward direction varying from about 1.2 m in the nearshore zone (-20m NAP), to about 0.7m in the mouth of the inlet, to less then 0.5 m at Lauwersoog.

The mean tidal range varies from 2.2 m near the inlet, to ca. 2.4 m at station Lauwersoog. The maximum current velocities varies from ca. 1 - 1.5 m/s in the tidal channels, to ca. 0.2 - 0.4 m/s in the shallow areas of the flats and sandy shoals. The mean tidal prism of the basin after partial closure of the basin Friesche Zeegat is about 195 million m³.

Dominant winds generally occur during autumn and winter and are from west to northwest, with an average velocity varying from about 7 m/s near the coast to circa 5.5 m/s in the tidal basin.

The physical conditions of tide and waves indicate that the shore line along the Friesche Zeegat inlet system can be classified as a mixed energy to tide-dominated shore line, in which the combined influence of tidal and wave processes is responsible for the coast line morphology (Hayes, 1979; Fitzgerald et al., 1984)

5. RESULTS

HYDRODYNAMIC CHANGES

Partial closure of the basin of the Friesche Zeegat in 1969 has reduced the tidal basin area by about 40% (from 210 to 120 million square metres). The tidal prism was reduced from 320 million m³ to 195 million m³. The tidal current velocities in the basin were reduced by about 20-40%. Harmonic analysis of the vertical tide (at station Lauwersoog) indicate that the closure has not significantly affected the amplitude and phase of the dominant components M2 and S2 as well as the overtide M4.

MORPHOLOGICAL CHANGES

Since 1970, the observed changes in sediment volume of the inlet system indicate a significant positive sediment budget for the tidal basin and a significant negative budget for the outer delta. As a consequence of the changes in hydrodynamic conditions, the deeper parts of the tidal channels including the inlet are silting up. The major part of the sediment surplus consist of silt. Sofar little change has been observed in the elevation and extent of the tidal flats.

The observed geomorphological development suggests that on an intermediate time scale, the changes will continue at decreasing rates.

EQUILIBRIUM CONDITIONS

Using data of Wadden Sea inlets, Gerritsen and De Jong (1985), the equilibrium relation between cross-sectional area of the inlet gorge and tidal prism was determined. A similar exercise was carried out for the outer delta's of these inlets and the basin volumes. Observations in the Friesche Zeegat show a gradual return of the inlet cross-sectional area, volume of outer delta and basin to their equilibrium values. As state in the previous section the rate of adaptation appears to decrease in time.

SEDIMENT TRANSPORT

Analysis of the velocity and sediment concentration measurements are still in progress. In the analysis emphasis is on the type of sediment that is imported and the mode of transport (suspended versus bed load).

6. FORMULATION OF AN LSCM

A common assumption in formulating LSCM's for tidal inlet systems is that after a change in the driving forces or geometry, the system will return to some form of equilibrium. When in equilibrium unique relationships exist between the gross characteristics of various morphological units and those of the hydrodynamics. An example is the relationship between cross-sectional area of the inlet and tidal prism. For the Wadden Sea various equilibrium relationships were determined using information of those tidal inlet systems that are in equilibrium. From these relationships and observations since closure, it follows that outer delta, inlet and basin of the Friesche Zeegat are again approaching equilibrium conditions. The time scale of adaptation is several decades.

Short of prescribing the time scale of adaptation of the various morphological units, simulating the behaviour of the system with an LSCM during the transition period requires knowledge of sediment transport mechanisms. An attempt to delineate these mechanisms using long term measurements in the gorge of the inlet is described.

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GEOMORPHOLOGICAL ANALYSIS AND MATHEMATICAL MODELLING OF SEDIMENTATION AND COASTLINE STABILITY OF THE TIDAL INLET GRÅDYB

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1. INTRODUCTION TO THE AREA

Esbjerg Port - the largest in Denmark by extent of operations - is located in the Danish Wadden Sea in the SE part of the North Sea. The access to the port is through the tidal inlet *Grådyb* and a 12 km long tidal channel, cf. Fig. 1. The annual net SE-ward longshore transport along the coastline of the *barrier peninsula Skallingen*, in the order of 1 mill. m³/year, has caused the formation of a large bar crossing the tidal inlet with natural depths of only 3-5 m. The present dredged access channel over the bar has a width of 200 m and a minimum depth of 9.3 m ACD. The mean spring/neap tidal ranges are 1.7 m and 1.2 m, respectively. Storm surges of up to approximately 4 m are experienced.

The annual maintenance dredging in the bar area amounts to approximately 1.2 mill. m³/year. Most of the dredged material is disposed south of the channel in such a way as to reduce interference with the barrier system dynamics and thus prevent beach erosion at the southern barrier island Fanø.

During the last 20 years, the southern spit of Skallingen has receded approximately 1500 m, which has caused *moderately increased wave impact* on the otherwise protected coastline of the mainland. During the same period several washover fans have formed on the back barrier, especially on the northern and middle part of Skallingen.

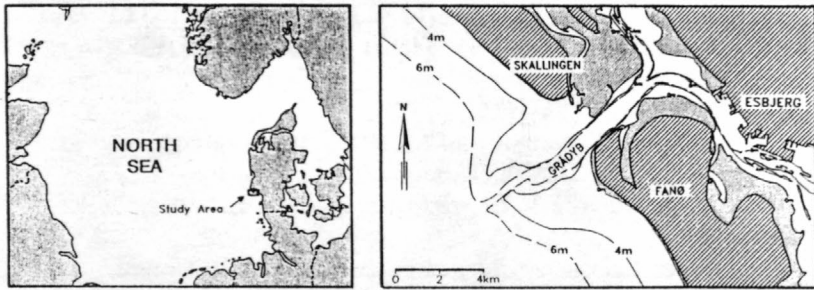


Figure 1: Location chart showing the North Sea, The Port of Esbjerg, and the tidal inlet Grådyb.

2. OBJECTIVES OF THE STUDY

The main objective of the study has been to investigate whether the maintenance dredging is carried out optimally. The influence of alternative channel alignments as well as introduction of structures have been studied.

Secondary objectives were to study the stability of the barrier peninsula Skallingen and its possible interaction with the navigation channel and to study developments in wave and water level conditions inside the tidal inlet.

3. METHODOLOGY AND MAIN FINDINGS

3.1 Overall Programme

The field programme included wave, current and tide recordings as well as sediment- and water sampling. Then, a 3-tiered layout of the investigation was followed in accordance with the study objectives; the geomorphological analyses and engineering analyses using advanced mathematical models were run in parallel supporting each other by their alternative study approach.

3.2 Morphological baseline studies

The investigation of the coastal geomorphology indicates an increased shoreline retreat during the last two decades. Basically, natural phenomena are responsible for this development. Beside a slowly rising sea level, a turn in the prevailing wind direction has occurred and there has furthermore been a significant increase in the number of gales after the relatively mild meteorological conditions during a period of more than 50 years, which finished around 1975.

For the time being Skallingen retreats faster than before, which results in a narrowing of the barrier width. But this is to be expected as a natural response to a higher impact of waves and high water levels.

It was demonstrated that breakthroughs of the shore parallel dune ridge and the successive development of washover fans were very soon replaced by foredune formations and built-up by small dunes on the distal inner margin of the fans. Concerning the total material budget of the barrier body, examinations of washover profiles demonstrated a slightly positive net sediment balance. Artificial enforced areas, however, unambiguously displayed a net sediment loss caused by the high and steep dune cliffs.

Historical coastline analysis of 300 years data shows that Skallingen has continuously been migrating landward at a mean rate of about 3 m/y. In spite of a recent shoreline retreat of more than 4 m/y, no imminent threat of destruction is estimated if the barrier is allowed to adapt dynamically to its changing environment, and if human interfering does not occur (such as groyne buildings and means for damping eolian processes).

The experienced shortening of Skallingen is to be attributed to the combined effect of natural barrier retreat and continuous maintenance of the navigation channel. The picture is somewhat confused by the construction and existence of a number of small groynes in the spit area. To some extent the retreat may be a delayed effect of a gradual increase in the Grådyb navigable depth from 6.7 to 9.3 m during the sixties.

3.3 Sediment Transport and Dynamics in the Bar Area

With respect to the channel, it was concluded from the mathematical modelling that no permanent relief of the maintenance dredging task

may be achieved. Minor and temporary effects are possible, for example by the establishment of a reservoir in the bar area North of the navigation channel; - or by an alternative, straight alignment South of the present navigation channel, which for historical reasons (no dredging during World War II) has two bends, cf. Fig. 1.

The southern channel alignment is the only alternative which is attractive with respect to both the morphology in the area and maintenance dredging requirements. In case an increased navigation depth is desired in the future, a southern alignment was recommended in order to minimize possible strain on the bar area and the Skallingen spit to the North.

3.4 Wave- and Water Levels Inside the Grådyb Inlet

Water level conditions were shown to be largely unaffected by morphological developments in the study area.

In the area surrounding the inlet channel, a noticeable increase in wave heights has occurred during the last 25 years, attributable partly to the shortening of the Skallingen spit and partly to the rough weather in the period. Overall, and particularly along the coast, the latter is clearly the major effect.

Despite the increased wave impact in some areas along the coast of the main land, the water level surges are the main threat in all areas and safety factors for protective structures instituted by legal requirements and professional codes of practices have not been undercut.

4. ORGANIZATION

The study was carried out jointly by the Danish Hydraulic Institute and the Institute of Geography, University of Copenhagen on behalf of the State Port Authority, Esbjerg, and the Coastal Authority under the recently established 'Coast-Port Collaboration', the main goal of which is to coordinate the interests of the ports and the coast administrations along the highly exposed littoral North Sea coast.

Subsequent to the described investigation a further study has been made on the environmental impact of deepening the Grådyb navigation channel to as much as 14.3 m.

GEOMORPHIC RESPONSE TYPES ALONG BARRIER COASTLINES: A REGIONAL PERSPECTIVE

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Long-term monitoring of shoreline position enables researchers to better understand process-response relationships that shape regional coastal morphology. Certain geomorphic response types are ubiquitous for different barrier beach settings, whereas other types are unique and appear to operate only in selected areas. Depending on temporal and spatial scales, beach response is driven by: (1) short-duration, high-energy, low-frequency processes (e.g., storms, tectonic activity), (2) long-duration, lower-energy, higher-frequency processes (e.g., normal wind waves and astronomical tides), (3) relative sea level rise, or (4) a combination of all factors. Based on quantitative documentation of historical changes in shoreline position in response to natural and human processes, seven response types were developed for classifying barrier coasts, including: 1) *in-place narrowing*, 2) *landward rollover*, 3) *in-place break up*, 4) *lateral accretion*, 5) *progradation*, 6) *beach/headland erosion*, and 7) *dynamic equilibrium*. This paper addresses the spatial distribution of these geomorphic response types and identifies similarities as well as differences among Louisiana, Mississippi, and Georgia barrier coasts.

Historical shoreline positions between the 1840s and 1991 form the extensive data base in which geomorphic response types were developed for the Louisiana, Mississippi, and Georgia barrier beach systems. In Louisiana, low-profile, washover-dominated barrier shorelines retreat landward at average rates that commonly exceed 10 m/yr (McBride et al., 1992). This coastal setting is influenced by numerous storm events, diminishing sediment supply, intense human disturbance, and high rates of relative sea level rise (~ 1 cm/yr). In contrast, the Mississippi barrier island coast is undeveloped with no structured inlets and is characterized by medium-profile, moderately stable barrier islands with an average retreat rate of about 1.7 m/yr (Byrnes et al., 1991). In Mississippi, lateral island migration to the west controls shoreline evolution along Dauphin, Petit Bois, Horn, and Ship Islands as a result of dominant longshore sand transport processes, whereas cross-shore change in shoreline position is the primary mechanism by which beaches on Cat Island respond to incident processes. Compared to Louisiana and Mississippi, Georgia is characterized by high-profile, stable barrier islands that are short and wide, with moderate human disturbance. Shoreline change rates along the Georgia coast have averaged about ± 1 m/yr over the last century (Griffin and Henry, 1983; Byrnes and Hiland, in press). The Georgia barriers tend to have stable central shorelines with most fluctuations occurring adjacent to tidal inlets.

The seven geomorphic response types are based primarily on the direction and magnitude of shoreline movement along seaside and bayside shorelines in response to incident processes. Figure 1 is a generic model of geomorphic change as a function of

shoreline movement. *In-place narrowing* (A) characterizes islands that experience seaside and bayside erosion. As a result, the central core of the island remains stationary. This response type has been documented for barrier islands along the northern Gulf of Mexico and the U.S. Atlantic coasts (Leatherman, 1983; McBride et al., 1991). *In-place narrowing* tends to evolve into either *landward rollover* (B) or *in-place break up* (C). *Landward rollover* is characterized by movement of the seaside and bayside shorelines in a landward direction as a result of washover processes (Leatherman, 1979; Byrnes and Gingerich, 1987; McBride et al., 1993). *Landward rollover* appears to evolve from *in-place narrowing* once an island has reached some critical width that enables washover processes to maintain and translate the barrier island landward. Shorelines that experience *landward rollover* tend to be lower in profile and migrate landward at high rates, some as high as 30 m/yr.

In-place narrowing can also evolve into *in-place break up*, a type of barrier island response described for certain deltaic barrier islands in Louisiana (McBride et al., 1991). As an island narrows, it becomes more susceptible to breaching during storms. Due to inadequate sediment supply, the breaches are unable to close and become permanent inlets that trap littoral sediment. Therefore, instead of maintaining continuity, the island starts to break up and deteriorate in place as inlets continue to cut and widen (Figure 1b). Washover processes are important but are not able to translate the barrier island system landward, because an ever-increasing amount of sand-sized sediment is lost from the system and stored in the tidal inlets. This characteristic differentiates these kind of deltaic barriers from other typical Gulf and Atlantic coast barrier islands.

Progradation (D) defines a coast where the shoreline is translating seaward in response to an excess in sediment supply. The bayside shoreline tends to be stable in these settings. In contrast, *beach/headland erosion* (E) indicates a shoreline that is undergoing landward and/or lateral movement due to the removal of sediment, thus providing a sediment source to downdrift beaches. *Lateral accretion* (F) is characterized by shore-parallel sediment transport. This response type is associated with the ends of spits and beach ridges that form adjacent to tidal entrances. Typically a zone of erosion is located updrift of these sites. Finally, *dynamic equilibrium* (G) defines a shoreline that exhibits long-term stability. Although shoreline movement fluctuates about a mean, the magnitude of change associated with this particular response type is not significant enough to detect an erosional or accretional trend.

In general, all seven geomorphic response types mentioned above are found along Louisiana's barrier shoreline. The three most common types are *landward rollover*, *lateral accretion*, and *in-place break up*. The Mississippi barrier island coastline is dominated by *lateral accretion*, *dynamic equilibrium*, and *beach erosion*, whereas in Georgia, *progradation*, *lateral accretion*, and *dynamic equilibrium* are the primary response types. Geomorphic response types documented in this study are expected to be applicable to other barrier shorelines around the world.

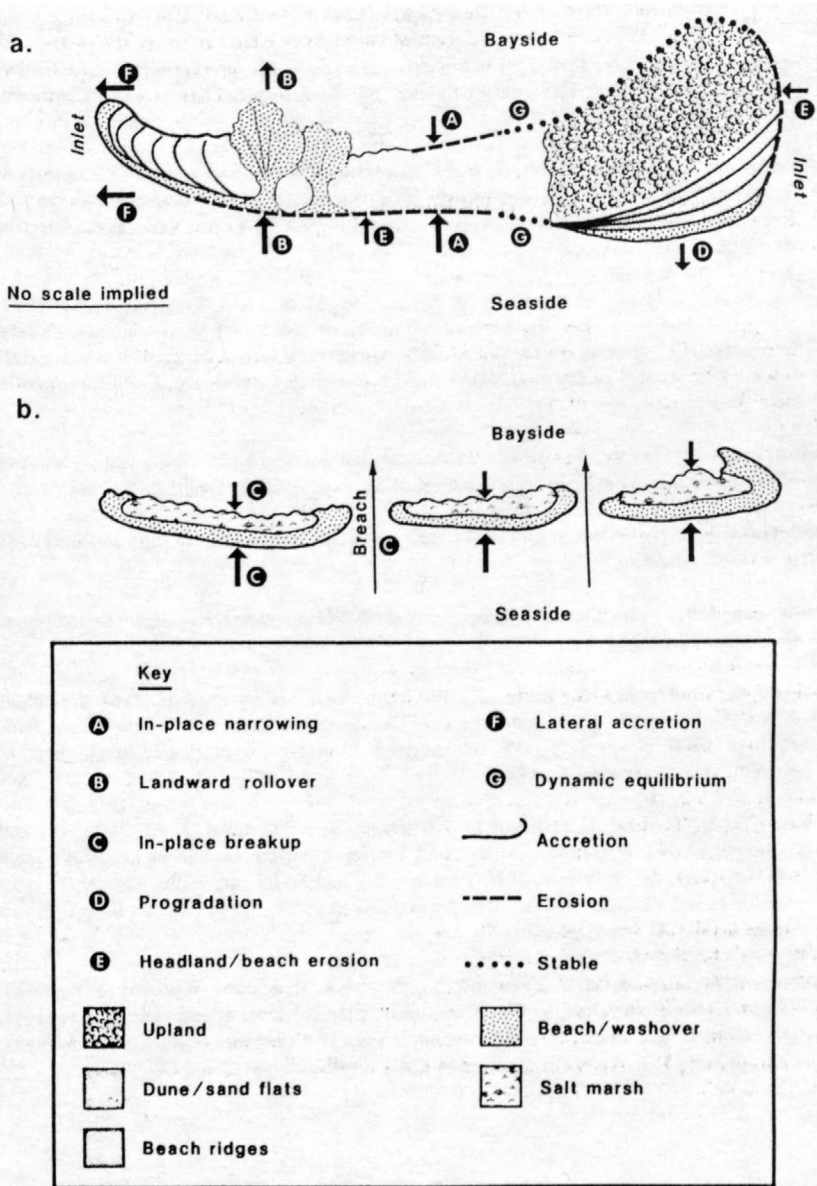


Figure 1. Vector representation of shoreline movement along barrier coastlines.
 a. Generalized geomorphic response type model. b. Component of the model depicting evolution from *in-place narrowing* to *in-place breakup*.

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Background Erosion Determined by Three-Way PCA Method

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1 Introduction

The evaluation of long-term shoreline recession due to background erosion is affected by short-term processes, such as seasonal changes and by small scale processes such as longshore rhythmic formations or cross-shore sand movements. In order to estimate those long-term processes, it is necessary to separate the spatial and temporal scales of variability of a beach.

Principal Component Analysis (PCA), also known as Empirical Orthogonal Eigenfunction (EOF), is a technique of linear statistical predictors which can be used to objectively separate the spatial and temporal scales of variability of a beach.

The method allows us to represent a large number of data variables by a few spatial, $c_n(s)$ and temporal $f_n(t)$ empirical orthogonal eigenfunctions which describe most of the variance of a data set $y(s, t)$ by:

$$y(s, t) = \sum_n f_n(t) c_n(s) c_n \quad (1.1)$$

where c_n is a normalizing factor. The eigenfunctions are ranked according to the percentage of the variance defined as the mean square value (MSV) of the data they explain, so that the first eigenfunction explains most of the MSV of the data.

This technique has been previously applied to cross-shore beach profile data: Winant *et al.* (1975), Aubrey (1978), Dick and Dalrymple (1984), Medina *et al.* (1991), and to longshore profiles data: Losada *et al.* (1991), Liang and Seymour (1991).

Recently, Medina *et al.* (1992), presented a Three-Way PCA Method in which both cross-shore and longshore variations are retained as well as time. Although standard EOF may be correct for some particular applications, the utilization of a particular profile or bathymetric contour line for evaluating background erosion is inadequate when two dimensional movement of sand is expected. In this case

beach profile data are 'contaminated' by longshore sand movement information and a method which can separate those spatial modes of variability must be applied.

In this paper, the background erosion of Playa Castilla (Huelva, Spain) is analysed by means of the Three-Way PCA method.

2 Bathymetric Data Set

Playa Castilla is located in the Southwest Coast of Spain (Province of Huelva), between the Guadiana and Guadalquivir rivers (fig. 1). The beach extends over 25 km between Masagón and Matalascañas and field data obtained during the last 30 years indicated that the coastline was receding at a rate of 1.5 m/year (Fernández *et al.*, 1990). An artificial nourishment of the beach of more than 1,500,000 m³ of sand was pumped to the updrift edge of the beach forming a protruding area of about 2 km length and 115 m width.

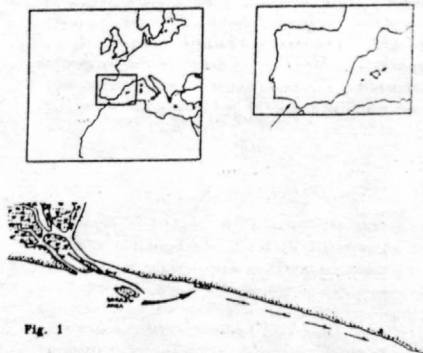


Fig. 1

In order to study the beach evolution through time and space, a field measurement Program start-

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ing in 1989 is being carried out. One of these sets of data is bi-monthly beach transects each 500 m over 25 km along the beach up to a depth of -10 m.

3 Analysis Method

Historically, PCA has been carried out for data which depend on two dimensions. If more dimensions were involved, data aggregation or other techniques were used to reduce the problem to a two dimensional problem. Solutions for three-way data were first proposed by Tucker (1966) and extended by Kroonenberg and DeLeeuw (1980) and Berge *et al.* (1987).

A detailed discussion of the method can be found in the paper by Kroonenberg and DeLeeuw (1980), in brief, one seeks a factorization of a three data matrix Y , such that:

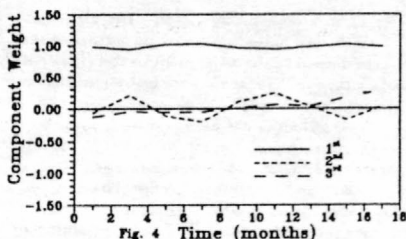
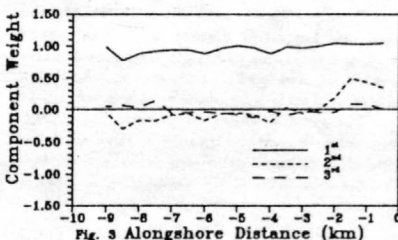
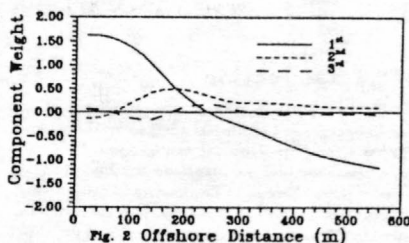
$$y(x, y, t) = \sum_{p=1}^P \sum_{q=1}^Q \sum_{r=1}^R [c_p(x) f_q(y) g_r(t) c_{pqr}] \quad (3.1)$$

where the coefficients g_{ip} , h_{jq} and e_{kr} are the elements of the columnwise orthonormal matrices G , H , E , respectively, and c_{pqr} are the elements of the so-called three-mode core matrix, C . The matrices G , H and E have a similar interpretation as the two-mode eigenvectors. The core matrix, however, is no longer a diagonal matrix of eigenvalues. The solution of the problem is based on the observation that the optimal C matrix can be expressed uniquely and explicitly in terms of the data and the component matrices for the three modes. The latter components matrices are optimized by an alternating least squares algorithm.

4 Results

When the three-way PCA method is applied to the bathymetric data set, the temporal and spatial (longshore and cross-shore) scales of variability are separated. In figures 2, 3 and 4, the first three cross-shore, alongshore and temporal eigenfunctions are shown. Observing figure 4, it can be seen that the first temporal eigenfunction is almost constant in time and, consequently, is associated with the mean bathymetry of the beach. The second temporal eigenfunction shows a seasonal dependence. The third temporal eigenfunction shows a net trend. This trend is related with the background erosion

and allows us to estimate the shoreline recession at different locations of the beach.



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Large-Scale Behaviour of Poland's Coastal Areas

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1 Field Data Collection and Monitoring System

Routine data has been collected at the PAS Institute of Hydro-Engineering IBW PAN Coastal Research Station, situated on the Polish coast of the Baltic Sea some 80 km from Gdańsk. The coastal zone belongs to multi-bar dissipative ones, with an average slope of 1–1.5% and sand quartz of grain size $D_{50}=0.22$ mm and density $\rho_s=2650$ kg/m³. Usual maximum storm waves have $H_s=3.5$ –4.0 m and $T_s=7$ –8 s at the seaward boundary of the surf zone ($h \approx 7$ m), primarily from N–NW sector.

Along with parameters of wind, waves, currents and other hydrologic factors, topographic features have been measured there since 1983 on a 2.7-km beach and nearshore zone extending some 800 m from shoreline. The beach profiles have been arranged every 100 m. The first systematic and mutually compatible records of beach and shore topography date back to 1964, and echosoundings plus tachimetry are continued until now.

In addition to Lubiato, a few other sites have also been logged routinely by the IBW PAN staff, such as Bulgarian Black Sea or Senegalese coast off St. Louis, where one-year monitoring scheme aimed at complex shore protection along tens of kilometres of *Lange de Barbaries* consisted i.a. of measurements of waves, currents, sediment transport and coast topography and bathymetry change.

A separate database stems from an extensive programme of retrieval and analysis of historic coastline changes relating to the Baltic coast as long ago as 13th century. Various segments of the present Polish coast, both sandy dunes and semi-cohesive cliffs, are then amenable to examination of the extent and rate of erosion, and some other quantitative and qualitative characteristics of long-term shore behaviour. Particular attention is drawn to the unique spit feature of 30 km long Hel Peninsula, which borders the Gulf of Gdańsk and is subject to centuries long regression and transgression due to combination of longshore and cross-shore sediment transport factors, partly induced by anthropogenic activities.

Hence we have insight into a considerable bulk of reliable data stretching over a reasonably long span of time.

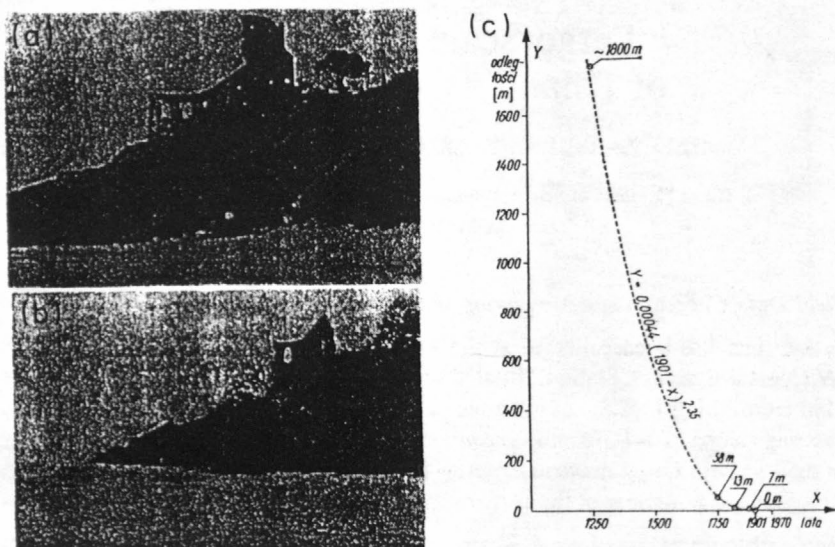


Figure 1. Ruins of church at Trzęsacz (western Polish coast) in 1923 (a) and 1971 (b) and the retrieved erosion rate (c).

2 Raw Data, Primary Estimates and Inherent Errors

Several examples of reported shoreline migration can be given. Spectacular ruins of a church at Trzęsacz, now at the edge of cliff, belong to the structure raised some 1800 m from shoreline in 1250. Continuing erosion forced the parishioners to give up their sanctuary in 1874. About 1923 there was still one wing of the ruins hanging over the cliff (Fig. 1a) while since about 50 years later (Fig. 1b) until now the erosion has come to very low figures. From overall historical evidence one can draw conclusions on the decreasing rate of erosion (Fig. 1c).

Two more cliffs on the western coast and three other ones on the east are analyzed on the background of cartographic material stemming primarily from 19th century. Erosion rates are reported and anthropogenic effects are blamed responsible for the irreversible loss of land (to the benefit of quite stable bar systems at some locations).

Topographical transformation of the Szczecin Bay (Stettiner Haff) since 1720 has been spurred by political changeover between Sweden and Prussia, who shaped their own independent waterways through the bay towards the Baltic Sea (Gulf of Pomerania). Dramatic changes in hydraulic gradients have led to a cascade of hydrographic and lithodynamic events.

Other cases are illustrated more fully in the final paper.

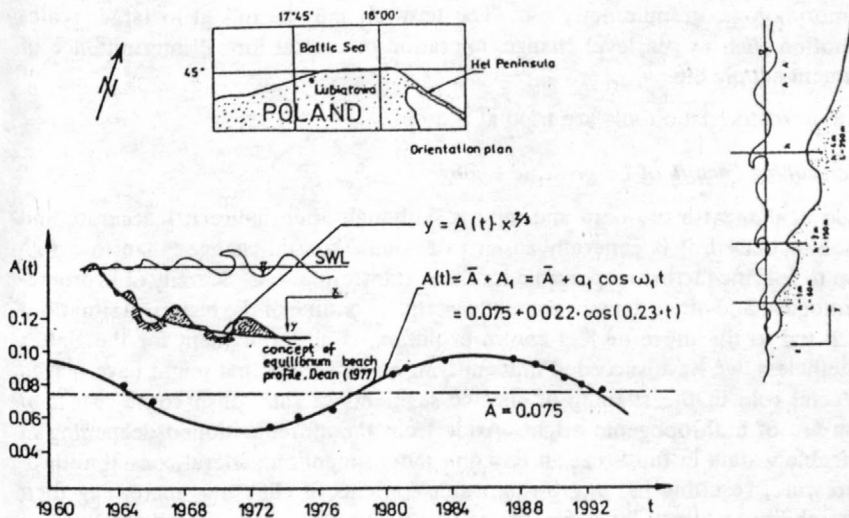


Figure 2. Temporal Variability of Parameter A of Dean Profile

In every analysis of large-scale coastal data, particularly those resulting from cartographic materials, utmost attention should be paid to errors involved in both measuring techniques, data processing algorithms and analytical methods. Various statistical estimates can be provided depending on the type and scale of physical processes.

3 Large-Scale Change Assessment Methods

For our database at hand we have attempted different approaches. *Statistical* and *spectral* tools have permitted us to identify the typical scales and their contribution to the overall changes. *Inter alia*, the celerities of alongshore propagation of macrocusps have been assessed to vary from 2 km in Senegal to 1 km at Lubiatowo. Some change modes were found to correlate with others and some have appeared mutually independent.

Empirical orthogonal functions (EOF) have been harnessed to identify cross- and long-shore locations of the most pronounced seabed changes, most intensive sediment transport rate etc. The concepts of *Dean profile* and *EOF* have been combined in yet another approach where the parameter A in the shore profile $y = Ax^{2/3}$ is postulated as a function of time — $A = A(t)$. Its two primary components are found to vary as in Figure 2. The component \bar{A} is regional and depends on site

geomorphology, granulometry etc. The term A_1 may be linked to larger scales of motion such as sea level change, migration of coastal forms, intermittence of sediment supply etc.

