WASHOVER TRANSPORT ON SANDBARS

Literature study

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1. INTRODUCTION
2. ORIGIN OF ISLANDS

2.1 General

Barrier beaches (islands and spits) are found along coasts all over the world. These elongated, narrow landforms are composed of sand and other loose sediments transported by waves, currents, storm surges and wind. The term "barrier" identifies the structure as one that protects other features, such as lagoons and salt marshes, from direct wave attack of the open ocean. Also human interference in coastal zones (reclamation projects, harbour development plans, coastal protection) require a clear view in barrier beach morphology. Therefore due to this intense development pressure on barrier beaches, there has been a surge in research over the past twenty years to determine their geologic and ecologic characteristics in order to more clearly define barrier carrying capacities and susceptibility to storm damage.

Since they are produced by one of the fundamental coastal processes, viz. the displacement of material, accumulation forms provide an excellent indication of coastal dynamics. For an accumulation to form, material must be moved to a given locality and deposited there. To explain the origin of forms it is therefore necessary to ascertain how the material moves and the reasons for any cessation or weakening of this process (Zenkovich, 1967).

In paragraph 2.2 the development of coastal features will be discussed and paragraph 2.3 describes the different forms and types of these features.

2.2 Barrier Island Development

2.2.1 General

The type of barrier island found in a given geographical region reflects the geological history of that particular region, sediment supply, oceanic forces, and relative sea-level changes. Man has to be cautious as information about the development of the barrier islands is only preserved where the coastline has not undergone major retreat (Balkema, 1988).

2.2.2 Four main theories of barrier island genesis

Barrier coast similar to those of the North Sea and the East Coast of the United States are found in various parts of the world. As in the American literature (Leatherman, 1982), four different basic concepts of barrier island genesis have been considered for both the North Sea coast and the East Coast of the U.S.: (1) the drowning of coastal dune ridges; (2) the formation of spits, which were later breached by the sea; (3) island development from emerging shoals and (4) barrier islands formation caused by river deposits.

1. Drowning of coastal dune ridges:

Hoyt (1967) proposed that as sea level rose, dune ridges on the seaward
edge of the mainland became the barrier islands. The lower areas behind these ridges were flooded, creating lagoons (Fig. 1). This appears to be the major means of barrier formation along the southeast Atlantic and perhaps the Gulf Coasts.

2. Formation of spits, which were later breached by the sea: Littoral transport:

Gripp (1944) promoted the idea that the chain of islands along the North Sea coast had developed from a chain of spits. Coast parallel sand transport would have created the barrier. However, within this area major spit formation is only observed on the Isle of Sylt. There is an imminent danger that the existing spits will be breached, both being heavily eroded on their western sides. Whether such a breach during a storm tide would actually lead to a separation of the spit from the island, or whether the newly formed gap would be rapidly closed again by coast-parallel sand transport remains an open question (Balkema, 1988). Along the northeast coast of the United States, like the North Sea coast, spit accretion by littoral drift appears to be the characteristic manner in which barrier islands have formed. Barrier beaches form as sand is transported from a source, such as sea cliffs and associated beaches, toward a region of accretion and sedimentation in open water. The result may be a spit: an elongate, fingerlike ridge composed of beach sediment that is attached at one end to the mainland but terminates in open water. Some barriers clearly grew laterally as long spits and were converted into barrier islands with inlet truncation (Fig. 2 and 3).

3. Island development from emerging shoals: Onshore transport:

The third major mechanism for barrier island formation is by submarine bar upbuilding (Fig. 4). Ordemann (1912) appears to have been the first to argue that the islands had formed as a result of emerging sandbanks. Behrmann (1922) also favoured this concept.

In the course of his geological mapping of the East Frisian Islands along the Dutch coast, Keilhack (1925) came to the same conclusion that the island chain had been built up by sediment supply from the sea. From the formation of the Isle of Memmert he invoked the following model of barrier island formation:

1. Material from the nearshore bottom is transported landward by waves forming offshore bars near the shore.
2. Increasing sand supply causes the shoals to rise until they finally emerge during low tide.
3. If accumulation continues, they eventually remain dry even during ordinary floods.
4. Under the influence of the wind, parts of the former shoal finally grow above spring tide level, forming an island.

4. Barrier islands formation at the mouths of major rivers:

Barrier island may also form at the mouths of major rivers which deposit large quantities of sediment in the ocean. For instance, small barriers have developed on the Mississippi River delta and on small deltas along the west coast of Taiwan through wave erosion and reworking of these riverine sediments. Some rivers are cut off some 80 years ago. Barriers islands which were formed by sediment deposits from
those rivers are eroded and deformed heavily by waves, currents and wind, suffering great sand losses at the distal end of the barrier.

A similar type of barrier beach can develop where rivers empty into the sea without forming large deltas. Along the East Coast of the U.S. barrier beaches can be found near the mouth of some major rivers.

Marshy platforms or cheniers can originate in a low-energy environment, where a large portion of sediment in transport is silt, clay and very fine sand. Ridges are created when storm surges push sandy/shelly sediment landward from the shoreline for deposition on the marsh surface.

Modern (Holocene) barrier islands can also be welded onto mainland features, which are called coastal plain remnants. Thus the core of these sea island consists of Pleistocene sediment which were deposited prior to the last Ice Age. These sections of the coast were cut off from the mainland by sea-level rise and are now isolated as islands.

Adding to this complexity, barriers have been reshaped by waves and currents since their initial formation and migrated landward in response to a rising sea level. Although an island may have developed by any of the above mechanisms, subsequent changes due to landward migration have resulted in major alterations to its original geomorphic structure.

2.3 Barrier Types

2.3.1 General

In this chapter two forms of classification of barrier island will be discussed. The first classification is based upon the examination of the geomorphic expression of coastal features (Leatherman, 1982). Discussing the different types of barrier islands, a classification can be made regarding the island morphology as a function of tidal and wave regime (Hayes, 1975).

2.3.2 Bay Barriers

Barrier beaches, that are connected at both ends to headlands, are called bay barriers. These barriers are typically small and common along highly indented coastlines, where rock or glacially-formed headlands jut into the sea. Bay barriers usually do not have permanent inlets, but temporary openings can occasionally occur. Bayhead barriers are found in the upper reaches of a bay and form where the adjacent cliffs are highly resistant to erosion. (Fig.5). While some sandy sediments may be supplied from sea cliff erosion, much of the material necessary to form the barrier feature is derived from other sources, such as in-place biological (shell) production, upland runoff and adjacent rivers.
Barrier beaches that connect headlands together along the outer reaches of an embayment are called baymouth barriers (Fig. 6). These barriers form where headlands are easily erodable and contain sufficient quantities of sand and gravel for barrier construction. Baymouth barriers originated as spits merging from adjacent, eroding headlands, the end result being a fairly straight shoreline. Baymouth barriers are common in microtidal areas, where low tidal range inhibits inlets except immediately following breaching by a major storm. With the restriction of salt water input, fresh water or brackish marshes will develop.

A tombolo can be regarded as a special kind of bay barrier, but probably deserves recognition as a separate feature (Fig. 7). Tombolos attach an island to the mainland following spit growth. The island, which may be erodable or nonerodable, serves to anchor the barrier. Tombolos are most common along the Northeast glacial coast of the United States, where the attached islands are erodable and provide sand for barrier development.

2.3.3 Barrier Spits

Barrier beaches that are attached at one end to a source of sediment - mainland or large island- and extend into open water, are called spits. Barrier spits are formed as longshore currents move abundant sand and gravel from eroding cliffs into open water. Since these sediment-laden currents are directed straight along the shore, material is deposited offshore of shoreline embayments. With the continued delivery of sand in conjunction with overwash and wind, a barrier spit finally forms.

Spits elongate in the general direction of the littoral drift and therefore represent the direct movement of sand along a beach. Barrier spits can develop into bay barriers if they grow completely across a bay, or bay barriers become spits if a permanent inlet is created. Leatherman (1986) divided spits into four types based on their overall shape and position relative to each other: simple spits, double spits, recurved spits and convex spits.

Spits which are relatively straight and narrow are called simple spits (Fig. 8). Examples of simple spits include Ogunquit Beach, Maine, and Sandy Neck, Massachusetts.

A double spit develops when there is significant longshore sand transport in both directions along the shoreline (Fig. 9). The inlet between the two spit segments is often unstable, migrating laterally in response to the prevailing direction of net littoral drift. If the inlet closes, the double spit can be converted into a baymouth barrier. The inlet usually remains open at all times since the enclosed bay dictates tidal exchange with the open ocean.

Recurved spits are essentially simple spits which are recurved or significantly bent shoreward (Fig. 10). The distinctive recurved dune ridges represent lines of accretion, indicating successive barrier growth. Recurved spits form when the barrier is growing seaward from a headland terminus such that the barrier is subject to waves from a wide range of directions. In some cases spits can form at the downdrift end of a barrier island; these spits can exhibit large recurvatures, such as Fishing Point at the southern end of Assateague Island, Virginia.

Complex spits are uncommon and only form when both sides of a mainland peninsula are eroding and subject to large enough waves to produce significant longshore sand transport (Fig. 11). The only documented
example of a complex spit is the Province Lands formed as a large recurved spit followed by the growth of an enclosed simple spit. The extremely large size of this barrier spit and the presence of multiple dune ridges on the recurved flank make it the most stable of all such features.

2.3.4 Barrier Islands

Barrier islands, which occur primarily on coastal plain shorelines located on the trailing edges of continents and on marginal seas, vary in morphology in response to the interaction of tidal range and wave energy effects. Barrier islands are restricted to those coastal plain shorelines with tidal ranges less than approximately 4m. This conclusion is based on a study of coastal charts of the world, conducted at the Defence Research Laboratory, University of Texas (in 1963-64; discussed by Hayes 1965 and Hayes and Kana, 1976). The lack of extensive barrier islands in areas with large tides has been pointed out by several other authors (Prince 1955, Gierloff-Emden 1961 and King 1972). Gleaser (1978) noted that "only 10% of the world's barrier islands are present along coastlines where the tide range exceeds 3 m".

In the following text the two most important parameters involved in barrier island formation and coastal morphology will be discussed.

Following the work of W.A. Price (1955), it is concluded that the most important control of the geomorphology of depositional coasts is the type and amount of hydraulic energy expended within an area. Furthermore, the two energy factors of most significance are wave energy and tidal current energy, which can be related directly to tidal range. Davies (1964) classified shorelines on the basis of tidal range (see table 2.1).

Table 2.1: Shoreline classification

<table>
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<tr>
<td>microtidal coasts</td>
<td>0-2 m</td>
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<tr>
<td>mesotidal coasts</td>
<td>2-4 m</td>
</tr>
<tr>
<td>macrotidal coasts</td>
<td>&gt; 4 m</td>
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In areas of average marine wave conditions, coasts with small tidal ranges (microtidal) are usually dominated by wave energy, and coasts with large tidal ranges (macrotidal) are usually dominated by tidal currents and tidal-level fluctuations. Coasts with intermediate tides (mesotidal) show influences of both waves and tides and are thus termed mixed-energy coasts (Hayes, 1965).

The reason for the emphasis on tidal range is the fact that the effectiveness of wave action diminishes (i.e., waves cannot break in a concentrated area for a long period of time), and tidal current activity increases as the vertical tidal range increases. Small waves are most effective on coasts with small tidal ranges. On the other hand, in areas of small waves, a smaller tidal range is required to produce tidal-influenced morphology than on coasts with medium or high wave energy.

An other variable which plays a principal role in determining barrier expression is sand supply. Utilizing Hayes' (1979) micro-meso classification coupled with sediment supply conditions yields four
basic types of islands:

- Microtidal transgressive
- Microtidal regressive
- Macrotidal transgressive
- Macrotidal regressive.

Transgressive indicates sand deficiency and the tendency for shoreline erosion. Regressive denotes a past period of accretion as evidenced by multiple dune ridges. Past accretion does not preclude present or future erosion; a regressive barrier can be converted to a transgressive one with the demise of sufficient sand supplies (Fig. 12).

For macrotidal conditions (tidal ranges above 4m), barriers are not expected, although it is possible that some small, ephemeral features may exist. In actually, barrier island morphology depends upon tidal energy in relation to wave energy so that exact boundaries cannot be set by knowledge of tidal range alone. Macrotidal shorelines are also characterized by a broad extensions of intertidal flats and salt marshes. Sand deposits are usually restricted to linear sand shoals, or tidal current ridges, in the offshore areas.

**Microtidal barriers**, corresponding to wave-dominated shorelines, are expected along sandy coastlines with a tidal range less than 2 metres. Leatherman (1979) describes that there is a significant difference between microtidal barrier island morphology and macrotidal barrier island morphology. These differences are outlined in table 2.2, derived from a study of barrier island formation at the East Coast of the United States.

Microtidal transgressive barriers (Fig. 13) in areas of medium wave energy are the least stable and most vulnerable to storm-induced changes of the four island types. Tidal inlets and tidal deltas are of relatively minor significance, in contrast with mesotidal barriers, where tidal inlets and tidal deltas are large and significant. These islands, which are long and narrow, are characterized by low-lying topography and numerous washovers, indicating deficient sediment supply and relative rapid shoreline retreat. Figure 14 is an example of a microtidal regressive barrier.