Some new stochastic tools are tried at present.

4 Controlling Factors of Large-Scale Change

While dealing with old data and reports, although often indirect, inaccurate and somehow biased, it is generally easier to evaluate coastal changes than to assign them to specific factors responsible for those transformations. Scarcity of hydrometeorological and other data, or even of descriptive outline of the historical situations which led to the more or less known evolution, is all too evident for the Baltic. Nonetheless, we have succeeded in identifying some factors that might have played a crucial role in the shaping of specific segments of the Polish coast. Most of them are of anthropogenic origin. Aside from the aforementioned deepening of navigable waters in the Szczecin Bay one should mention general coastal anthropopressure, resulting i.a. in growing water contents of cliffs thus increasing their vulnerability to attack by storm surges and waves (tides being negligible). Construction of harbours and coastal engineering structures at some locations is quite obviously attributable to lee erosion caused by a variety of hydrodynamical factors, in the scales of decades at least. This is illustrated by the cases of Władysławowo (a port built for geopolitical reasons between the World Wars), Kołobrzeg, Ustka and a dozen of other harbours.

On the other hand, it is much more difficult to identify clear climatic causes of coast evolution. A series of disastrous storms and storm surges at the turn of the century is thought to have brought about specific configuration of the semicohesive seabed off Jastrzębia Góra (E coast). Some indirect evidence is presented to support the hypothesis on that linkage.

Wave climate variability and sea level change are tracked back for some coast segments in an effort to prove they are primary causes of large-scale evolution, on which yearly, seasonal and diurnal changes are superimposed.

5 Modelling Approach

The concept of beach and shore equilibrium has been analyzed as to whether it can be applied to large-scale modelling. Coastal climate has been quantified with respect to the controlling factors in various time scales. Characteristic beach and shore features are discriminated as to their weight in the measurement and prediction of coastal changes.

A mathematical model has been derived from the concepts of mass conservation and long-term dynamic equilibrium, for a control volume encompassing various types of shore profiles. Its present version, sensitivity and other aspects will be presented in the final paper.

LARGE-SCALE TRANSFER OF SAND DURING STORMS: IMPLICATIONS FOR MODELING AND PREDICTION OF SHORELINE MOVEMENT

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Updrift beach and shoreface erosion are components of the sediment budget equation that are difficult to quantify because much of the dynamic profile is subaqueous and seldom is the depth of closure reached during beach surveying. Beach and dune erosion is becoming a more important source of sand to stabilize downdrift beaches as the primary sand sources such as rivers, offshore bars, and tidal deltas are depleted or artificially reduced. Large-scale stability of the shore is partly controlled by intense storms that transport large volumes of sand moderate distances in brief periods. During these events, the balance of sand supply to nearby beaches is dramatically altered. Because of the temporary imbalances in sand supply, it is difficult to predict the post-storm location of erosion or deposition or to anticipate the volumes of sediment involved. Consequently, the ability to model and to predict future shoreline positions based on short-term data may be seriously impaired.

In 1983, Hurricane Alicia eroded more than 1.5 million m³ of sand from the beaches and dunes along 40 km of western Galveston Island and northeastern Follets Island, Texas (fig. 1). Within 2 years after the storm, about 60% of the sand eroded from Galveston Island could be accounted for in washover deposits (12%) and sand returned to the beach in the form of beach aggradation and dune reconstruction (48%). The remaining sand was considered to be either deposited on the inner shelf and lost from the littoral drift system or transported to the southwest by storm currents and deposited on the shoreface. Sedimentological studies of the inner continental shelf conducted immediately after the storm suggested that most of the subaqueous sand was still on the shoreface and not on the inner shelf (Morton, 1988). Beyond this inference, no data were available to locate the former beach sand or to map its alongshore distribution.

Comparisons of beach profiles and aerial photographs of Galveston and Follets Island since 1983 show a range of beach responses including continuous erosion, partial recovery and stability, and continuous accretion. The site of continuous accretion (Follets Island) is where some of the storm-emplaced sand has been transported onshore for the past nine years. This site is located about 10 to 20 km southwest from the areas of maximum beach erosion. Before Hurricane Alicia, the beach of Follets Island was stable (short term) to moderately erosional (long term) depending on the length of historical record analyzed. Since Alicia, the beach of Follets Island has accreted more than 60 m and aggraded between 0.75 and 1.4 m.

Shoreline recovery following Hurricane Alicia occurred in four time-dependent stages. The first stage of recovery, which lasted for about one year from the fall of 1983 to the fall of 1984, involved berm deposition. During this stage of recovery, Galveston Island beaches between the seawall and a point about 8 km east of San Luis Pass experienced rapid berm accretion at a rate of 10 to 22 m³/m/yr and beach runnels formed. Throughout the second phase of recovery, berm crests aggraded and continued to advance seaward, but sand was also transferred from the forebeach to the backbeach. During the fall and winter of 1984/85, storms moved sand landward from the berms, filled in the runnels, and aggraded the backbeach. For the next year, the backbeaches continued to aggrade and the shorelines advanced again. The shifting of profile configuration from a berm runnel with low backbeach elevations (stage 1) to a level berm configuration (stage 2) was profound for the recovery of these beaches because foredune reformation could not occur until the berms widened and raised a sufficient amount (Morton and Paine, 1985).

The third and fourth stages of recovery, which involved foredune reformation and stabilization, lasted two to three years until 1987/88. During this period, backbeach elevations continued to increase through eolian deposition and foredunes developed at the location of the post-Alicia erosional escarpment. Later the dunes were colonized by dune vegetation and they continued to expand above and seaward of the post-Alicia erosional escarpment. During recovery stages 3 and 4, the rate of sand accumulation slowed to about 6 m³/m/yr and was largely in the form of eolian deposition.

Four years after Alicia, continuous foredunes were present and rose about 1.5 m above the backbeach at most of the monitoring sites. However, some of this dune accumulation was aided by placement of sand and organic debris scraped from the beach. At the end of the four-year post-storm recovery period, the beaches between the seawall to about 8 km east of San Luis Pass regained about 49% of the sediment

eroded by Hurricane Alicia. However, beaches on Galveston Island within about 8 km of San Luis Pass did not undergo post-Alicia recovery stages. In 1987, four years after Alicia, these beaches had 8 m³/m less sand than measured during the first post-Alicia survey in December, 1983.

Since Hurricane Alicia, the beach at profile 8 on Follets Island (fig. 1) has undergone the same general stages of recovery as the beaches on Galveston Island except the Follets Island site has continued to accumulate sand as a result of beach accretion, beach aggradation, and dune formation. This is the only beach that has systematically gained volume during each monitoring period since 1985. Much of the sand that continues to accumulate on Follets Island was eroded from Galveston beaches and the ebb-tidal delta at San Luis Pass during Alicia. Strong alongshore currents during the storm transported the sand to the southwest and deposited the sand on the shoreface where it has continued to supply the accumulation on Follets Island.

The post-storm recovery period lasted about four years and then the beaches began responding to other more dominant processes and changes in sand supply. All the beaches on Galveston Island experienced lows in sand volume around 1988/89. The high incidence of wave energy during Hurricanes Gilbert and Chantal caused the wide-spread erosion. After the 1988/89 low in sand volume, Galveston beaches evolved in several ways. The eastern section within 5 km of the seawall eroded, and by 1993 had 17 m³/m less sand than after Alicia. This erosional beach response is at least partly caused by the lack of updrift sand supply, which is limited by the Galveston seawall and groin field. Beaches to the west alternated between accretion and stability beginning with profile 2, which accreted, profile 7, which remained stable, profile 3, which accreted, profile 5, which remained stable, and profile 4 near San Luis Pass, which showed the greatest amount of accretion.

The alternating pattern of beach stability and accretion along Galveston Island reflects the development of new or alteration of existing shoreline rhythms since 1989. The wave length of the rhythms is not discernible from the profile data because of the wide profile spacing. These rhythmic features involve volumetric variations of as much as 40 m³/m of sediment.

Beach responses adjacent to San Luis Pass (profiles 4 and 8) are affected by the interaction of the beach with the shoals and tidal channels of the ebb-tidal delta. The large and continuous accretion of the beaches on Follets Island within a few kilometers of the Pass may be caused by episodic ebb-channel switching, shoreward movement of ebb-tidal delta sand during Hurricane Alicia, shoreward movement of Alicia-

emplaced shoreface sand eroded from Galveston beaches, or various combinations of these sand sources. Beaches on the Galveston Island side of San Luis Pass are probably interacting with marginal flood channels.

The rapid large-scale alongshore transfer of sand during Hurricane Alicia (hours) and the prolonged systematic onshore transport of sand from the shoreface to the beach at Follets Island (decade) are poorly understood. Nevertheless, if numerical models of shoreline movement and predictive capabilities are to improve, then changes in beach volume at the myriameter spatial scale and decadal temporal scale must be resolved.

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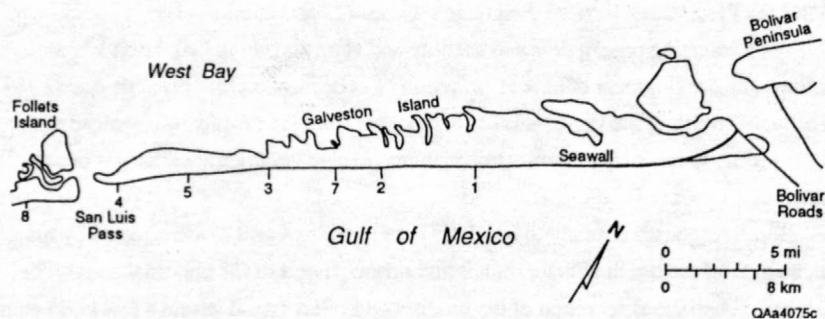


Figure 1 - Location of study area and beach profiles.

COHESIVE SHORES AND LARGE SCALE COASTAL EVOLUTION

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There are two primary shore types which are susceptible to considerable evolution through erosion: sandy shores and cohesive shores. A recognition of the fundamental differences in the erosion processes for these two shore types is essential for understanding the large scale coastal evolution of cohesive shores.

Sandy shores are generally distinguished by an inexhaustible supply of beach sediment. Large scale coastal evolution is primarily related to net erosion or deposition and can be linked to both alongshore and cross-shore gradients in transport integrated over a long period of time.

In contrast, a shore is defined as cohesive when a cohesive sediment substratum (such as, glacial till, glacio-lacustrine deposits, soft rock or other consolidated deposits) occupies the dominant role in the change of the shoreline shape (e.g. through erosion). In other words, under the sand beach there lies an erodible surface which plays the most important role in determining how these shorelines erode, and ultimately, how they evolve in the long term. Often the only visual distinction of a cohesive shore (above the sea or lake level) is the existence of a backshore bluff or cliff. The bluff may be as low as 0.3 m, in the form of a wave cut terrace, or as high as 50 m or more. Typically, the maturity of the vegetation on the bluff face provides some indication on whether the shore is indeed eroding. Wherever shoreline or bluff recession is a concern, there will be little in the way of vegetation on the bluff face. At the base of the bluff there is often a sandy beach, and offshore, there may even be sand bars. Other evidence pointing towards the cohesive shore classification may include the exposure of the cohesive layer beneath the beach during severe storms or the exposure of the cohesive substratum in the troughs between the offshore bar crests.

A cohesive shore erodes and recedes because of the permanent removal and loss of the cohesive sediment (both from the bluff and the lake or sea bed). The sand cover may come and go (depending on the season, water level and storm activity), but the erosion of the cohesive layer is irreversible. The cohesive layer is often a glacial deposit and derives its strength from the cohesiveness of the clay in its composition and/or through the compression it was subjected to during the period of glaciation. Once this material is eroded by the action of waves, it cannot reconstitute itself, the cohesive form is lost forever. Furthermore, any beach sand that may be a by-product of the erosion of the cohesive sediment is usually quickly swept away (glacial tills are generally comprised of less than 30% sand and gravel, and often less than 10%). Most cohesive shores are subject to strong alongshore drift which has the potential to remove large quantities of sand during storms and does not allow stable and protective beaches to build up. Where very large amounts of sand (in the order of 200 m³/m) do build up over the cohesive substratum - either at a change in the shoreline orientation or at a large

obstruction such as a natural headland or a harbour breakwater (groynes will only have a minimal influence) - the erosion of the cohesive shore and the backshore bluff will be arrested. Where this occurs, the shoreline reverts to a sandy shore classification since the presence of the underlying cohesive material is no longer an important factor in the evolution of the shore.

The critical point to understanding the evolution of cohesive shores is that the bluff recession could not continue without the ongoing downcutting of the nearshore lake or sea bed. The long term average rate at which the bluff or shoreline recedes on a cohesive shore must be governed by the rate at which the nearshore profile is eroded or downcut.

Cohesive shores make up a large part of the Great Lakes shorelines with some continuous segments of over 100 km in length (i.e. the north central shore of Lake Erie). Other examples of cohesive shores include: a large proportion of the North Sea coast of England; sections of shoreline on each of the U.S. ocean coasts; and along the Black Sea coast. Many other examples throughout the world, including erodible rocky coasts, are cited by Sunamura (1992).

Planning for the future evolution of cohesive shores is conventionally based on an extrapolation of historic erosion rates (from the comparison of recent and historic aerial photos and surveys). Improvements to the predictive capability for the large scale and long term evolution of cohesive shores must recognize the following constraints: 1) the importance of variations in the erosion resistance of the cohesive material (both across the shore and along the shore) - large headlands or points consist of more erosion resistant till compared to adjacent shores which erode at a greater rate; 2) the future movement and distribution of the overlying cohesionless sediment and the interaction with adjacent sandy shores - points or headlands can also be attributed to the deposition of sand and the protection of the underlying cohesive material from any further erosion; and 3) changes to the wave and water level conditions.

Considering these constraints, Nairn (1992) describes the application of a numerical model for the prediction of cohesive shore evolution over periods of up to 100 years for sites on the Great Lakes. The background of the numerical model development is discussed in the following paragraphs. The fundamental erosion processes on cohesive shores have also been recently studied in a unique series of wave flume experiments with undisturbed glacial till samples (Bishop et al, 1992). Kamphuis (1987) has also described an approach to the modelling of cohesive shore evolution for the Great Lakes and a quasi-3D model application has been presented by Sir William Halcrow and Partners (1991) for sections of the North Sea coast of England. An example of the result of this type of modelling is shown in Figure 1, which provides a comparison of observed versus predicted profile evolution over a 83 year period for a site on the north shore of Lake Erie (see Nairn, 1992 for details). The shoreline moved inland about 100 m over this period.

A numerical model has been developed to simulate the processes that occur on a sandy beach profile subjected to wave action (see: Southgate and Nairn, 1993; Nairn and Southgate, 1993). These processes include: 1) random wave transformation (including shoaling, breaking and decay); 2) hydrodynamics (including orbital velocities, undertow and longshore currents); 3) sediment transport (bed load, sheet flow and suspended load in the cross-shore and alongshore

directions); and 4) profile adjustments due to gradients in the cross-shore transport pattern. Each of the processes is evaluated at approximately 200 finite difference calculation points across the profile. This process based numerical model developed for sandy shores has been adapted for application to cohesive shores. Within the input description for the profile characteristics, a second profile shape may be specified corresponding to the cohesive substratum, thus allowing for a description of an overlying sand veneer. The sediment transport and profile change routines pertaining to the overlying cohesionless sediment are modified to address the possibility of supply limited sediment transport. For example, where the sand cover is very thin, the predicted potential cross-shore transport at a specific location may be unattainable if the gradient in transport between two adjacent calculation points implies more erosion than is possible due to the presence of the cohesive layer. In these cases, the potential transport rate is reduced to the supply limited level.

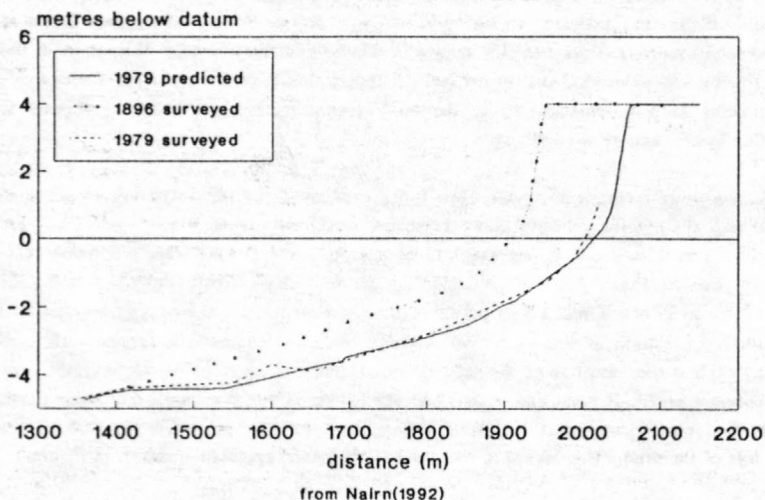
The downcutting process of the cohesive sediment is a more complicated and less understood process than the transport of the overlying sand. A conceptual model of the downcutting process was presented by Nairn et al (1986). It has been postulated by other researchers that shear stress at the bed created by orbital velocities along with the presence of sand as an abrasive agent may be responsible for irreversible downcutting of the cohesive surface. However, the shear stress at the bed due to orbital velocity decreases in an onshore direction through the surfzone due to the diminished wave heights in shallower water. This is at odds with the fact that downcutting must increase closer to shore in order for the profiles to maintain a similar shape as they recede shoreward (as is generally known to occur from a review of historical profile data at numerous sites - see Figure 1). Nairn et al (1986) suggested that additional downcutting mechanisms in the surfzone include the breaking induced turbulence which is able to penetrate to the bed as well as the shear stress attributable to the undertow velocity near the bed. Both the generation of turbulence and the undertow may be quantified by the rate of wave energy decay across the profile for the waves which are broken. Therefore, in the present model, downcutting is attributed to two driving forces: shear stress at the bed caused by orbital velocities (for unbroken waves) and wave energy dissipation in the surfzone (for broken waves). Owing to the lack of understanding on the details of the complex erosion process at a microscopic scale, the two erosion mechanisms must be related to downcutting by two empirical coefficients.

Since cohesive shore recession rates are quite low compared to the rapid dynamic response of sandy beaches (long term cohesive shore recession rates vary from 0 to 5 m/yr, but are typically less than 1 m/yr), it is important to be able to perform long term predictions (e.g. from many years to several decades). Considering the difficulty of realistically simulating the changes to the sandy beach over a long period of time and recognizing the heavy computational demands, the application of the numerical model to cohesive shores is generally performed without an explicit consideration of the sand cover. In these types of model applications, only a single cohesive profile is input and output (the sand cover is not modelled). For these cases, the calibration coefficients must therefore also account for the protective as well as the abrasive role of the sand. This approach was used in the model application shown in Figure 1.

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Figure 1 - Numerical Model Results
Port Bruce, North Shore of Lake Erie
Profile Evolution 1896 to 1979



COASTAL EVOLUTION AND ACCELERATED SEA-LEVEL RISE

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SUMMARY

Experience assessing shoreline vulnerability to accelerated sea-level rise in areas with limited data is discussed. Improvements for future studies are suggested.

INTRODUCTION

The prospect of accelerated rates of sea-level rise has focussed attention on Large Scale Coastal Evolution more than any other issue. National assessments of vulnerability to sea-level rise have already been conducted in a number of countries with many future studies proposed. These studies are being conducted within the framework of the Intergovernmental Panel on Climate Change (IPCC), Coastal Zone Management Sub-Group and are a first step towards the formulation of integrated coastal zone management plans (IPCC, 1992). While such studies focus on human vulnerability to sea-level rise, they require estimates of shoreline response to sea-level rise as input. In many parts of the world, there is limited basic data on the coastal zone. Therefore, it is only appropriate to apply simple response models. Experience determining possible land loss in studies of Senegal, Nigeria, Venezuela, Uruguay and Argentina (Nicholls and Leatherman, 1993) is presented, together with suggestions for improvements to future studies.

VULNERABILITY ANALYSIS

A vulnerability analysis (VA) constitutes an 'estimate of the degree of loss or damage that would result from the occurrence of a natural phenomenon of given severity'. The primary purpose is to develop reconnaissance level understanding of the vulnerability of different coastal areas to sea-level rise. Our studies focus on land loss; both erosion and inundation; other impacts of sea-level rise are not considered quantitatively. The future magnitude of sea-level rise is uncertain. Therefore, various measures of the vulnerability of the coastal land, its use and its population are estimated for a range of future sea-level rise scenarios and possible response options, assuming the present pattern and level of coastal development (Nicholls *et al.*, 1993). This gives the magnitude of impacts and the possible costs for each response option.

Physical data on the coastal zone in the countries studied were generally limited, including existing rates of shoreline change, detailed topography and bathymetry, sediment characteristics, wave climate and wetland accretion rates. This data limitation reflects conditions in many parts of the world. Therefore, we utilized aerial videotape-assisted vulnerability analysis (or AVVA), which utilizes oblique aerial videography and occasional ground truthing and existing maps and charts to characterize and classify the shoreline into similar units from the perspective of land loss response to sea-level rise. Then appropriate land loss models were applied as described below.

EROSION MODELING

To estimate erosion due to sea-level rise, the Bruun Rule was utilized. This geometric concept appears simple and can be readily applied given onshore and offshore boundary conditions. However, as discussed below, there are a number of practical and conceptual problems which need careful consideration. The form after Hands (1983) is preferred, as it includes an overfill ratio:

Nicholls - LSC

$$R = H/L * S * G \quad (1)$$

where: R - shoreline recession; H - active profile height; L - active profile width; S - sea-level rise; G - overfill ratio, related to grain size.

The landward boundary is easily defined, but the seaward boundary presents more fundamental problems. The annual depth of closure (d_i) can be estimated using wave climate in the model of Hallermeier (1981). However, the depth of closure over longer time scales is less obvious and existing guidance is limited. Hands (1983) noted the vital role of time scale in the determination of depth of closure. His long-term observations (125 years) in Lake Michigan suggest that the long-term depth of closure is about twice the short-term (5 years) depth of closure. Taking 20 years of hindcast wave data from Ocean City, Maryland and considering it on a year-to-year basis, Anders and Hanson (1990) found that d_i averaged 5 m, but annual values ranged from 3.7 to 7.3 m. Further, Hallermeier's (1981) formulation for d_i can be generalized to a time-dependent depth of closure by considering the significant wave height exceeded for 12 hours over any period (cf. Stive *et al*, 1992). Based on all this evidence, a simple approximation to estimate the depth of closure over the time scale of a century (d_{L100}) appears reasonable:

$$d_{L100} = 1.75 * d_{L1} \quad (2)$$

where:

d_{L1} is equivalent to d_i

Thus, these two depths provide two estimates of the seaward boundary condition and hence, a low and a high estimate of beach recession, which will embrace the likely shoreline response to a given sea-level rise scenario. Given the limited wave data that is available, longshore interpolation and extrapolation of the available closure depths, including utilization of the coastal classification developed as part of AVVA, was essential.

The problem of a reference depth is also critical. Hallermeier (1981) referenced d_i to low water. However, at Duck, NC, the observed depth of closure shows best agreement with Hallermeier's model if it is referenced to the peak water level during the storm (Birkemeier, *pers. comm.*) Applying this procedure in areas with a significant tidal range would often place the depth of closure above low water, which is physically unreasonable. Therefore, d_{L1} and d_{L100} were referenced to a depth of one meter above low water. This is consistent with the observations at Duck and allows for the influence of tidal range.

The Bruun Rule is based on the concept of profile equilibrium. Having defined boundary conditions, the profile system is defined by the active profile height and width. On shorelines with an apparent excess of sediment, such as much of Senegal, this assumption has some validity. However, on shorelines which appear starved of sediment, such as parts of Venezuela, where the offshore gradient *increases* with depth, the validity of this assumption is less certain. In such cases, it can be hypothesized that sediment is episodically lost to deeper water and these shorelines are already erosive. Hence, the Bruun Rule may underestimate the likely recession.

Lastly, the composition of the eroded material is often uncertain. If some of this material is silt or clay, the recession will be greater than if it is all beach grade sand. In the absence of data to the contrary, it was assumed that the material was 100% sand. Using the form of the Bruun Rule as shown in Eq. (1) makes such an assumption explicit. In the case of pocket beaches, the sediment supply is limited and the recession is controlled by the (effectively) unerodible rock surface beneath the sand, whose geometry is generally uncertain.

INUNDATION MODELING

Inundation will impact low-lying areas, particularly coastal wetlands. For non-wetland areas, a simple drowning concept is appropriate. However, for wetlands, including mangroves, this model is inappropriate as they exhibit a non-linear response to sea-level rise (Stevenson *et al*, 1986; Ellison and Stoddart, 1991). For small rises in sea level, wetlands can maintain their elevation, relative to mean sea level, by sediment/biogenic input. However, above some threshold rate of sea-level rise the wetlands begin to progressively lose elevation relative to sea level. If this continues, the constituent plants decline and ultimately die, and the former wetland rapidly converts to open water.

The critical knowledge to model this response is the appropriate threshold rate of sea-level rise. Unfortunately, while values are available in the literature, their application to other areas is uncertain as they are site-specific. In the absence of better guidance, a threshold rate of rise of 0.18 m/century was assumed (the present rate of eustatic sea-level rise). (Almost) total loss was generally assumed for a rise of 1 m. For intermediate scenarios, losses were linearly interpolated. Wetland migration was not evaluated.

While these assumptions are basic and simple, they are physically reasonable. More realistic models will require more information on wetland dynamics and more sophisticated modeling techniques (Capobianco *et al*, 1993a).

RESULTS

The results presented are for a one-meter scenario. Assuming no protection response, more than 90% of the land loss would be expected to be due to inundation in Venezuela, Argentina, Senegal and Nigeria. This land primarily represents coastal wetlands. Over 3% (>6,000 km²) and 2% (>18,000 km²) of the national land area of Senegal and Nigeria, respectively, could be lost. The exception is Uruguay, where the likely land loss is relatively small (94 km²) and erosion is the dominant process causing loss of 71 km². While erosion only threatens relatively small areas, they are often developed, including important tourist infrastructure. Protection of these areas by beach nourishment could necessitate significant investment, if such a response was chosen. The implications of land loss due to inundation is less clear, mainly because of the uncertain relationships between coastal resource availability and the existence of coastal wetlands. Active reclamation of coastal wetlands is also a problem. Lastly, most of the coastline in the countries studied remains undeveloped (>78%). Therefore, there is a great opportunity for effective integrated coastal zone management, including land use planning.

IMPROVEMENTS FOR FUTURE STUDIES

These types of analyses indicate land vulnerable to loss, given a rise in sea level. However, they can only be considered reconnaissance estimates. A number of simple improvements are under consideration, although their application may be hampered by the lack of data.

- (1) Given the importance of depth of closure for the Bruun Rule, and other long-term cross-shore modeling approaches, better guidance on its variability with time and the influence of tidal range is essential.
- (2) The Bruun Rule assumes an instantaneous response to sea-level rise and ignores the shoreface, which is a potential source or sink of sand. A two-part active zone/shoreface model with a lagged response would be more realistic (*cf.* Stive *et al*, 1990). In areas with limited data, behavior-based modeling may be most useful (Capobianco *et al*, 1993b).
- (3) On shorelines with inlets, the associated estuaries are sinks for sand given sea-level rise. This 'indirect' effect of sea-level rise may be more important in terms of shoreline recession than the processes described by the Bruun Rule (Stive *et al*, 1990), and should be included.

(4) Sea-level rise is only one factor influencing shoreline position. All coastal processes, including human influence and other impacts of climate/global change such as changing storm frequency or direction, or changing sediment supply should be included. If coastal zone management plans are to be developed such models will become essential, including interactive modules which allow the exploration of different response strategies (Capobianco *et al*, 1993a). Such models are also applicable to coastal wetlands.

CONCLUSIONS

Simple modeling approaches can give useful indications of the likely response to sea-level rise. However, while such approaches are suitable for VA, they are clearly inadequate for integrated coastal zone management, which will require a new generation of models.

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MODELING SHORE-NORMAL LARGE-SCALE COASTAL EVOLUTION

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Beaches and other coastal systems are generally at, or near, dynamic equilibrium. However, it is widely recognized that the terms of the process/response equilibrium differ according to the relevant time scale. We have examined the controls of coastal system development caused by long-term (decades to centuries) equilibrium of laterally-averaged shore-normal sediment dynamics in the coastal ocean by developing a series of numerical models. These models permit us to evaluate the role of changes in the four main regime variables (the nearshore and coastal ocean bottom profile shape, the sediment supply, the hydrodynamic climate, and the sediment grain size availability) on the long-term large-scale stability of shorelines.

We have elected to avoid attempts to extend event-scale sediment dynamics to large time scales because the balance between the regime variables results from small differences between large changes. Instead, we represent long-term processes directly based on parameterizations from detailed event-scale models of boundary layer and sediment dynamics. Our first long-term model is solved analytically for equilibrium conditions resulting in a representation of the interdependence of the variables and of the relative control each variable has on the overall equilibrium. A finite-difference numerical code has been developed to represent non-equilibrium conditions. Other formulations that require numerical solutions have also been developed. It was found that although the analytical solution could produce realistic behavior of the coastal ocean bottom profile response to the independent variables, it was not capable of dealing with systems characterized by several grain sizes. The subsequent models overcome this difficulty.

The models are calibrated for conditions of specific locations on the present Atlantic and Gulf Coasts of the U.S. Figure 1 shows modeled profiles (lines) compared to measured profiles (points) for U.S. East and Gulf Coast locations.

MONTAUK

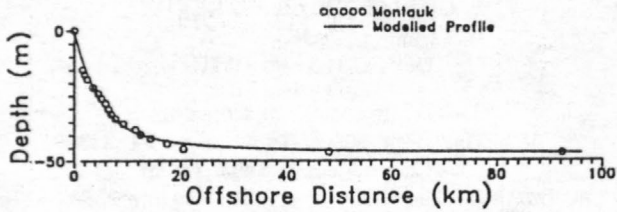


Figure 1a

MORGAN CITY

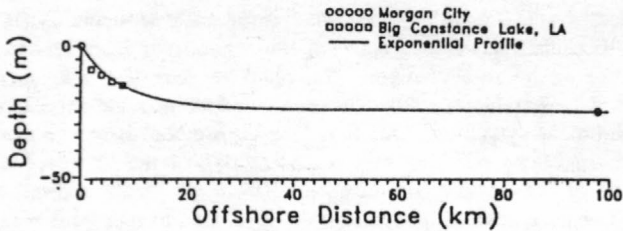


Figure 1 b

One test was performed to determine that the model correctly replicated an initial equilibrium profile after a sudden 5-m rise of sea level. Figure 2 shows that this was successful. The time to re-establish equilibrium under these conditions is a function of water depth. Figure 3 the computed depth changes over time at points whose initial depths were 5, 30, and 60 m. In this very simple representation the adjustments require from decades to millenia according to the water depth.

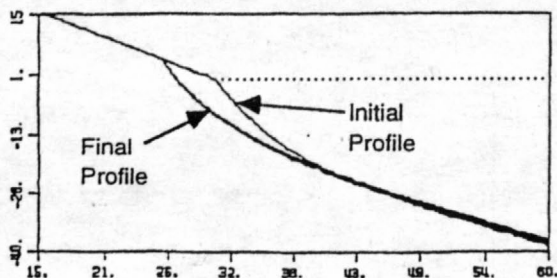


Figure 2

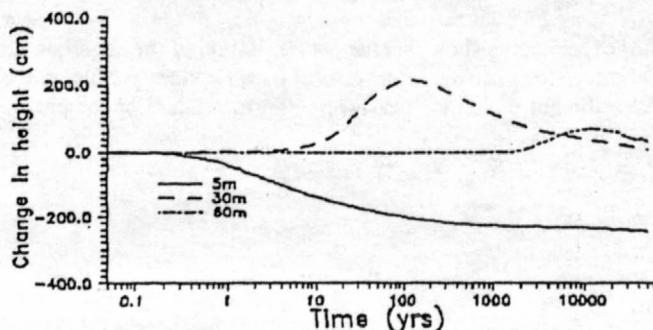


Figure 3

The results show that an increase in hydrodynamic forcing, as can develop if climate change causes increased storm activity, results in a deepening and steepening of the bottom profile. A shift from stable to rising sea level results in a flattening of the bottom profile. Coarser sediments are associated with deep and steep profiles. An increase in sediment supply results in a steepening of the coastal ocean bottom profile and a building out of the shoreline. Figure 4 shows the result of a profile adjustment resulting from halving the sediment supply rate at the shore as might occur with the damming of a river or the opening of an inlet updrift of the point of the profile.

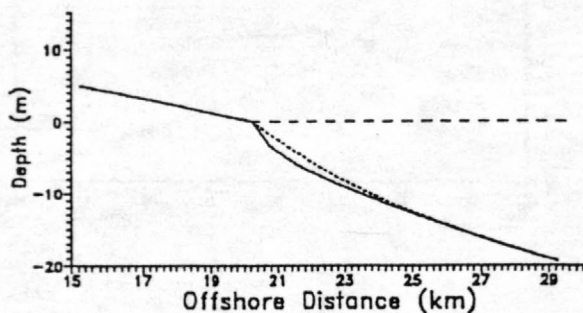


Figure 4

The numerical experiments show that the net translation of the shoreline results from the combined effects of changes in the coastal ocean bottom profile, due to changes in the overall sediment dynamics, and geometric translation of the profile.

LONG-TERM MORPHODYNAMICAL BEHAVIOUR OF THE EAST FRISIAN ISLANDS AND COAST

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The morphological behaviour of the East Frisian Islands and Coast has been well documented with respect to tidal low and high water line, position of the dune foot and extension of supratidal salt marshes since 1650. Additionally there is even information with lesser accuracy concerning former situations on certain areas available (fig. 1). On that basis a qualitative model of tidal inlet and barrier island migration was evaluated by LUCK [1977] explaining the morphological development as a cyclic process (fig. 2) which was supplemented by NIEMEYER [1990] considering hydrodynamical impacts causally. This natural development has been superimposed by large scale

human impacts interfering with these processes in the course of the last centuries. The two most important were

1. artificial acceleration of resedimentation of the medieval storm surge bays at the mainland coast for land reclamation purposes

and

2. fixation of four of the six tidal inlets separating the East Frisian barrier islands in order to protect holiday and health resorts which developed there since the middle of the 18th century.

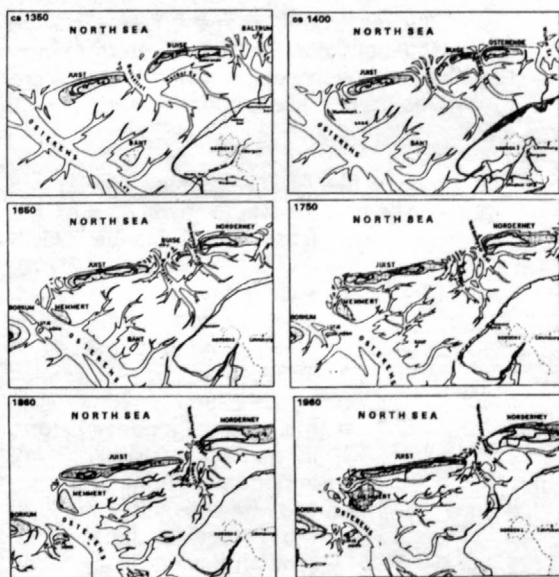


Fig. 1: Tidal inlet migration in the area of present inlet Norderneyer Seegat since 1350 by HOMEIER [NIEMEYER 1990]

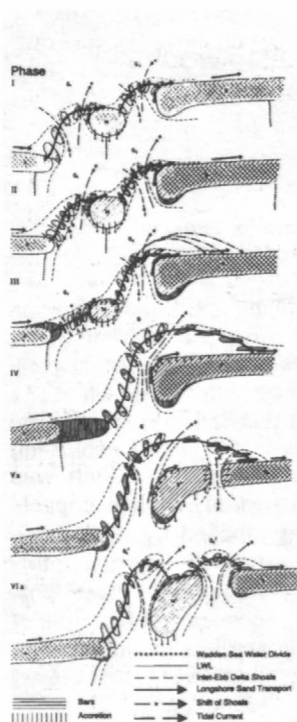


Fig. 2: Qualitative model of tidal inlet and barrier island migration [LUCK '77]

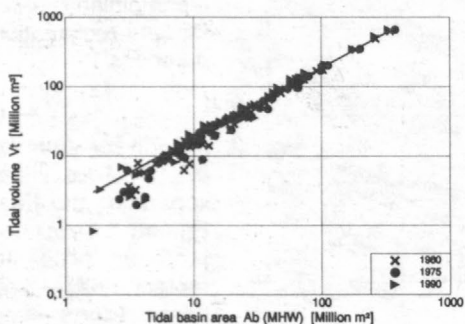


Fig. 3: Relation between tidal volume and basin areas; East Frisian Wadden Sea (surveys 1960, 1975, 1990)

The enclosure of medieval storm surge bays has lead to the following consequences: reduction of basin area, of basin volume, of tidal basin volume and of ebb delta volume in order to provide the basin's requirements for sedimentation. All these changes provoked additional changes of local wave climate in the basin leading to a further increase of sedimentation, resulting reduction of tidal volume and again sedimentation until a new equilibrium stage has been achieved [NIEMEYER 1991]. An example of consequences of the enclosure of a medieval storm surge bay becomes evident regarding the reduction of basin area of the tremendous relative reduction of the basin area of the tidal inlet Harle (fig. 4, H), which additionally lead to a relative increase of the basin area of the westward situated tidal inlet Otzumer Balje (OB). Coincidentally the barrier islands experienced corresponding changes of their length. The decrease of sums of percentages of the area of all considered tidal basins (fig. 4) makes evident that the losses due to the resedimentation of former storm surge bays has not been compensated by the aforementioned relative shifts of basin areas of tidal inlets.

The fixing of the tidal inlets has decelerated dramatically the longshore evolution of the East Frisian islands [LUCK 1977] and consequently resulting changes in the Wadden Sea tidal basins and at the mainland coast since 1860 (e. g. fig. 4). Coincidentally land reclamation in the areas of medieval storm surge bays has been only of importance in one of the six tidal basins since then. In the course of time up to now an evident stabilization has occurred leading to a dynamical equilibrium, which at least is expected to be only temporarily.

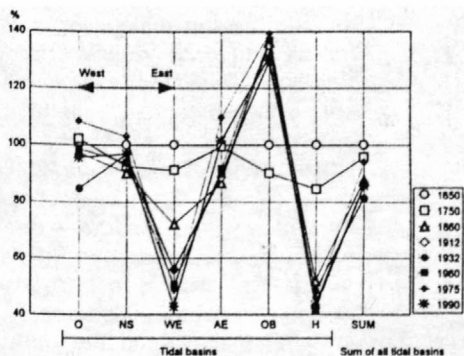


Fig. 4: Changes of total tidal basin areas since 1650 in percentage

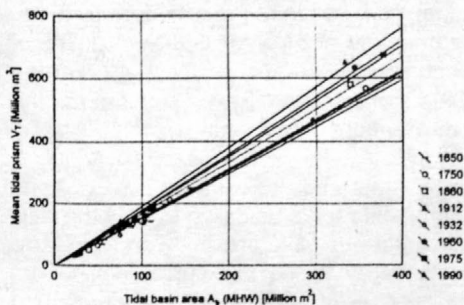


Fig. 5: Relation between tidal volume and total tidal basin areas since 1650

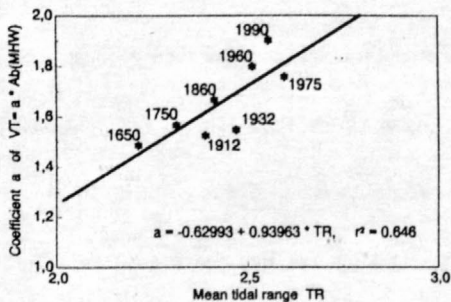


Fig. 6: Coefficient of tidal volume / tidal basin area relation versus tidal range since 1650

For that reason nowadays morphological changes occur on much lower length and time scales which is regarded as a mid-term dynamical equilibrium being expressed by a number of well-established semi-empirical relationships (e. g. fig. 3) which have been evaluated on a data basis with sufficient density for that purpose. That method for a functional parametrization of morphodynamical processes has proved as a success for coastal research and for coastal engineering planning since decades [O'BRIEN 1931; EYSINK 1991; NIEMEYER 1991].

In order to get a deeper insight into those documented long-term morphodynamical processes occurring at the East Frisian coast since 1650 (fig. 1 + 4) these relationships have also been used. For that purpose a procedure has been developed to apply them also to morphological data sets with lower density than present surveys. On the basis of the established relations significant parameters like basin volume, tidal basin volume, cross-sectional area of the inlet gorge and mean level height of intertidal areas could be derivated for "dynamical equilibrium conditions". Though that assumption is incorporated and a comparable "dynamical equilibrium" for the situations between 1650 and 1860 could not be assumed the relation e. g. between tidal volumes and

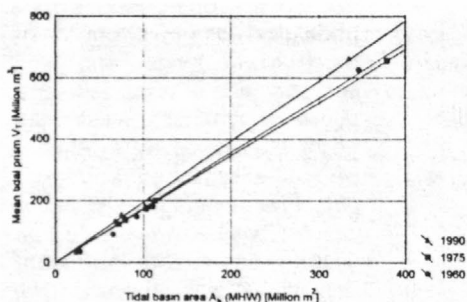


Fig. 7: Relation between tidal volume and tidal basin area; differentiated for the surveys of 1969, 1975 and 1990

emphasis will be laid on finding a reliable explanation and detecting the mechanism steering that process. A first explanation of the discrepancy to the results of EYSINK [1991] is that the only use of data for mid-term developments: As tidal ranges vary to a much lesser extent than for the long-term time scale, statistical scattering is of the same or even bigger order of magnitude than the structural differences due to tidal range variation.

The solution to this problem is of high importance because these relations provide the basis for conceptual or empirical model being up to now the only promising tool for morphodynamical modelling of future mid- and long-term development. A successful treatment will therefore deliver a valuable basis not only for current problems with structural erosion but also for the forecasting of prospected long-term morphodynamical processes due to changing boundary conditions like an expected acceleration of sea-level rise or in storm surge frequency.

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intake areas of the basins are for all situations rather strict. But they differ in their functional shape with a tendency to higher tidal volumes in relation to intake areas for increasing tidal range (fig. 5 + 6). In respect of the experience gained from mid-term processes [EYSINK 1991] this result is not trivial. Due to the fact that this tendency is also evident if the data sets with higher density of 1960, 1975 and 1990 are used (fig. 7). For that reason further

COASTAL BEHAVIOR OF INLETS AND BARRIER ISLANDS ALONG TIDE-DOMINATED SECTIONS OF LARGE SCALE COASTAL COMPARTMENTS

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Contrasts in inlet morphodynamics at tide-dominated and wave-dominated coasts result in substantially different long-term behavioral characteristics of barrier islands. Wave-dominated sections of coast are often associated with major headlands that provide large sediment sources for spits, baymouth barriers and long narrow barrier islands. In the mid Atlantic area, these areas are major components of large scale coastal compartments. Tide-dominated portions of these compartments receive depleted quantities of longshore drift and are associated with small headlands separated by low-order streams and tidal watersheds.

During transgression, coastal behavior along wave-dominated sections of the coastal compartment is primarily controlled by bluff retreat, lateral spit-progradation and inlet migration, by-passing and overwash. At tide-dominated coasts, lateral spit-progradation/inlet migration and by-passing are generally inhibited, and island systems are less dependant on updrift sediment sources.

The tide-dominated barriers of the southern Delmarva Peninsula are on the headlands of necks that are located between deep tidal channels. Each headland is a source of sediment for individual barrier islands. Littoral sediments eroded from headlands tend to migrate away from the headlands and toward the inlets at the ends of islands. Spits and beach ridges on the ends of the islands illustrate intermittent pathways of island shift during sea-level rise and transgression. However, exchange of sediment to adjacent islands is inhibited by deep inlet channels that are swept by strong inlet jets extending far onto the shoreface. Thus sediment that has not been displaced landward by overwash or spit processes may be swept offshore well beyond the littoral system. It is believed that inlet circulation may spread some of the shoal sediments back toward the island shoreface through flood-dominated channels adjacent to ebb deltas. This idea is reinforced by high resolution seismic reflectors illustrating a thin carpet of sediment below the shoreface. While the carpet forms a band that generally parallels the shoreline, it broadens and thickens adjacent to inlet deltas. Below the carpet, a proto-ravinement crops out at the base of the shoreface.

Thus, where the large scale behavior of wave-dominated barrier islands may be dependent on

updrift sediment sources from inlet by-passing, barrier islands along tide-dominated coasts may be a self-contained systems. The islands are only the subaerial portion of ashoreface sediment package that is tied to interfluvial headlands. Sediments eroded from the headlands are distributed toward the ends of the ends of the islands and jetted offshore onto massive tidal deltas. From the deltas, sediments are dispersed laterally over the shoreface. Over transgressive time scales, the pathways of these systems are controlled by the thalwegs of low-order antecedent river channels that separate the interfluvial headlands. The bifurcating nature of drainage channels in low-order watersheds may cause barrier island to lengthen or shorten independent of gradients in the littoral transport system. Thus, the formation of inlets (controlling island length) is not determined by storm break points, but by antecedent channel location and density.

It is clear that management strategies for wave- and tide-dominated sections of large scale coastal compartments must be different. At wave-dominated coasts, managers must contend with the affects of littoral transport gradients on naturally migrating islands and inlets. At tide-dominated coasts, inlet position may be relatively stable, and beach erosion and island migration are not only controlled by littoral transport gradients but also by subaqueous mid-island headlands and inlet circulation patterns.

CONTROLS ON SANDY SPIT DUNE EVOLUTION: A COMPARISON OF BUCTOUCHE SPIT, NEW BRUNSWICK, CANADA, AND LONG POINT SPIT, ONTARIO, CANADA

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Buctouche Spit is a 12 km long spit on the micro-tidal Northumberland Strait shore of northeastern New Brunswick that has evolved over the past 2000 years (Figure 1). Historically, sediment supply to this spit was $\sim 56\,000\text{ m}^3\cdot\text{a}^{-1}$. A sediment supply of this magnitude allowed the spit to extend at a rate of $\sim 4\text{ m}\cdot\text{a}^{-1}$ into 4-5 m of water. Current sediment supply to Buctouche Spit is $\sim 7\,640\text{ m}^3\cdot\text{a}^{-1}$ and it appears to have ceased growing since the middle of this century due to the removal of distal end sediments by ebb tide currents. Long Point spit is a 40 km long spit on the north shore of essentially tideless Lake Erie that has evolved over the past 4 000 years (Figure 1). Sediment supply to the spit is on the order of $500\,000\text{ m}^3\cdot\text{a}^{-1}$ and it is currently extending at a rate of $3\text{--}4\text{ m}\cdot\text{a}^{-1}$ into 30-35 m of water.

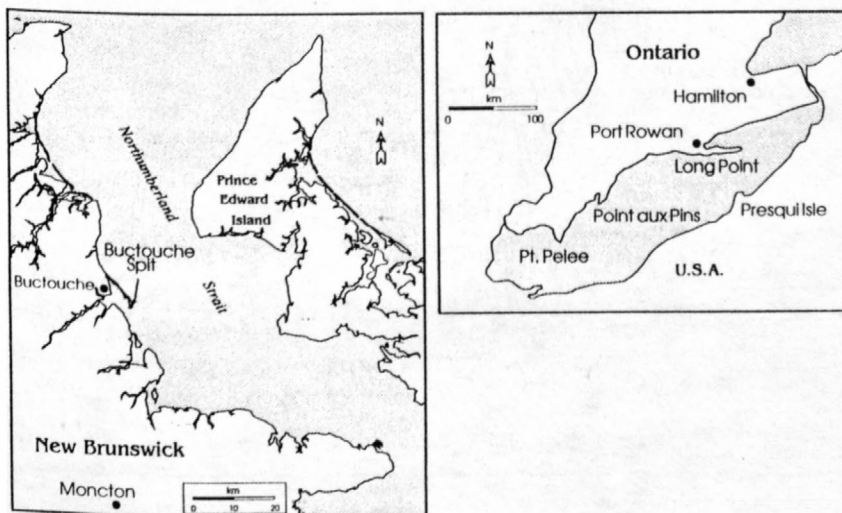


Figure 1: Locations of Buctouche Spit, New Brunswick and Long Point spit, Ontario.

The two spits are being studied using a variety of information including: stratigraphic profiles determined from cores and boreholes, aerial photographs (shorter-term temporal change), ^{14}C and luminescence dating techniques (longer-term temporal change), wave refraction modeling, and wind and wave climate data. Rates of growth and patterns of morphologic change have been established for each spit and comparisons are being undertaken using the aforementioned data. Preliminary results indicate that many general similarities exist between these two features but that there are several aspects, such as the orientation of distal end dunes (Figure 2), that are distinctly different.

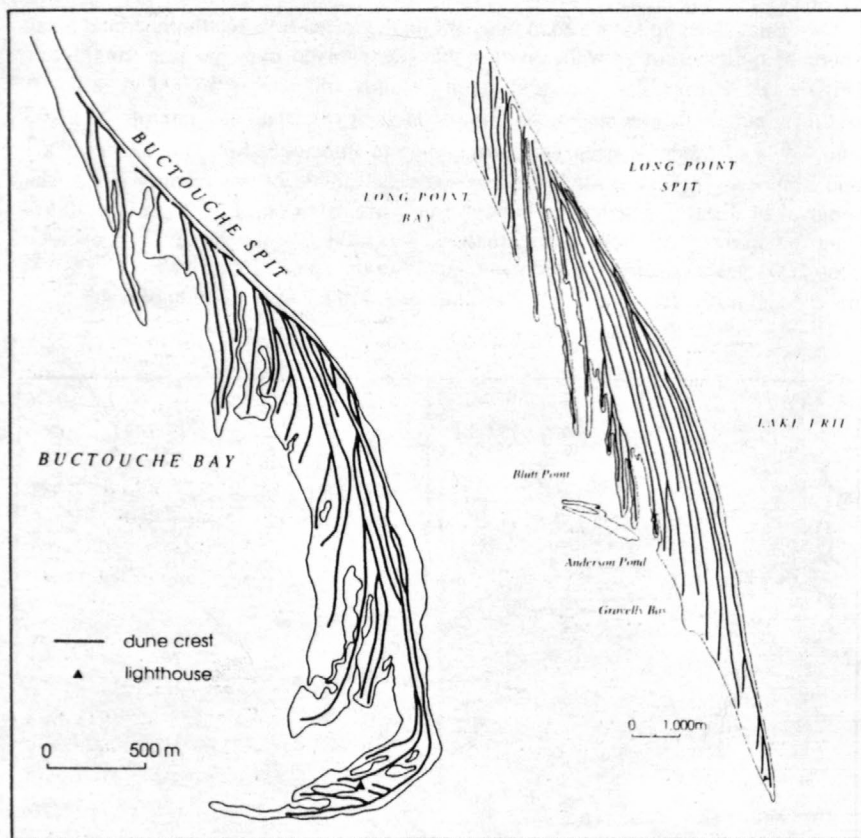


Figure 2: The distal ends of Buctouche Spit, New Brunswick and Long Point Spit, Ontario. In order to facilitate comparison, the distal ends of the two spits have been oriented so that their long axes are parallel, and the image of Long Point has been inverted so that the bays are on the same side.

Both spit systems have a narrow proximal section, characterized by a single foredune ridge that is prone to overwash and inlet formation, and a wide distal section, characterized by a progradational dune field. Buctouche Spit has a broad depositional spit platform and increasingly complex dune recurves at the distal end. Long Point has a proportionately narrower depositional spit platform and recurvature of the shoreline and dunes at the distal end has almost completely ceased (Figure 2).

The differences and similarities between these two spit systems can be explained in terms of a number of factors that control dune and spit evolution including: rate of relative sea (lake) level change, volume of sediment supply, wave refraction patterns, wave climate, wind climate, the morphology and depth of the basins into which each spit is prograding, and the rate of recession of the updrift mainland in each case. The purpose of this poster is to focus on differences between two related aspects of distal end dune development: (i) dune alignment/recurvature, and (ii) the scale and morphology of the progradational dune field.

The development of dune recurves is controlled by wave refraction patterns (which are controlled by basin characteristics), wave climate, and rate of spit elongation. The term elongation is used here to describe spit growth in the direction of a spit's long axis, while the term progradation is used to describe the overall growth of the distal end, both parallel and normal to the long axis. The degree of dune recurvature at Buctouche Spit is greater than at Long Point due to the shallow basin into which Buctouche Spit is growing. Wave refraction is thus a significant control at Buctouche Spit. At Long Point, water depths of 30-35 m mean that wave refraction is a relatively minor control and as such little recurvature is seen (Figure 2). In fact, as Long Point continues to extend into deeper water, wave refraction becomes less of a control and the degree of recurvature is seen to decrease. In contrast, the degree of recurvature at Buctouche Spit is increasing as the rate of spit extension declines. The slower extension rate allows for a greater cumulative effect of infrequent waves from the E and SE which transport sediment across the spit end toward the bay (Figure 1).

Differences in the scale and morphology of the progradational dune field are a function of: volume of sediment supply, rate of distal end progradation, and the relationship between wind and wave climate. The distal end dunes at Buctouche Spit are 2-3 m high and have spacings of 30-100 m. Because of the low sediment supply at Buctouche Spit, the major dunes are essentially wave built recurves covered with aeolian sediments. In comparison, the distal end dunes at Long Point are large. They are 5-10 m high and have spacings of 100-400 m. These dunes are formed primarily by aeolian transport from relatively wide prograding beaches to the foredune. The progradation of Long Point's distal end dunes is also aided by the fact that the dominant wind and wave directions are both SW, while at Buctouche Spit the dominant wind direction is WSW but the dominant wave direction is N or NNE (Figure 1).

Numerical Predictions of Radiation Stresses Throughout
The Southern California Bight

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Radiation stresses, the momentum fluxes due to the presence of surface gravity waves, are the principle driving forces for important coastal processes such as wave setup and longshore currents. Estimates of these stresses are of practical value to coastal engineers, and in situ measurements of radiation stresses are routinely collected at various coastal sites within the Southern California Bight. However, the offshore bathymetry in the Bight is so complex that wave spectra and radiation stresses are spatially variable on scales of only a few kilometers and measurements at one site are not readily applied to other locations.

A method which uses a deep water directional buoy and selected shallow water measurements to make Bight-wide estimates of coastal radiation stresses is described. The method uses a spectral refraction model, and is limited to the low frequency bands associated with swell waves from distant sources. Verification of the prediction scheme through comparisons to measurements at several stations (not included in forming the predictions) is presented. Finally, the Bight-wide estimates of longshore radiation stress will be discussed within the context of traditional descriptions of littoral cells in Southern California.

A METHOD OF ESTABLISHING MESOSCALE (DECADAL TO
SUB-DECADAL) DOMAINS IN COASTAL GRAVEL BARRIER
RETREAT RATE FROM TIDE GAUGE ANALYSIS.

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This paper examines the possibility by which the relationship between coastal shoreline response and sea-level change for gravel-based coastal barriers may be characterised by time domains of differential process forcing conditions that exist at a decadal or sub-decadal scale. At this initial stage the use of the term "domain" to indicate discrete time zones is preferred, rather than the use of the term "period" which might be seen as implying temporal rhythmic oscillations in process that have yet to be established.

As a first approximation, gravel barriers appear to migrate landwards by rollover as sea-level rises: barrier retreat rate correlates with short-term sea-level rise rate (Orford *et al.*, 1991). This relationship is however so strongly conditioned by other variables e.g. storm intensity, sediment supply and basement geometry, that the relationship between migration and sea-level rise is hard to specify beyond a first approximation as well as appearing time-dependent over a sub-decadal scale. It is important to recognise the response between barrier behaviour and sea-level rise is dependent on the stage in the life term of the gravel barrier being forced by a transgressive sea. It is only when the barrier shows a single consolidated ridge and is swash-aligned, can the role of sea-level rise be regarded as primary in the forcing of the barrier retreat. At other stages of barrier life-term stretching over 500 to 2000 years (Carter *et al.*, 1990) the role of sea-level is co-dominant, if not subordinate to sediment supply as the principal control on barrier architecture and evolution.

Story Head barrier in eastern Nova Scotia, is a well

* deceased

documented swash-aligned, single-ridge gravel barrier (Carter *et al.*, 1990). It is suggested that since 1945 the barrier has undergone major structural change with the barrier becoming detached from its basement boulder/cobble frame, which has been left in a drowned nearshore position (Forbes *et al.*, 1990). Migration rates of the reduced remnant Story Head gravel barrier, deduced from computer-rectified air photographs and maps, show evidence of migration fluctuation segmented into terms of slow change (c. 2 m a^{-1}) punctuated by shorter terms of fast change ($>15\text{ m a}^{-1}$). The role of sea level in this process is uncertain as it is primarily storm magnitude and frequency which control barrier migration rate through washover. However, the meteorological and wave climate data for the analysis of such domains are insufficient to carry the analysis over a time length in which the potential temporal scale of domains may be established for Story Head barrier. The Halifax (Nova Scotia) tide gauge which is c. 30km from Story Head offers a continuous data set (1920-1990) for water level analysis which can be examined for evidence of process forcing domains. These domains can then be used on an independent basis to examine the response variation in barrier behaviour.

Water level analysis is based on identification of positive surges after detrending for sea-level rise rate scaled to the length of the tide gauge record (3.8 mm a^{-1}). The presence of short-term sea-level rise of an interannual nature persists but is measured on a scale which at most is an order of magnitude less than surge measurement. Surges over any given period can be characterised to generate a scale of forcing magnitude and frequency commensurate with storm activity to which gravel barriers respond. Orford *et al.*, (1992) showed how forcing magnitude can be used to identify domains of relative stability and instability for a retreating gravel barrier where the domain boundaries were set independently by dates of photographic evidence of retreat. This paper establishes domain boundaries identified solely in terms of variability of the forcing magnitude and frequency on an annual basis rather than depending on randomly timed air-photograph evidence. As barrier retreat is dependent on crestal overwashing, forcing characterisation of surges at high water positions only is appropriate. The two indices of forcing magnitude (b_1) and forcing frequency (b_0) are used together (cluster analysis) to

establish 'event' types (by dendrogram partitioning) which can characterise each year of the gauge record. A Markov analysis of

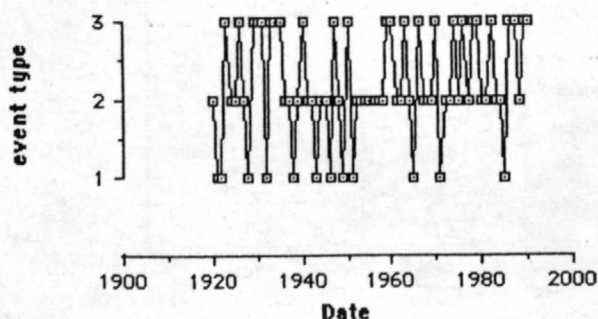


Fig.1 Yearly characterisation of surge magnitude and frequency in terms of events
Event 1: low surge potential Event 3: high surge potential

first-order nth step (year) event transitions shows that only a 6-step and a 22-step are significantly different from a random expectation. This suggests that forcing domains may exist at both 6 year and 22 year scale (decadal and sub-decadal). Fig.2 shows variation in the forcing magnitude coefficient over the tide gauge record, smoothed on an eleven year basis. The 22-year domain structure shows at least two distinct periods of low and high surge forcing that approximates to two periods of stable and unstable barrier behaviour as observed by barrier migration rates. The main domain transition position around the mid-1950's fits the dating of the major disturbance to Story Head barrier (Forbes *et al.*, 1990). The relatively tranquil period during 1935-55 may well have generated barrier architecture of a morpho-sedimentary nature that helped the barrier detachment process occurring at the next domain transition point. Short term variation of a sub-decadal nature may account for the variation of migration rate once the barrier is set in an unstable mode. The interplay of decadal and sub-decadal process operates on the background of antecedent conditions as indicated by the memory inherent in the sedimentary response system. Mesoscale domains at the decadal to sub-decadal

resolution should be regarded as being central to the explanation of Story Head barrier behaviour during the last sixty years.

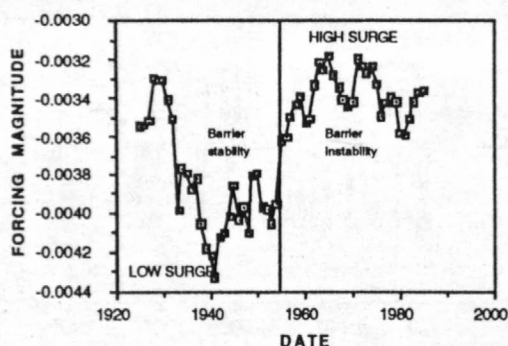


Fig 2. Mesoscale phases of barrier stability and instability at Story Head based on the 11-year smoothed forcing magnitude time series from the Halifax tide gauge record.

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SEA LEVEL CONTROLS ON MISSISSIPPI RIVER DELTA BUILDING DURING THE HOLOCENE TRANSGRESSION

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ABSTRACT

Geologic studies of the Mississippi River delta plain and continental shelf reveal the occurrence of several relative sea level stillstands during the last stages of the Holocene transgression. During periods of rapid relative sea level rise, individual stillstand delta plains undergo coastal submergence and land loss as their lower alluvial valleys fill with transgressive sediments. During periods of relative sea level stability, this trend of transgression and submergence is reversed. The river begins to fill its lower alluvial valley with lacustrine delta complexes behind the high-stand shoreline. With time, bayhead deltas build out into shallow coastal bays forming delta complexes that evolve into shelf-phase delta plains on the continental shelf. The timing of delta formation is tied to sea level stillstands when the rate of rise dropped below a critical threshold value that allow river delta development. In the Gulf of Mexico, one can observe river deltas in all stages of evolution, since sea level reached the current highstand about 3,000 years ago. The lacustrine and bayhead delta complex stage can be observed in the Atchafalaya basin in Louisiana. Other examples of bayhead deltas can be found at the Trinity River in Texas, the Pascagoula River in Mississippi, the Mobile River in Alabama, and the Escambia River in Florida. In contrast, the Mississippi River has filled its alluvial valley and built a large shelf-phase delta plain composed of a series of delta complexes.