**Mesotidal shorelines** are characterized by barriers with a short, stunted nature and by the abundance of tidal inlets. Mesotidal barrier island can also be subdivided into transgressive (Fig. 15) and regressive types (Fig. 16), reflecting sediment supply. Tidal inlets between mesotidal barriers tend to be more stable then those in a microtidal setting. As a rule, the ebb-tidal deltas are larger then the flood-tidal deltas (the opposite is true for microtidal areas). FitzGerald (1977) suggested that the ebb dominance of the main channel at Prince Inlet, South Carolina, results from greater inlet efficiency at low water than at high water, which shortens the time of ebb flow and lengths the time of flood flow, hence producing stronger ebb currents. The large tide necessitates many breaks through the barrier chain to allow daily exchange of water between the enclosed bays and the open ocean.

Table 2.2 shows the main barrier characteristics comparing microtidal and mesotidal barrier islands.
Table 2.2: Barrier island characteristics (after Leatherman, 1976)

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<th>Mesotidal</th>
</tr>
</thead>
<tbody>
<tr>
<td>Length</td>
<td>long (30-100 km)</td>
<td>stunted (3-20 km)</td>
</tr>
<tr>
<td>Shape</td>
<td>elongated hot dog</td>
<td>drum stick</td>
</tr>
<tr>
<td>Washover features</td>
<td>abundant: washover terraces &amp; washover fans numerous</td>
<td>minor: beach ridges or washover terraces; washover fans rare</td>
</tr>
<tr>
<td>Tidal inlets</td>
<td>infrequent</td>
<td>numerous</td>
</tr>
<tr>
<td>Flood-tidal deltas</td>
<td>large, commonly coupled with washovers</td>
<td>moderate size to absent</td>
</tr>
<tr>
<td>Ebb-tidal deltas</td>
<td>small to absent</td>
<td>large with strong wave refraction effects</td>
</tr>
</tbody>
</table>

Three other types of barrier structures deserve attention: capes, sea islands and cheniers.

Barrier islands that project into the sea to form a right-angled shoreline are called cuspate forelands or capes (Fig. 17). Cuspate forelands typically consist of transgressive and regressive sections.

Sea islands, like cheniers, are principally found in specific regions along the U.S. coastline. These large islands consist of two parts: a recent or Holocene outer barrier and a Pleistocene core (Fig. 18). The outer barrier may be welded onto the Pleistocene core or separated from it by intervening salt marshes and tidal creeks. Only the outer barrier structure behaves in a dynamic manner. The island core is Pleistocene in age (approximately 100,000 years old), is quite high (often above the 100 year and sometimes above the 500 year floodplain level), and converted by a mature, thick continental soil which supports a climax maritime forest. Thus, the Pleistocene core, which can be quite large in areal dimensions, is actually a very stable feature and is merely an emersed piece of the mainland.

Cheniers, which are the lowest and most vulnerable to storm surges of all barrier types, are generally strand plains, instead of true barrier structures (Fig. 19). Cheniers consist largely of organic deposits (salt marshes) with ridges of sand and silt overlaying this basal structure (Leatherman 1979, 1982).
3. THE COASTAL SEDIMENTARY ELEMENTS OF BARRIER ISLAND SYSTEMS

3.1 General

The fundamental elements in descriptive definitions of barrier islands usually include the terms elongate, narrow land formations, composed of unconsolidated materials, lying parallel to the shoreline, and separated from the mainland coasts by lagoons and/or bays. Operational definitions used by a variety of authors and working groups (Berryhill et al., 1969; Cromwell, 1971) may impose other limitations on the definitions such as "generally being less then 10 m above sea level". Other operational restrictions that have been used include a length to width ratio of less than 10:1, and commonly being subjected to wave, tidal and wind energies. While the above definitions are generally geomorphically oriented, several authors have attempted to define barrier islands as being composed of several major depositional environments. One of them is G.F. Oertel (1984). He considered the barrier island as the focal element of a much larger system that is termed the barrier island system.

The barrier island system consists of seven major coastal environments. The seven environments are also required elements needed to impose the designation "barrier island" to a littoral sand body. The elements are (see Fig. 20):

1. Mainland
2. Backbarrier lagoon
3. Inlets and inlet deltas
4. Salt marshes and tidal flats
5. Washover fans
6. Barrier flats
7. Shoreface.

3.2 Mainland Element

The mainland element of the barrier island system establishes an island as a barrier and is a necessary requirement for the designation of barrier island. Three major characteristics of the mainland determine how this element interacts with the barrier island system; these are:

1. lithology
2. slope
3. drainage.

3.2.1 Lithology

Lithological characteristics of the mainland strongly affect the rate of lagoonal expansion by influencing the retreat rate of the mainland shoreline of the backbarrier lagoon. Studies on the erosion (shoreline recession rate) of the esturine shoreline in the Albemarle-Pamlico region of North Carolina (Bellis et al., 1975) illustrate how shore type influence the erosion rates. Mean erosion rates for five counties bordering the mainland shores of the lagoon ranged from 0.3 to 1.4 m yr⁻¹. However, local recession rates of 0.6-0.9 m yr⁻¹ where typical of many areas and rates of over 6.1 m yr⁻¹ had been reported. Bellis et al. (1975) showed that exposed, low shorelines of sand and clay were less susceptible to erosion than marshy shorelines (0.9-2.4 versus 1.8-6.1
m yr⁻¹, respectively). Under similar energetic conditions it would be anticipated that indurated material along the mainland shore of backbarrier lagoons would be far less susceptible to erosion.

### 3.2.2 Slope

Regional slope is also a very important factor controlling the recession rate of lagoonal and barrier island shorelines. In backbarrier lagoons wave energy is generally considerably lower than in the open seas. Therefore, recession of mainland shores is largely controlled by submergence. Assuming a constant rate of sea-level rise during periods of barrier island evolution, the rate of shoreline recession would be: (1) constant at constant slopes; (2) increase at decreasing slopes and (3) decrease at increasing slopes.

The mainland element of the Virginia (USA) barrier island system has an average slope of approximately 0.5° and the estimated local rate of sea-level rise is about 30 cm yr⁻¹ (Hicks, 1972; Frommer, 1980). These conditions would produce a recession rate of 34.4 cm yr⁻¹ due to submergence.

### 3.2.3 Drainage

Drainage characteristics have a greater influence upon the shapes of evolving barriers than on the rates of development. Coastal areas with trellis drainage pattern (i.e., relict coastal ridges and swales) often produce barrier systems with well-developed shore-parallel trends. Coastal areas with well-developed dendritic drainage topography are more apt to have headland barrier systems than have wetlands and estuaries with shore-normal trends.

While the influence of lithology, slope and drainage pattern are readily apparent along the mainland shore of a barrier lagoon, the overwhelming influence that the mainland surface has on a retreating barrier system has only recently been given attention by several coastal workers (Kraft, 1971; Kayan and Kraft, 1979; Halsey, 1979).

### 3.3 The Backbarrier Lagoon

Lagoons or bays are large bodies of open water that are protected from oceanic forces by a barrier beach; fresh water input is generally limited to land surface runoff. The most obvious characteristics of backbarrier lagoons are surface-area changes related to ebbing and flooding tides (Oertel and Dunstan, 1981).

Two endpoints exist: - **open-water** lagoons - **expandable tidal** lagoons.

Regardless of tidal stage, open-water lagoons are bodies of water that have a relatively constant water surface area between the main land and the barrier island. Albemarle and Pamlico Sounds in North Carolina are examples of open-water-type lagoons. Water-surface areas of expandable lagoons increase by greater than 15% between low and high tide. Many barrier lagoons illustrate water-surface areas (and prisms) that vary significantly during a tidal cycle. Surface-area variability usually results from partial or complete emergence of the seabed during low water stages of the tide (Fig. 21 and 22). Emergence is related to
tidal water levels with respect to a variety of intertidal lagoonal
features (marshes, oyster reefs, mangroves, tidal flats, flood deltas,
swash berms, and bars, etc.). The coastal marshes of Georgia (USA) are
examples of well-developed expandable tidal lagoons whose water surface
areas increase by greater than 50% between spring low and high tides
(Oertel, 1984).

3.4 Inlets and Inlet Deltas

Separating the individual islands of a barrier chain are tidal inlets
through which tides flow to backbarrier lagoons. These features are
necessary to establish the barriers as islands rather than barrier
spits or bay barriers. Inlets are the primary means by which sand is
transported landward across a migrating barrier beach system. They open
and close in response to changing conditions and may migrate long
distances along a barrier shoreline. An inlet may approach equilibrium
conditions if a balance is achieved between tidal inflows and outflows
that scour a channel, and the longshore transport of sand that tends to
close an inlet.

Inlet stability is related to the strength of the longshore current
versus tidal jet flushing capacity. Inlets tend to close unless there
is substantial outflow of water from a major river. When open, an inlet
acts as a complete or partial barrier to longshore sand transport.
Depending on equilibrium conditions, an inlet may trap sand or
naturally bypass a large portion of the longshore sediment transport
(littoral drift). Under the conditions of small tidal flows and high
rates of littoral drift, an inlet will eventually close.

Tidal barrier inlets have bathymetrical and sedimentological patterns
that result from the interaction of (1) inlet currents; (2) backbarrier
energy and (3) open sea energy sources. Two types of barrier inlets can
be considered: open-ended barrier inlets and Tidal barrier inlets.

When the lateral distance between barrier islands exceeds the length
of the island by several multiples, inlet flow will be more characteristic
of the circulation of adjacent backbarrier lagoon or the open sea
rather than an opening between to different bodies of water. These
open-ended barrier inlets have distinct sedimentation and bathymetric
patterns that are often related to spit development.

Tidal barrier inlets, or fluvial barrier inlets are associated with
river systems and are characterized by multiple distributary channels
that diverge and converge around shoals and islands in the entrance
throat area. Also characteristic for tidal barrier inlets is that the
inlet width is less than several multiples of an island length (Oertel,
1984).

With a rising tide, sand that has been moving along the beach as
littoral drift is interrupted by the tidal current; a portion of this
material will be transported through the inlet for deposition in the
bay. This sedimentation produces an extensive flood tidal delta that
exhibits a deltaic pattern when fully developed (Fig. 23). Figure 24
and 25 show two tidal inlets of the Algarve Barrier Islands, Portugal.
On the seaward side of the inlet are a series of shoals called the ebb
tide delta, marked by breaking waves. The ebb tidal delta is formed
from the sand that moves through the inlet with the ebbing tide. Oertel
(1975b) and later Hubbard et al. (1979) used the general configuration
of ebb tidal deltas as a means of classifying different types of ebb
deltas, by relating delta configuration to the relative magnitudes of
the forces of dynamic diversion (Fig 26).
3.5 Salt marshes and tidal flats

As a barrier spit grows into open water, it provides protection for ecosystems that require quiet water, such as tidal flats and salt marshes. These intertidal environments are built from sands, muds, and clays transported through tidal inlets or across the barrier for deposition in the enclosed bay. Through time the sea bottom is eventually elevated extending the intertidal zone and providing new substrate for salt marsh expansion. Marsh grasses gradually stabilize this surface and spread slowly outward into the bay.

When inlets open, sandy shoals develop in the bay. These shoals, which have a characteristic deltaic pattern, provide substrate for new salt marshes after the inlet closes or migrates downdrift. Marsh that grows close to active inlets or accreting spits is most productive due to continual flushing by tidal currents. Also major washover deposits can serve in this capacity, carrying sand directly into a lagoon.

3.6 Washover fans

Washover fans are created by the flow of water through the primary dune line with deposition of the sand on the barrier flats, marsh, or into the lagoon, depending on the storm magnitude and the island width (Fig 27). Overwash is defined as any swash uprush that crosses a dune line (or storm berm, if no dunes are present). Usually overwash occurs during coastal storms, but a marginal event can occur during extremely high spring tide conditions at low areas in the barrier dune line. Storm overwash can occur through narrow dune caps, over wide sections with low dune topography, or over an entire stretch of a barrier island (Fig. 28). The shape and dimension of the washover deposit is determined by the backshore topography and the volume of sand being introduced by the overwash surges (Fig. 29). Large overwash fans can terminate in the lagoon or bay behind the barrier beach. If scour of the barrier proceeds to an elevation below mean sea level during the overwash period, an inlet can result. Overwash generally occurs where the dune system has been weakened, either naturally by blown outs, or artificially by man.

Where the barrier is wide with extensive barrier flats or salt marshes, washovers rather than inlets will generally occur. Overwash surges quickly lose their initial, high velocities due to percolation and frictional effects as the bore of water crosses the barrier flats. Hence erosion is minimized and sand deposition occurs in the fan (Fig 27) as the surge energy drops to a low level.

For a section of the Louisiana coastline to the east of the Isles Dernieres, Ritchie and Penland (1988) defined four general barrier island morphologies: washover flats, washover terraces, dune terraces, and continuous dunes (Fig. 30).