The objective of the paper is to examine the Holocene history of the Mississippi River delta plain and continental shelf in light of trying to understand the relationship between sea level and large scale coastal evolution. The continental shelf of the Mississippi River delta plain contains an 18,000-year record of sea level changes and coastal evolution. By examining high resolution seismic profiles, deep cores, and shallow vibracores combined with new radiocarbon dating results, it is our intent to develop a better understanding of the sea level change thresholds required to drive the development of different styles of coastal evolution observed in the Mississippi River delta plain.

An overview of geologic and oceanographic assumptions required by coastal models.

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Quantitative sedimentology is increasingly being applied to the practical problems of predicting sand behavior related to coastal engineering projects. Such application of numerical models is questionable in our opinion because 1) major assumptions in all the numerical models are invalid in part because well established principles of nearshore oceanography are ignored, 2) appropriate efforts to document the success or failure of real world sand behavior predictions are routinely not instituted and 3) the customer (the public) is not informed of the great uncertainties inherent in the prediction of sand behavior by numerical models.

Most numerical models require some combination of four major types of "non-process" input data from which all model results are calculated. These are: (1) **Q or quantity of sand moved.** No one has ever measured **Q**. All methods used for estimating or calculating **Q** are indirect and are inaccurate to varying degrees. In some applied situations the standard numerical approach has produced estimates that are clearly at odds with field evidence (eg. Carolina Beach, NC), or resulted in sand movement in the wrong direction (eg. Topsail Island, NC and Murrels Inlet, SC).

(2) **Wave Climate.** Waves are generated by winds. Weather is a chaotic phenomenon, thus waves generated by the weather must be treated as a system in chaos. This is never the case in numerical models. Most numerical models involve some sort of average wave height of monochromatic waves approaching from some average direction. The models, as applied, cannot handle multiple, interacting wave sets approaching simultaneously from many directions. The actual data, if available, are derived from measurements taken offshore from a shoreface whose bathymetry and sediment cover, on both a large scale and small scale (offshore bars), is not precisely known and across which wave refraction

and wave dampening cannot be accurately predicted. The point is that a replenished beach may well be destroyed or severely damaged by a storm the day after its emplacement. If the numerical model cannot account for this possibility, then it is inadequate, and the public is being deceived.

(3) **Grain size.** The grain size of the shoreface is normally used as the basis for determining the shape of the shoreface (equilibrium profile). It is also the basis for predicting volume of sand movement by wave orbital interaction with the sea floor. Most numerical models characterize the shoreface using a single grain size. A few models may attempt some rudimentary prediction of the variation of grain size across the shoreface. In reality, most shorefaces have a complex sediment cover of highly variable grain sizes. The variation in grain size across the shoreface is neither regular nor predictable.

(4) **Local shoreline retreat/accretion rate.** This parameter is usually better known than any of the other major unknowns and in many cases a reasonable number is available.

In addition to problems with the above "non-process" input data, there are a number of "process-based" assumptions underlying all nearshore numerical models that are not met in any real world situation:

(1) It is assumed that the processes transporting sand on a beach can be described using very few parameters (mostly those listed above). Furthermore, empirical equations derived to describe beach behavior elucidated at one location, or in one region, are assumed to be applicable at other locations. We believe that this is not true. There are a very large number of factors that govern nearshore sediment transport, and while they all may operate to some degree at every beach in the world, the relative importance of each factor is different at every beach.

(2) It is assumed that sand is moved exclusively by wave orbital interaction with the seafloor. This is not true. Evidence of bottom current activity is overwhelming including direct measurements, theoretical predictions and observation of nearshore derived sediment bodies on distal continental shelves, ancient and recent. Interaction of currents with sediment suspended by wave

activity is a certainty. Types of bottom currents include storm surge ebb, wave set up and set down, wind set up and set down, turbidity, tidal and several others.

A large variety of other incorrect assumptions of greatly varying importance are used in the models. Among these are:

- (1) Offshore bars, other bedforms, and beach state do not affect beach response to storms.
- (2) There is a closure depth beyond which no significant amount of sediment is transported.
- (3) Underlying geology plays no role in the control of shoreface profile shape.
- (4) The beach is a two dimensional system.

We believe that the assumptions behind all currently used numerical models are too weak to justify the use of the models in predicting the behavior of sand at engineering projects. Until field-based, probabilistic numerical models are developed, the statistical approach should be used. This requires detailed and long term observations of specific beaches, similar to those that have been carried out on the Gold Coast, Australia and on the Dutch coast.

Theoretical Stability of Equilibrium Beach Profiles

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Introduction

The goal of this research is to determine whether "self-organization" (in this case, the coupled response of morphology and time-averaged flow) is a likely mechanism involved in the formation of large scale, nearshore topography. Prediction of the variety of both regular and irregular morphologies (Figure 1) that form in response to changing wave conditions has been a long term goal of nearshore researchers; however, even the simplest equations describing the relationship between gradients in sediment transport and changes in bed elevation are highly nonlinear, and thus, there are no analytical solutions for time-dependent beach response. So far, the approaches to this problem have been to (1) set the time-averaged sediment transport to zero and solve for possible topographies that meet this condition (these are equilibrium models, e.g. Bowen, 1980; Bailard and Inman, 1981; Holman and Bowen, 1982), or (2) formulate numerical models to let topography evolve from specific initial conditions (these are time-dependent models, e.g. Boczar-Karakiewicz and Davidson-Arnott, 1987; Howd and Holman, 1987; Roelvink and Stive, 1989; Werner, 1993). The equilibrium models do not indicate whether arbitrary or near-equilibrium conditions actually evolve towards equilibrium. Time-dependent models typically address only a limited range of initial conditions, and do not indicate the range of parameters that lead to equilibrium vs. unstable behavior (an exception is Howd and Holman, 1987). We investigate the general stability of equilibrium profiles in order to determine the likelihood of beaches reaching a steady-state, and determine the importance of "self-organization" in forming large scale, nearshore topography.



Figure 1. Time exposure of nearshore region near Newport, Or., February 1993. Bright, linear bands represent waves breaking over sand bars and at the shoreline.

Methods

Using Bagnold's sediment transport formulations, the interaction between the fluid motions and topography can be described using sediment continuity, which relates the time derivative of the bed elevation, h , to the horizontal derivative of the sediment flux, Q . Because sediment flux can be formulated as a function of beach slope, $g(\frac{\partial h}{\partial x})$, and velocity (Bowen, 1980), which is itself a function of water depth, $f(h)$, this system can be represented in two spatial dimensions as a nonlinear partial differential equation:

$$\frac{\partial h}{\partial t} = \frac{1}{1-v} \frac{\partial Q}{\partial x} = \frac{1}{1-v} \frac{\partial}{\partial x} [f(h) g(\frac{\partial h}{\partial x})]$$

where, v , is the void ratio (porosity) of the sediments. We will systematically investigate the stability of this system with a perturbation approach, using simple expressions for the velocity field and sediment transport. These results apply to near-equilibrium conditions. Numerical models must be used to address the far-from-equilibrium behavior of this system.

Summary

Although important qualities of actual sediment transport processes are ignored by considering simple velocity and sediment transport formulations, the behavior of the resulting nonlinear systems may give us insights to the behavior of actual beaches. Equilibrium solutions to several simple models of the coupled nearshore morphology-flow system have been compared to actual topography, but we note that the equilibrium profiles predicted by the simplest models are typically monotonic, while beaches are, often, more complicated. To account for more complex topography, an investigation of the stability of equilibrium profiles may suggest that growth of perturbations occurs under some conditions, and that barred topography results from the self-organization of an evolving profile, not necessarily from more complex flow systems.

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FOREDUNE MIGRATION AND LARGE SCALE NEARSHORE PROCESSES

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The alongshore variation in foredune development and its morphological expression is an example of a large scale coastal feature. Studies by Fisher (1984), Terwindt and Battjes (1991), and Psuty (1990), amongst others, point to the periodic variation in foredune characteristics and relate these changes to similar periodicities in the beach and nearshore environments. Allen and Psuty (1987) and Psuty and Allen (1987) have shown a dimensional relationship between nearshore circulation cells and alongshore undulations in the foredune crestline. The conclusive alongshore association of nearshore energy variation and foredune modification has yet to be established but this study points to several areal relationships that occur at a variety of scales and may help explain the large-scale pattern of foredune development.

The study area is the Fire Island barrier island, approximately 50 km in length. It is a low mesotidal barrier (mean tidal range of 1.3 m), situated on the southern shore of Long Island. Long term records of shoreline change (Taney, 1961; Leatherman and Allen, 1985) show that the island has been retreating for more than 150 years, except for the accretionary western terminus. The erosional condition has resulted in the inland transgression of the foredune form, but neither the form nor its rate of migration has been constant. Starting in 1976, detailed aerial photographic analysis and field surveys have been gathering data on the variation of foredune configuration and the variation of alongshore variability in crestline displacement. These data depict landward and seaward migrations of the foredune crestline.

Because the foredune response is a conservative expression of coastal change (it reacts to high energy events that affect the upper beach, and it retains the modification until the next high energy event), the alongshore crestline configuration preserves the spatial pattern of foredune scarping and mobilization. An analysis of this pattern suggests several scales of causative events responsible for foredune migration and coastal evolution.

Digitization of the dune crestline interpreted from large scale aerial photography in 1976, 1981, 1986, and 1992 provides files that are analyzed arithmetically for direction and magnitude of change and statistically to estimate periodicity of alongshore patterns and nesting of different hierarchies of waveforms. We also have data from post-storm aerial photography of the December 10-15 1992 nor'easter and from differentially processed kinetic Ground Positioning System (GPS) traverses along the dune toe and approximate Mean High Water shoreline that will be used to identify spatial patterns involved with more frequent changes such as individual storm effects. Annual beach profiles complete the data base by providing vertical change measurements, increasing temporal resolution at selected locations, and adding precision to the crestline position.

Each data file includes the curvilinear trend of the island and a series of smaller undulations. Decomposition of the data by harmonic analysis and subtraction of the fundamental island curve highlight the presence of a series of wave forms with a length of about 2800 m and an amplitude of 15-20 m. Spectral analysis yields other significant forms with wave lengths of 800 to 900 m, at 460 m, and at 200 m. Less prominent but commonly-found forms have lengths of 270 m, 170 m, 135 m, and 115 m. Explanations of the presence of the major wave forms are complicated by such factors as antecedent topography, anthropogenic processes, nearshore processes, and offshore topography.

Comparisons of crestline positions at different times depict the changes and

spatial patterns but there are many methods to accomplish this, depending on the question. Based on 243 sample transects, 63% of the crestline moved landward more than 1.5 m, 21% moved seaward more than 1.5 m, and 16% had no significant directional change between 1976 and 1986. The population distribution of dune displacement magnitudes for this period is bimodal with maxima at 0 to +1.5 m (seaward) and at -4.5 to -6 m. Both Park and community-managed foredunes show evidence of inland migration despite many attempts to stabilize the crest or even build it seaward.

The pattern of displacement suggests a series of sinusoids that vary from inland displacement, to stability, to seaward displacement, back to stability, and this pattern is repeated several times along the length of the island. Each of the five-year intervals presents a similar pattern, but spatially displaced. Field data, especially after storms, reveal that the beach zone and the adjacent nearshore area display a periodic pattern of cut and fill. Nearshore circulation cells in the classic mode of Wright, et al.'s (1986) rhythmic bar and beach morphodynamic association characterize the shore zone. Some of these morphodynamic forms interact with the foredune and scallop the face of the dune. Numerous variables combine to define the degree and length of scarping and the subsequent imprint on the dune crestline. Some information on the spacing of the circulation cells does exist and is interfaced with the crestline data. Field observations suggest that some of the cells are relatively permanent whereas others are shifting alongshore. Each of these situations affects the foredune differently and adds to the complexity of the alongshore variation. Extending over distances of two-to-ten kilometers, this sinusoidal pattern is an imprint of the larger scale processes operating continuously to cause foredune migration in an alongshore periodic framework.

Conclusions

The patterns of coastal change along Fire Island as expressed by the variation in foredune crestline migration are caused of a hierarchy of factors

operating at different spatial scales: coastal orientation is a result of continental and regional tectonic processes acting over hundreds of kilometers, island curvature at about 30km is controlled by inherited Pleistocene topography and relative sea-level rise, locationally-stable erosional arcs in the island trend at 5-10 km may be induced by shoreface ridges, and migrating sinusoids spaced at hundreds of meters to 3 km which result in alternating sequences of dune retreat, stability, and even growth, are associated with nearshore process interacting with the nearshore bars and beach geometry, especially during storms. Superimposed upon this ordered hierarchy are stochastic effects due to controls of antecedent dunal topography and to human activities, such as accelerated erosion in communities due to the presence of buildings and beach structures causing localized sediment starvation of the beach/dune system.

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LONG-TERM DEVELOPMENT OF PACIFIC ATOLL ISLET SHORELINES

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Atolls are islands consisting of an annular reef rim encircling an inner lagoon. They occur in tropical oceans and may be up to 60 miles across, although most are much smaller. Atoll islets, the small islands that occur on the annular reef rim of atolls, are entirely Holocene landforms and are perched on older reef limestone foundations. They are composed of reef-derived detritus of sand- through gravel-size and their geologic development is unique because it is intimately related to biologic calcification within reef systems. Maximum elevations are generally less than 5 m above sea level and any potential sea-level rise (either man-induced or natural) could pose serious environmental consequences for the inhabitants.

Critical processes in the evolution of the islets are revealed by examination of islet geology and morphology, and a basic knowledge of the processes affecting the coasts. Four dominant processes, each operating on different time scales and energy levels drive the evolution of the islets:

1. Holocene sea-level history. Atoll islets require a reef foundation near sea level for their development. The upper vertical limit of reef growth is limited to the approximate mean low water level because most framework corals require constant submersion. Variations in the mean sea-level position through time result in a shift of the vertical limits of reef growth. Although a falling level is not necessary for islet formation, a sea-level drop (< 2 m) as postulated for much of the Pacific since the mid-Holocene, could be responsible for an exposed reef foundation conducive for greater islet stability.

2. Large storms. Once a suitable foundation is available, a mechanism is needed to deposit sediment above normal tide levels. Large storms, such as tropical cyclone Bebe, a 100-year storm which struck Tuvalu in 1972, are responsible for transporting large amounts of reef material and forming large deposits on the upper reef surface. These storm deposits form the nuclei around which islets later develop. The magnitude, frequency, and direction of the storms play an important role in the location and amount of islet coverage on reef rims.

3. "Normal" Tradewind Conditions. As soon as storm-derived sediment is deposited on the upper reef surface it undergoes reworking and modification. The atolls studied in the Pacific lie in the tradewind belt where easterly winds are common throughout most of the year. Water circulation in the lagoons, through the reef passages, and over the reef rim are largely controlled by the tradewind patterns. Islets commonly exhibit a morphologic and compositional asymmetry as a result of windward/leeward differentiation under the tradewinds.

4. ENSO Reversals. During periods of strong ENSO events (El Nino - Southern Oscillation), reversals in the "normal" tradewind pattern may exist for extended periods (several months or more). ENSO events typically occur on an approximate decadal periodicity. Islet shorelines which are developed on leeward coasts are exposed to windward conditions often resulting in significant shoreline reorientations and modification of islet plan shape.

Islet development, as a percent of the total area available for islet formation, appears to vary according to the size of the atoll (Fig. 1). The smaller atolls tend to have more continuous islet development whereas islets on larger atolls occupy a much smaller percentage of available reef top and tend to be dispersed and fragmented. There are several explanations for this relationship. On smaller islands there is a greater concentration and convergence of islet building processes; for example, a storm will affect a larger percentage of the total island in a small atoll. There is also a higher ratio of productive reef to reef top in smaller islands suggesting there is more sediment available for islet formation. Table I illustrates carbonate production and lagoonal infilling rates for three hypothetical islands of different-sizes. The smaller reef island undergoes relatively rapid lagoon infilling resulting in abundant sediment available for islet development whereas in the large open atoll the lagoon remains a significant sediment sink..

Effects of the above processes can be recognized in the islet geology. Any predictive models used to describe the behavior of atoll islets must take into account the relative effects of each of these processes -- they are all responsible, to varying degrees, for present islet configuration. For example, to consider the consequences of a "Greenhouse" induced sea-level rise on atoll islets, one must consider the ability of coral reefs to grow vertically, changes in storminess which could lead to more storm deposits, and latitudinal shifts in weather bands leading to greater or lesser reworking of storm deposits.

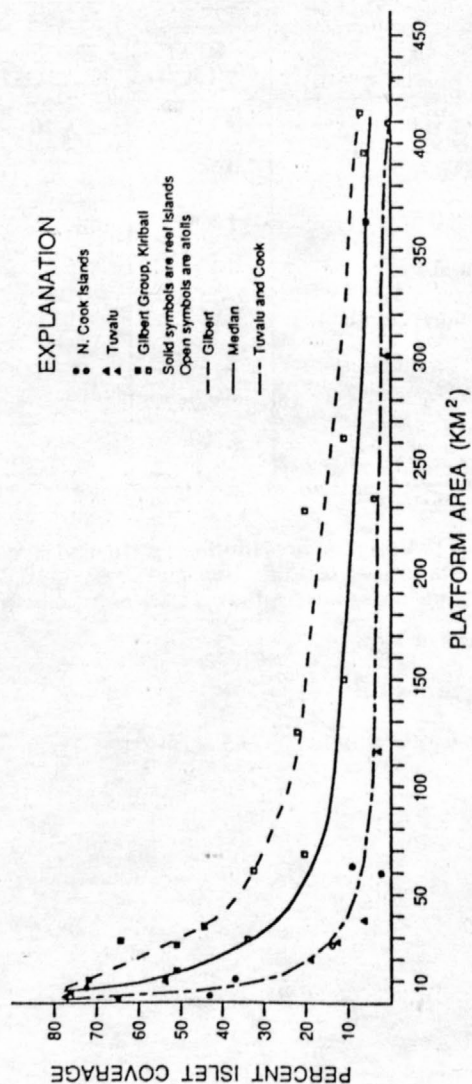


Figure 1. Plot of platform area (reef rim and lagoon if present) versus percent of the platform area covered by islets for atolls and reef islands in the Cook Islands, Tuvalu, and the Gilbert Group of Kiribati. Reef islands are lagoonless low-lying carbonate islands where any initial lagoon has been infilled.

	REEF ISLAND	ENCLOSED ATOLL	OPEN ATOLL
Platform Size (km)	3 X 3	5 X 10	10 X 20
Rim Area ¹ (km ²)	9	26	56
Lagoon Area (km ²)	1	24	144
Rim/Lagoon Ratio	9.0	1.1	0.4
Lagoon Volume ² (m ³ x10 ⁶)	40	960	5,760
Rim CaCO ₃ Production ³ (m ³ x 10 ³ /yr)	36	104	224
Years to Fill	1,100	9,300	25,700

TABLE I. Variation in platform size, rim and lagoon area, lagoon volume, total rim carbonate production, and time required to fill a lagoon with rim derived sediment for three different sizes of hypothetical islands.

¹ Average width of 1 km.

² Average depth of 40 m.

³ Average rim CaCO₃ production of 4 kg/m²/yr

INFLUENCE OF INHERITED GEOLOGIC FRAMEWORK UPON
BARRIER ISLAND MORPHOLOGY AND SHOREFACE DYNAMICS

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ABSTRACT

Passive margin coastlines with limited sand supplies, such as much of the U.S. Atlantic margin, are significantly influenced by the geologic framework of older stratigraphic units that occur within and seaward of the shoreface. Many east coast barrier islands are actually perched barriers in which the underlying, pre-modern sediments determine the three-dimensional morphology of the shoreface and have a strong influence on modern beach dynamics and sediment composition. Perched barriers consist of thin surficial beach sands that overlie older, eroding stratigraphic units with highly variable compositions and geometries. Along many parts of the coastal system, stratigraphically controlled bathymetric features also occur on the inner continental shelf. These features modify incoming energy regimes effecting patterns of sediment erosion, transport, and deposition on the adjacent shoreface. Holocene sea-level rise has produced a modern transgressive shoreface, barrier island, estuarine, and fluvial sequence of coastal sediments that are being deposited unconformably over irregularly preserved remnants of pre-existing stratigraphic sequences. Thus, modern barriers are often perched on top of highly dissected lithostratigraphic units with erosional geometries and compositions that range from tight peat and mud to indurated sandstones and limestones. These variations in the underlying geologic framework result in different responses to the erosional forces and thus determine the detailed geometry and morphology of the barrier islands.

In North Carolina, many barrier islands are perched on and controlled by the pre-existing Cenozoic sediments. The surficial geology of the North Carolina coastal zone can be subdivided into two distinct provinces. North of Cape Lookout and the associated Cape Lookout High, the surficial geological framework is defined by a thick Quaternary sequence (50 to 70 m) that fills a regional depositional basin parallel to Albemarle Sound and designated the Albemarle Embayment. The coastal zone south of the Cape Lookout High is dominated by Tertiary and Cretaceous units with only local occurrences of Quaternary sediment units preserved. The more lithified, offlapping stratigraphic sequences of Tertiary and Cretaceous units wrap around the Carolina Platform High, a major basement structural feature that occurs between Cape Fear and Cape Romain, and crop out across much of the continental shelf in Onslow and Long Bays. Superimposed upon this regional structural and stratigraphic framework is a major drainage system and associated interstream divides that further modify the inherited paleotopography upon which some modern barrier islands are perched.

Five general categories of shoreface systems occur along the North Carolina coast; each is characterized by a distinctive set of coastal processes and shoreface responses. Headlands are morphological features that rise above the active ravinement surface and are composed of semi-indurated to indurated, Pleistocene or older sediment units. 1. Subaerial headlands are characterized by the active incisement of a wavecut platform with an associated perched strandplain beach.

2. Submarine headlands are submerged morphological features that have been incorporated into the modern shoreface and upon which the barrier-estuarine system is perched. Ancient sediment units crop out on the eroding shoreface and commonly occur on the inner-shelf as paleobathymetric highs in front of the modern shoreface and thus modify incoming wave climates.

Shorefaces without headland associations can be subdivided into three general categories. 3. Transgressive shorefaces are composed of tight Holocene peat and mud sediments that extend from the modern estuaries, under the barrier sands, and crop out within the surf zone and upper shoreface. The steep shoreface is often characterized by an irregular geometry and dominated by discontinuous and highly scarped dune ridges with abundant washover fans. 4. Regressive shorefaces are composed of unconsolidated Holocene sands and occur along barrier island stretches that have an adequate sediment supply and are often associated with headlands and cape structures. They are dominated by concave progradational geometry with accretionary beach ridges, and are relatively stable. 5. Channel-dominated shorefaces are composed of unconsolidated Holocene gravel, sand, and soft mud sediments occurring along barrier island stretches dominated by inlet or fluvial channel deposits.

Thus, the basic structural and stratigraphic characteristics of any barrier island complex influence the resulting barrier island morphology, inlet development, beach dynamics, and commonly prevent the concept of an equilibrium profile from being realized. Examples of perched barriers associated with different geologic situations within the extensive North Carolina coastal system follow.

SUBAERIAL HEADLANDS

The coastal areas of Fort Fisher, Kure Beach, and Wilmington Beach are dominated by an extensive eroding subaerial headland. This mainland peninsula is composed of Pleistocene sediment units with no barrier island and estuarine system. Rather, the coastal system consists of a wave-cut platform in the Pleistocene units with a thin and highly variable strandplain beach perched on the platform. Linear outcrops of Pleistocene coquina and associated lithologies occur irregularly on the beach north of Fort Fisher and dramatically control the geometry of this entire coastal system. These rock outcrops are almost shore-parallel and extend southward across the shoreface onto the inner continental shelf as a submerged 'groin-like' feature. This extensive sequence of coquina outcrops act as a groin in the forebeach and as an offshore barrier on the inner shelf. Similar geologic controls, resulting from the presence of a subaerial headland, occur on segments of the North Carolina shoreline south of Cape Fear including portions of Yaupon and Long Beaches.

SUBMARINE HEADLANDS AND PALEOTOPOGRAPHIC HIGHS

Pleistocene deposits have been drilled by many researchers under numerous shorefaces along the Atlantic coast from Long Island to Florida. These coastal areas are often characterized by a seaward thinning and fining veneer of modern shoreface sands resting disconformably on Pleistocene or older strata. The modern sand veneer is ephemeral and easily removed from the shoreface during storms, during which time the older strata crop out on the shoreface and undergo erosion. These older units not only control the geometry of the shoreface, but also are an important source of 'new' sediment to the modern beach system.

Kitty Hawk to Cape Hatteras. The coastal system from Kitty Hawk to Cape Hatteras is migrating up onto the interstream divide that occurs between Albemarle Sound and Pamlico River. This divide controls the morphology of Pamlico Sound, occurs in the subsurface below this portion of the Outer Banks, and forms extensive paleotopographic shoal structures with Pleistocene cores on the inner continental shelf. The paleotopography of this interstream divide controls changes in the orientation of the barrier islands and the occurrence of minor cape structures that occur on the barrier beach at the towns of Rodanthe, Avon, and Buxton. For example, Wimble and Kinnakeet Shoals are a series of ridges oriented at oblique angles to the barrier, have significant relief, and rise well above the present ravinement surface. High-resolution seismic data and drill records demonstrate that major Quaternary units crop out to produce these scarped hardbottom features.

Nearshore topographically high features such as Wimble and Kinnakeet Shoals can have dramatic impacts upon the energy regime effecting the adjacent beach area through wave refraction and wave setup. The mini-capes that occur at the towns of Rodanthe, Avon, and Buxton are separated by cusped-shaped barrier beaches. Bathymetric charts suggest fairly steep and deep beachface profiles occur directly off the headlands with the actual morphology probably being controlled by outcropping units in the beachface. Whereas, in the cusped portion of the barriers adjacent to the headlands, beachface profiles are relatively broad and shallow. Long-term average annual erosion rates correlate with the offshore bathymetry and shoal structures. Not only are the irregularities in the shoreline profile opposite the north end of each shoal system, but the only areas of shoreline accretion occur opposite the major portion of the shoal structure with shoreline recession on both the updrift and downdrift reaches. Similar patterns exist along the coast north of Oregon Inlet to the Virginia line.

Onslow Beach to Topsail Island. The New River Inlet coastal area forms a seaward protruding bulge in the coastline of central Onslow Bay. This shoreline bulge is the product of a submarine headland composed of indurated limestones of the Oligocene Silverdale Formation. The Silverdale Fm crops out at sea level in the New River estuary mouth, occurs extensively on dredge spoil islands of the Intracoastal Waterway behind Topsail Island and Onslow Beach, and forms a series of paleotopographic ridges on the inner shelf on either side of New River Inlet. These prominent arcuate submarine rock scarps occur just seaward of the lower beachface, have up to 5 meters of relief above the surrounding sea floor, and extend well above the ravinement surface. The hardbottom scarps are subparallel to the beach and intersect the shoreface on both Topsail and Onslow Beach, subdividing each of these barriers into major physiographic segments. The rock ridges continue under the barriers and back-barrier estuarine systems, where they have been mapped by core drilling, and parallel the geomorphic pattern of older Pleistocene beach ridge systems.

The presence of these subsurface headland features have produced dramatically different coastal orientations for the different physiographic segments, as well as extremely different coastal characteristics and processes. For example, on Onslow Beach, the northern segment occurs on the seaward side of the Oligocene ridges and is characterized by cusped shoreline geometry with wide beaches, a recurved accretionary beach ridge, a nearly continuous dune ridge, and shoreline accretion rates that average 2 m/yr. The southern segment occurs on the landward side of the Oligocene ridges and is characterized by a narrow beachface with abundant rock gravel on the beach,

a single discontinuous scarped foredune ridge, presence of major washover fans, and current erosion rates up to and exceeding 6 m/yr.

TRANSGRESSIVE SHOREFACE

Transgressive shorefaces are often composed of tight peat and mud sediments, often containing ancient *in situ* tree stumps, that extend from the modern estuaries, under the modern barrier sands to crop out in the surf zone and upper shoreface. A significantly large portion of the North Carolina barrier islands are underlain by such a sequence of estuarine deposits. These fossil units are all Holocene in age, formed in back-barrier estuarine environments, and now occur under the beachface in response to Holocene sea-level rise and barrier island migration.

Transgressive estuarine deposits crop out in the surf zone intermittently along major portions of the barrier islands from Nags Head to Buxton and from Drum Inlet to Cape Lookout. At West Topsail Island, a 0.5 meter thick peat crops out periodically in the surf zone that can be traced laterally around New River Inlet to a modern back-barrier salt marsh. Underlying the peat is a tight gray clay of unknown thickness. Storm erosion produces large boulders (up to 0.7 meters across) of peat and clay, along with Oligocene rock fragments from the offshore scarps; these gravels represent a significant input of 'new' post-storm beach sediment.

REGRESSIVE SHOREFACE

Only local and relatively small segments of the North Carolina shoreline are presently characterized by regressive shoreface conditions. These areas generally occur on the flanks of headland features and represent temporary episodes of coastal progradation that usually alternate with episodes of longer-term truncation as the headland recedes. However, during episodes of shoreface regression, these shorefaces are relatively stable, are characterized by concave progradational geometries, beach ridge accretion, dune ridge development, and have the potential for approximating the idealized "profile of equilibrium". The northern segment of Onslow Beach, along with portions of Hatteras Island, Ocracoke Island, Shackleford Banks, and Bald Head Island are examples of this type of system.

CHANNEL-DOMINATED SHOREFACE

An estimated 50% of the barrier beaches of the Outer Banks have been occupied by inlets during the recent past. These portions of the Outer Banks are underlain by thick accumulations of either inlet fill sands and gravels or fluvial--estuarine sequences of channel-fill sediments. Over 70% of the barrier islands south of New River Inlet have channel-dominated shorefaces as a result of inlets migrating along the island during the last several centuries. For example, the entire Wrightsville Beach barrier island is a channel-dominated shoreface which is underlain by inlet fill sands and gravels that have formed in response to a series of rapidly migrating Holocene inlets of very recent age. Where channel-dominated shorefaces occur, there are often morphological responses of the beach system to wave and current energy that are quite different than either transgressive or regressive shoreface barriers.

LONG TERM WAVE HEIGHT AND BEACH PROFILE CHANGES, NARRABEEN BEACH, AUSTRALIA

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1 Introduction

Wave height at Sydney has been recorded by Waverider buoys since April 1971. On nearby Narrabeen - Collaroy Beach monthly beach profiles have been surveyed at six locations, since April 1976. The two data sets now 22 and 17 years in length permit a view of both long term wave conditions together with beach response along Sydney's longest beach, including the impact and recovery from major wave erosion events. It is the aim of this paper to examine the relationship between wave climate and beach profile changes along a beach with a longshore decrease in breaker wave height.

Narrabeen Beach is 3.6 km in length, is bordered by prominent headlands and rock reefs, and faces east exposing it to the full force of north east and east waves. The dominant south east waves arrive unimpeded at the northern end of the beach, but owing to a headland extending 1.5 km out from the southern end, decrease to 30 % of H_o along the southern 1 km. The beach is composed of fine to medium sand. Mean tide range is 0.95 m with a spring range of 1.6 m.

Five shore perpendicular beach profiles were established along the beach in April 1976. The northern three (1, 2 and 4) are exposed to waves averaging 1 to 1.5 m and have a modal transverse bar and rip beach type. The southern two (6 and 8) receive waves averaging 1 m and 0.5 m and are modally transverse bar and rip and low tide terrace respectively. All five profiles been surveyed at low tide at approximately monthly intervals. Each profile commences at a permanent benchmark. The elevation of the sand is recorded from the bench mark at 10 m intervals, across the shoreline to variable distances into the surf zone. The data set in this analysis runs to March 1993 and includes 198 monthly surveys.

2 Wave climate

Sydney has a mean deepwater wave height (H_o) of 1.6 m ($sd = 0.24$ m), period (T_s) = 8 sec ($sd = 0.43$), with waves arriving from the south east (40 %), east (41.5 %) and north east (18.5 %). Mean monthly wave height peaks in February-March and June ($H_m = 1.7$ m), with two troughs in April-May and August-September ($H_m = 1.4$ m). Mean annual wave height and power show a weak periodicity since 1971, with higher annual means in the mid 1970's and mid 1980's (Short and Trenaman, 1992).

3 Beach response

3.1 Temporal response

During the 17 year survey period major wave erosion events, together with seasonal variation in wave height produced oscillations in beach width and volume. These are evident in Figure 1 which shows all profiles recovering from extreme erosion events in 1975 (Thom and Hall, 1991), peaking in the early 1980's, followed by erosion in the period 1984-1988, then general recovery into the early 1990's. The net trend over the survey period is accretion (Figure 1).

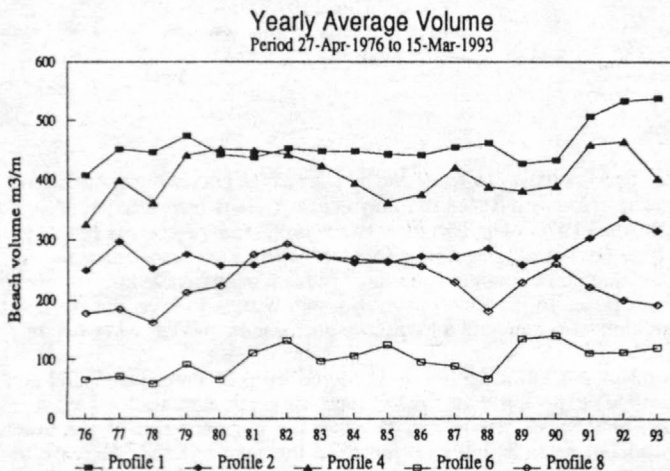


Figure 1. Average annual beach volume along Narrabeen Beach (1976-1993), for profiles 1, 2, 4, 6 and 8.

3.3 Seasonal trends

Figure 2 plots the monthly mean beach width for each profile. It illustrates two trends. Profiles 1, 2 and 4 show a synchronous increase in width into late summer (February to April), decreasing in early winter (May to July), then increasing into late winter - early summer (August to November). These trends reflect the seasonal wave climate (Short and Wright, 1981; Short and Trenaman, 1992).

Profiles 6 and 8 display a different trend. They have a similar late summer peak, then a long decrease in width through to September for profile 8 and December for profile 6. This delayed recovery during a period of generally lower waves can be attributed to their still lower breaker wave height hence slower recovery. Furthermore the greater variation in width at profile 8 (Table 1), can be explained by its location 200 m from the southern headland, which traps sand during summer north east wave conditions, while rapid erosion occurs during higher winter southerly waves.

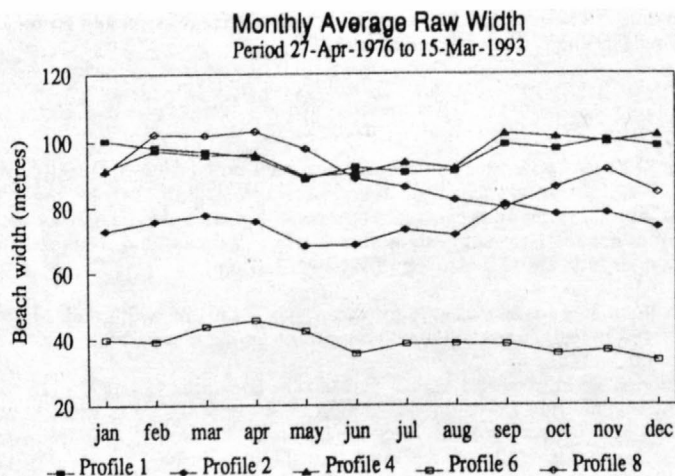


Figure 2. Average monthly beach width and volume for the period 1976-1993 for profiles 1, 2, 4, 6 and 8, Narrabeen Beach.

Table 1. Morphometric properties for profiles 1, 2, 4, 6 and 8, Narrabeen Beach, 1976-93.

Profile	1	2	4	6*	8
n	198	197	197	196	197
mean volume (m^2)	142	132	175	94	145
sd volume	34	31	46	32	47
cv volume	0.24	0.24	0.26	0.34	0.32
max volume	255	214	286	181	261
min volume	70	37	44	36	43
mean width (m)	55	54	65	39	71
sd width	13	12	16	11	18
cv width	0.24	0.22	0.24	0.29	0.25
max width	94	93	100	73	124
min width	28	24	22	14	27
volume change (m^3/m)	52	33	39	51	40
width change (m)	4.8	0.8	-1.5	11.2	-1.2

* profile 6 is backed by a seawall which limits volume and width.

3.3 Spatial response

The temporal trends are not uniform or synchronous longshore. Table 1 lists the mean and variance in both profile volume and width. Beach width varies on average up to 73 m and volume by $193 m^3/m$. Absolute changes in width and volume is greatest at the lowest energy profile 8, and at the central high energy profile 4. Neither of these profile have closure at 5 m depth (AHD). Indicating the beach sand is exchanging with the nearshore zone. Other evident suggests this exchange extends

out to 15 m depth. The mechanism for the seaward exchange is related to megarips during high wave (>3 m) conditions (Short, 1985).

4 Conclusions

- Net beach width and volume accretion has occurred on Narrabeen Beach between 4.76 and 3.93. Total beach accretion averages 2.8 m, or 43 m³/m and 155 560 m³/m. However, as the Narrabeen Beach sand budget is at best balanced, this net increase suggests a short term (17 year) gain, rather than the expected longer term trend (balanced to slightly negative).
- The beach fluctuates as much as 96 m (mean max 73 m) in width and 243 m³/m (mean max 193 m³/m) in volume in response to changing wave conditions.
- Severe beach erosion occurred over a period of a few months in 1975, 1978 and 1986. The magnitude of erosion varied along the beach, but was greatest at the lower energy section, owing to megarip locations. The erosion was followed by slower beach recovery taking up to 2 years on the higher energy section of beach and 5 years on the lower energy section.
- Non closure of some profile envelopes at 5 m depth indicates sediment exchange with the nearshore zone, which in part explains the slow beach recovery.
- Seasonal trends in wave height and power are reflected in seasonal (monthly) beach volume, indicating an expected inverse relationship between wave height/power and beach width/volume..
- The high degree of both temporal and spatial variance in beach width and volume indicates the need for long term (> 20 year) monitoring of shorelines to understand both the scale and mechanism for on-offshore and alongshore sediment exchange, as well as long term shoreline trends.

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The Effects of Wave Event Sequencing on Long-Term Beach Response

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1. Introduction

The long-term prediction of coastal morphodynamic behaviour is an increasingly important issue in coastal engineering design, where the performance of natural systems and artificial shore protection schemes need to be predicted over timespans of decades.

The understanding of such long-term behaviour is still very much at an early stage. The inherent *long-term morphodynamic response* of coastal systems is strongly non-linear, with possible limitations on predictability through chaotic behaviour. The *hydrodynamic driving forces* are a combination of systematically varying medium-term processes (tidal and seasonal variations), systematically varying long-term processes (sea-level rise, climatic changes), and stochastic processes at all timescales (meteorologically induced events, especially wave conditions). The generic types of morphodynamic response, their relationship to the hydrodynamic driving forces, and the relative importances of the different types of driving force and their timescales are all poorly understood at present.

Attempts at *modelling long-term coastal behaviour* are at a correspondingly early stage. The best-established approach is that of one- and n-line beach planshape models which have been used commercially with some success. However, they employ substantial a priori assumptions about the relevant long-term hydrodynamic driving forces, and assume deterministic and predictable morphodynamic responses. Many relevant physical processes are tied up in site-specific calibration factors. More recent modelling possibilities are provided by medium-scale profile and area models. Although these models could (in principle) be run for long time periods, the validity of the long-term predictions would be questionable even if the models were well-verified at the medium timescale. Such procedures would also ignore questions of inherent predictability and stochastic input. A more considered way forward would be through suitable input data reduction ("input filtering") and the formal averaging over short and medium timescale processes ("process filtering") so that the models operate directly at large timescales. Establishment of appropriate methodologies for addressing problems of predictability and stochastic input would also be needed.

Another major problem is the scarcity of suitable *long-term data* to help understand processes and verify models. Ideally, beach and seabed levels at a site would need to be measured over several decades, but with a sufficiently fine temporal resolution to resolve individual meteorological events, so that the effects of medium-term variability on long-term trends can be ascertained. Simultaneous measurements of hydrodynamic forcing conditions are also needed.

The present work considers the importance of medium-term stochastic variations in wave conditions on long-term morphodynamic trends. Measurements from a long-term data set are analysed, and modelling methodologies using medium-scale profile models are suggested.

2. Analysis of Field Data

Measurements of beach levels have been made on a series of 18 upper-beach profile lines along a 20km stretch of the Lincolnshire coastline in England (Figure 1). The profile lines are approximately 100m long with measurements taken at about 10m intervals at a time separation of about 4 weeks, over a 30-year period between 1959 and 1990. The data have been analysed to determine linear trends of beach level at each point on each profile over the full 30-year period and over constituent sub-periods of 5 and 10 years duration. Examples of data output are shown in Figures 2, 3 and 4, with a summary of beach level trends over the 5, 10 and 30 year periods at one site in Table 1.

The following general observations are made on the data:

- (a) At each point on each profile, beach levels fluctuate strongly over the medium term (typically by 0.5 to 1m over periods of 4-12 weeks) but show only a very gradual 30-year linear trend (typically between +1cm/yr and -3cm/yr. + denotes accretion and - erosion).
- (b) The 10-year and 5-year linear trends are progressively larger than the 30-year trend. The example in Table 1 shows 10-year linear trends between +9cm/yr and -13cm/yr, and 5-year trends between +14cm/yr and -32cm/yr.
- (c) At a few locations (eg Figure 4) there is a sudden severe erosion event in which beach levels drop by 2-3m over a period of a few weeks with only partial recovery over the next ten years.

Although further analysis needs to be done, some tentative conclusions can be drawn:

- (a) The variations in linear erosion/accretion trends between the 5, 10 and 30 year periods suggests a substantial contribution from the stochastic nature of the seabed response. At the 5 and 10 year periods this would seem to be dominant. At the 30 year period, contributions of the stochastic and systematic responses to the linear long-term trends are of similar magnitude.
- (b) There are various sources of the stochastic response. One possibility is the variation in average wave conditions between successive 5 or 10 year spans. This can be checked from wave measurements. A second possibility, and perhaps the most important, is the effect of sequencing of wave conditions (chronology) within each of the timespans. A third possibility is the inherently chaotic response of the seabed, although this would tend to be masked by the random sequencing of wave forcing conditions.

3. Computational Modelling of Chronology Effects

The foregoing data set provides evidence (at least circumstantial) for the important role of wave event sequencing in determining long-term erosion/accretion trends over 5, 10 and even 30 year timespans. It would be possible to test this effect by using computational models that reproduce reliably the morphodynamic response at medium timescales (days, weeks) and which can be run continuously for large timescales (months, years and decades). The method would involve selecting suitable measured or synthetic wave data lasting months or years and rearranging the events (identifiable storms etc) in a large number of different sequences. Each sequence would then be run through the

model. This is currently being undertaken with HR's COSMOS-2D profile model, and it is hoped to present some initial results at the conference.

If wave chronology has a significant, even dominant, role in determining long-term beach response trends, practical studies would need to provide answers in a statistical rather than a fully deterministic manner. A thorough study would need to consider a large number of possible wave sequences within a given overall climate to determine the probabilities of occurrence of *average* beach levels. This statistical information would be provided in addition to deterministic information on long-term trends resulting from systematic variations in hydrodynamic forcing, plus the short-term beach response to design storm events.

4. Acknowledgements

Separate parts of this work were funded by the UK Ministry of Agriculture, Fisheries and Food, and by the Commission of the European Communities Directorate General for Science Research and Development under contract no. MAS2-CT92-0027 as part of the G8 Coastal Morphodynamics research programme.

The Lincolnshire data analysis was originally carried out for the clients Posford Duvivier and the National Rivers Authority, Anglian Region. Alan Brampton and Penny Curl performed the analysis work at HR Wallingford.

Table 1 Beach Level Trends at Trunch Lane (+ denotes accretion and - erosion. Units are cm/yr)

Dates	Chainage (m) from landward end		
	15.0	50.0	90.0
MAY-APRIL			
1960-1965	+6	+6	-11
1965-1970	+14	+4	+13
1970-1975	-7	-8	+9
1975-1980	-4	-3	-12
1980-1985	-29	-32	-23
1985-1990	-31	-17	+6
MAY-APRIL			
1960-1970	+9	+4	-1
1970-1980	-1	-6	-2
1980-1990	-8	-13	-6
OVERALL			
1960-1990	-2	-3	-2

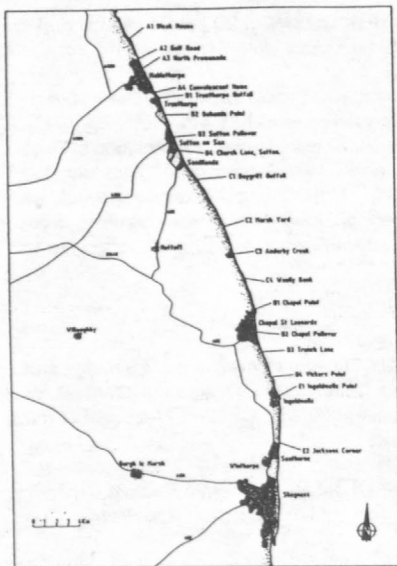


Fig 1 Location Map. Lincolnshire Coastline, UK

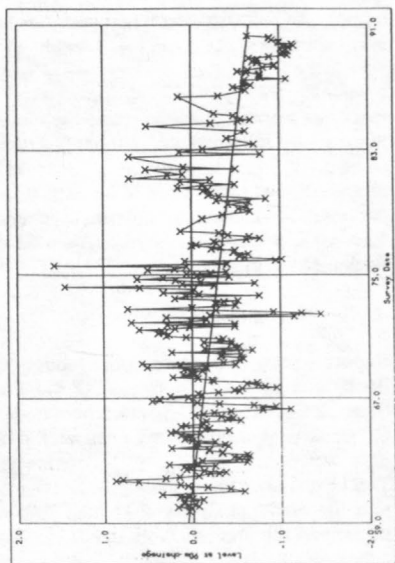


Fig 2 Time History of Beach Levels at Vickers Point (90m Chainage Position)

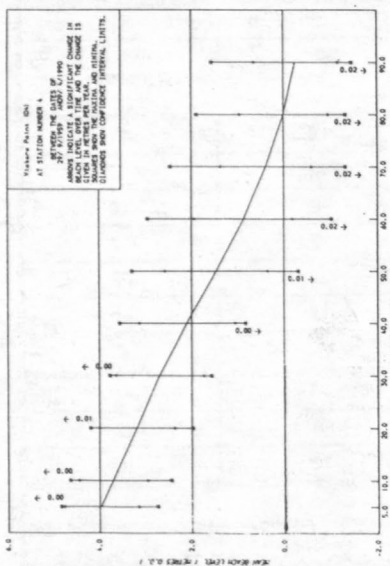


Fig 3 30-Year Mean Beach Profile at Vickers Point

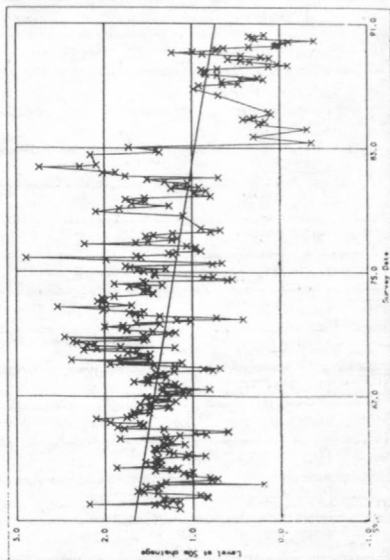


Fig 4 Time History of Beach Levels at Trunch Lane (50m Chainage Position)

REGIONAL IMPACTS OF INLET ENGINEERING AND BEACH REPLENISHMENT AT FENWICK AND ASSATEAGUE ISLAND, MARYLAND

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INTRODUCTION

Analysis of several diverse engineering and scientific studies have been combined to evaluate the effect that recent inlet engineering, beach restoration, and storms have had on regional coastal evolution of Assateague and Fenwick Islands, Maryland. Urbanized shorefront development has occurred on Fenwick Island, in contrast to the largely undeveloped nature of Assateague Island. The use of comprehensive data sets has allowed detailed examination of coupling shoreline change trends, barrier island morphology, and inlet shoal evolution to recent inlet jetty rehabilitation, beach replenishment, and three large storm events. Ebb shoal growth and natural sand bypassing are now providing sand to a portion of the sediment starved downdrift beach. Evidence indicates that the nourished beach is providing storm protection to upland development on Fenwick Island, while the north end of Assateague is still impacted by large scale washover processes.

Prior to August 1933, Assateague and Fenwick Islands were part of a continuous barrier spit connected directly to a coastal headland along the Delaware-Maryland-Virginia (Delmarva) coastline. Numerous shoreface-attached and other linear shoal features are present in the nearshore (Figure 1). The 1933 hurricane breached the barrier spit creating the present Ocean City Inlet. Shortly after formation, local interest decided to keep the inlet open for commercial and recreational purposes. Because the southerly prevailing longshore drift would have probably closed this breach, as in 11 previous storm-induced spit breaches (Truitt, 1968), the U. S. Army Corps of Engineers (USACE) was authorized to stabilize this inlet. Construction of two jetties and the development of an ebb-tidal delta interrupted pre-inlet longshore sediment transport patterns. An 2-km-long updrift fillet developed on the north side Ocean City beach, while progressive erosion occurred on downdrift Assateague Island beaches.

FENWICK ISLAND - BEACH REPLENISHMENT

Starting around 6 km north of the inlet fillet on Fenwick Island, there had been long term average erosion of around 0.6 m/yr (Dolan et al., 1980). Continued erosion of the beach threatened the shorefront infrastructure with damage due to storms. In order to provide storm protection, a major beach replenishment project was completed in 1991 along 11 km of Fenwick Island shoreline (Stauble et al., 1993). The project consisted of three separate construction events. The initial fill, sponsored by the State of Maryland during the summer of 1988, placed 2.1 million cu m of recreational beach fill. In March 1989, a series of storms caused removal of an average 67 cu m/m of sand from the subaerial beach, but all of the sand was accounted for in the nearshore area between the -1 and -6 m depth contours. The second phase of the project was constructed by the USACE to enhance storm protection during the summers of 1990 and 1991. The 1990 portion of the project extended from 3rd St. to 100th St. The remainder of fill was placed between 100th St. and the Delaware State line during the summer of 1991. A total of 2.9 million cu m of fill was placed during these two summers of the Federal fill. Two large extratropical events in October 1991 and January 1992 impacted this second fill, removing an average 112 cu m/m from the subaerial beach and depositing 96 cu m/m of sand on the upper shoreface out to the -6 m contour. From analysis of 4.5 years of profile surveys, a depth of closure has been identified around this -6 m contour. Seaward of this depth less than 0.15 m of elevation change has occurred.

Two wave gages were in operation in 10 m of water seaward of the project and provided useful wave and surge information. Variability of erosion along the length of the beach fill, which had "hot spots" of erosion, was the result of two shoreface-attached sand ridges. Maximum erosion of the fill occurred in the vicinity of shoal attachment to the shoreface. The resulting loss of sand from the beach appears to have been deposited on the shoreface and downdrift at the north jetty of the inlet and on the ebb shoal. The beach at the updrift end of the project exhibited a steep non-barred profile, while the south end of the project had a flat, nearshore bar profile.

ASSATEAGUE ISLAND - DOWNDRIFT INLET EFFECTS

As the northern jetty and ebb-shoal trapped littoral material, shoreline recession rates along northern Assateague Island changed from pre-inlet conditions of around 0.9 m/yr to greater than 12 m/yr (Leatherman, 1984). Historic shoreline change indicated that the inlet erosional influence extends around 9 km to the south of the inlet. Since 1933, the shoreline and dunes of northern Assateague have eroded landward until washover penetrated across the full width of the island, depositing sand into Sinepuxent Bay. Barrier width has remained dynamically constant as the entire island form has migrated landward one island width. As the barrier has been retreating landward, a washover flat area devoid of any dunes or vegetation was created between 1 and 4 km south of the inlet. After a November 1961 storm, a temporary inlet formed within this low lying area 2 km to the south of the present inlet. This breach was closed by placement of 1.2 mill cu m of material, but the island elevation remained low at this location. Further to the south, within 9 km of the inlet, the primary dune line was bisected by frequent overwash channels and washover fans developed at low points between the dunes. During severe storms, this part of the island was often inundated with sand transported inland onto the back barrier

and to the bay. South of this 9 km long segment, the shoreline has been relatively stable, with a higher, more continuous primary dune line and less overwash.

Major storm processes continue to play an important role in washover and barrier island change. The October 1991 and January 1992 storms have highly modified the north end of the island. The first storm, the "Halloween" northeaster, was larger in size and intensity, and longer in duration, but was not as damaging in respect to beach erosion and overwash as the January northeaster. The second storm was more severe due to the easterly direction of wind and resulting higher storm surge (Jensen and Garcia, 1993). The combined effect of both storms resulted in most of the low dunes and narrow overwash channels being replaced with a large washover flat deposit that extended from the ocean to the bay along almost the entire northern 9 km of the island. The only exception was adjacent to the south jetty, where vegetation had colonized sand placed there during maintenance dredging of the navigation channel.

OCEAN CITY INLET- ENGINEERING AND EBB SHOAL EVOLUTION

Analysis of inlet shoal volume data suggest that the Ocean City Inlet ebb shoal volume increased steadily from inlet formation in 1933 to 1962, slowed from 1962 to 1978, and again increased over the period 1978 to 1990 (Underwood and Hiland, in press). This ebb shoal growth is allowing bypassing of littoral materials to northern Assateague from natural longshore drift, with possible additional input from the beach fill. The ebb shoal has accumulated some 6.1 million cu m over its 60 year history, which averages to a rate of growth of 102,000 cu m/yr. The distal end of the ebb shoal has now welded itself to the shoreline around 500 m south of the inlet. The shallowest portion of the shoal is on the southern side, forming two parallel crests. Bypassing of sand may result in possibly reducing future shoreline recession and landward island migration, at least in the vicinity of the inlet. The growth of the inlet ebb-shoal resulted in localized wave refraction and northerly littoral drift along this northern 500 m of Assateague Island. Movement of littoral materials into the inlet throat and subsequent shoaling of the ship channel were halted by sand tightening of the south jetty, resulting in beach accretion within this area. The rate of erosion for the northern 6 km of Assateague Island has slowed from around 9 m/yr to around 6 m/yr. Storm impacts however, continue to overwash the low portions of northern Assateague Island south of this inlet shoal influence. An understanding of the entire system is needed to assess barrier island responses before any management alternatives or additional coastal engineering can be put in place. Permission was granted by the Chief of Engineers to publish this information

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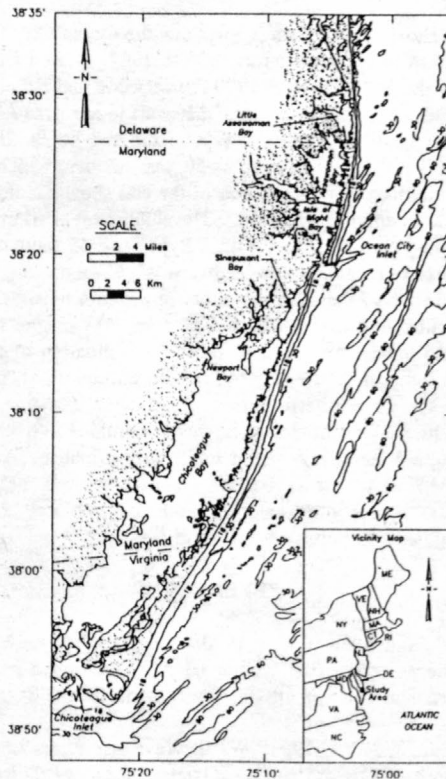


Figure 1. Location map of Fenwick and Assateague Islands, with Ocean City Inlet and nearshore shoals

SHOREFACE PROFILE EVOLUTION ON THE TIME SCALE OF SEA-LEVEL RISE

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ABSTRACT

A representation of shoreface profile behaviour on the time-scale of sea level rise is presented using the concept of hinged rigid panels. Compared to its initial formulation (Stive et al., 1990) the model was extended with a dynamic component (depending on the morphological state of the profile) besides the "autonomous" component (depending on large-scale transport processes which are assumed to be primarily a function of the water depth). We have verified our model in the context of the studies undertaken for the Coastal Defence Policy Analysis of the Netherlands (1989; cf. De Vriend et al., 1993a), using long term observations over the last century. Although it is concluded that the principal shoreface evolutions are reasonably well represented, it must be stated that the approach is still in its infancy. There are some important unanswered questions yet and the validity of the concept needs to be checked more extensively.

INTRODUCTION

The demand for long-term morphological predictions needs to be met with the knowledge we have available at this moment, viz. short-term process-knowledge based on first physical principles and empirical knowledge based on observations in a large number of similar systems. For a long time, these two sources of information were separated: a study was approached either from the empirical end or from the process-based modelling end. In either case important resources were ignored, which is not quite recommendable in an area like this (cf. Terwindt and Battjes, 1990).

One step towards integration is formed by the diagnostic utilization of process-based models, to hindcast and analyze a morphological system's past behaviour. Combined with empirical knowledge from similar cases, these models can then be used to predict future evolution. Another way is to integrate physical knowledge and measured data in data assimilation.

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

Note that the term "behaviour" has a restricted meaning in this context: it does not refer to the morphological evolution in all its complexity, but only to certain aspects, e.g.

within a given range of space and time scales. For instance, if we consider long-term behaviour, this implies that short term fluctuations as such (not their residual effects!) are ignored. A behaviour-oriented model therefore represents only these aspects of the coastal behaviour, whence it is less generally applicable than a process-based model.

In essence, the approach is not entirely new. Multiple coastline models, for instance (see Bakker, 1968, or Perlin and Dean, 1983), make use of an empirical relationship between the cross-shore transport and the profile slope. In a continuum representation this boils down to a diffusion model, which is rather popular for behaviour-oriented coastal profile models (for instance, see De Vriend et al., 1993b).

BEHAVIOUR-ORIENTED LARGE-SCALE MODELLING OF A COASTAL STRETCH

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

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

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

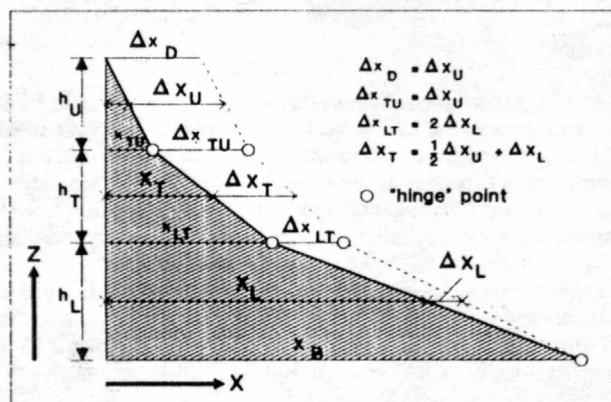


Figure 1 Large-scale coastal behaviour model: subdivision of the profile in panels (U = active zone, T = middle shoreface, L = lower shoreface)

The lower part of the shoreface is much less dynamic, though not immobile at this timescale (cf. De Vroeg et al., 1988). If we assume the inner shelf to be at rest, it seems reasonable to fix the position of the lower end of the shoreface. Thus we have the following cross-sectional units (see Figure 1):

the dune area, which is assumed to be morphologically separated from the other

parts of the profile, and to act as a sink of sediment transported by wind over the top of the dune front,

- the active zone, from the top of the dune front to approximately 5 m waterdepth, which moves as a rigid body along with mean sea level,
- the middle shoreface, a transition zone between 5 m and 12 m waterdepth, which is "hung up" between the active zone and the lower shoreface, from 12 m to 18 m waterdepth, which tilts about its foot point, and
- the inner shelf, below 18 m water depth, which is assumed to be morphologically at rest at these timescales.

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

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

In the full conference paper a table will be included giving an overview of the general behaviour and the possible displacements of the various cross-shore units. It also gives an indication of the underlying processes.

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

We have verified our model in the context of the studies undertaken for the Coastal Defence Policy Analysis of the Netherlands (1989; cf. De Vriend et al., 1993a), using long term observations over the last century. It is concluded that the principal shoreface evolutions are reasonably well represented.

DISCUSSION

The panel model was applied to two representative cells of the central part of the Netherlands coast, which is a (now) weakly erosive Holocene based coast, without important coastal inlet interruptions. The time span considered concerns the evolution over the last century. As far as we know, the external conditions for this evolution are relatively gradually changing, such that assuming constant conditions does not alter the model hindcast principally. We feel that an important check and extension of the panel model should be found when trying to model the shoreface profile behaviour over longer and/or other time spans. For instance, both on the Netherlands coast (Beets et al., 1992) and the French Aquitaine coast (Froidefond and Prud'homme, 1991) there has been an important, relatively sudden or at least with a clear change-of-trend, phase in construction of Younger Holocene Dunes somewhere between 1500 and 400 years BP. An important question remains which external conditions ("Little Ice Age"?) and which initial shoreface profile state we have to

assume, but we are attempting to hindcast these evolutions with a panel-type model based on most likely scenarios.

In an interesting critical review paper about shoreface profile modelling (Pilkey et al., 1993) the concept of shoreface equilibrium profile is criticised, pointing out some pertinent weaknesses behind the assumption of the Bruun equilibrium concept when applied to the whole of the shoreface. Although the panel method acknowledges most of these critics, by taking into account that the equilibrium profile only exists in the upper shoreface, that there exists no depth-of-closure on the shoreface, and that there exist net current effects along- and cross-shore, there remain important aspects, such as the influence of underlying geology, sediment grading and most importantly the validity of the assumption of an "equilibrium" shape of the upper shoreface (what is the role of bars here?). These aspects form part of our research.