Revegetation of barren washover flats progresses rapidly after the dune line has been restored. From field inspection it is sometimes difficult to identify old washover areas, but their existence can often be verified by using historical aerial photography (Leatherman, 1982).
3.7 Barrier Flats

Barrier flats are located on the lagoonal side of barrier islands and have negligible relief (generally less than 2 m above mean water level). These areas may result from the destruction of dune areas by overwash and inlet processes (Fig. 31).

Dunes are the major topographic feature of the interior of northeast barrier beaches. Therefore, barrier flats are quite restricted in areal extent. In some cases, dunes and salt marshes share a common border, and no true flats exist, as is the case on parts of Nauset Spit, Cape Cod, Massachusetts. However, mid-Atlantic and southeastern U.S. coast barriers are more typically broad and low and have extensive flats.

Barrier flats covered with grassland or meadow vegetation are adapted to frequent sand burial and flooding by overwash. With decreasing overwash frequency, either because of increasing dune growth or lack of storms, barrier flats will support thickets, followed by woodlands and finally forest.

3.8 Shoreface/Beach

A beach can be defined as an accumulation of wave-washed, loose sediment that extends between the outermost breakers and the landward limit of wave and swash action. A generalized beach profile and related terminology are shown in figure 32. The beach consists of the backshore and foreshore. The backshore, which includes the berm, is subject to wave action only during storm conditions. The foreshore is the intertidal area that lies seaward of the berm crest.

Where waves break offshore, submerged ridges or bars of coarse sediment are present. Most waves break at the inner bar that parallels the shore of a barrier island, but some smaller waves may not break until reaching the water's edge. Bars, by definition, are submerged at all times. A longshore feature that becomes exposed at low tide is called a ridge. The trough between the ridge and beach berm is termed a runnel. Ridges and runnels are commonly found along mesotidal coasts.

Waves are the prevailing source of energy delivered to the shoreface and are primarily the agents responsible for shaping an "equilibrium" upper shoreface profile. In general, high waves with short periods cause the beach to erode, and the berm sand is shifted offshore. Low waves with longer periods, primarily in summer, move sand from the bar and return it to the berm. The average wave height varies on a seasonal basis, resulting in typical "winter" (high energy) and "summer" (low energy) profiles. The winter profile is characterized by a narrow backshore, flat foreshore, and a large offshore bar. The summer profile has a wide backshore, a steep foreshore, and a well-developed berm.

With breaking, waves are translated into swash, which is a thin sheet of water flowing over the beach face. Swash consists of two motions: (1) uprush, created by initial wave break and (2) backwash, which is that portion of the water that returns seaward. The swash process results in sediment accretion, creating a beach terrace known as a berm (Fig. 32). The crest of a berm may rise several meters above mean sea level. The height limitation for a berm is determined by the size of the waves; large swell waves can build substantial berms. Severe storms, however, often result in beach erosion with berm flattening.

Although adjustment to an equilibrium profile takes place continuously with the tidal cycle and the seasonally, large-scale changes are most
Washover Transport on Sandbars

evident during a coastal storm. With intense storms, the barrier dunes may be eroded and breaches by overwash surges. Beach changes take place within a very short time interval, lasting from only one-half a tidal cycle to several days. Swell usually follows sea conditions as a low pressure cell (coastal storm) moves along and/or offshore. The swell moves the sand back onshore, often by the process of ridge and runnel migration.

From the berm crest to the toe of the primary dune is the zone termed the beach backshore. This area is rarely subject to hydraulic (water) action, except during extremely high spring tides or coastal storm activity (Fig. 32).

Sand accretion can only proceed as high as the storm surges can reach without the development of dunes. Vegetation is critical in this building and stabilization process; otherwise, the sand would continue moving as it does in the desert. After sand is deposited on the beach backshore by waves, the winds take over and either move the sand inland or back out to sea, depending on the prevailing wind direction. Wind velocities of 12 mph or greater are capable of moving fine, dried beach sand (Leatherman, 1982).

According to Leatherman (1979), coastal dunes play at least three major roles during storms, functioning as:

(1) sand reservoirs
(2) energy dissipators
(3) barriers to storm waves and swash.

The position and height of barrier dunes depends on a number of factors. The height of the dune is controlled primarily by sand size and wind velocity. Seaward dune vegetation growth is controlled by wave action and soil salinity. The seaward dune position also depends upon the frequency of storms eroding the dune face and the rate at which these scarps can heal by wind transport and vegetation growth.

Sand transported over the dune crest results in landward movement of this feature. On eroding shorelines, dune system survival depends upon the relative rates of dune migration and shoreline retreat. During periods of increased shoreline erosion, the dunes may be eroded and breached during storm conditions.
4. PROCESSES OF BARRIER ISLAND EVOLUTION

4.1 General

The long-term behavior of a barrier island depends principally upon the rate of sea level rise, sand supply, and sea energy (Fig. 33). Human intervention can also be an important factor when modern technology is applied. Sand supply is often the key to a barrier island's evolution. With a relative constant supply of sediment, the island can maintain itself in place and build upward with sea level rise (Fig. 34). With an excess sand supply, the barrier can actually accrete seaward.

Barrier islands are subjected to erosional and depositional processes. The most general of these processes are the rise of sea level, a process that results in barrier island transgression, or onshore migration. Other processes include tidal-inlet migration which results in reworking of large portions of barrier islands. Storms play an important role in washover sedimentation, windblown (aeolian) sand transport, inlet formation, and bay infilling. Flood-dominated inlet sedimentation is responsible for partial infilling of shallow water lagoons. Flood-tide deltas weld onto updrift portions of adjacent barriers resulting in wide, thick accumulations of sediments.

Nummedal (1983) pointed out that washover deposits and the sediments in the migrating inlet channels remain a part of the landward and upward moving barrier island as they adjust to rising sea level. The washover and inlet deposits are the predominant material in all the mid-Atlantic barriers. These deposits remain in the barrier complex although they are frequently found in the lagoons behind the islands as submerged parts of the migrating barriers. If this "rollover" process of barrier island transgression were not true the islands would have disappeared long ago (Fig. 35).

In the following paragraphs five major barrier island evolution processes will be discussed.

4.2 Sea level rise

As pleistocene glaciers melted and sea level rose, the river valleys were flooded and much of the riverine sediments have been trapped in coastal estuaries instead of reaching barrier shorelines. With no new sediment or sediment loss, a barrier island must retreat landward up the coastal plain in order to maintain a constant elevation and prevent submergence (drowning) with sea level rise (Fig. 36). Sand from the ocean side of the barrier is transported by water and wind toward the backside of the island so that the whole landform gradually changes its location.

For coasts with a broad, gentle sloping surface, a small rise in sea level will result in a dramatic horizontal retreat of the land (Fig. 36). For a given rise in sea level, a barrier island must migrate landward a distance of two or three orders of magnitude of this value (vertical rise), as shown in figure 36. From the evidence available, it is clear that the barriers located on the coastal plains of the U.S. are migrating landward over the long term as sea level rises.

Thus, rising sea level is the primary driving force for landward movement of the shoreline. Through geological time sea level has always been rising or falling relative to the land surface. The last major
change in sea level occurred during the most recent Ice Age, when sea level was approximately 100 meters lower than at present. The volume of water in the sea increased very rapidly after ice melt intensified about 15,000 years ago, with sea level rising at a faster rate then during the past 5,000 years. Presently, the rate of sea level rise, relative to land, averages 40 cm per century along the Northeast coast of the U.S. Where the land is also subsiding, relative sea level rise has been much greater.

While melting of the polar ice cap continues to be a major cause of rising sea level, the rate of rise is expected to accelerate in the future due to increasing global temperatures, driven by higher levels of carbon dioxide in the atmosphere. Due to the "greenhouse effect", higher average global temperatures could result in as much as a 2 m rise within the next 100 years, which could be catastrophic for major cities located in low-lying coastal areas (e.g. Washington D.C., London).

Orford et al. (1991) determined the relationship between sea level rise and gravel barrier migration for the Story Head barrier, Nova Scotia, Canada. The results are shown in appendix 1.

It is clearly that the submergence of barrier island due to various processes along the coast has the same effect on the migration of barrier islands as sea level rise.

4.3 Storm Events

Sea level rise by itself would not be sufficient to drive the barrier beach system landward. The major amount of sand transport occurs in a quantum fashion during extreme events. Storms that effect the coastline fall into two basic categories: (1) hurricanes and (2) "Normal" storms.

Near the centre of a hurricane, winds may gust to more than 200 miles per hour. While intensive winds are the most obvious threat to life and property and do much damage, massive storm surges are by far the greatest cause of deaths and destruction on barrier beaches. Six thousand people died on a barrier island - Galveston Island - during a 1900 hurricane surge which still stands as the worst natural disaster in U.S. history.

A storm surge is a super-elevated mound of water that sweeps across the coastline near the area where a hurricane passes or makes landfall. Surge and hurricane driven waves act in deadly combination to hammer the shore, sweeping across any low-lying coastal areas. Barrier islands being the outer shorelines, are forced to absorb the storm brunt, reducing the force of wave and wind before they reach the mainland.

Ritchie & Penland (1987) investigated the impacts of storms and hurricanes on the Caminada-Moreau barrier headland, along the deltaic coast of South Louisiana. This coast is a storm-dominated, Microtidal environment with a mean diurinal range of 36 cm. Normally, the passage of frontal systems elevates water levels up to 90 cm, 10 to 30 times per year (Boyd and Penland, 1981). These storms often produce waves 2 to 3 m high; the local average wave heights are about 60 cm. Tropical storms (winds over 63 km/h) have a recurrence interval of 1.6 yr., and hurricanes (winds over 118 km/h) have a longer recurrence interval of 4.1 yr. (Sympon & Lawrence, 1971). These storms raise water levels considerably higher that the more common cold front passage, normally 2 to 7 m above mean water level. (Table 4.1). During these storms the frequency and intensity of overwash is a function of the water level
Washover Transport on Sandbars
elevation produced by a combination of storm surge, wave set-up, wave
run-up and astronomical tide settings against the variable, pre-
extisting elevations of the coastline. Measurements of overwash events
has shown that the regional overwash threshold is 1.42 m above mean sea
level, which will produce overwash along 72% of the coastline an
average of 15 times per year; at a level of 2.50 m most of the
coastline will be inundated.

Table 4.1: Potential overwash elevations in coastal
Louisiana (modified from Boyd and Penland, 1981)

<table>
<thead>
<tr>
<th>Event</th>
<th>Overwash elevation (m)</th>
<th>Frequency</th>
</tr>
</thead>
<tbody>
<tr>
<td>Minor cold front</td>
<td>1.42-1.73</td>
<td>10-15 per yr</td>
</tr>
<tr>
<td>Major cold front</td>
<td>1.73-2.02</td>
<td>5-10 per yr</td>
</tr>
<tr>
<td>Force 1 hurricane</td>
<td>1.73-2.92</td>
<td>1 per 8 yrs</td>
</tr>
<tr>
<td>Force 2 hurricane</td>
<td>2.66-4.46</td>
<td>1 per 10 yrs</td>
</tr>
<tr>
<td>Force 3 hurricane</td>
<td>3.90-7.00</td>
<td>1 per 32 yrs</td>
</tr>
</tbody>
</table>

A generalized conceptual model for sand dune development along the
Caminada-Moreau coast in Louisiana has been set up by Ritchie and
Penland (1987) (Fig. 37), as it relates to the role of storms and the
return period of hurricane impact. The model represents an increasing
volume of supratidal sand storage leading to dune development and
revegetation, increasing the stability of the barrier shoreline. Major
storms correlate to hurricanes and minor storms correlate to cold
fronts. Figure 38 shows the effect of hurricane Bob (1979) on the
cross-sectional area of a barrier island along the Caminada-Moreau
coast.

4.4 Inlet Dynamics

Inlets are important for barrier island migration by developing flood
tidal deltas. A portion of sediment moving along a coast is swept
through an inlet and into a lagoon by the flood currents (Fig. 23, 24
and 25). Since bay water is less turbulent than that along the ocean
shore, sand settles out to form a tidal delta. Some sediment is moved
back out to sea as the current reverses on the ebb tide, but usually
the net change result in bay sediment deposition. The growth rate of a
flood tidal delta is a measure of the amount of sand being trapped from
the littoral drift by the inlet. Flood tidal deltas serve as platforms
upon which salt marshes can develop once an inlet closes or migrates
downdrift (Fig. 23 and 39).

There is evidence that a cycle of inlet formation, marsh development,
and overwash occurs on most barrier beaches as they migrate landward
(Leatherman, 1982). An inlet will occur at a narrow point along a
barrier beach, then migrate downdrift until it closes or reaches the
end of the spit. (Fig. 40). With time, the barrier will migrate over
older inlet or overwash deposits.

It is known that a given volume of water in a tidal prism tends to
maintain a constant size cross-section in a tidal inlet or in a series
of inlets. The situation seems analogous to that for rivers, where the
section area is related to the stream discharge. For tidal inlets,
there appears to be a balance between the scouring action of the tidal
currents that keep the channel open, and the longshore transport of beach sand that tends to close them (Johnson, 1919), see also figure 40. Both O'Brien (1969) and Jarrett (1976) have shown a bivariate relationship between cross-sectional area of channel and tidal prism of the form:

\[ A = nV^m \]

Where

- \( A \) = minimum cross-sectional area of the inlet channel measured below M.S.L.
- \( V \) = volume of tidal prism between spring tides
- \( n \) = constant
- \( m \) = constant.

The values of the constants for Atlantic coast inlets are \( m = 1.05 \) and \( n = 3.04 \times 10^4 \) when \( A \) and \( V \) are in m\(^2\) and m\(^3\) (USACE CERC, 1984a). For Oregon Inlet, Jarrett (1976) lists the spring tidal prism as \( V = 133 \times 10^6 \) m\(^3\) and the channel cross-sectional area as \( A = 6190 \) m\(^2\) with a hydraulic radius of 4.1 m.

4.5 Overwash processes

The second principal method by which sediment is transported across the barrier island is overwash. Overwash may be considered as a higher magnitude extension of overtopping, in that the volume of swash that moves over the crest is sufficient to create a unidirectional flow on the backslope, eroding the upper back beach, lowering the crest, and depositing material on the landward side (Carter & Orford, 1981).

The magnitude and frequency of overwash depends on the following elements:

- barrier exposure and orientation
- shore-parallel variation of the beach crest line.
- frequency of major storms
- wave energy
- tidal range
- beach sediment
- ecological response of vegetation to the overwash process

In paragraph 5.2.2 examples of the quantification of overwash compared with the total sand loss of a barrier island which is exposed to the above mechanisms will be discussed.

Normally overwashing would occur during high-magnitude, low-frequency storms where swash flows create and occupy narrow channels through beach or dune ridges, which act as conduit transferring sediment into characteristic fan-shaped sinks. This process provides an important mechanism for shoreline migration. Where tide ranges are low (microtidal) and storm frequency high, overwash is a regular event on barrier islands without naturally high or artificially-stabilized dunes. Conversely, where tide ranges are large (mesotidal), the chance that a storm will arrive at the highest tide levels and overtop the barrier is considerably less, and hence overwash is an infrequent event (Leatherman, 1982).

The ecological response to overwash is different on Northern and Southern barrier beaches (Leatherman 79). Godfrey (1969) made an overwash model based on Core Banks, North Carolina, that shows the ecological response to overwash on a retreating barrier island (Fig. 41).
4.6 Aeolian processes

Overwash sedimentation is episodic, occurring during discrete storm events of varying magnitude and frequency. **Aeolian processes** (wind-blown processes), which operate on a more continual basis, can deposit or erode sand from washover fan surfaces. Leatherman (1976a) concluded that aeolian processes were slightly more dominant than overwash in accounting for changes in sediment budgets along northern Assateague Island, Virginia-Maryland, U.S. during the mid-1970's. Leatherman also observed that aeolian deflation between overwash events was significant enough to result in net lowering of fan surfaces by deflation over the period of his study. Similar high deflation rates were observed by Fisher and Stauble (1977) on North Assateague following a period of excessive offshore winds during the winter of 1976-1977.

Although mean wind velocity and frequency of windy days is important for deflation, the variability in wind seems to be less important than the magnitude and frequency of fan wetting events. Periods of high accumulation can occur when washover fans remain wet, correlating with periods of frequent overwash and precipitation. Studies of the relative role of overwash and aeolian processes on washover fans (Kochel and Wampfler, 1988) showed that the aeolian erosion of fan surfaces is greatly effected by how wet the fan remains, due to the combination of overwash flooding and precipitation, because wet fans are more resistant to deflation.

4.7 Littoral Drift

The movement of sand along a beach is accomplished by longshore currents. As waves approach a beach at an angle, a current is generated along the shore, resulting in sand movement in this direction.

Barrier beaches can form if sand is transported from a source, such as an eroding glacial headland, for deposition into open water. The longshore transport of sediment by wave action tends to lengthen the beach in the direction that waves generally strike a coastline. These currents flow parallel to shore, and thus tend to build relatively straight sediment deposits along the coast, even if the shoreline is indented to form a bay (Fig. 42). The configuration of the coast can be extremely varied and projections need not necessarily be found at bay mouths, but the flow of material may be slowed down at any point where the direction of the coast changes sharply.

Spit formation through the mechanism of littoral drift is the principle process shaping barrier beaches. As a spit builds across open water the early stages are dominated by overwash that regularly moves sand from the ocean to the bay. At the same time, tidal currents are moving huge quantities of sediment around the end of an accreting spit. This material is deposited in the quiet waters in the lee of the protective barrier.

4.8 River sediment discharge

An important factor in the study of deltas is that a large amount of material in suspension may be brought down by rivers in time of flood and deposited offshore. This material builds up the bottom and hinders the production of a profile of equilibrium by wave action. If there is relatively little alluvial material, the waves are able to move all the material shorewards and thus cause the coast to advance seawards. Small
Washover Transport on Sandbars

River mouth bars often form in the shallow water in front of deltas. Deltas that are growing actively and advancing into relatively deep seas are fronted by a shallow zone as described above. This zone, which is usually fairly narrow, is bounded by a steep slope (perhaps 20°). Examples include the deltas of the Danube (fig. 43), and the Rhône. Submarine furrows, often smoothed and gently sloping, are characteristic features in the shallow water off former and existing river mouths. After a rise of sea level they may be regarded as submerged river courses, but their true nature can usually be established (Zenkovich, 1958).

The generally held view that the bars always develop from alluvial material is not true. On the contrary, they are often composed entirely of beach material, and more often on a mixture of both. Barrier island which are formed by the deposition of rivers, and are laying in the longshore current zone, have an erosional pattern which varies from almost zero at the attached end of the island to a maximum erosion seaward, where the wave entrance is most oblique, and thus the longshore current, is maximum. When there is not enough sediment supply, these barrier spits rotate due to the eroding processes landward, and the sediment is deposited at the lee side at the end of the barrier, forming a backbarrier lagoon (Wai San Ting sandbar at the west coast of Taiwan, figure 44).

The coastal plain, near the river mouth, has morphological characteristics related strongly to the hydrodynamic regime which is present at the mouth of the river. Paragraph 3.4 figure 26 showed the various forms of ebb tidal deltas on which barrier islands can form. They depend on the relative magnitudes of the forces of dynamic diversion of the hydraulic conditions. When barrier islands or spits form on these ebb tidal deltas, there shape depends on the major current direction and the supply of sand. When the tidal current is almost equal in ebb and flood direction, the river sediment is spread out like a fan over the sea bed, with a characteristic "umbrella shape".

Not only alluvial material from the river serves as building material for the delta, also sediment transported alongshore can be a large supply accumulating on the delta. Even if there is a strong coastal flow of material, not all of it will pass a river mouth. The river water will slow down the flow of material; waves will break somewhat farther from the shore, and their effect on the coast and on the bar will be weaker than that on adjacent sectors. If there is an unidirectional movement of material, some of it may be held in front of the river mouth and give rise to accretion in the form of a triangular foreland. It is also possible that river waters force back the beach material that moves towards the river mouth and carry it to the sea floor. Waves cannot return this material and pile it up, since they are weakened when they encounter the river current. Wave continue to displace material along the bottom, and a submarine spit is built up in the zone where the effect of the waves on the bottom is strongest. A bay-mouth bar is semicircular because it forms along the edge of the fan where the river water flows out, the tidal current is active, and the waves break (see also figure 44; the deltaic coast of Yunlin).

Figure 45 shows a depositional model of a transgressive barrier shoreline (the Caminada-Moreau barrier headland coast of South Louisiana) in the Mississippi River delta plain (fig. 46). Rapid coastline retreat, due to subsidence, sediment deficiency, and overwash activity, is characteristic of the entire coastline. The barrier consists of three main forms; an erosional headland and two flanking barriers with recurved spits.
5. SEDIMENT TRANSPORT PROCESSES AND BARRIER ISLAND MIGRATION: A QUANTITATIVE AND QUALITATIVE ANALYSIS.

5.1 General

Barrier islands and barrier spits are deformed by environmental processes (wind, waves, current). These environmental processes are the driving forces for the morphological processes (sediment transport by water) and aeolian processes (sediment transport by wind). The morphological and aeolian processes on their term occasion the geodynamic processes (barrier migration, deformation) of barrier islands. In this context, knowing the sediment transport quantities due to the different mechanisms, is essential.

In the former chapters the main element of barrier islands and spits and the main environmental conditions are examined, even as the processes of barrier island evolution. In this chapter the sediment transport processes are discussed which cause barrier migration and barrier erosion. An attempt is made to quantify the contribution of the different "morphodynamic" mechanisms, discussed in chapter 4.

5.2 Quantification and qualification of sediment transport rates due to environmental processes acting on barrier islands.

5.2.1 Qualitative analysis

Processes by which barrier island are deformed vary from place to place. As discussed in chapter 2 barrier islands form in areas where the tidal range does not exceed the +4m M.S.L. Also the sediment supply is of great importance, for it is the building material, or with a negative sand balance, forms the basis for the decay of coastal elements. The sediment sources are related to each other and a qualitative flow diagram of the coastal sediment budget can be produced to illustrate the contextual position of the sand involved in the beach/dune changes (Fig. 47).

Especially for washover fans, there are a large number of process variables likely to exert influence on sediment budgets. Figure 48 summarizes the trends, resulting from multivariate analysis of variables, likely to affect sediment budgets and the data acquired from twelve surveys (Kochel and Wampfler, 1989). The different factors affecting the sediment budgets on washover fans (fan surface dynamics) are:

- Overwash frequency
- Overwash magnitude
- Rainfall frequency
- High wind frequency
- Relief near the fan
- Depth to the water table.

The scale tips towards accretion or erosion depending on symmetry of factors toward their end values.
5.2.2 Quantitative analysis

To produce a quantitative analysis of sediment transport rates of different barrier islands around the world, a comparison can be made between the sediment transport rates/retreat rates of these barrier islands due to the different environmental processes. However, making a proper comparison is not easy. The several studies of barrier island use different units and methods when determine barrier island migration.

An approach of quantifying sediment transport mechanisms will be presented by the discussion of a few case studies of barrier island migration along four different coasts around the world.

Table's 5.1 and 5.2 shows five different mechanisms which cause a lost or an accumulation area for dunes on Old North Beach, at the east coast of the United States. The ocean has the largest effect on dune losses, but it is also the main element that restores the dune damage (sand balance). Hardly no area is lost due to the effects of the ocean (waves, currents, tides). The second largest mechanisms causing sediment transport is washover transport. At a washover event a lot of material from the dunes is transported to the backside of the material, causing the loss of 21% of the dune area. The 5% gained area from washover can be explained by the sand transported with the backflowing water to the sea after the sealevel lowered. The net result of lost or gained area for dunes concerning washovers is a loss of dune area of 16% (in this situation), which is the largest net sediment loss of the five environments and showing that the contribution of washover events to dune losses (and thereby barrier migration) is significant.

For salt marshes a different situation occurs. Here washover is the main element conserving the amount of area for salt marshes. New dunes (and washovers) developed landward of the "1868-features", primarily on washovers and on salt marshes that had been buried by overwash. Twelve percent of the salt marsh loss between 1868 and 1978 was attributed to bayside erosion. Tidal currents behind Old North Beach do not generally move large volumes of sediment or erode cohesive salt marsh peat. Most of the salt marshes developed on old washovers (51%) and by rhizome extension into shallow Peasant Bay (30%).

<table>
<thead>
<tr>
<th>Lost to or gained from:</th>
<th>% lost area</th>
<th>% gained area</th>
</tr>
</thead>
<tbody>
<tr>
<td>ocean</td>
<td>63</td>
<td>61</td>
</tr>
<tr>
<td>washover</td>
<td>21</td>
<td>5</td>
</tr>
<tr>
<td>salt marsh</td>
<td>10</td>
<td>35</td>
</tr>
<tr>
<td>bay side</td>
<td>10</td>
<td>0</td>
</tr>
<tr>
<td>terminus</td>
<td>1</td>
<td>0</td>
</tr>
</tbody>
</table>

Table 5.1: Location of areas lost or gained for dunes on Old North Beach, U.S. (from 1868-1978, after Leatherman and Zaremba, 1986)
An other way of showing the influence of sediment transport processes effecting the migration barrier islands is to compare the landward sediment transfer due to tidal inlets and overwash (table 5.3). Three islands at the south east coast of the United States were studied. In this mesotidal area tidal inlets are numerous and cause the largest amount of sediment transport to the backside of the barriers. Overwash events are less numerous, but they also carry a lot of sediment to the backbarrier lagoon, causing the barrier to migrate.