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THEORETICAL CONCEPTS OF PARAMETERIZATION OF COASTAL BEHAVIOUR

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Sandy shorelines are very dynamic. Rapid adaptations of the morphology occur if the driving forces and the morphology are out of balance. These adaptations of the morphology are coupled with large amounts of sediment transport and result in a large gross sediment budget.

Over a longer time scale, the adaptations of the morphology tend to reach a state of dynamic equilibrium. This means that the original and final beach state are almost the same, irrespective of the progress of the adaptations. However, the non-linear character of the sediment transport processes, and the difference between the quick responses of the driving forces and the slow adaptations of the morphology may prevent a state of dynamic equilibrium and induce a small deficit in the sediment balance. This deficit in the sediment balance is the net sediment budget and may show a certain trend in time. It is this trend which determines the behaviour of the coastline.

The deficits in the sediment balance are scale-dependent as illustrated in figure 1.

On the small scale the morphological changes may be rather large, reflecting the storm to fair weather or seasonal variations, but the position of the coastline shifts around a mean value and shows a trend in time. This trend of the mean position of the coastline is superimposed on the trend in a higher scale and so on. The transition between scales is gradual.

The selection of the scale of interest in a certain coastal problem is arbitrary. However, once selected, this scale establishes the scale of the morphodynamic entities and the related scale of predictions of coastal developments.

An example of different entities (from small to meta) is given for the Dutch coast:

- the coastal cell, which is made up of the beach, a crescentic inner nearshore bar and rip system, and an outer nearshore bar;
- a set of coastal cells;
- the coastal stretch, which is made up of several sets of coastal cells;
- the coastline, which is a succession of coastal stretches.

On this premise the small-scale coastal behaviour has a spatial scale of the

dimensions of the cell, the meso scale that of a couple of cells, the large scale that of the coastal stretch, and the meta scale that of the coastline. Herein, the longshore scales are varying over larger ranges (1 to 100 km) than the cross-shore scales which may never be beyond the dunes and lower shoreface. The selected morphodynamic entities also determines the time scales involved. The behaviour of the morphodynamic entities along the coast may differ. There is no general spatial and temporal scale of the entities. The scales are site-specific and not always related in the same way between different entities.

The specification of the physical processes for the various entities is hardly possible. There is the uncertainty of having identified all acting processes. Besides, it is doubtful that the interaction between hydrodynamic processes and morphological responses are adequately known. This is partly due to the lack of knowledge about the sediment transport processes. Therefore, a schematization of the real world is needed. This schematization is executed by means of parameterizations. The parameters correlate variables of the steering forces and the net responses over a certain time frame. This implies that every defined scale has its own parameterization, which highlight the development of the net sediment budgets in time. The selection of these variables may rely on the present knowledge of the processes in the coastal zone.

The location of the variables in the parameter might be a problem. The hydrodynamic processes in the nearshore zone are showing large gradients, especially in cross-shore direction. The location of the parameter must be chosen and may be preferentially placed at the seaward boundary of the morphodynamic entity in order to avoid the rapidly changing hydrodynamic variables in the entity itself (both spatial and temporal). Besides, it is not the gross budget, but the net budget which is of importance to the coastal development. This implies that with every larger scale step the boundary is located further offshore.

After the specification of the structure of the parameter, the variation in the numerical values of the parameter alongshore and/or in time must be interpreted. In fact, two possible causes can induce this variation:

- an external forcing, which implies that the spatial differences in behaviour are generated by the spatial differences in the external forcings (the seaward boundary conditions, the variables in the coastal parameters);
- an internal steering mechanism, where the external forcings remain the same, but that the internal difference within the coastal system (e.g. slope) produce the different behaviour.

The decision which of the two possibilities operates along a certain coastal area is difficult to make and may rely on an analysis of appropriate data sets of hydrodynamic variables as well as morphological ones at the specific site. However, the variations in the time domain are often coupled to the differences in external forcings, whereas the spatial differences are often coupled to internal differences.

The aspect of the transitions between the various scales in coastal behaviour is finally discussed.

The interrelations between the entities are often difficult to determine. The transition between the scales in figure 1 is gradual and is often not so obvious, because the differences in scale dimensions of the entities are too large. This means that the ratio of cycle times and frequencies within two distinct entities are too different.

The problem of upscaling the results of the morphodynamic behaviour of a lower entity to a higher entity may be solved in two ways (figure 2):

- using the input data of the lower entity and integrate the coastal parameters of these series over the successive higher entity; this means that all the small-scale variations in process-variables are included to find the morphodynamic behaviour at a higher scale;
- defining a new coastal parameter at the successive higher entity, neglecting the coastal parameters of the lower entities.

An example of the first approach is presented by Larson and Kraus (1992). They use a small-scale parameter, derived for the prediction of erosional and accretional events at the beach, to understand the nearshore bar behaviour over a decade.

The second approach implies that each entity has its own autonomous development. The interactions between the entities are unknown, because of the differences in scale dimensions. The applied driving process-variables in the new coastal parameter are crudely schematized. This approach is often the first step in analyzing the multi-scale coastal behaviour. An example of this approach is presented in Kroon (1994), using a storm number in describing the nearshore bar behaviour over a decade.

The theoretical considerations presented above proved to be rather helpful for the set-up and the analysis of measurements in the coastal zone.

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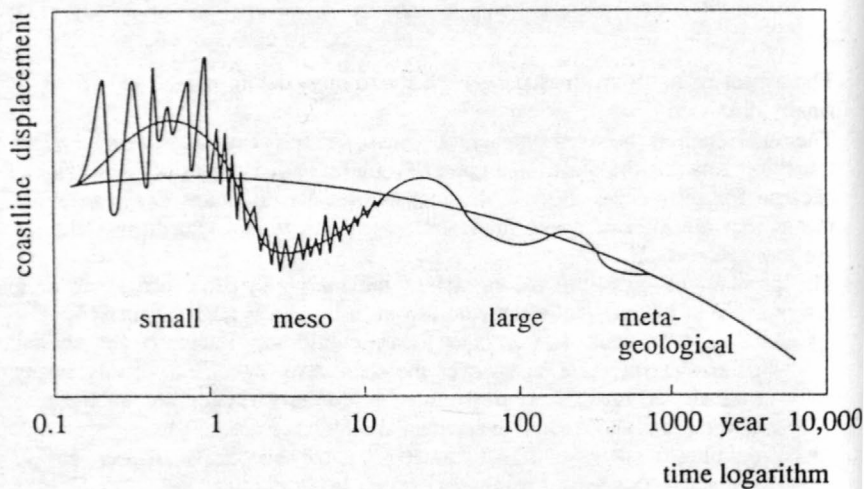


Figure 1. Scales of variations in coastline position

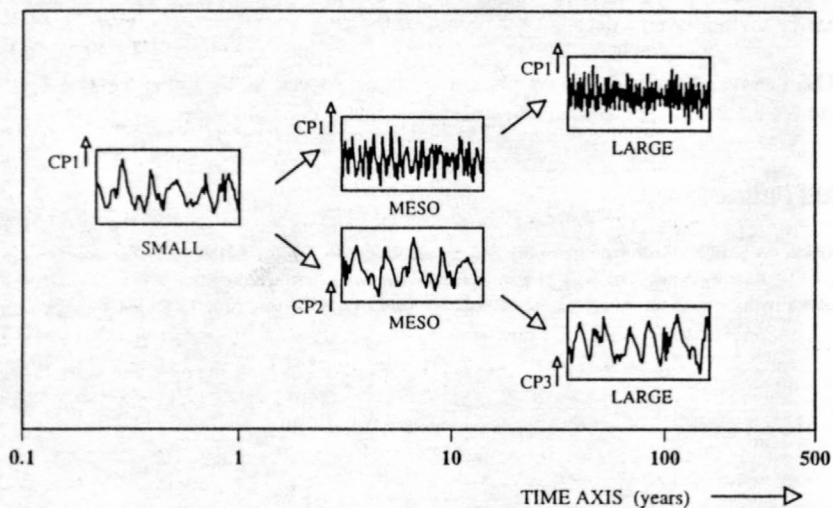


Figure 2. Approaches to scale transitions
CP = Coastal Parameter

NUMERICAL MODEL OF THE PROPAGATION OF LONGSHORE SAND WAVES

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INTRODUCTION

Longshore sand waves (LSW) are longshore wave-like movements of the shoreline, and their formation was first described in detail by Evans (1939). Bruun (1954) noted the significance of LSWs as migrating large-scale morphologic features. Although LSWs are not well understood, it is known that their length can range from tens of meters to kilometers, and their speed of motion is on the order of km year⁻¹. LSWs, therefore, hold implications for both understanding nearshore morphology change and for conducting engineering activities in the coastal zone (Bruun 1954, Verhagen 1989). Some creation processes of LSWs are known. For example, energetic sediment discharges at inlets and rivers and limited-extent beach fill placement can create sand bodies observed to move alongshore. These bodies eventually decay or diffuse alongshore and offshore, but, during their lifetime, they are identifiable as distinct morphologic entities. A LSW can impound sand on its updrift side and cause erosion on its downdrift side, even as it moves (Grove et al. 1987).

Longshore sand waves are one manifestation of "collective movement" of sediment (Sonu 1968) in which a sand body maintains morphologic identity during a life which may reach years. Other examples of collective movement of sand bodies are inlet shoals, longshore sand bars, and transverse sand waves and dunes at inlet entrances and navigation channels. Collective movement should be distinguished from movement of individual sand particles. Kraus and Horikawa (1990) estimate that typical longshore sand particle speeds are on the order of 100 times greater than the speed of LSWs, which is probably in the range of 0.5 to 4 km year⁻¹ (Grove et al. 1987, Inman 1987). The speed of LSWs has been postulated to be inversely related to their length l as $l^{-4/5}$ (Sonu 1968). The speed must also be related to the strength and duration of the hydrodynamic forcing.

Field observations have provided some quantitative information on collective sand motion, but field results are most useful when integrated into an overall quantitative model of morphology change. A quantitative model such as a numerical model of coastal morphology change must be capable of describing sediment motion at two

scales; relatively rapid movement of sand particles and relatively slow collective movement of sand bodies. As a first step in developing this approach, it is convenient to consider LSWs, which undergo simple two-dimensional motion. Larson and Kraus (1991) have already shown that analytic solutions for shoreline change can be modified to describe approximately collective movement of LSWs.

In the present study, the authors developed a numerical model of shoreline change that is capable of describing the collective movement of LSWs in addition to particulate movement as given by a common longshore sand transport rate formula. The model does not possess the limitations of analytic solutions, and, does not suffer from the problem of predicting overly rapid diffusion of sand waves that is obtained in analytic solutions and standard numerical models of shoreline change. Numerical and physical accuracy of the model was checked by testing for sand conservation and wave "transportivity," meaning that the LSW should be convected in the direction of the driving hydrodynamic forces (longshore current). The numerical model, results of sensitivity tests, and an application to LSWs observed along the shore at Southampton, Long Island, New York, are described in this study.

PROCEDURE

Numerical model

Mathematical modeling of shoreline change has become a routine engineering technique for predicting the long-term (order of months and years) evolution of the shape of the shoreline (e.g., Hanson and Kraus 1989). Two basic geomorphic assumptions underlie such models; 1) permanency of beach profile shape (implying existence of a profile equilibrium shape), and 2) existence of a depth of closure of sediment movement. The equation governing shoreline change, which is an expression of mass conservation under the stated and other assumptions, can be modified in phenomenologic fashion for the presence of a LSW moving with velocity V (Inman 1987, Larson and Kraus 1991) by addition of an advective term as

$$\frac{\partial y}{\partial t} + V \frac{\partial y}{\partial x} = \frac{1}{D} \frac{\partial Q}{\partial x} \quad (1)$$

in which shoreline position y is a function of time t , distance x alongshore, depth of closure D , longshore sand transport rate Q , and velocity of the LSWs.

Under further simplifying assumptions, principally that Q is proportional to the incident wave angle and that the waves are constant, Eq. 1 reduces to the one-dimensional diffusion-advection equation

$$\frac{\partial y}{\partial t} + V \frac{\partial y}{\partial x} = \epsilon \frac{\partial^2 y}{\partial x^2} \quad (2)$$

in which ϵ acts as a diffusion coefficient and is a function of D and amplitude of Q . Larson and Kraus (1991) showed that for constant V , a simple change of variables transforms Eq. 2 to the pure diffusion equation describing movement of a sand body with characteristic velocity V that tends to diffuse and lose identity. A problem

appearing in analytic solutions and in simple numerical models of shoreline change is that a LSW diffuses much too rapidly in comparison to observation.

To address these and other problems, in the present study, Eq. 1 was solved numerically, which allows use of a realistic water wave model and general time series of wave input. Both central differencing and upwind differencing were tested in explicit solution schemes to discretize the advective term. The forward differencing scheme preserved transportivity of the wave while conserving mass and was selected. A key feature for preserving the identity of the LSW, i.e., reducing the unphysical rapid diffusion, was recognition of the wrapping of waves around the LSW by wave refraction over curved offshore contours accompanying the LSW. Therefore, a "contour correction" (Kraus and Harikai 1983) technique was incorporated in the wave model.

Another key feature of the numerical model is the assumption that the velocity V of the LSW is proportional to the longshore water discharge accompanying the longshore current. Nasner (1974) found that the migration speed of transverse sand waves in the upper part of a weakly tidal estuary was proportional to the inflow or fresh-water discharge, and Kraus and Dean (1987) obtained good correlation between the particulate longshore sand transport rate and the longshore discharge of water on an Atlantic Ocean beach. Connection of V with the discharge is more reasonable physically than connection with only the longshore current, and contains all information associated with seasonal and other cyclical patterns of the waves moving the sand and the LSW.

Field Data

Field data were sought with which to test predictions of the numerical model. Controlled aerial photographs were obtained showing LSWs along the beach at Southampton, Long Island, New York. Five sets of photographs were obtained for the 16 month interval September 1991 to January 1993. Of these, four sets at the same scale are being digitized for quantification of the LSW size and movement. Scanning techniques will be required to form digital images of the fifth set of photographs.

These LSWs originate as ephemeral ebb-tidal shoals created by either artificial or natural opening of Mecox Bay through a short-lived inlet that pierces the narrow sandy beach. The inlet temporarily opens as the bay accumulates fresh water runoff, through storm action, or through digging of a channel. The shoal eventually migrates landward and welds to the shore, as documented by Zarillo and Smith (1986) and may then become a LSW. Periodic openings of Mecox Inlet are manifested by as many as 11 probable LSWs that appear in the photographs. The waves tend to move westward, in the predominant direction of littoral drift, where they merge with the updrift impoundment fillet at the east jetty of Shinnecock Inlet.

The times at which the photographs were taken are known. This information, combined with beach profile data from the site (Zarillo, personal communication,

1992) and estimates of astronomical tide level, enable digitized shoreline positions to be transferred to the same vertical datum for tracking of LSW shape and position.

In the presentation of results, predictions of LSW movement and shape calculated with the numerical model will be compared to the measurements. Qualitatively, the authors have observed that the model predicts higher rates of LSW movement in the winter, which agrees with behavior observed in the photographs.

ACKNOWLEDGEMENTS

The authors would like to thank Mr. Aram Terchunian, of First Coastal Corporation, Westhampton, Long Island, New York, for bringing the longshore sand waves at Southampton to our attention and for informative discussions. We also thank Dr. Gary Zarillo, Florida Institute of Technology, for providing beach profile data and an unpublished report of a study he performed at Mecox Inlet.

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LONG-TERM EVOLUTION OF THE KASHIMANADA COAST

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Kashimanada is an open sea located on the east of the middle of the Japan island arc (Fig.1). It has a long monotonic coastline of 90 kilometers characterized by beaches of fine sand. The sediment is supplied at the southern end by the Tone river with the greatest basin in Japan, and at the northern end by the Nakagawa river. As the term "nada" in Japanese means a sea having strong and high waves, the waves are relatively high on the coast with a mild bottom slope.

At first sight the depth contours appear to be fairly monotonic so that waves are incoming nearly normal to the shoreline for the natural beaches to be kept stable. However, the surveys on the shoreline changes revealed that on the natural-beach condition there used to be some patterns in which the shoreline continued to advance in some part and to retreat in another. Fig.2 illustrates the shoreline changes from 1903 to 1961 on the natural-beach condition. The parts at 20 to 30 km and 35 to 45 km exhibited distinct advancement while there was a small portion showing retreat between the two parts.

These features turned out to be predominantly due to northerly waves propagating over a shallow area off the northern beach (Fig.3). Note that in the rest of the coast, 0 to 15 km and 50 to 70 km, even qualitative features of shoreline changes are not quite consistent with those in another duration.

Statistics of wave data are available in the north Pacific Ocean for 10 years from 1964 to 1973. The data were originally obtained by visual observations on the voluntary ships navigating in the area. The parameters of incoming waves in the analysis are determined by referring to the offshore-wave statistics and bathymetry. Thus three wave directions are choosed: 10, 20, 30° northerly from E with a wave height of 2.75 m and a period of 10 seconds.

Wave propagation is calculated with a grid size of 200 m in on-offshore direction and of 300 m in alongshore direction over the real depth (no smoothing). Smoothing is only once made at a reference depth of 6 m, to provide input for the one-line model on the shoreline evolution. Calculation of the shoreline changes shows the same characteristics as the obsevation (Fig.4). The longshore-sediment-transport rates thus calculated indicate that it tended to accumulate at 20 km from both directions (Fig.5).

Since 1961 three ports have been constructed on the coast: the Oarai port for fishery and transportation in the north of the coast, the Kashima industrial port in the middle, and the Hasaki fishery port at the southern end (beyond the area of analysis). Fig.6 illustrates the observed shoreline changes from 1961 to 1990. Erosion is apparent on the north of the Kashima port and on the south of the Oarai port, whereas erosion in the area between 23 and 33 km might be due to some errors in the survey.

Judging from the directions of longshore transport (Fig.5), it is known that the beach material has been

lost from the downdrift beaches of the ports. With the parameters established the shoreline changes are calculated after the port constructions (Fig.7). They show the same qualitative features as the observed ones.

With the knowledge on the cause of erosion on the north of the Kashima port (Figs.5 and 6), it is now interesting to investigate what would have occurred had it been constructed at 20km. Fig.8 illustrates the 'imaginary shoreline changes'. As expected it is shown that no erosion would have occurred had the Kashima port been constructed at 20km where sediment accumulates from both directions.

As a countermeasure to conserve the Kashimanada coast, it is proposed to divide the coastline into a number of segments by manmade headlands and to stabilize each smaller beach independently. Fig.9 illustrates the shoreline changes after such measures executed. The resulting shoreline changes are small.

It is concluded that it is indispensable in designing coastal structures to understand the coastal processes as the whole coastline with sufficient knowledge of the incoming waves and offshore bathymetry. The model thus established is expected to help in designing the future development of the Kashimanada coast.

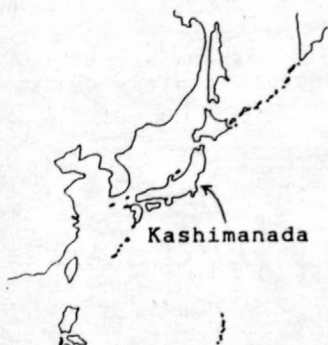


Fig.1 Location of the Kashimanada coast

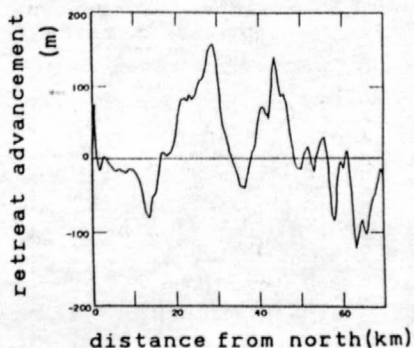


Fig.2 Shoreline changes from 1903 to 1961

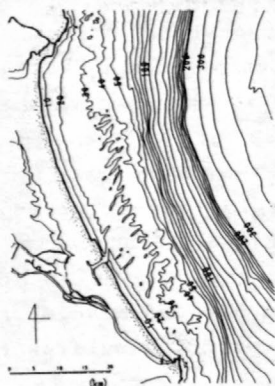


Fig.3 Bathymetry of Kashimanada

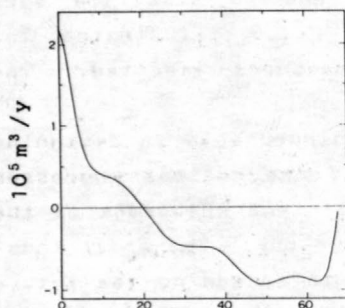


Fig.5 Annual longshore-transport rates

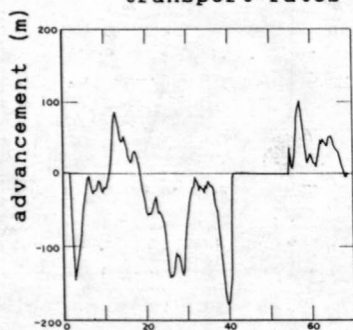


Fig.6 Shoreline changes from 1961 to 1990

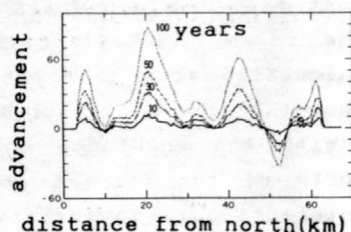


Fig.4 Shoreline changes as a natural beach (calculation)

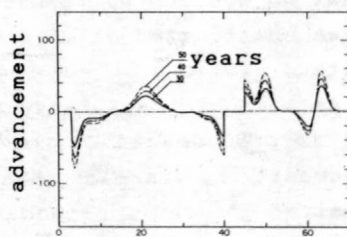


Fig.7 Shoreline changes after port construction

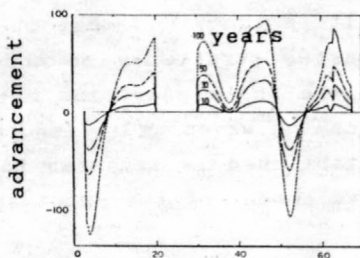


Fig.8 'Imaginary shoreline changes' with the Kashima port relocated

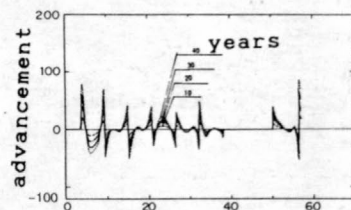


Fig.9 Shoreline changes with manmade headlands

WAVE CLIMATOLOGIES OF THE AUSTRALIAN SOUTH EASTERN COAST

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1 Introduction:

Instrumentally recorded wave data have been analysed for eight locations, spanning 10 degrees of latitude (28° S to 37° S), and covering 1000 kilometres of the Australian south-eastern open ocean coastlines (Brisbane to Eden Figure 1.). Offshore of Sydney, wave data has been collected since 1971, this represents one of the longest continuous wave data sets available. Of the remaining seven sites, five have been in operation since the mid to late 1970's. The remaining two sites began recording data in the early 1980's.

The availability of this data base allows for the first time, a comprehensive analysis and description of the regional wave climatologies of the Australian south-east coast. The characteristics of the Sydney regional wave climates have been detailed by Trenaman (1985), Trenaman & Short (1987) and updated in Short & Trenaman (1992). A summary of the coastal wave climate of New South Wales was presented by Webb & Kulmar (1989). However, until present, there has not been a detailed site by site description of the regional wave climatologies for the Australian south-east coast.

The primary characterisations of each wave climate site are a product of the seasonal variation in the range of the meteorologically driven wind forcing on the ocean surface, and to a lesser known degree, the prevailing regional oceanographic conditions (particularly the sea surface temperature variability).

Anticyclonic high pressure systems dominate the weather systems around the Australian continent. Cyclonic low pressure systems are responsible for the storm systems in the Coral and Tasman Seas. Three cyclonic systems have been categorised, they are (i) the Tropical Cyclone, (ii) the East Coast Low, and (iii) the Mid -latitude cyclone. All of these systems produce waves which impact on the coast.

2 Wave Climates

The wave climate sites have been divided into four separate regions based on a latitudinal gradation from north to south. The first region is the Far North Coast

and contains the Brisbane and Byron Bay sites. The second region is the Mid-North Coast which includes Coffs Harbour and Crowdy Head. The third region is the Central Coast, containing the Sydney and Port Kembla sites. The fourth region is the South Coast region which contains the Batemans Bay and Eden sites.

Preliminary data analyses for all sites show a remarkable similarity in comparative statistics of the long term means and standard deviation for Significant Wave Height (Hs), Maximum Wave Height (Hmax), Significant Period (Ts) and Peak Period (Tp)

Brisbane (Hs: 1.52[0.65], Hmax: 2.52[1.07], Ts: 7.66[1.38], Tp: 8.66[1.75])
 Byron Bay (Hs: 1.60[0.62], Hmax: 2.70[1.03], Ts: 7.94[1.35], Tp: 9.57[1.72])
 Coffs Harbour (Hs: 1.52[0.60], Hmax: 2.60[1.01], Ts: 7.58[1.36], Tp: 9.42[1.68])
 Crowdy Head (Hs: 1.67[0.65], Hmax: 2.88[1.09], Ts: 7.90[1.39], Tp: 9.76[1.76])
 Sydney (Hs: 1.59[0.64], Hmax: 2.70[1.07], Ts: 7.81[1.38], Tp: 9.78[1.75])
 Port Kembla (Hs: 1.53[0.63], Hmax: 2.60[1.06], Ts: 7.79[1.43], Tp: 9.61[1.88])
 Eden (Hs: 1.46[0.64], Hmax: 2.51[1.08], Ts: 7.41[1.44], Tp: 9.30[1.86])

Only when each site is considered on a monthly and seasonal basis, do the differences in the regional characteristics become evident. The major cause for the regional variation is based on the aspect of each site with respect to the major wave producing meteorological events.

3 Wave Generation.

Although the weather systems in the Australian region are dominated by the large anticyclonic high pressure systems, the storms which generate the largest waves on the Australian south-east coast are attributed to intense cyclonic low pressure systems.

A Public Work Department Report (anon 1983) documented the analysis of damaging storms on the N.S.W. coast between 1900 and 1980. The requisite for inclusion in the analysis was that the Hs \Rightarrow 2.5 metres. Most notable in the report, was that between 1900 and 1980, 242 storms were reported to have a significant wave height greater than 5.0 metres.

Three types of Cyclonic storm systems are capable of producing large waves at the coast.

- Tropical Cyclones
- East Coast Lows
- Mid-latitude Lows

The regions most affected by waves generated by Tropical Cyclones are the Far North Coast, the Mid-North Coast, and less frequently, the Central Coast. Very few large wave events recorded in the South Coast region result from Tropical Cyclones

East Coast Lows predominantly affect the Central and Mid-North Coast regions, the South Coast is exposed to waves generated by these events, though the wave heights rarely equal the heights recorded for the same storm further north. Wave data recorded in the Far North Coast regions, associated with East Coast Lows, tend to have smaller Significant Wave Heights and longer Significant periods. This

characteristic is caused by an increase in fetch distance between the region and the storm.

The Central and South Coast regions are most affected by storms produced by the Mid-latitude lows. The Mid-North Coast is less affected and the Far North Coast rarely records large waves produced by Mid-Latitude Lows.

Coupled with the Cyclonic storms, is the Anticyclonic systems, which traverse the mid-to-subtropical latitudes throughout the year. During the summer, the anticyclonic systems are located along the southern extent of the Australian continent. The frequent west to east passage of the anticyclone produces is characterised by a precursor known as a southerly front. Waves produced by this event are most noticeable in the South and Central Coast regions.

During the winter, the anticyclones are located further to the north, and tend to become a stable continental feature. Mid-latitude cyclonic activity in the Tasman Sea increases during this phase of the year. Local sea breezes produce a characteristic wave climate signature, and all four regions experience sea breeze activity during the summer months.

4 Discussion

The wave climates on the Australian south east coast are produced by a complex interaction of meteorological generating forces. The regional characteristics are a reflection of the composite of these wave generating events as they occur throughout a year and from year to year.

Further understanding of the complexities of the air-sea interaction, resulting in wave generation is required, especially in the areas of the modifying effect of sea surface temperatures on the atmosphere, and the extent of the influence of the El Niño/La Niña phenomena.

These areas are being investigated in work in progress.



Figure 1.

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Reduction Process of River Delta due to Natural Change of Its River Mouth — In the case of the Kurobe river —

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1. Introduction

Geologically speaking, river mouths change resulting to form a river delta which is normally fun-shaped. As seen in Figure 1, the river mouth of Kurobe river flowing into the Japan Sea has changed many times and is formed as a fun-shaped river delta, alluvial fan. The most latest location of the river mouth was near Furu-kurobe about 360 years ago. Since then the river mouth changed to the present position. The old shoreline was estimated by a coastal geomorphologist as shown in the figure by the dotted line curved convex offshore. The shoreline was examined by the property of sediment deposited. It is well-known that the coast where there was the old river mouth is one of the most seriously eroding coasts in Japan. One says that the coast is a fatefully erosive coast. It can be considered that this fact is due to a reduction process of river delta since the river mouth changed about 360 years ago. It is therefore recognized that the beach erosion of the coast has to be investigated as a reduction process of river delta due to the lack of sediment sources.

In this paper, referring to the previous investigation on the formation and reduction processes of the river delta of Kurobe river (Tsuchiya, Shibano, Suyama and Yoshimura, 1987), a mathematical simulation of the reduction process is first proposed

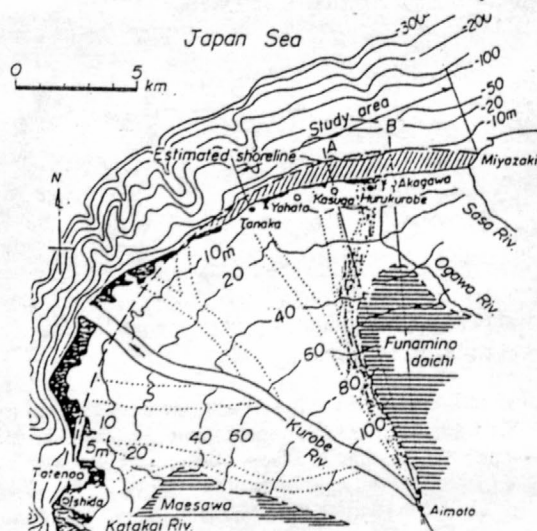


Figure 1. Alluvial fan of Kurobe river and its reduction shown by shaded area.

to hindcast the shoreline change for a period of 360 years and then compared with the actual shoreline change. Secondly, a methodology for stabilization of the coast is briefly discussed by making a series of stable sandy beaches formed by headlands

2. Estimation of Historical Shoreline Change

By use of geographical maps and characteristics of the present alluvial fan, the historical change of shoreline was estimated. The tidal change is only about 40cm on the coast. It can be assumed that no sea level change has taken place during the past 360 years. Extended the present land slopes on the maps to the present mean sea level to find the distance from the present shoreline to the intersecting location with the sea surface, the old shoreline was estimated as shown in Figure 2 showing a little foreshore deposition in the central area of the shoreline estimated. As shown in Figure 3, the predominant waves are coming from NW to NNW and NNE to NE directions, and the later one is the most predominant. It can therefore be understood that, as shown in Figure 2, the old shoreline near the old river mouth near Furukurobe point is inclined westary due to the longshore sediment transport in the west direction and this tendency is a little different from the geomorphologist estimation shown in Figure 1.