<table>
<thead>
<tr>
<th>Lost to or gained from:</th>
<th>% lost area</th>
<th>% gained area</th>
</tr>
</thead>
<tbody>
<tr>
<td>washover</td>
<td>43</td>
<td>51</td>
</tr>
<tr>
<td>bay side</td>
<td>12</td>
<td>30</td>
</tr>
<tr>
<td>dune</td>
<td>46</td>
<td>18</td>
</tr>
<tr>
<td>terminus</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

Table 5.3: Landward sediment transfer in different environments

<table>
<thead>
<tr>
<th>Barrier system</th>
<th>Tidal Inlets</th>
<th>Overwash</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cap Hatteras</td>
<td>72.4</td>
<td>14.2</td>
</tr>
<tr>
<td>Assateaque Island</td>
<td>82</td>
<td>12</td>
</tr>
<tr>
<td>Malpeque barr. syst.</td>
<td>52</td>
<td>24</td>
</tr>
</tbody>
</table>

In table 5.4 the results of a large barrier migration study at the west coast of Portugal (Algarve) are presented (mesotidal shoreline; tidal range 2-4m). This table shows the various dominant processes on an island by island basis, and separates front-side and lagoon-side processes. The lateral variation in importance of island processes is assumed to be due primarily to variations in island orientation relative to both wind and wave energy. Secondary factors include shelf morphology, island shape/elevation control on the vulnerability to erosion by spring tides, sediment grain size, proximity to inlets, and the impact of man due to channel dredging and jetty constructions. It appears that the vulnerability of barrier islands to washover events is direct related to the presence and length of a single dune ridge at the sea side of the barrier. No dune ridge means 100% vulnerable to washover at the frontside. At the lagoonside of the barrier islands three processes are considered: Overwash, incorporation of the recurved spit with the lagoonside, and tidal delta incorporation with the lagoonside. When no overwash occurs and no tidal inlets are present, all the sediment will be transported along the barrier, causing spit accretion at the end of the barrier. However when tidal inlets and washover events transfer sediment directly to the back side of the barrier, almost no sediment is left for the recurved spits to accrete.
Table 5.4: Comparison dominant processes affecting frontside and lagoonside barrier island

<table>
<thead>
<tr>
<th>Island</th>
<th>Total Length (m)</th>
<th>Frontside Processes</th>
<th>Lagoonside processes</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Overwash %</td>
<td>Single Dune Ridge %</td>
</tr>
<tr>
<td>Ancao</td>
<td>10,000</td>
<td>6</td>
<td>94</td>
</tr>
<tr>
<td>Barreta</td>
<td>6,850</td>
<td>48</td>
<td>52</td>
</tr>
<tr>
<td>Culatra</td>
<td>5,750</td>
<td>65</td>
<td>35</td>
</tr>
<tr>
<td>Armona</td>
<td>8,750</td>
<td>40</td>
<td>60</td>
</tr>
<tr>
<td>Tavira</td>
<td>10,250</td>
<td>-</td>
<td>100</td>
</tr>
<tr>
<td>Cabanas</td>
<td>4,500</td>
<td>100</td>
<td>-</td>
</tr>
<tr>
<td>Cacela</td>
<td>4,400</td>
<td>100</td>
<td>-</td>
</tr>
</tbody>
</table>

Table 5.5 shows the washover volumes per unit beach length all due to one same storm, also at the south east coast of the United States. The numbers show the variety of impact due to washovers.

Table 5.5: Washover volumes/unit beach length due to 1 storm, East Coast U.S. (after Leatherman, 1976)

<table>
<thead>
<tr>
<th>Barrier system</th>
<th>vol./meter beach (m³/m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Caffey Inlet, N.C.</td>
<td>2.7</td>
</tr>
<tr>
<td>Buxton, N.C.</td>
<td>38.2</td>
</tr>
<tr>
<td>Cape Cod, Mass. to Ludlam Isl., N.Y.</td>
<td>10.8</td>
</tr>
<tr>
<td>Jones Beach, N.Y.</td>
<td>24.9</td>
</tr>
<tr>
<td>Sandy Hook to Barnegat Light, N.J.</td>
<td>101.7</td>
</tr>
</tbody>
</table>
5.3 Barrier migration

5.3.1 Introduction

Barrier islands and barrier spits migrate towards the coastline due to several coastal processes, which were described in chapter 4. Some of them (overwash, sea level rise) play a more important role in the migration process than others (aeolian transport, on- and offshore transport by waves). In this paragraph the most important processes causing barrier migration are discussed.

5.3.2 Barrier migration: A spit, a detached spit and an offshore barrier island

In this paragraph the differences in migration rates between Barrier islands, spits and detached spits are examined, in terms of sediment supply, morphology and physical processes responsible for the landward migration of the coastal landforms. Héquette and Ruz (1990) studied the migration rates of different spits and barrier islands along the coast of the southeastern Canadian Beaufort Sea. Comparison of air photographs showed the following migration rates of three coastal features, between 1950 and 1985:

<table>
<thead>
<tr>
<th>Coastal landforms</th>
<th>Retreat rates (m/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Offshore barrier islands</td>
<td>3.1</td>
</tr>
<tr>
<td>Spits</td>
<td>1.7</td>
</tr>
<tr>
<td>Detached spits</td>
<td>2.0</td>
</tr>
</tbody>
</table>

Regression analyses of retreat rates as a function of longshore sediment supply revealed that sediment availability is a significant parameter controlling spit and barrier island landward migration. Analyses of retreat rates as a function of incoming deep-water wave power showed that barrier islands are retreating more rapidly than spits even if they are exposed to the same level of wave energy. Direct sediment supply from adjacent bluffs and the presence of coastal dunes effectively limit spit erosion and retreat. In the case of barrier islands, the lack of significant longshore sediment supply is favourable to a net sediment deficit and results in higher retreat rates. Spits are retreating in response to beachface erosion and storm overwashing where low crestal elevation permits swash incursion across the crest. Barrier islands are more often and more extensively overwashed and overwashing represents the main physical process responsible for their migration.

5.3.3 Barrier migration & sea level rise

Sea level rise per se is not a mechanism directly influencing barrier migration. Clearly, sea level is the passive plane upon which mechanisms occur and in turn generate the dynamic processes of barrier alteration.
Orford, Carter and Forbes (1989) studied the relationship between gravel barrier migration and sea level rise. They pointed out that the process through which gravel barriers migrate is rather different to the way sand barriers migrate. The dominant process by which gravel barrier migrate is through rollover. Beachface sediment is passed over the barrier crest by storm-wave activity. It then remains passively on the backslope before burial by subsequent storm-generated washover sediment. In this way the barrier shift landward, so that in time the backbarrier sediments will emerge through the beach face to be incorporated once more in the overpassing cycle. In general, the capacity for sediment return to the seaward face of a gravel system is radically below that of a sand system, where tidally maintained breaches in the barrier and washover fans are common and effective in cross-barrier sediment recycling. (Carter et al., 1989b).

The rate of barrier migration is dependent on that part of the spectrum of run-up volume that crosses the barrier crest (Fig. 49). The differential response of migrating seaward barrier and back-barrier shorelines is a measure of barrier volume change (per unit length of barrier) related to barrier elevation changes. If the barrier’s seaward shoreline retreats at a faster rate than the back-barrier shoreline then the barrier must be building up with crestal elevations rising through concentrated overtopping (Fig. 50). The reverse situation with the back-barrier edge migrating faster than the seaward face indicates that the barrier crest must be falling with overwash predominating. When migration rates are equal then crestal elevation is held despite migration.

The balance between barrier crest built-up due to wave overtopping and crest break down by wave overwashing dictates the rate of rollover and migration. The balance of overtopping to overwashing may be altered by the rate of sea level change assuming that the storm intensity remains constant. Any increase in mean sea level will change the overtopping/overwashing ratio toward overwashing. This ratio can only be held constant under rising sea level by the provision of sediment to build the crest up. It is likely with high rates of sea level rise that overwashing will predominate, causing a reduction in the barrier crest elevation and an increase in the back-barrier migration rate.

Inman and Dolan (1989) tried to describe the shoreline retreat due to two general beach changes can be considered:

1. Seasonal beach changes (by meteorological conditions)
   - Summer and winter profile (Fig. 51)
2. Long-term shoreline changes (sea level rise)
   - Shoreline changes due to local accretion/erosion.
   - Shoreline changes due to sea level rise.

For the case of constant sea level it is assumed that accretion or erosion cause the beach profile to be displaced a horizontal distance $\Delta x$ (Fig. 52a). Responsible for the local accretion/erosion is the change in the longshore current, driven by waves arriving the coast at an oblique angle, and caused by local features that disturb the flow field.

For rising sea level it is assumed that the profile is raised vertically a distance $\Delta z$, then moved horizontally a distance $\Delta x$ such that the volume eroded from the upper beach is equal to the volume deposited further offshore along the shore rise (Fig. 52b). The shoreline recession obtained by equating the onshore and offshore volumes during sea level rise is known as the Bruun rule (Appendix 1).
5.3.4 Barrier migration & tidal inlets

Landward displacement of a barrier ecosystem seems to depend largely upon inlet dynamics. Temporary and migrating permanent inlets provide the bases -that is, the large flood tidal deltas upon which the island environments are established. These substrates serve as platforms for marsh development and hence landward extension of the bay shoreline. Subsequently, wind-blown and overwash sediments can be deposited on top of this accretionary base. These latter two forms of sediment movement are responsible for the vertical growth of a barrier island.

In areas where the tidal range is great (2-4 m, mesotidal), tidal inlets and associated construction of flood deltas may account for most of the landward sediment transfers (Armon and McCann, 1979) compared to overwash, wind and littoral drift.

5.3.5 Barrier Migration & washover events

In microtidal areas (tidal range 0-2 m) overwash is the main process responsible for landward migration of barrier islands. Overwash has contributed to the migrational process, largely by making available fresh sand for dune development (vertical growth) and to a lesser degree by adding new sediment to the backshore. (lateral expansion). It is in this fashion that high areas (dunes) become low areas (washovers) in a cyclic fashion. Therefore, the occurrence of overwash has the effect of displacing the barrier landward, and different sections of the barrier represent various stages in this cyclic phenomenon.

In appendix 3 the overwash magnitudes will be estimated of a wave-dominated barrier spit (The Trabucador Bar, north eastern Spanish coast)

5.3.6 Barrier Migration & longshore sediment supply

Héquette and Ruz (1991) described in their study of spit and barrier migration in the southeastern Canadian Beaufort Sea the role of longshore sediment supply. In their study regression analysis showed a linear relationship between retreat rates and potential longshore sediment supply. If the retreating bluffs in the updrift direction of a spit provide a large sediment supply, coastwise propagation of the spit will occur with little erosion on the beachface. However, if the amount of sediment entering the system is decreasing due to a reduction of available sediments at the source (e.g. lower bluff height or bluff sediment too fine to remain in the littoral system), wave-generated erosion of the seaward side of the spit will occur. In that case, most of the sediments are transported alongshore leading to accretion at the distal part of the spit while the spit is progressively thinning as a result of beachface erosion due to the reduction of sediment input into the system. Foreshore and backshore erosion results in the landward migration of the spit crest and of the seaward edge of the fordune (if any) and thus in the landward displacement of the shoreline (Fig. 53).
6. THE WAI SAN TING SANDBAR

6.1 Introduction

In the former chapter elements of barrier islands and mechanisms by which barrier islands are affected were examined and will be applied in this chapter for a specific feature along the west coast of Taiwan: the Wai San Ting sandbar. The Wai San Ting sandbar is located along the mid-west coast of Taiwan (Fig. 54), between the Choshui River and Kuosheng Kang (Fig. 44 and 55), and is formed on an old relict river delta. The Wai San Ting is actually a barrier spit, detached from the mainland by a large tidal inlet. This transgressive barrier is moving (rotating) towards the coastline at a speed of 40 m/year at the seaward end of the barrier, measured from several satellite charts. This feature is chosen for examination for its function as a sea defence, protecting the coastline in the lee side of the sandbar. Estimations show that, about one century, when the Wai San Ting reaches the coastline, the poor protected coast behind the bar will then be exposed directly to the wave and currents, which can cause severe flooding in the overcrowded area behind the coast.

In this chapter a description will be given of the study area in general and the Wai San Ting sandbar in particular. Paragraph 6.3 describes the morphological development of the Yunlin coast and a qualitative hypothesis is given to define the different features of the Yunlin coast.

In paragraph 6.4 the different mechanisms (discussed in chapter 4) causing the morphological and geological changes of the Wai San Ting sandbar will be determined and in 6.5 an estimation will be made of the contribution of the different mechanisms to the total sediment losses of the bar.

6.2 Morphological description of the study area.

6.2.1 Yunlin Coast

The study area is bordered by the Haifeng bar in the north and the Wai San Ting sandbar in the south (Fig. 44). The distance between these two features is 15 km. The centre and the northern half of the region (Yunlin coast) is characterized by a 3 to 4 km wide shallow intertidal sand area in front of the present dykes and sea defence. The Wai San Ting bar, together with some other barriers extend from the seaward edge of the intertidal area in a SSW direction, the largest of these being that extending from Tai Hsi which is referred to as the Haifeng bar. Between Haifeng and Wai San Ting the coast has the nature of an embayment which is sheltered from the predominant wave direction by the Haifeng sandbar.

The huge but intermittent sediment inputs from the Choshui River in the north and the migration and reshaping of the sandbars along the shoreline presents a complex morphological scenario which increases the difficulty of making predictions of impacts of reclamation on the shoreline evolution.

6.2.2 Wai San Ting sandbar

The Wai San Ting sandbar is partly emergent at high tide with the +2m TD level being reached only in a small area close to the seaward end. The seaward slope of the sandbar experiences full exposure to the monsoon waves, particularly near the southern end of the feature where the wave attack is most oblique. Double and often triple submerged breaker bars off the north facing coast are visible on the satellite photograph. Several channels connect the foreshore with the backshore area. Particularly in the northern part, small washover deltas can be seen at the backshore side of these channels which indicates sediment transport through the channels to the backshore. The leeside of the sand bar is a gently sloping and wide intertidal area, protected from the waves and currents (Fig. 56).

The southwestern tip of the Wai San Ting and other sandbars and submarine features show a sharp turn to the north, often following a steep bank. This can be interpreted as the effects of the north flowing flood tide reshaping the ends of these features and/or as the former seaward extent of relict river deltas.

6.3 Morphological development of the Yunlin coast

6.3.1 Recent development

The Yunlin coast has experienced severe interference by man even through it is largely "undeveloped" in the industrial sense. Dykes and retaining structures have been erected and the "backshore" has been fixed by these structures along the entire length of the county. These dykes were built at or above the high water mark and today the seabed level at the toe of the dykes is mostly between -1 an +2m above TD. The shallow intertidal areas in front of the dykes and between the dykes and the sandbars are heavily exploited for oyster cultivation in particular between -2 and 0m TD.

Also in the shallow intertidal areas a number of very narrow deep (5 to 10m deep) channels are maintained by the local fishing boat traffic, tidal flows and small scale dredging efforts. The largest of these channels is situated off Santiaolun in the lee of the Haifeng sandbar. It is about 8m deep at its deepest point and crosses the Haifeng sandbar opposite the fishing harbour of Potzuliao. South of Potzuliao, the existing coast is sheltered from wave attack and tidal currents both by the Haifeng sandbar and the Wai San Ting sandbar. Haifeng acts as a very effective barrier to the NNW wave attack.

Wai San Ting is undeveloped except for a small lighthouse near the southern end where the elevation ensures dry land at high water over a small area. Many small boats use the channels which cut across Wai San Ting close to its northern end.

*Taiwan Datum
6.3.2 Coastline mobility

A reliable record of large-scale sandbar movements has been gained from satellite imagery and aerial photography. Although there is inherent inaccuracy due to the water level differences at the various observation times the gross movement and features are clear. A summary of the locations of the Wai San Ting sandbar in different years is given in Figure 57.

The following are clear trends from these figures:

1. It can be seen that from 1932 to 1972 the movement was dramatic and a complete reshaping of the Wai San Ting sandbar has taken place from the convex shape in 1932 to the concave shape in 1972. The average rate of movement of the northward face of the sandbar has been about 100 to 150m southwards over this period. At this rate the sandbar's northern face is retreating at a rate of about 10km per century. The continuation of this trend at the present rate would result in the retreat of Wai San Ting back to the shoreline in about 100 years. However more recently the rate has slowed.

2. From 1972 to 1992 the reshaping has been less dramatic and the main movement has been an apparent rotation of the Wai San Ting as a result of more severe retreat at the seaward end compared to the landward end. The landward end is situated in the lee of the Haifeng sandbar and is sheltered from the predominant wave attack. Some accretion can even be seen at the landward end between the 1990 and 1992 shorelines. The retreat of the shoreline at the seaward end has slowed down and a retreat of the northern shore by about 40 to 50m/year is apparent from the coastal overlays. This indicates that the reshaping prior to 1972 had reduced the transport potential by alignment of the sandbar to predominant wave action.

3. The striking feature of Haifeng is the rate of movement of the southernmost tip. The tip has moved in a S to SSW'ly direction by about 4 to 6km over the past 20 years (1972 to 1992), a rate of about 200 to 300m per year. This rate would mean that in the absence of further interference by man the tip would reach the present location of the northern side of Wai-San-Ting in about 30 years time. This would act as a "natural" nourishment for the northern and landward parts of Wai-San-Ting. However, there is uncertainty about the continued rate of movement, it may be expected that as the feature approaches Wai San Ting it becomes more dispersed and "moves" slower, perhaps affecting a complete merging with Wai San Ting over a much longer time scale, say 30 to 100 years.

4. The total surface area of Wai-San-Ting that is outlined on various map overlays appears to be fairly consistent from year to year and this suggests that the sandbar as a whole may be moving (eg. by washover flows and breakthrough) or that there is continual build up of fine material due to weaker transport from south to north in the lee of Wai-San-Ting.
6.3.3 Qualitative hypothesis: Yunlin coastal morphology, Wai San Ting

Although there is little quantitative certainty about the historic episodic sediment supplies to the coast by the several historically active rivers in the Yunlin region, it cannot be doubted that the majority of the Holocene coastal sediments between say the -20 m TD contour and the present coastline have been deposited by these rivers over thousands of years. Until 1911, with the construction of the dikes at Linnei, there could have been five active rivers in the catchment area extending over the whole of the lowlands of Yunlin and Chung Hwa county, viz. Mai Yu Cho Hsi, Choshui Hsi, Old Hu Wei Hsi, New Hu Wei Hsi and Peikang Hsi. The latter four were probably active along the coast of Yunlin, but an educated guess was made about their quantitative roles. In order to do so, the following hypothesis was used.

Observation of the present depth contour lines (Fig. 44) indicates that the primarily tide dominated depths are contained in the depth regions below -20 m to -25 m TD. The primarily wave dominated depths are found in the regions shallower than 10 m depth (e.g., note the parallel character of the -10 m TD depth contour relative to the shoreline position). Between these depths there is a transition region of a mix between tide and wave induced phenomena. The -10 m TD depth contours more or less instantaneously follow the coastline changes, in contrast with the -20 m TD depth contours. Based on this, it is proposed to adopt the hypothesis that the -20 m TD depth contour still reflects the seaward boundary of the outer deltaic systems formed by the now sedimentologically inactive rivers.

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In doing this it must be concluded that the largest relict delta was formed by the Peikang Hsi, the second largest by the Old Hu Wei Hsi and a minor delta - if any - by the New Hu Wei Hsi. The depth contours near and just south of the Choshui Hsi mouth confirm that this is a still active delta, although with a fairly weakly pronounced deltaic outbuilding (possibly connected with the sediment sinks due to strong subsidence and sandmining).

It may be observed that the position of these relict delta baselines is strongly skew towards the south. This may be explained not only on the basis of the relatively large net southward directed longshore transports, but also from the - to a certain degree - instability effects caused by the strongly oblique angle of dominant wave incidence to the coast (see also fig 45). The dominance of the incident waves from the north and north-northeast most probably implies that the longshore sediment transport is driven by waves which approach the average coastal orientation under angles of more than 45 degrees relative to the normal direction. In the case of a triangular outbuilding of the outer deltas (connected with the diverging river outflow, creating eddies on both sides, which are influenced by the interaction with the tides, see also figure 26), it is probable that the 45 degrees angle is surpassed by the coastline near the rivermouth. This leads to an instability in the shoreline evolution on the northern lob of the deltas. However, this instability will not grow unlimitedly, firstly since there exists a limit to the growth where the coastline meets the -20 m TD depth contour and secondly since it is expected that initially wave-asymmetry induced onshore transport and finally overwash mechanisms act as additional sinks to the system (see Stive, 1986, describing a similar effect in the Voordelta, The Netherlands). The combination of these mechanisms will promote a southward oriented skewness of the deltas and a net onshore transport and regression.
In summary, the transition of these relict subaqueous deltas from a primarily river dominated to a purely wave (from northern incidence) dominated system has led to the formation of barrier beaches on the northern fringes with larger oblique angles to the coast than the deltaic base, through a combination of longshore transport mechanisms, losses at the -20 m TD depth contour and overwash mechanisms, fed by onshore wave-induced transport mechanisms. This implies that the barrier beaches (sandbars) of Wai-San-Ting Chou and Haifeng Chou are attributed to the relict deltas of Peikang Hsi and Old Hu Wei Hsi.

Based on the above concepts of the large scale historical evolution of the Yunlin coast and in agreement with the geostatistical analysis of the longshore correlation lengths of sediment size and shoreline changes, it is concluded that the present coastal system near Peikang Hsi is a moderately interacting relict delta, characterized by the Wai-San-Ting barrier beach and intertidal bar system (Features F of figure 44).

The Peikang Shi relict deltaic system has to a varying degree degenerated to a shoreline feature because of the natural tendency of coastal systems to smooth out disturbances. Obviously, the largest of the three, the Peikang Hsi delta with Wai-San-Ting Chou, is still reflecting most clearly the initial transformation features as described with the historical conceptual model. The transitional state between these two systems is reflected by the relict Old Hu Wei Hsi delta.

An important question concerns the degree of interaction between the Haifang sandbar and the Wai San Ting sandbar. Connected with the instability mechanism there exists a relatively large southward directed, net alongshore sediment transport due to the dominance of northern wave directions. As long as this transport is not interrupted there may not exist large alongshore gradients, but interruptions will cause large gradients and thus large effects. In this respect, it can be remarked that the existing coast contains such interruptions and these may be the cause for the rapid erosion of local areas relative to others along the coast. The present interruptions are mainly natural such as the Haifeng tidal channel, Haifeng tip, Peikang tidal channel, Wai San Ting tip etc. The Haifeng system, due to its geometry, at present still effectively blocks the feeding of the Wai San Ting Chou along a surf-zone route.

This is a very important conclusion since it results in several points of "discontinuity" along the Yunlin coast where the transport and supply of sediment from north to south is completely disrupted. The result is that features such as the main portion of Wai San Ting are considered to be developing at present independently from the adjacent coastal supply situation.

An important phenomenon in the evolution of the "coastline" is caused by the relatively large subsidence which occurs in the Yunlin coastal region. The observations show that the maximum rates (up to 10cm/year) may even be close to or at the actual coastal fringe. Subsidence of the coastline effectively acts as a sink mechanism and can account for a considerable amount of volume losses over periods of years. These apparent losses will be slightly increased by the additional relative rise in sea level due to global warming which is predicted (Section 4.2) to be in the range of 3 to 10cm over the next decade.
6.4 Evolution of the Wai San Ting

6.4.1 Sediment transport mechanisms

The formation of the barrier beach and intertidal bar system Wai-San-Ting Chou is explained (paragraph 6.3.3) as the result of wave-induced longshore and cross-shore mechanisms acting on the shallow outer delta of the Peikang Hsi, after this river stopped supplying sediment to the delta. Initially, the longshore instability mechanism enhanced its oblique orientation to the shoreline, while the wave asymmetry driven cross-shore transports deepened and flattened its foreshore and thereby increased its height to intertidal levels. At that stage overwash and barrier break processes must have grown in importance, reinforcing the shallowness of the intertidal flats on the southern side of Wai-San-Ting Chou.

The mechanisms requiring examination when considering Wai-San-Ting in detail are the following:

- **Longshore transport:**
  Breaking waves transport sediment alongshore.

- **Offshore transport:**
  Storm and typhoon waves transport sediment on and offshore.

- **Effects of washover:**
  Waves transport sediment over top sandbar at high tide

- **Channels transport:**
  Tidal channels (natural and dredged) transport sediment to the back side.

- **Relative sea level rise:**
  Subsidence, sea level rise due to global warming and compaction of the deltas all serve to bring about net losses of material to the active coastal system.

- **Rip currents:**
  Transport material offshore in local accelerations, mainly due to the breaking out of longshore flows in the surfzone between bed features.

- **Wind transports:**
  Transports of sand by wind.

These mechanisms have been addressed in this Section and where possible, quantified. The Section ends (Section 6.4.9) with a quantified estimate of the present and future evolution of Wai San Ting. In Appendix 6 a sand balance of the Yunlin coast is been given.

6.4.2 Longshore transports

The results of the longshore transport computations, executed with UNIBEST_LT and MIKE21_ST, are presented on the next page. Along Wai San Ting three cross-sections (Fig. 58 & 59) were used to schematise the wave and current climate and the bathymetry, all of which change along the sandbar significantly.

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In summary the calculated longshore transports are:

- Section 7 0.2 million m³/year
- Section 8 0.8 million m³/year
- Section 9 1.2 million m³/year

These figures show an increase in transport rates moving north to south along the sandbar. The average gradient along Wai San Ting is about 0.1 million m³/km/year although the tidal channels and bathymetry effects can cause higher and lower transports within some areas.

The gradient towards the southwestern end of Wai San Ting is in response to the greater exposure to both waves and currents. The diffraction of waves around Haifeng and the shielding from strong currents at the landward part of Wai San Ting is the main mechanism for this gradient. It is expected that just north of Profile 7 the transport will be almost zero and still further north and east the transport will become negative, viz. north-going. To the southwest of profile 9, toward the tip of Wai San Ting the transport will increase further until the coast angle becomes so oblique that the transport drops.