3. Modelling of Reduction Process of River Delta and Its Result

In the calculation of wave transformation, both the wave ray method and mild-slope equation were used. Longshore distributions of wave height, wave vector and wave power along the coast were obtained both for the predominant wind waves and swell. The duration of the wind waves and swell of which the wave heights are higher than 2 m were evaluated from observed wave data. By use of one-line shoreline change prediction model in terms of the total rate of longshore sediment transport, the shoreline change was hindcasted for a period of 360 years.

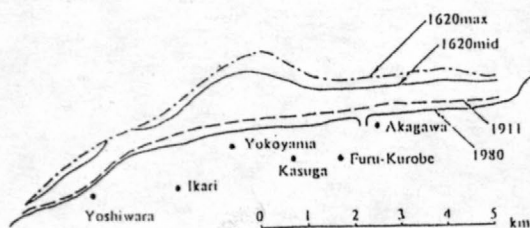


Figure 2. Historical change of shoreline around old river mouth.

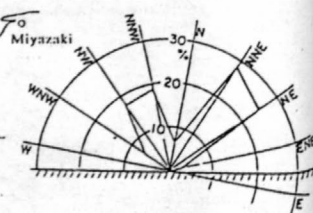


Figure 3. Wave power rose.

In the simulation of shoreline change the initial and boundary conditions are given as the initial shoreline 360 years ago and no sediment input from the old river mouth since the year, and the total rate of longshore sediment transport of 20,000 m³/year from the upcoast. An example of hindcasted shorelines along the coast is shown in Figure 4 where the dotted line shows the initial shoreline that the old shoreline estimated is

approximated and the others indicate the shoreline change at interval of 20 years. It can be considered that the shoreline has quickly changed and shown a tendency of approaching to a final one where no longshore sediment transport may exist. Further improvement to both the initial shoreline and predominant waves is needed to be able to compare with the actual shoreline change more closely and to predict the final shoreline in equilibrium if exist.

Based on the reduction process of river delta a methodology for stabilizing the coast can be essentially proposed by making either a series of statically or dynamically stable sandy beaches formed both by headlands and sand bypassing method if necessary. A practical methodology will be explained at the symposium.

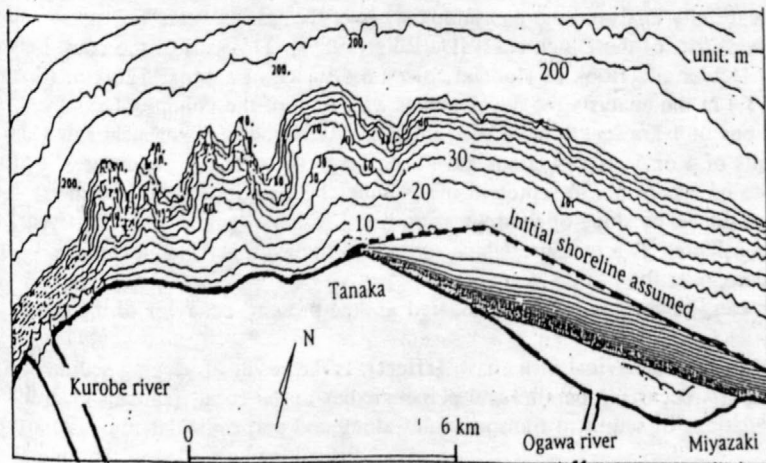


Figure 4. An example of hindcasted results of shoreline change in reduction process of river delta.

4. Conclusion

In studying the reduction process of river delta in the Kurobe river, the shoreline near the old river mouth around 1630 was estimated by use of geographical maps. By a mathematical simulation model for river delta reduction process, the shoreline change was hindcasted for a period of 360 years. Further improvement is needed, but the result of simulation can explain the reduction process by changing the river mouth naturally about 360 years ago.

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LARGE SCALE BEHAVIOR OF THE COAST OF THE 'CLOSED' PART OF THE DUTCH COAST

by:

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1 Introduction

Entire cross-shore profiles are yearly measured all along the Dutch coast since 1964 (over more than 1000 m; dunes and foreshore included). The spacing between the cross-sections is 200 to 250 m.

The Tidal Waters Division of the Public Works Department in the Netherlands has carefully analyzed the measurements over the last 25 years for the so-called 'closed' part of the Dutch coast [De Ruig (1989)]. This part of the coast between Den Helder and Hook of Holland covers a distance of approx. 120 km. (See Fig.1.) In the analysis the development with time of the volume of sand in sections of 1 km length have been studied. (A section of 1 km means that the results of 4 or 5 cross-sections have been taken together.) In most cases it turned out to be acceptable to represent the volume-change with time by linear regression. The slope of the regression line ['loss' or 'gain' in (m^3/km)/year] can be considered as a measure which represents the average behavior of the 1 km sections over the last 25 year.

This can, however, also be considered as 'the present' behavior of the coast.

'The present behavior' of a coast, [effect], is the result of varying sediment transports occurring parallel and perpendicular to the coast, [cause]. Quantitative calculations of sediment transport rates along and perpendicular to coasts are still

difficult to be made, so the logical sequence [cause] yields [effect] can often hardly be made in practice.

Especially the contribution of cross-shore transport is often less understood than the longshore contribution.

Combining a fair knowledge of the [effect] with rather poor insights in the [cause], nevertheless allows us to formulate scenario's of longshore- and cross-shore transport patterns which apparently yield the behavior of the coast. The different elements of promising scenario's are very useful topics for further research.

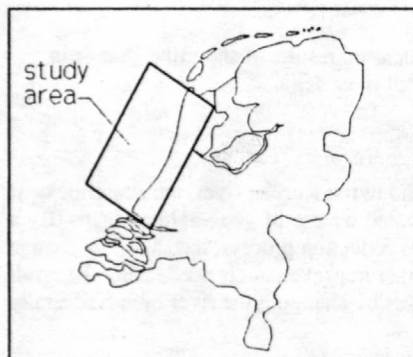


Fig.1 Study area

2 Yearly measurements

All yearly measurements along the Dutch sandy coast (approximately 300 km long) are stored in the so-called JARKUS database. (JARKUS: Dutch acronym for yearly coastal profile measurements.) Fig.2 shows an example of a single cross-section. For the aim of the present study the variations in the volume of sand within the profile with time is important. The notion 'volume of sand within the profile' calls for further specifications. In Fig.3 an example has been given; the volume of sand above the (horizontal) level of Datum -6m and seawards of a (vertical) line 'far enough' into the dunes has been given as a function of time. [The volume of sand is expressed in m^3/m .] 'Far enough' means that at least in the time span covered by the measurements (approx. 25 year) no appreciable amounts of sand have been transported by wind to or from the part of the profile landward of that selected line. According to Fig.3 the volume of sand within the profile varies with time. Yearly 'jumps' of 300 to 500 m^3/m often occur; that means an average sedimentation or erosion over the whole profile of 0.3 to 0.5 m. In figures like in Fig.3 sometimes more or less periodical fluctuations seem to occur; period of fluctuations several years. In the present study these fluctuations are disregarded. 'The present' behavior of the profile (average behavior over the last 27 year) is represented by the fitted line in Fig.3 (least square fit). In the example of Fig.3 the slope of the fitted line [gain of 24 (m^3/m)/year] is a measure of the behavior of that cross-section.

Fig.3 holds for a single profile; similar plots are found and similar considerations can be hold if coastal sections of 1 km are taken into account. [Cf. the study of the Tidal Water Division; De Ruig (1989).]

3 Large scale behavior of the coast

From the erosion and accretion figures of the 1 km long sections of the coast, the present behavior (average behavior over the last 25 years) of these sections, but also of the entire coast is determined; [effect]. To know the [cause] of this

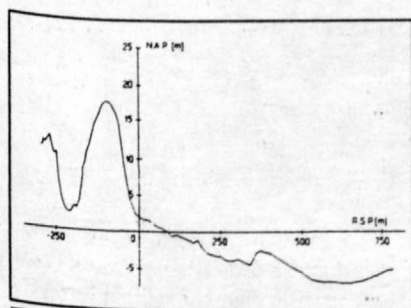


Fig.2 Cross-shore profile

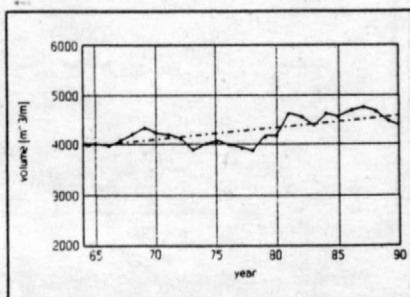


Fig.3 Volume of sand within profile

behavior is very important for a coastal zone manager. In Fig.4 the three sediment transport vectors have been indicated which are ultimately responsible (are the [cause]) for the behavior of a section of 1 km of the coast. ('behavior' = $S_L - S_R + S_C$; all sediment transport parameters expressed in adequate units; e.g. m^3 per year across the boundaries; 'behavior' expressed in gain or loss per section of 1 km per year.) As long as the transport vectors are unknown, we have one equation with three unknown parameters, so not to be solved. For the entire coast from Hook of Holland to Den Helder (118 km), we have 118 equations with 237 unknown parameters!

The aim of the present study was to quantify as reliable as possible the 237 unknown parameters. The transport vectors parallel to the coast have been calculated by taking into account the wave climate along the Dutch coast and the orientation of the coast. The direct results of the longshore transport calculations are considered as first estimates; some variations in the calculated values are not excluded and taken into consideration.

Next the coast has been divided in several sections; each section is considered more or less as a unit. (E.g. between the harbours along the Dutch coast and around the harbours.) For each section various scenario's have been formulated. In a scenario the estimated longshore transport distribution along the coast was the starting point (S_L - and S_R terms); the cross-shore component (S_C) was next 'adjusted' in such a way that the apparent behavior of the coast was achieved. The various elements of a scenario were judged with respect to consistency and general knowledge of the Dutch coast. A distinct longshore transport distribution along the Dutch coast and an apparent cross-shore transport contribution was ultimately found.

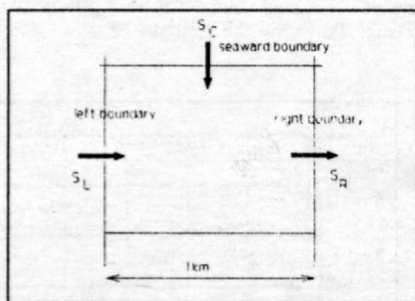


Fig.4 Sediment transports

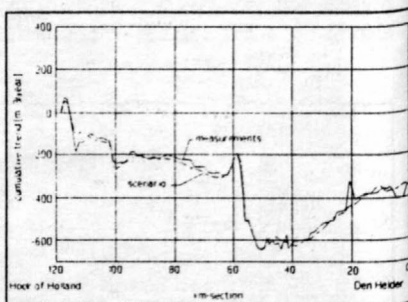


Fig.5 Comparison between measurements and a promising scenario [Due to a specific sign-convention in this study, a negative sign means a gain of sand!]

The ultimate result is shown in Fig.5. Starting in Hook of Holland (left side of Fig.5) with zero, the cumulative losses and gains of sediment (from the 1 km long sections) have been plotted as a function of the position along the coast. (Notice that the entire 118 km long coast gains apparently about 350,000 m³ sand per year; the 'gap' near Den Helder at the right side of Fig.5.) The ultimate results of the 'best' scenario's for the several sections have been plotted in Fig.5 as well.

4 Discussion

The value of this study is certainly not that all the S_L , S_R and S_C terms as found are true. The main value of this study is that it yields topics for challenging and promising research. Several questions can be formulated; some examples:

- Is it true that along the Dutch coast at some positions onshore directed transports do occur, while at other positions apparently (probably) offshore transports take place?
- Are we able to quantify these different cross-shore transport rates?

(In the sections of the coast where groin systems have been built in the past, the calculated longshore transport rates had to be reduced considerably (with 50 %) in order to achieve a consistent picture.)

- Is it true that a groin system has such a large effect on the longshore transport?

5 Acknowledgements

The study has been sponsored by the Tidal Waters Division of the Public Works Department in the framework of the 'Coastal Genesis' project. Mr. Jauk Stroo has developed and carefully judged the various scenario's as a part of the requirements for his Master's of the Delft University of Technology [Stroo (1991)].

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Morphodynamic Modelling For A Tidal Inlet In The Wadden Sea

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A dynamic model for morphological development in a tidal inlet, "Het Friesche Zeegat", in the Wadden Sea along the northern coast of the Netherlands (Fig.1), is set up to obtain more understanding of the behaviour of the morphological system around tidal inlets.

The geometrical schematisation of the model is such that the main features of the system are represented, whereas details are determined by the shape of the computational grid rather than by the actual geometry (see Fig.2 for the situation just before the closure of the Lauwerszee in 1969).

The model is based on DELMOR, a program package for morphological development, developed at DELFT HYDRAULICS (Wang et al, 1992). The program consists of three modules, the flow module, the sediment transport module and the bed level module. For the flow module a quasi-3D approach is applied. First the depth averaged flow field is calculated from the tidal flow model TRISULA. Then the secondary flow in the lateral direction due to curvature and Coriolis's force is calculated. In the longitudinal direction a logarithmic velocity profile is assumed. The

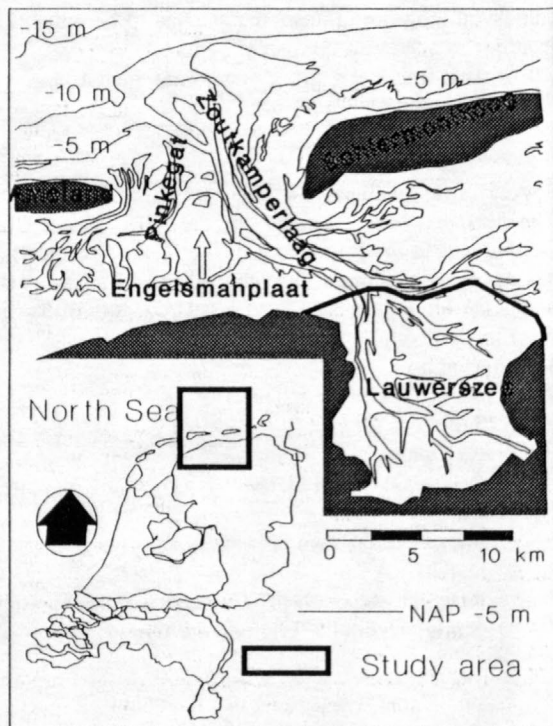


Fig.1 Study area: "Het Friesche Zeegat"

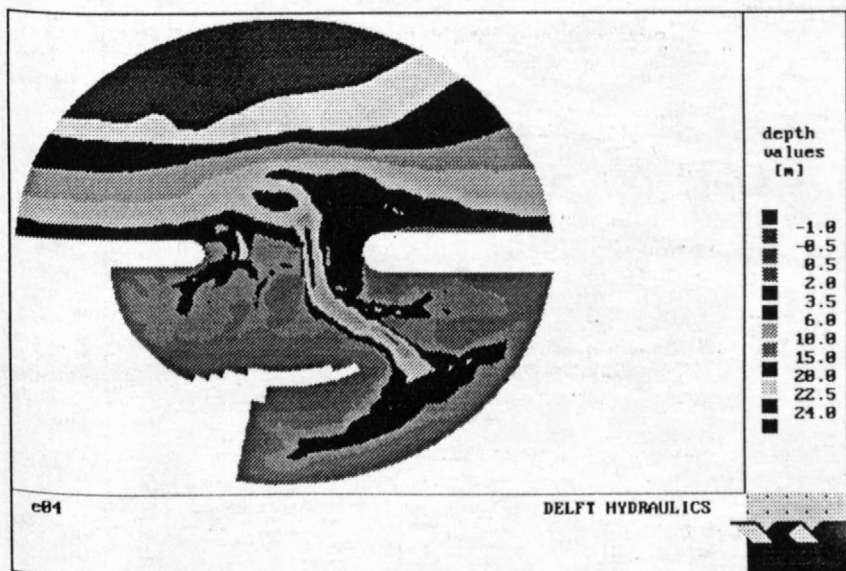


Fig.2 Schematised geometry just before the closure of Lauwerszee

sediment transport is divided into bed load transport and suspended transport. The bed load transport is calculated from a transport formula (e.g. Van Rijn 1984) whereas for the suspended transport a quasi-3D model based on an asymptotic solution of the convection-diffusion equation for the sediment concentration is applied. This quasi-3D model was first developed by Galappatti and Vreugdenhil (1985) and later extended by Wang (1989). The bed level module is simply based on the mass-balance of the sediment.

Computations of the flow, the sediment transport, the initial bed level change rate as well as long-term morphological development have been carried out. The first results of the computations have already been presented by Wang et al (1991).

The model has been compared with an initial model with a much finer computational grid (about 10 times more grid points), which is set-up by Steijn and Hartsuiker (1992). For the case that the wave influence is neglected the two models produce almost the same patterns of the residual flow field and the net sediment transport field. This justifies the relatively rough schematisation and computational grid in the present model.

The tidal inlet under consideration has undergone significant morphological changes since 1969, due to the closure of the Lauwerszee (see Fig.2 and Fig.3). The model results show that the closure has a strong impact on the net sediment transport pattern. After the closure the magnitude of the net sediment transport rate is decreased because of the decreased tidal volume. In the region near the gorge of the inlet the net sediment transport in the channel Zoutkamperlaag turns from export to import due to the closure. The long-term simulations show that the closure causes significant sedimentation in the basin, which agrees with the field data since 1970 (Biegel, 1991).

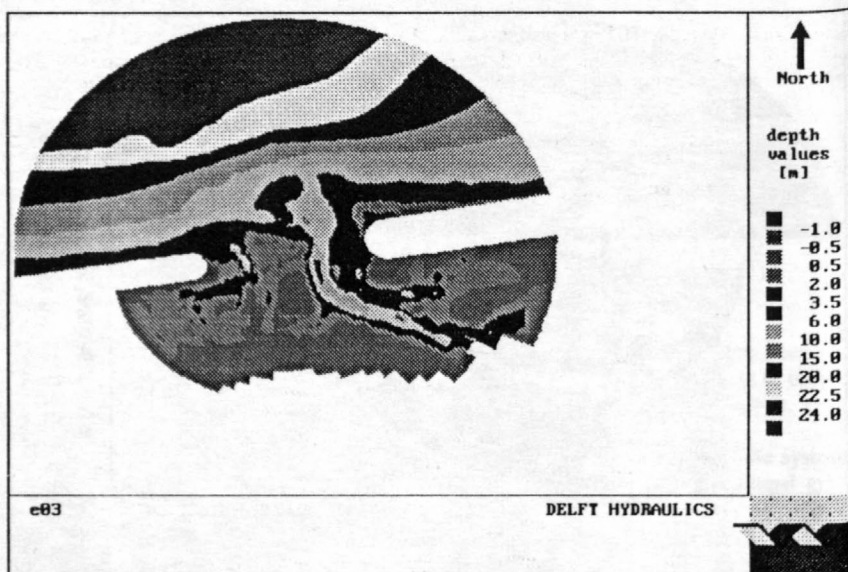


Fig.3 Situation in 1987

The model is still not able to reproduce the natural bathymetry in detail. Especially in the outer delta area the agreement between the computational results and measurements is poor. Apparently simplifications such as neglecting the influence of waves still make the model unsuitable as an accurate prediction tool. However, as a research tool the model has proven very useful.

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DEVELOPMENT AND APPLICATION OF A GENERIC MODELLING SYSTEM FOR COASTAL PROCESSES

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INTRODUCTION

The increase in the level of education and expertise in mathematical modelling during the last decade has lead to dramatic changes in the form and method of application of the models. The typical user profile has changed from the hydraulic modelling specialist who was deeply involved in the development of the model to users who see the models as tools to solve their problems. This new group of users must still be professional engineers and scientists, but are not modelling specialists.

For computational hydraulics specialists, the real challenge has been the development of accurate, efficient and reliable numerical solutions to the hydrodynamic and advection dispersion equations. The hydraulics is of prime interest if the study concerns, for example, storm surge, layout of structures to minimize hydrodynamic losses or wave disturbance in harbours. However, more frequently the prime interest is, for example, the resulting sediment transport or the environmental impact of polluted discharges. The scientific interest in these problems is the development of quantative descriptions of the transport capacities of sediments and the biological/chemical interactions, while the numerical solution is a relatively simple matter.

The user's concern with these scientific and numerical problems is limited. He must simply be confident that the processes are adequately described and that the model produces accurate and reliable results. The physical problem which he must study is, however, not simple and he must manipulate a whole sequence of models before he obtains the desired result. Fig. 1 illustrates the models which could be used in a coastal sediment transport study.

The integrated, generic modelling system, MIKE 21 (Fig. 2), which is the subject of this paper, is designed to make the procedure of application and interaction of models easier for the user. The paper describes the structure of MIKE 21 and its application to the MAST II Coastal Morphodynamics research project.

MIKE 21

MIKE 21 is a comprehensive 2-D modelling system for estuaries, coastal waters and seas. It is an integrated, generic software package for modelling of flows, water levels, waves, sediment, pollutant transport and water quality. The fields of application are illustrated in Fig. 2.

The modules in MIKE 21 are:

PP	-	Pre- and post-processing	ST	-	Sand transport
HD	-	Hydrodynamics	MT	-	Mud transport
AD	-	Advection-dispersion	WQ	-	Water quality
BW	-	Boussinesq waves	EU	-	Eutrophication
EMS	-	Elliptic mild-slope waves	HE	-	Heavy metals
PMS	-	Parabolic mild-slope waves	SAW	-	Oil and chemical spill
OSW	-	Offshore spectral wind-waves	PA	-	Lagrangian particle advection-dispersion.
NSW	-	Nearshore spectral wind-waves			

APPLICATION TO MORPHODYNAMICS RESEARCH

The MIKE 21 modelling system is specially designed for ease of application to 'routine' investigations. However, its modular structure and total integration of all components has also proved to be very valuable in research because of the ease of manipulation of the modules. This will be illustrated here with the Coastal Morphodynamics research project under the CEC MAST II programme.

One of the objectives of the project is to develop techniques for long term predictions of morphological changes at coasts. The state-of-the-art is that it is only possible to determine rates of changes of bed levels under a limited number of conditions and then to use these to make a "guestimate" of long term development. In this project, the feasibility of using a total, timestep-by-timestep, integration of the three physical processes has been investigated, ie. waves, hydrodynamics and sediment transport.

Basically, the morphological modelling is the solution of the continuity equation for sediments

$$\frac{\partial z}{\partial t} + \frac{1}{1-n} \frac{\partial q_x}{\partial x} + \frac{1}{1-n} \frac{\partial q_y}{\partial y} = 0$$

where z is the bed level, n the porosity and q_x and q_y are transport rates in x and y directions.

The modular structure of the system makes it possible, for example, to include wave-current interaction both in terms of current refraction and hydraulic roughness and to enter the calculated $\partial z/\partial t$ directly into the hydrodynamic module. This leads to a consistent treatment of the phenomena in all process modules.

The sequence of computations is shown in Figure 3. The computation timesteps in the three modules may vary in accordance with the timescales of the processes. As has been mentioned elsewhere, the morphological timestep for stability is surprisingly short, of the order of 1 hour or less, which is the main justification for embarking on research into this technique.

The deformation of the sea bed behind an offshore breakwater was chosen as the case to be studied. Some results are shown in Figures 4-7 and will be commented upon in detail in the presentation.

ACKNOWLEDGEMENTS

This work was carried out as a part of the G8 Coastal Morphodynamics research programme. It was funded jointly by the Danish Technical Research Council (STVF) and the Commission of the European Communities, Directorate General for Science, Research and Development, under contract no. MAS2-CT92-0027.

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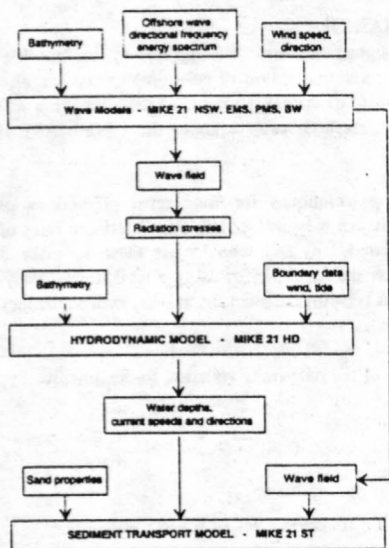


Fig. 1 Modelling procedure for sediment transport computations

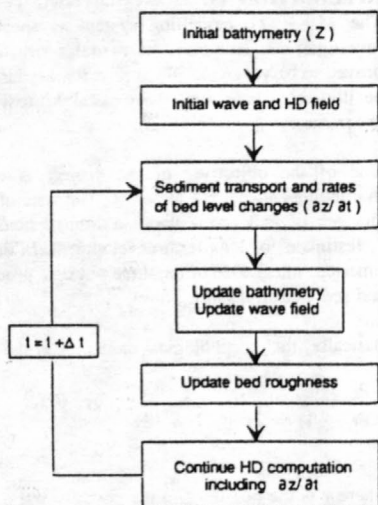


Fig. 3 Principle flow chart for morphological modelling

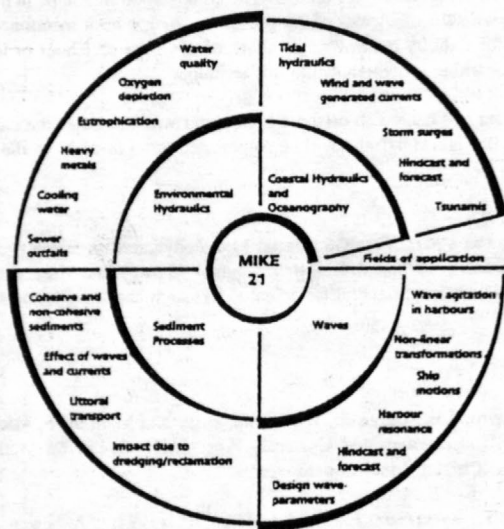


Fig. 2 The MIKE 21 Modelling System

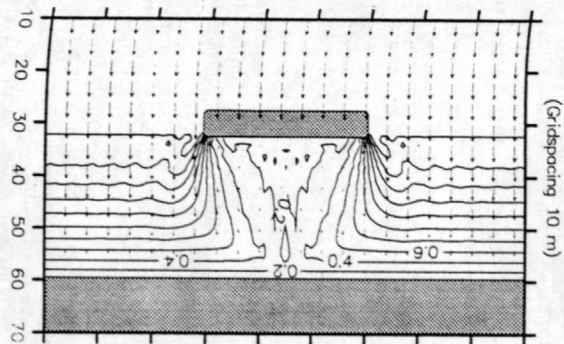


Fig. 4 Initial wave field calculated with M21 PMS

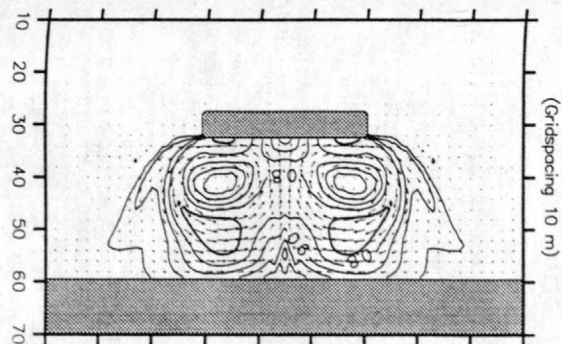


Fig. 5 Initial flow field generated by wave radiation stresses above and calculated with M21 HD.

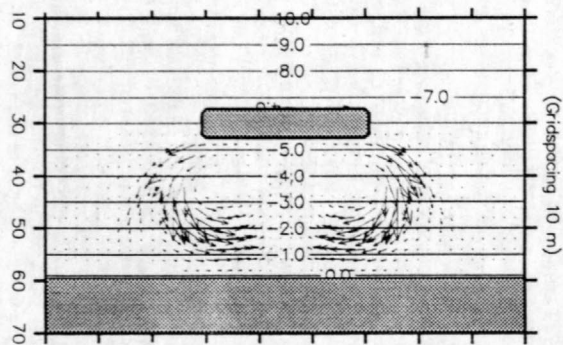


Fig. 6 Initial bathymetry and sediment transport field computed with M21 ST.

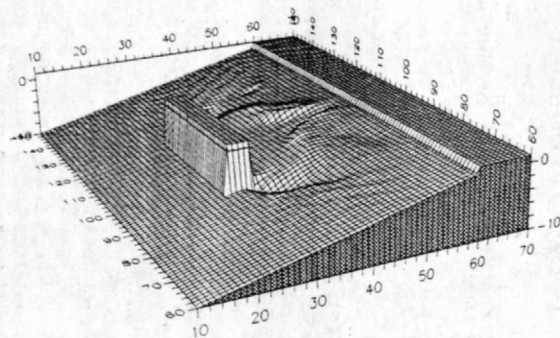


Fig. 7 Bathymetry after 48 hours

THE ANALYSIS OF COASTAL PROFILES FOR LARGE-SCALE COASTAL BEHAVIOR

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ABSTRACT

A large data set of coastal profiles was used to determine the large-scale behavior of the Central Dutch Coast (115 km). The data set was analysed by combining the empirical eigenfunction technique and a moving window approach. Five regions with a specific Large-Scale Coastal Behavior (LSCB) appeared, separated by quite sharp boundaries. Profile shape and profile behavior are clearly related. The actual size of LSCB-regions varies considerably, just as the time span over which a long-term trend in coastal behavior becomes visible.

INTRODUCTION

LSCB addresses coastal developments on a time scale which is assumed to be in the order of decades and a length scale which is assumed to be in the order of kilometers to tens of kilometers. It is hypothesized that areas in which coastal profiles show similar large-scale developments ("LSCB-regions"), are controlled by specific forcing processes and/or boundary conditions (Terwindt and Wijnberg, 1991). The location and nature of boundaries between the LSCB-regions may give clues about which are the controlling processes and/or boundary conditions for LSCB along the Central Dutch Coast. Therefore, one of the prerequisites for the understanding of LSCB is an integrated description of morphological developments in time and space.

A large data set of coastal profiles (the JARKUS data set) was used to determine the size of areas with similar developments and the nature of the boundaries between these areas. The coastal profiles are measured from the fore dune to approximately 1 kilometer seaward at every 250 meter distance along the Central Dutch Coast (115 km). The profiles are measured once a year, over a total time span of 27 years.

METHOD OF ANALYSIS

In the development of coastal profiles one can differentiate between an overall shift of the profile and a change in the shape of the profile. The shifting of the coastal profile is represented by the behavior of +1m-NAP contour. This level approximates high water level along the Central Dutch Coast. The development of the shape of the profile is analysed relative to this (moving) reference, so only the part of the profile below +1m is considered.

Several functions have been proposed to describe the coastal profile. However, most descriptions does not account for a barred topography, which does exist along the Central Dutch Coast. In Terwindt and Wijnberg (1991) the empirical eigenfunction approach was pointed out as the most promising method for schematizing both the bar system and the mean profile.