6.4.3 Crossshore transports

Another transport mechanism is crossshore transport whereby effects such as undertow, wave asymmetry and wave grouping (wave surges) can cause the transport of material in a crossshore (onshore-offshore) direction. Typically, these effects are most severe during high storm waves and result in material being eroded from the upper part of the crossshore profile and dumped on the lower part. The subsequent action of tides and non-storm waves may then remove such material alongshore or to deeper water.

Computation with the simulation model Unibest-TC were made on profiles between 7 and 9 (Fig. 59). Some loss due to the typhoon sequence was found over the -8 m contour at the more southerly profiles, but this is only in the order of 50 m³/m. In the other sections the losses during the typhoon cycle as computed with Unibest-TC are negligible.

On the basis of the indicative computations of the cross-shore transport and inspection of data, it appears that the overall offshore losses are not significant. The most likely conclusion is that offshore losses do not contribute significantly to the erosion of Wai San Ting.

6.4.4 Washover

During high water large parts of Wai San Ting submerge. When the water level at the front (north) side of Wai San Ting is higher than the back side then the washover flows will be set up. In figure 60 the water levels on either side of the sandbar have been plotted for a winter spring tide and neap tide. It can be seen that, assuming the sandbar becomes flooded at elevations above +0.5 m TD the water level on the north side is always higher than that on the south side (the same is true for summer tidal conditions). This leads to strong sheet flow from the north to the south over the inundated parts and there is no return flow from south to north. The water level difference in the model is
due to a difference in phase and tidal amplitude on either side of the sandbar. In reality wind and wave set-up would cause an even higher water level on the north and more washover. The vector plot in figure 61 demonstrates the washover effect (in the model the topography of Wai San Ting was estimated as flat at an elevation of +0.4 m TD).

The flow over the crest will carry sediment from the seaward side and from the crest across to the southern side. In appendix 4 a rough estimate has been made to quantify the washover transport.

With a water depth of 0.4m on top of the bar the sediment transport is about 220 m³/yr, while a water depth of 0.5m at the crest of the bar causes a sediment transport of 220,000 m³/yr! These estimates indicate that the effect of washover can be significant, and the waves (and the corresponding wave setup) can increase this washover current velocities even more. Over a length of 10 km the estimates suggest that the contribution of washovers could be in the order of 1 to 2 million m³/yr. However, this is a very rough estimate and it should be considered as a first indication only that significant losses due to washovers under daily tidal conditions can occur. It should be noted that no extreme events have been considered. It can be expected that one extreme event with high surges and high waves can contribute significantly to the yearly washover.

6.4.5 The effect of tidal channels

Through Wai San Ting several channels can be observed. Though the channels are shallow, they can act as sediment sinks. The largest of these channels is about one third of the way from the land towards the seaward end. The landward directed flow in this channel can transport sediment from the seafacing slope of Wai San Ting to the landward side of Wai San Ting. The wave set-up at the seafacing (and no set-up at the landfacing side) of Wai San Ting might also contribute to this "sediment sink" during conditions with high waves. No detailed information is available for the crest and the channel. The depth of the channel is estimated as -2 m TD. Appendix 5 shows a rough estimate of the sediment transport capacity by channels (see also paragraph 4.4 for the calculation of cross sectional areas of inlets (tidal) channels.

The results indicate an order of magnitude for the transport capacity through a channel of 1000 to 1500 m³/yr. per m width of the channel. For channels of several hundreds of meters this could contribute to the loss with 0.1 to 1 million m³/yr per channel. This implies that the transport capacity of the largest channel through Wai-San-Ting (if the width is estimated at 500 m) could be roughly estimated at 0.5 million m³/yr. This implies that all the longshore transport coming from the north (which is also in the order of 0.5 million m³/yr at the location of this channel) can be absorbed completely through the channel. This would increase the transport gradient and erosion tendencies on the seaward side towards the middle of Wai San Ting.

Though the above is a rough estimate, it indicates that the channels can significantly contribute to the sediment loss.

It should be realized that the sink capacity of the channels is only determined by the tide. The difference in tidal level over the bar determines the flow velocity and the corresponding transport through the channel. This results in a sediment sink capacity of the channel during every high water. These channels are therefore basically tidal features and they are a sediment sink independent of whether there is
Washover Transport on Sandbars

a longshore transport capacity along the seafacing side of Wai-San-Ting or not. Also, because of this, losses due to the channels (and overwash) are suppleted not only by surfzone sediments but also by sediments from up to 10 m depth.

6.4.6 Subsidence

At the northern, landward end of Wai San Ting the recorded subsidence is 0.10 m/year. Also the rate of sea level rise due to global warming can add up to 1cm per year in the worst scenario. These effects cause a relative sea level rise which acts as a "sink" in terms of a net loss of volume from the active profile. It is not known how the subsidence rate changes along Wai San Ting but it is expected to decrease significantly in seaward direction (see also paragraph 4.2).

For a conservative estimate, if it is assumed that the 0 m TD contour subsides by 0.10 m/year and that the subsidence decays with depth to zero at the -10 m TD depth contour (an average distance of 2 km away) the alongshore volume loss is (2,000*1*0.1*0.5 = ) 100 m^3/m/year. This figure is about 10% of the longshore transport rate. Though this analysis is rough it serves to show that the "sink" due subsidence and sea level rise is rather small relative to other possible losses.

6.4.7 Remaining processes

The remaining process have not been quantified, but they are expected to have only a limited effect on the overall sediment balance. The effect of rip-currents is probably limited to the upper regions (above the -10 m TD level). They can locally redistribute the sediment but will probably not cause significant losses to the deeper regions. This effect does therefore not affect the sand balance of the upper 10 m. Transport by wind (aeolian transport) can result in some losses, but these losses are expected to be small. The percentage of time that certain areas are dry is small. The percentage of time that this coincides with strong winds is even smaller. In addition, the dry areas only have a limited surface area. Relative to the large transport capacities of the other mechanisms, the contribution of wind transports is believed to be not significant.

6.4.8 Wai San Ting sand balance measurements

Survey data:

The Institute of Harbour and Marine Research and Technology (IHMRT) conducted surveys of the southern part of the seafacing slope of Wai-San-Ting in October 1990, October 1991 and October 1992. The original data were transformed to a 50 by 50 m grid. This data set was used for further analysis. Two methods were applied, viz. the analysis of separate profiles and the analysis of surveyed areas. From this dataset 10 profiles with a spacing of 1 km along the coast were derived and the change of the profile volume was determined for the periods 1990-1991 and 1991-1992.

The plotted profiles are presented in figure 58. The locations of three of the ten analyzed profiles are indicated in figure 59. From these plots an average yearly landward shift of the profiles between 40 and 100 m/yr can be estimated. It can be observed that this shift takes
Washover Transport on Sandbars

place over a profile depth of more than 10 m.

The survey levels were based on a LLW datum which at Wai San Ting can be taken at about -1.0 m TD for the purposes of the following descriptions.

For the interpretation of the survey results it should be mentioned that an error may have been introduced by the tidal correction that has been applied. No specific details about the corrections are known, except that the corrections were based on measurements carried out in a shielded and shallow area near Putai. The modelling has shown that the water levels at Putai can vary at high and low water by about 0.2 to 0.3 m from those at the seaward side of Wai San Ting. The impact of such an error could be significant (ie average error 0.10 m over the tide can account for a discrepancy of 1 million m$^2$ over an area of 10 km by 1 km). The magnitude and the direction of the error in a specific point depends on the phase of the tide at the moment the measurement was taken.

Results of survey analysis:

The balance results based on the profile data are presented in Appendix 6. The results for each year both indicate a loss from the upper 10 m of ±7.5 million m$^3$/yr, from which ±50% is lost from the section between 0 and -5 m LLW (-1 and -6 m TD) and 50% from the section between -5 m and -10 m LLW (-6 and -11 m TD). The changes in individual profiles as presented in Figures 58 show some large variations between the two periods 1990-1991 and 1991-1992. Areas with higher and lower erosion rates can therefore not be indicated accurately. There seems to be a significant interaction between the upper and the lower parts of the profiles. The balance results based on the area analysis agreed well with those of the profiles analysis and acted as a good check that the calculations were correct. The analysis of the areas also suggests a deficit in of 7 to 7.5 million m$^3$/yr for the 10 km stretch between 0 and -10 m LLW (-1 to -11 m TD).

From satellite pictures and old coastlines it can be roughly estimated that the average retreat of Wai San Ting over the last decades can be estimated at 40 to 90 m/yr. For the last decades the retreat seems closest to the 50 m/yr.

From the present location of the -10 m TD contour it can be concluded that the -10 m TD contour has followed this retreat to a large extent although in some local areas there are deviations from this (eg gullies, breaker bars etc.). It can also be concluded that the -20 m TD contour has not responded at all. The closure depth will therefore be somewhere between the -10 m and -20 m TD depth contour. If the closure depth is chosen in between, say at the -15 m TD contour and a deformation of a parabolic profile with a retreat of 50 m/yr at the LW-line is considered, then a sediment loss results of 0.67*15*50 = 500 m$^3$/m/yr. For a 10 km stretch this results in a loss of 5 million m$^3$/yr. Though these coastline changes can only be estimated with a limited accuracy, it can be concluded that the measurements and an evaluation of the historical development indicate the same order of magnitude for the sediment deficit of the southern part of Wai-San-Ting. Figure 62 gives an overall view of transport processes acting on the sandbar.

6.4.9 Summary - Wai San Ting present and future development

In summary the Wai San Ting sandbar is considered to be a decaying
delta formation of the (old) Peikang River. The Wai San Ting shoreline is retreating over its central and southern portions at a rate of about 50m/year at present. The northern end is sheltered by Haifeng and is retreating much slower, with even some signs of accretion.

Losses have been estimated at about 5 to 7.5 million m$^3$/year over a 10km stretch near the southern end, based on measurements and analysis of historical charts and satellite images. Based on the quantifications of the several major mechanisms acting over the 10 km section, it is estimated that the longshore transport losses through the southern end, southward losses through the channels which cross the sandbar and washover losses during high water and storms are each responsible for about one third of the volume loss. Minor mechanisms are offshore transport (undertow and rip currents), subsidence and compaction, wave asymmetry induced onshore transport, wind transport etc.

The Haifeng barrier will merge with Wai San Ting within several decades and while this will act to feed the northern end of the sandbar it will mean that the central part of the sandbar will be more exposed to waves and currents and the risk of separation of the southern and northern parts will increase.
CONCLUSIONS
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LIST OF LITERATURE

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Two general beach changes can be considered:

1. Seasonal beach changes (by meteorological conditions)
2. Long-term shoreline changes (sea level rise).

**Seasonal changes:**

In paragraph 3.8, two beach profiles were examined: the winter and the summer profile. In his study of the Duck Pier, North Carolina, Birkemeier (1984) identified, in agreement with studies elsewhere, a seasonal offshore movement of material from the beach and inner bar during winter with a gradual return to the beach face during summer. He also found that the largest and most rapid changes in the profile occurred during storms. The inner bar (Fig. 51) moved offshore during the storm and onshore again following the storm. The outer bar responded more slowly and the summer profile required about six months to recover following a storm.

The seasonal changes in the shore zone are both 2 and 3-dimensional in response to cross-shore and longshore water motions. The beach responds to the interaction between incident and lower frequency waves by forming crescentic bars (Bowen and Inman, 1971) and cusps (Inman and Guza, 1982). Thus the inner and outer bars shown in figure 51 (winter) may represent parallel longshore bars, or crescentic bars. The transformation from longshore to crescentic bars and back to longshore bars are known to be associated with the occurrence of storms and the return to lower waves (e.g. Short, 1979; Wright & Short, 1983).

**Long-term shoreline changes:**

In addition to the seasonal beach changes, there are two types of longer term shoreline changes that are essential in determining the rate of shoreline recession. These are the shoreline changes due to local accretion/erosion and those associated with sea level rise. With sea level rise constant, local net accretion (or erosion) of the shoreline, \( \Delta x \) (Fig. 52a). While, following Bruun's (see next page) rule (Schwartz, 1967; Bruun, 1983), both a shoreline recession and a redistribution of sand in the shorezone are assumed to be associated with sea level rise, \( \Delta z \) (Fig. 52b).

For the case of constant sea level it is assumed that accretion or erosion cause the beach profile to be displaced a horizontal distance \( \Delta x \) (Fig. 52a). For rising sea level it is assumed that the profile is raised vertically a distance \( \Delta z \), then moved horizontally a distance \( \Delta x \) such that the volume eroded from the upper beach is equal to the volume deposited further offshore along the shore rise. The shoreline recession obtained by equating the onshore and offshore volumes during sea level rise is known as the Bruun rule.

The volume change is given by

\[ q_v = \Delta x Z \]

where \( Z \) is the vertical distance over which the erosion occurs. In this case \( Z = h + c \), where \( c \) is the height of the beach berm crest above MSL.
When sea level rises a vertical distance $\Delta z$ the beach profile is adjusted upward such that the area between the two profiles is given by the product $\Delta z X$, where $X$ is the horizontal distance over which the adjustment occurs, Bruun assumed that the equilibrium beach profile would remain constant, and that a redistribution of material would occur such that the amount eroded from the upper beach is equal to that deposited offshore. The resulting shoreline recession becomes

$$\Delta x = \Delta z X / Z = \Delta z / \tan \theta$$

where $\tan \theta = Z / X$ is the extended beach slope. This relation is known as Bruun's rule of material associated with this redistribution of the profile is given approximately as

$$q' = \Delta x Z / 2 = \Delta z X / 2$$

where a factor of about $\frac{1}{2}$ enters because the erosion occurs only over a portion of the total distances $X$ and $Z$ (Inman & Dolan, 1989, Fig. 52b).