The seaward extension of the bar system, the number of bars and the bar spacing vary alongshore. As a consequence, it is impossible to schematize the various bar systems with one common set of morphological meaningful empirical eigenfunctions. This implies that empirical eigenfunction techniques that simultaneously take into account the alongshore and temporal variation in the whole profile data set (Ostrowski et al., 1991), will fail to schematize the various bar systems. Therefore, the profile data has been analysed in small subsets.

Along the full length of the coast, a "window" with an alongshore width of 1,1 km and a temporal extent of 27 years has been moved alongshore with a step size of 1

km. Within the window, all available profile data are summarized with 3 empirical eigenfunctions which have been computed according to the method described by Winant et al. (1975). The size of the window is large enough to get reliable eigenfunction shapes and small enough to avoid predetermination of boundaries of LSCB-regions.

The use of complex principal component analysis (Liang and Seymour, 1991) was rejected, because the results are hard to visualize. Without visualization it is very difficult to get an overview of the meaning of the eigenfunctions and the spatial and temporal coherence of the results.

RESULTS

The first empirical eigenfunction (E1) is the mean profile function, which usually explains over 97% of the correlation between cross-shore depth values. About 65-70% of the depth variation that is not explained by E1, is explained by E2 and E3. E2 and E3 are both multiple bar functions, which differ by the positions of the bars. These positions are about 90 degrees out of phase between E2 and E3.

When the temporal weighting of the observed profiles on the (locally defined) empirical eigenfunctions is plotted against the alongshore location, 5 well defined regions appear. When the position of the +1m-contour relative to its time-averaged position is plotted as function of the alongshore location (representing the horizontal shift of the profile) the same 5 regions appear but the boundaries are generally more transitional than those based on profile shape behavior.

The 5 regions differ in: (a) time-averaged profile shape [steepness, concavity], (b) dimensions of the bar system [height of bars, seaward extension of bar system], (c) behavior of the bar system [time-scale and spatial variability], (d) behavior of the +1m-contour [presence/absence of trends and cyclicities], (e) behavior of the mean profile steepness [time scale]. So, profile shape and profile behavior are clearly related.

The length scale of these regions varies from 5 to 40 km. Within the regions more gradual changes may occur. However, these changes are smaller than the differences between regions.

An example of regional differences in LSCB is shown in figures 1 and 2. The boundary between region A and B is remarkably sharp. This boundary is actually located at the harbour moles of IJmuiden (built in 1865-1879 and extended in 1962-1967 to a total length of about 2.5 km). The two regions, which are larger than shown, differ most clearly in the dimensions (fig. 1b) and behavior (fig. 2c, 2d) of the bar system. In region A, the cross-shore bar configuration schematized by E2 (or E3) returns about every 15 year. In region B this return period is only 4 to 5 year (fig. 2c, 2d). The return period is related to net offshore migration of the bars. Bars vanish at the seaward end of the profile and new bars develop near the shoreline. This cyclic profile behavior occurs quite in phase alongshore in region B, while in region A it is more variable. Further, the mean profile shape and its behavior (resp. fig. 1a and fig. 2b), and the shoreline behavior (fig. 2a) differ between the two regions.

DISCUSSION

If we use a decade as the time-frame for LSCB, then 5-year cycles will filter out but the 15-year cycles will not. Linear trend-fitting on time series of shoreline position and profile characteristics, in order to detect LSCB, will be influenced by these cyclicities if the time series spans only a few decades. The fitted trends are very likely not the real long-term trends.

These results have rather important implications for the analysis of LSCB. Although the order of magnitude of time scales and spatial scales assumed to characterize LSCB are correct, considerable variations occur in the actual size of regions with similar developments and the actual time span over which trends in large-scale developments will become visible.

Aside from the statistical influence of the time-scale of LSCB-variability on the determination of the long-term trend, it is more important to know whether a causal relation exists between the two. For example, whether the amplitude and time-scale of

LSCB-variability influence the stability of long-term tendencies in LSCB. However, at this moment it is unclear whether such relationships exist.

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- Winant, C.D., D.L. Inman, and C.E. Nordstrom, 1975. Description of seasonal beach changes using empirical eigenfunctions. *J. of Geophys. Res.*, 80(15), p. 1979-1986.

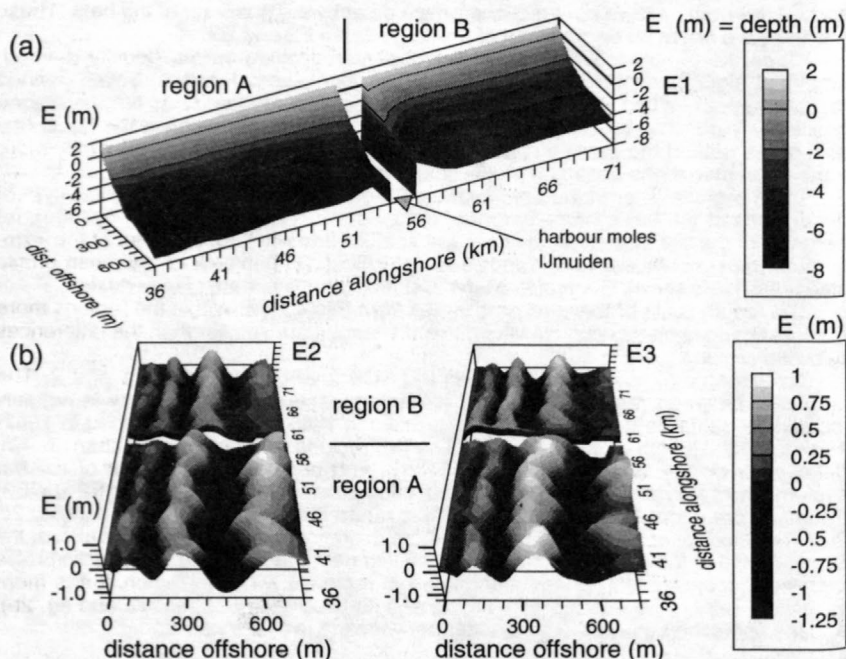


Figure 1: Alongshore variation of average profile characteristics, represented by locally defined, cross-shore empirical eigenfunctions (E = cross-shore eigenfunction weighting). (a) mean profile functions E_1 , (b) bar functions E_2 and E_3 .

Figure 2 (next page): Variation of profile characteristics as a function of time and alongshore location (shaded areas represent missing data). (a) Horizontal shift of the profile, represented by the shift of the +1m-contour relative to its time-averaged position, (b) Variation in profile steepness, represented by the deviation from the local mean depth at 750 m from the shoreline, reconstructed from the local cross-shore eigenfunction E_1 and the temporal weightings on E_1 . (To be continued)

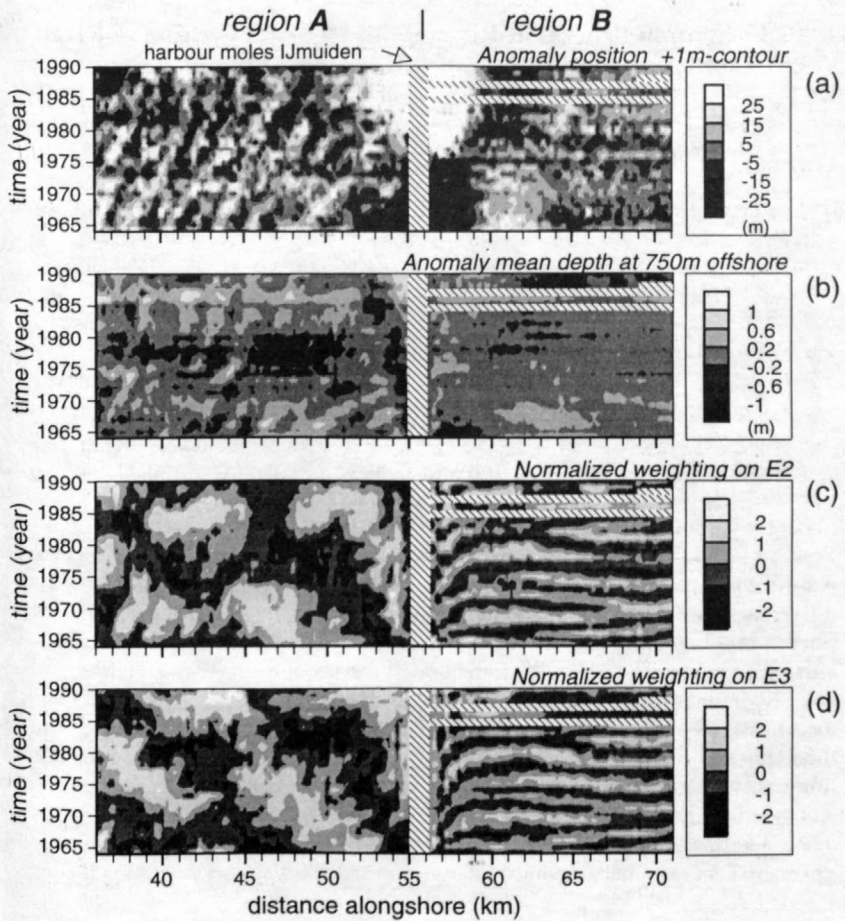


Figure 2 (continued) : (c),(d) Variation in bar positions, represented by the temporal weightings on the locally defined cross-shore eigen-functions E2 and E3. A large positive weighting (light grey), means that the bar topography defined by E2 (E3) is well represented in the profile observed on time t . A large negative weighting (dark grey) means that the bar topography in the profile observed on time t is opposite of that defined by E2 (E3). At locations where E2 (E3) defines bars, the observed profile has troughs, and troughs in E2 (E3) are bars in the observed profile. If the weighting is near zero (middle grey), the bar topography defined by E2 (E3) is not represented in the profile observed on time t . In that case the bar topography is defined by E3 (E2).

Formation Process of River Delta by Short Cutting of River — In the Case of Shinanogawa River Flowing into Japan Sea —

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1. Introduction

The Ohkozu diversion channel from the Shinanogawa river, the longest river in Japan, was completed in 1925 for preventing the downstream area of the river from flood disasters. The river mouth delta of the diversion channel has been formed and developed due to the sediment input from the Shinanogawa river as shown in Fig. 1. It can be observed the quick development of the river mouth as well as sediment deposition near the harbor on left-hand (west) of the river mouth. In the period of 1977-1989, nine times of depth sounding have been conducted as well as sediment sampling by boring at three points in the foreshore which was formed by sediment accumulation from the diversion channel on the beach before sediment supply which is called the original beach in this paper. The changes in shoreline position have also been measured with enough accuracy to discuss the formation processes of the river mouth delta.

Shoreline change simulation is carried in terms of the one-line model which is modified to be applicable to simulate the river delta formation process by considering the sediment flushed into an area deeper than the depth of sediment initiation. First the original beach topography is reproduced by using the Dean's equilibrium beach profile and the initial shoreline position.

The unknown parameter of Dean's formulation ($h = Ax^{2/3}$) is estimated by core sample of boring data in which obvious difference in sediment diameter between the original

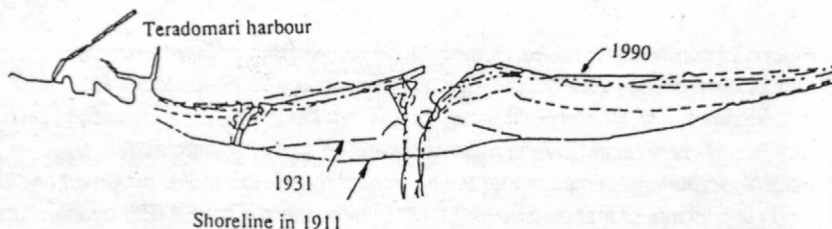


Fig. 1 The river mouth delta of the Ohkozu diversion channel

beach and newly formed one. Secondly, the hindcast of the shoreline change is performed after specifying the incident wave condition with observed wave data for six years and the compared with measured shorelines. Finally 100-year prediction of the shoreline change is performed to help the future development plan in the coast.

2. Estimation of Beach Topography and Sediment Input from the Channel

Figure 2 shows the boring positions and grain size distributions of core sample at three points, BNo1, 2 and 3. Notation P-1-6 in the figure of grain size distribution indicates the data of the 6th layer at Point No.1. As mentioned by thick solid and broken lines in the figure, the layer where grain size suddenly changes. This is the interface between original and newly formed beaches.

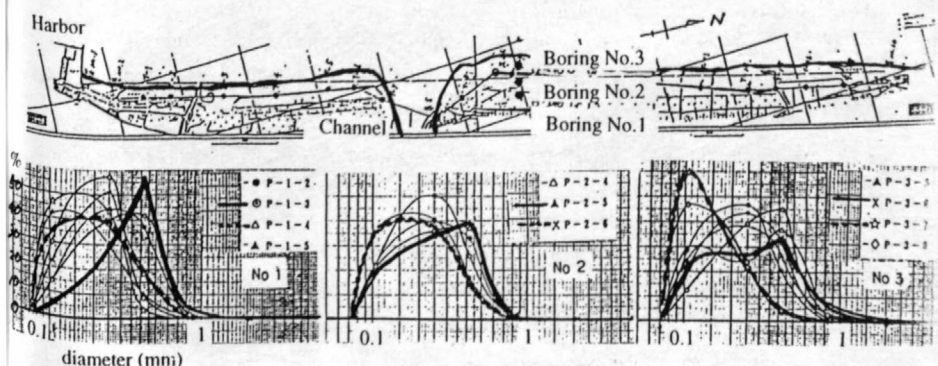


Fig. 2 Boring positions and grain size distributions of core sample

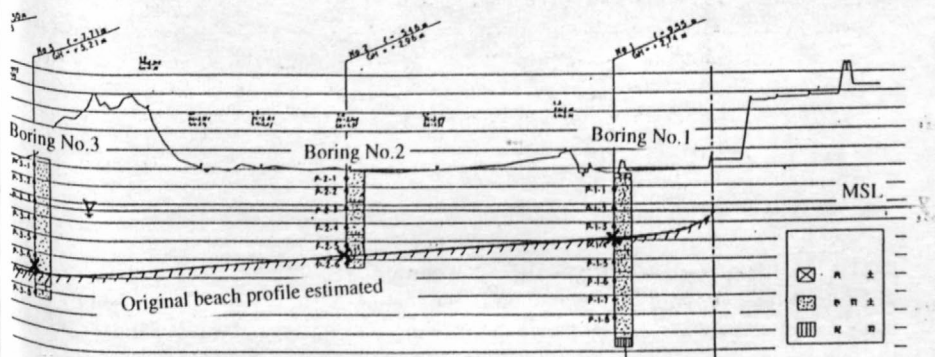


Fig. 3 Vertical distributions of the sampling points and estimated beach profile

Vertical distributions of the sampling points and estimated beach profile is shown in Fig. 3, in which the interface estimated from the grain size distribution is indicated by thick cross mark. It can be recognized that the estimated profile with the parameter

$A=0.08$ is fairly good reproduction of the original beach profile.

By using the estimated profile and shoreline before construction of diversion channel, we can estimate the topography of the original beach as in Fig. 4(a) which may be the basis to estimate total amount of sediment deposition from the channel. Figure 4(b) is measured present profile over-drawing on the original one. An accumulation curve of sediment input from the channel estimated by the budget in sounding charts is shown in Fig. 5. The averaged sediment supply is $910,000 \text{ m}^3/\text{yr}$ by distributing 41% in the west (Teradomari) and 51% in the east (Nozumi) coasts.

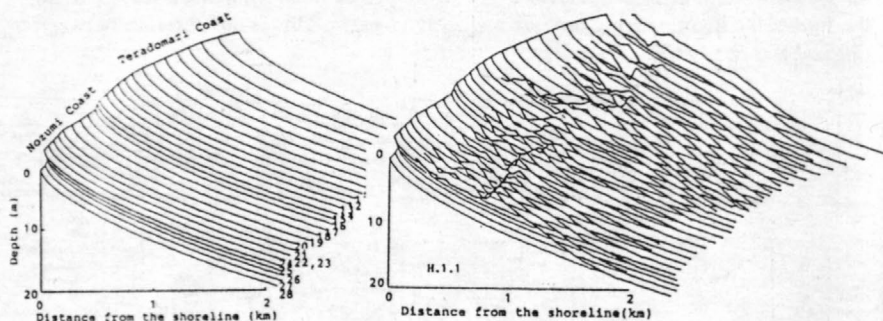


Fig. 4 Estimated topography of the original beach and present topography

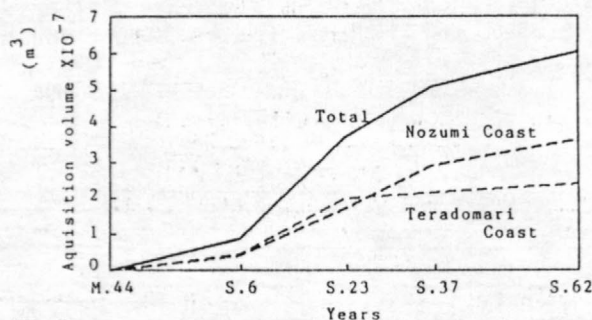


Fig. 5 An accumulation curve of sediment input from the channel

3. Mathematical Model for River Delta Formation

The model for monsoon wave condition in the middle Japan sea was established by using the observed wave data as shown in Fig. 6 in which the typical wave characteristics (direction, height H , period T and duration) due to monsoon storm is modeled. Considering the total energy flux in one year, we estimate 11 model storms a year.

For considering the loss of sand which may go out into deeper area of the sea, a modified one-line model is employed which has SINK term to model loosing sand as is schematically shown in Fig. 7. Figure 8 is the computed result by the model for predicting the shoreline in 2087. Boundary conditions in the computation are fixed at left-hand and free passing of sediment at right-hand of boundaries. It can be observed that

shoreline positions at the checking years of 1931 and 1987 are within allowable prediction range. Then the computed shoreline might not be far from the actual one if there is no changing in sediment input for the period. The predicted longshore sediment transport rate by the numerical model is $750,000 \text{ m}^3/\text{yr}$ which is less than that estimated by sounding charts with $160,000 \text{ m}^3/\text{yr}$.

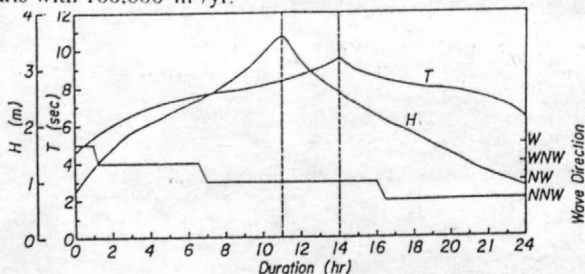


Fig. 6 Monsoon wave condition

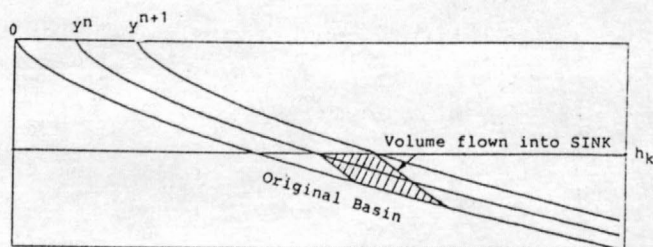


Fig. 7 SINK term to model loosing sand

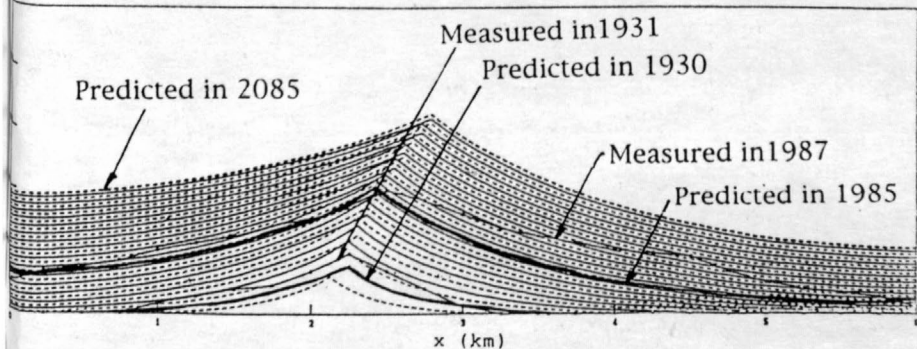


Fig. 8 Shoreline change predicted for 165 years in the river delta formation

4. Conclusions

The formation processes of the river mouth delta which was artificially formed has been investigated by both measured data and numerical simulation by the modified one-line model. It was confirmed that the method presented here was one of effective methods to predict long-term beach changes due to river mouth delta formation.

Appendix A

Large Scale Coastal Behavior '93: Schedule

**November 16-19, 1993: The Don CeSar Beach Resort
St. Petersburg, Florida**

Oral Sessions: Del Prado Meeting Room

Monday, November 15

**17:00 - 19:00 Registration: Don CeSar Lobby (Outside Del
Prado Nov. 16,17,18)**

19:30 - 20:30 Ice Breaker Reception (Boardwalk)

Tuesday, November 16

08:00 Conference Opening

**08:10 Keynote Address: The role of mathematical model-
ling in large-scale coastal morphology
de Vriend, H.J.**

Session 1: Cross-Shore Behavior

**08:40 Nearshore bars and large scale coastal behaviour
Kroon, A., and Hoekstra, P.**

**09:00 The energy spectrum of shoreline and sand bar
morphology
Holman, R.A., Lippmann, T.C., and Birkemeier, W.A.**

**09:20 Large-scale behavior of Poland's coastal areas
Mielczarski, A., Pruszek, Z., and Zeidler, R.B.**

**09:40 The analysis of coastal profiles for large-scale
coastal behavior
Wijnberg, K.M., and Terwindt, J.H.J.**

10:00 Discussion Period

10:20 Break

Session 2: Cross-Shore Modeling

**10:40 Behavior-oriented models applied to long term
profile evolution
Capobianco, M., de Vriend, H.J., Nicholls, R.J., and Stive,
M.J.F.**

**11:00 Prediction of cross-shore sediment transport at
different spatial and temporal scales
Larson, M., and Kraus, N.C.**

**11:20 Theoretical stability of equilibrium beach profiles
Plant, N.G., and Holman, R.A.**

**11:40 Modeling shore-normal large-scale coastal evolu-
tion
Niedoroda, A.W., Reed, C.W., and Swift, D.J.P.**

12:00 Discussion Period

12:20 Lunch (Pavilion)

Session 3: Long-term Monitoring

- 13:20 Long term wave height and beach profile changes, Narrabeen Beach, Australia**
Short, A.D., and Hall, W.
- 13:40 Interannual changes in bed elevation: Outer surf zone, Duck, NC 1981-1991**
Howd, P.A., Birkemeier, W.A., and Stockdon, H.
- 14:00 Regional seafloor changes near St. Marys Entrance, Georgia/Florida and their influence on shoreline response**
Byrnes, M.R. and Hiland, M.W.
- 14:20 A new method of investigating temporal and large-scale spatial shoreline trends**
Fenster, M.S., and Dolan, R.
- 14:40 Large scale behavior of the coast of the "closed" part of the Dutch coast**
van de Graaff, J.
- 15:00 Discussion Period**
- 15:20 Break**

Session 4: Geologic Controls

- 15:40 Cohesive shores and large scale coastal evolution**
Nairn, R.B.
- 16:00 Coastal development in Suriname at different temporal and spatial scales**
Augustinus, P.G.E.F
- 16:20 Sea level controls on Mississippi River delta building during the holocene transgression**
Penland, S., McBride, R.A., Suter, J.R., Boyd, R., Williams, S.J.
- 16:40 Influence of inherited geologic framework upon barrier island morphology and shoreface dynamics**
Riggs, S.R., and Cleary, W.J.
- 17:00 Discussion Period**

18:00-19:30 Poster Session (Pavilion)

- P 1 Geomorphological analysis and mathematical modelling of sedimentation and coastline stability of the tidal inlet Gradyb**
Mangor, K., Lintrup, M.J., Bartholdy, J., Nielsen, J., Nielson, N.
- P 2 Sediment budget for the west coast of Taiwan, a comparative study between results from numerical models and bathymetric maps**
Jensen, A., Hedegaard, I.B., Qian-Ming, L.
- P 3 Background erosion determined by three-way PCA method**

- Medina, R., Losada M.A., Losada I.J., Vidal, C.
- P 4 Foredune migration and large scale nearshore processes**
 Psuty, N.P., and Allen, J.R.
- P 5 Large scale high resolution coastal surveying with the SHOALS system**
 Lillycrop, W.J., Parson, L.E., and Estep, L.L.
- P 6 Sorting differentiation processes along the Israeli Mediterranean shore as indicated by grain-size populations**
 Hartmann, D.
- P 7 The effects of wave event sequencing on long-term beach response**
 Southgate, H.N.
- P 8 Long-term development of Pacific atoll islet shorelines**
 Richmond, B.M.
- P 9 Geomorphic response types along barrier coastlines: a regional perspective**
 McBride, R.A., and Byrnes, M.R.
- P 10 Seasonal erosion/accretion cycles in a littoral cell**
 Chesser, S.A.
- P 11 Wave climatologies of the Australian southeastern coast**
 Trenaman, N.L., and Short, A.D.

Wednesday, November 17

Session 5: Longshore Modeling I

- 08:20 Reduction process of river delta due to natural change of its river mouth—in the case of the Kurobe River**
 Tsuchiya, Y., and Yamashita, T.
- 08:40 Long term evolution of the Kashimanada Coast**
 Tomaru, T., Saitoh, K., Tago, Y., and Shibata, M.
- 09:00 Large scale response of the Nile River delta system**
 Dean, R.G., Fanos, A.M., and Khafagy, A.A.
- 09:20 Formation process of river delta by short cutting of river—in the case of Shinanogawa River flowing into Japan Sea**
 Yamashita, T., Saito, M., and Tsuchiya, Y.
- 09:40 An overview of geologic and oceanographic assumptions required by coastal models**
 Pilkey, O.H., Young, R.S., Thielier, E.R.
- 10:00 Discussion Period**
- 10:20 Break**

Session 6: Longshore Modeling II

- 10:40** Numerical modeling of the propagation of long-shore sand waves
Thevenot, M.M., and Kraus, N.C.
- 11:00** Numerical predictions of radiation stresses throughout the southern California bight
O'Reilly, W.C., and Guza, R.T.
- 11:20** Large-scale sediment movement in a macrotidal high-energy coastal area: the French coast of the English Channel
Levoy, F., and Avoine, J.
- 11:40** Offshore bathymetry and time step in shoreline response numerical models
Hanson, H., and Kraus, N.C.
- 12:00** Discussion Period
- 12:20** Lunch (Boardwalk)

Session 7: Inlet Behavior

- 13:20** Coastal behavior of inlets and barrier islands along tide-dominated sections of large scale coastal compartments
Oertel, G.F., and Foyle, A.M.
- 13:40** Natural and anthropogenic inlet-induced shoreline change in Southeastern North Carolina
Cleary, W.J., and Hosier, P.E.
- 14:00** Regional impacts of inlet engineering and beach replenishment at Fenwick and Assateague Island, Maryland
Stauble, D.K., Underwood, S.G., Byrnes, M.R., and Hiland, M.W.
- 14:20** Long-term morphodynamical behavior of the E. Frisian islands and coast
Niemeyer, H.
- 14:40** Field measurements and large scale morphological changes in a tidal inlet system in the Wadden Sea
Louters, T., van de Kreeke, J., and Biegel, E.J.
- 15:00** Discussion Period
- 15:20** Break

Session 8: Inlet Modeling

- 15:40** Predicting morphodynamic change from tidal residual vectors at a large tidal inlet, Tauranga Harbour, New Zealand
Healy, T., Bell, R., and de Lange, W.
- 16:00** Techniques for long-term morphological simulation under tidal current action
Latteux, B.

- 16:20 **Mathematically shaping of tidal inlets**
 Bakker, W.T., and de Vriend, H.J.
- 16:40 **Morphodynamic modelling for a tidal inlet in the Wadden Sea**
 Wang, Z.B., Louters, T., and de Vriend, H.J.
- 17:00 **Discussion Period**

Thursday, November 18

Session 9: Effects of Sea-level Rise

- 08:20 **Quantification of morphological changes of some supratidal sands in the German Bight (Germany)**
 Hofstede, J.L.A.
- 08:40 **A method of establishing mesoscale (decadal to sub-decadal) domains in coastal gravel barrier retreat rate from tide gauge analysis**
 Orford, J.D., Carter, R.W.G., and McCosky, J.
- 09:00 **Coastal evolution and accelerated sea-level rise**
 Nicholls, R.J.
- 09:20 **Simulation modelling of LSCB: Parametric scaling, markovian evolution and site idiosyncrasies**
 Cowell, P.J., Roy, P.S., and Jones, R.A.
- 09:40 **Shoreface profile evolution on the time scale of sea-level rise**
 Stive, M.J.F., and de Vriend, H.J.
- 10:00 **Discussion Period**
- 10:20 **Break**

Session 10: Morphologic Behavior

- 10:40 **Erosion and accretion of the Ebro delta coast: a large scale reshaping process**
 Jimenez, J.A., Valdemoro, H. I., Sanchez-Arcilla, A., and Stive, M.J.F.
- 11:00 **Controls on barrier spit evolution: a comparison of Buctouche Spit, New Brunswick and Long Point Spit, Ontario, Canada**
 Ollerhead, J., and Davidson-Arnott, R.G.D.
- 11:20 **Long-term morphodynamic evolution of beaches and barriers: Examples from paraglacial coasts**
 Carter, R.W.G., Forbes, D.L., Orford, J.D., Jennings, S.C., Shaw, J., and Taylor, R.B.
- 11:40 **Evolution of a barrier island and recurved spits on a macrotidal coast**
 Bristow, C.S., Horn, D.P., and Raper, J.F.
- 12:00 **Discussion Period**
- 12:20 **Lunch (Pavilion)**

Session 11: Morphologic Modeling

- 13:20 Two-dimensional numerical modeling of large-scale deltaic sedimentation
Letter, J.V., Donnell, B.P., and Powell, N.J.
- 13:40 A dynamical model for tidal offshore sand banks and sand waves
Hulscher, S.J.M.H., de Swart, H.E., and de Vriend, H.J.
- 14:00 Tidal schematisation in morphodynamic area models
Chesher, T.J.
- 14:20 Development and application of a generic modeling system for coastal processes
Warren, R
- 14:40 The Historical evolution of the coasts of Venice, Italy
Broker, I.
- 15:00 Discussion Period
- 15:20 Break

Session 12: Sediment Budget

- 15:40 The sediment budget on rugged coasts: special considerations
Everts, C.
- 16:00 A sediment budget for Saco Bay Maine, and an evaluation of the long- and short-term geologic and oceanographic processes
Dickson, S.M., Kelley, J.T., Belknap, D.F., Fink, L.K., Barber, D.C., Fitzgerald, D.M., van Heteren, S., and Manthorp, P.A.
- 16:20 Large-scale transfer of sand during storms: Implications for modeling and prediction of shoreline movement
Morton, R.A., Paine, J.G., and Gibeau, J.C.
- 16:40 Theoretical concepts of parameterization of coastal behaviour
Terwindt, J.H.J., and Kroon, A.
- 17:00 Discussion Period
- 17:20 Closing Comments
- 19:00-22:00 Banquet Dinner (Boardwalk)

Friday, November 19

- 08:00-17:00 Field Trip (Optional)

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