Table I: Shoreline change between beach recession $\Delta x$, sealevel rise $\Delta z$, and sand volume $q$ for profiles at Duck, North Carolina.

<table>
<thead>
<tr>
<th>Profile</th>
<th>$X$ (km)</th>
<th>$h$ (m)</th>
<th>$Z^*$ (m)</th>
<th>$\Delta x$ (m)</th>
<th>$q'$ (m$^3$/m)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>a. Sealevel constant</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Summer</td>
<td>0.8</td>
<td>7.5</td>
<td>10</td>
<td>1.0</td>
<td>10</td>
</tr>
<tr>
<td>Winter</td>
<td>1.0</td>
<td>10</td>
<td>15</td>
<td>1.0</td>
<td>15</td>
</tr>
<tr>
<td>Extended</td>
<td>1.8</td>
<td>12</td>
<td>17</td>
<td>1.0</td>
<td>17</td>
</tr>
<tr>
<td>Extended</td>
<td>4.5</td>
<td>18</td>
<td>23</td>
<td>1.0</td>
<td>23</td>
</tr>
<tr>
<td><strong>b. Sealevel rise (0.004 m)$^*$</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Summer</td>
<td>0.8</td>
<td>7.5</td>
<td>10</td>
<td>0.3</td>
<td>1.6</td>
</tr>
<tr>
<td>Winter</td>
<td>1.0</td>
<td>10</td>
<td>15</td>
<td>0.3</td>
<td>2.2</td>
</tr>
<tr>
<td>Extended</td>
<td>1.8</td>
<td>12</td>
<td>17</td>
<td>0.4</td>
<td>3.6</td>
</tr>
<tr>
<td>Extended</td>
<td>4.5</td>
<td>18</td>
<td>23</td>
<td>0.8</td>
<td>9.0</td>
</tr>
</tbody>
</table>

$^*$ $Z = h + c$ where $c$ is height of berm crest above MSL.

$^*$ As volume-equivalent factor in a, $q'$ has dimensions of m$^3$/m$^2$.

$^*$ Sealevel rise of 0.004 m/yr = 40 cm/century, as at Portsmouth, Virginia.
APPENDIX 2: The erosion rate of barrier islands; Sánchez-Arcilla and Jiménez (1993)

The rate of erosion of an barrier island depends on several elements, such as the environmental conditions (wind, waves, currents), sediment supply and morphological elements (nearshore slope, grainsize, crest height, barrier width etc.). Morgan and Stone (1985) derived a "storm wave susceptibility quotient" for islands along the Florida coast, which gives the tendency of the barrier to eroded. Higher values of the index indicate a higher vulnerability. The index is given by the dimensional relationship

\[
ES = \frac{Z_c}{B_h^2 + \sqrt{\frac{(X_w + X_b)}{100}}}
\]

where
- \( ES \) = erosion susceptibility index
- \( B_h \) = barrier height
- \( X_w \) = barrier width
- \( Z_c \) = depth at top longshore bar in front of barrier
- \( X_b \) = distance between top longshore bar and the shoreline,

and all variables are given in metres (see figure 62 for definitions). Although all variables involved are of morphological character, \( Z_c \) and \( X_b \) are representative of wave energetics because the former limits the incident wave height, and the second determines the surfzone width in which energy is dissipated after breaking. Table II shows ES values along the Trabucador Bar (Spain, see figure 63 for profile locations) obtained from morphological characteristics surveyed before the storm (July 1990). Profile 11 near the breached section presents the highest ES value along the barrier, indicating the highest erosion potential.

<table>
<thead>
<tr>
<th>Profile</th>
<th>( B_h ) (m)</th>
<th>( X_w ) (m)</th>
<th>( Z_c ) (m)</th>
<th>( X_b ) (m)</th>
<th>ES (-)</th>
</tr>
</thead>
<tbody>
<tr>
<td>P09</td>
<td>1.11</td>
<td>120</td>
<td>1.97</td>
<td>285</td>
<td>0.61</td>
</tr>
<tr>
<td>P10</td>
<td>1.02</td>
<td>190</td>
<td>1.71</td>
<td>110</td>
<td>0.62</td>
</tr>
<tr>
<td>P11</td>
<td>1.05</td>
<td>120</td>
<td>2.35</td>
<td>210</td>
<td>0.81</td>
</tr>
<tr>
<td>P12</td>
<td>0.87</td>
<td>130</td>
<td>1.40</td>
<td>135</td>
<td>0.59</td>
</tr>
<tr>
<td>P13</td>
<td>1.08</td>
<td>190</td>
<td>1.47</td>
<td>70</td>
<td>0.53</td>
</tr>
</tbody>
</table>
Washover Transport on Sandbars


Under the combined action of water level rise and storm waves a barrier can be overwashed. As the barrier degraded so the intensity of overwashing increased. To obtain an initial estimate of the overwash magnitude, the storm run-up was calculated using the expression due to Mase (1989):

$$ R = H_s a \xi^b $$

where

- $R$ = run-up height (measured vertically from S.W.L)
- $H_s$ = deep-water significant wave height
- $a, b$ = empirical coefficients
- $\xi$ = Iribarren number, surf similarity parameter (Battjes, 1974), which is given by

$$ \xi = \frac{\tan \theta}{\sqrt{H_s / L_0}} $$

where

- $L_0$ = deep-water wave length
- $\tan \theta$ = beach slope.

(This slope was obtained from the profile surveyed right before the storm, and it was defined as the ratio between the trough (in the outer bar/trough system) plus berm heights and the horizontal distance between trough and berm. This slope definition correspond reasonably well with the actual profile slope shorewards of the trough).

The run-up height obtained for the storm wave conditions can be presented in two heights, $R_{\text{mean}}$ and $R_{\text{max}}$, the mean and the maximum run-up values respectively. Although the run-up expression was derived from laboratory experiments under normal wave action, the expected run-up will be quit similar to the estimated one, because the nearshore angle of wave incidence was less than $30^\circ$. From experiments of run-up in breakwaters, Owen (1980) obtained that whenever the wave angle is less than $30^\circ$ run-up heights are not influenced by the angle. In any case, results show that run-up during storm action was important. Taking into account that during the storm the water level was 0.40 m above its normal position (due to storm surge) and 0.70 m if the wave set-up is included, the top of the barrier happened to be located at about 0.60 m above the mean water level (without wave set-up) or 0.30 m if the wave set-up is considered. This height is well below the run-up height estimation, showing that during the first stages of the storm (before barrier breaching), an important cross-barrier flow existed. This flow was able to start eroding the back-barrier. The removed sediment was transported towards the bay, where it was deposited forming a semi-submerged bar landward of the breach.
One problem to be faced is how to quantify the importance of this process. Sediment losses due to longshore transport gradient and offshore transport were estimated in the previous sections around 11,000 m³, i.e. 15% of the total loss. On the other hand, approximately 60,000 m³ of sand was deposited on the shelf behind the barrier during the storm. Thus, under the described storm conditions, most of the sediment eroded from the exposed part of the barrier was transported towards the inner bay, showing that under these conditions overwash processes were dominant.
APPENDIX 4: Estimation of washover transport of the Wai San Ting sandbar.

A rough estimation of the contribution of washover transport can be made assuming the following:

- **Time during which washover occurs:**
  
  3 hours per 12 hours = 25% of a day.

- **Conditions in the area have been roughly schematized as follows:**
  
  50% of the year: tide, no waves
  50% of the year: tide and waves with a wave setup of 0.1 m.

  Based on inspection of the water level differences $\Delta z$ at both sides of the Wai San Ting:
  
  - 10% of the year: washover due to only tide: $\Delta z = 0.15$ m
  - 10% of the year: washover due to tide and waves: $\Delta z = 0.25$ m.

- **Velocities can be determined with a Chezy formulation:**
  
  $$V = C \cdot V(h^1)$$
  $$C = 18 \cdot \log(12 \cdot h/r) = 45 \text{ m}^3/\text{s}$$

**Table III: Velocities of overwash current**

<table>
<thead>
<tr>
<th>water level diff. $\Delta z$ (m)</th>
<th>slope of overwash depth $i$ (-)</th>
<th>overwash depth $h$ (m)</th>
<th>current $V$ (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>No waves 0.15</td>
<td>0.00008</td>
<td>0.4</td>
<td>0.25</td>
</tr>
<tr>
<td>Waves 0.25</td>
<td>0.00013</td>
<td>0.5</td>
<td>0.35</td>
</tr>
</tbody>
</table>

If waves on the crest are assumed as $H_s=0.25$ m, $T_p=6$ s (depth limited) then the following transports can be computed with the Bijker formula:

**Table IV: Sediment transport calculation due to overwash**

<table>
<thead>
<tr>
<th>$h$ (m)</th>
<th>$V$ (m/s)</th>
<th>$H_s$ (m)</th>
<th>$T_p$ (s)</th>
<th>$s$ (m$^3$/m/hr)</th>
<th>hrs/yr (m$^3$/m/yr)</th>
<th>$S$ (m$^3$/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.4</td>
<td>0.25</td>
<td>0.00</td>
<td>0</td>
<td>0.002</td>
<td>850</td>
<td>2,000</td>
</tr>
<tr>
<td>0.5</td>
<td>0.35</td>
<td>0.25</td>
<td>6</td>
<td>0.300</td>
<td>850</td>
<td>220,000</td>
</tr>
</tbody>
</table>
APPENDIX 5: Calculation of sediment transport capacity of a barrier channel

The approach for the estimate of the transport capacity of such channels is similar to the approach for washover. As a result the greater depth, velocities increase although the water level differences will cause flow in both directions. However the width of the flood (south to north) flow at low water stages will be much smaller than the width and extent of the flow during ebb and high water stages. Also at high water stages the water level differences are greatest.

The duration of 3 hours has therefore been taken as representative of the period during which significant transports from north to south can occur.

The velocities can be determined again as:

**Table V:** Velocities of flow through tidal channels

<table>
<thead>
<tr>
<th>Δz (m)</th>
<th>i</th>
<th>h (m)</th>
<th>v (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>no waves</td>
<td>0.15</td>
<td>0.00008</td>
<td>2.9</td>
</tr>
<tr>
<td>waves</td>
<td>0.25</td>
<td>0.00013</td>
<td>3.0</td>
</tr>
</tbody>
</table>

and the corresponding transports as:

**Table VI:** Sediment transport calculation by channel flow

<table>
<thead>
<tr>
<th>h (m)</th>
<th>V (m/s)</th>
<th>$H_s$ (m)</th>
<th>$T_r$ (s)</th>
<th>s (m$^3$/m/hr)</th>
<th>hrs/yr</th>
<th>S (m$^3$/m/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.9</td>
<td>0.7</td>
<td>0.00</td>
<td>0</td>
<td>0.2</td>
<td>850</td>
<td>170</td>
</tr>
<tr>
<td>3.0</td>
<td>0.9</td>
<td>0.25</td>
<td>6</td>
<td>1.3</td>
<td>850</td>
<td>1105</td>
</tr>
</tbody>
</table>

Since the situation with wave setup mainly determined the high transport capacity through the channel, the order of magnitude of the sediment transported in a seaward direction will be an order of magnitude lower.
APPENDIX 6: Yunlin Sand balance

Table VII: Summary of Yunlin coast sand balance

<table>
<thead>
<tr>
<th>Description</th>
<th>Longshore transport</th>
<th>Subsidence</th>
<th>River bed load input</th>
<th>Channels in sandbars</th>
<th>Washover of sandbars</th>
<th>Total Loss/Gain</th>
</tr>
</thead>
<tbody>
<tr>
<td>Haifeng south tip</td>
<td>-0.3 to -0.6</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>-0.1 to -0.2</td>
<td>-0.60 +/-0.20</td>
</tr>
<tr>
<td>Wai San Ting</td>
<td>-0.1 to -0.12</td>
<td>0</td>
<td>0</td>
<td>-0.1 to -0.12</td>
<td>-0.15 to -0.2</td>
<td>-0.40 +/-0.04</td>
</tr>
</tbody>
</table>

It is an oversimplification to represent Wai San Ting as a single cell as there are local gradients which can differ strongly from the averages presented here. However the overall retreat of 40m/year for the central and southern parts compares well with analysis and measurements.
FIGURES
Figure 1: Hoyt’s theory of barrier island formation by drowning of a mainland beach/dune ridge.
Figure 2: Barrier island formation by spit accretion and inlet breaching (from Hoyt).

Figure 3: Cape Cod has evolved through erosion of glacial sediments and accretion of coastal deposits (after Davis).
Figure 4: Development of an island by upbuilding of a submarine bar.
Figure 5: Bayhead Barrier.

Figure 6: Baymouth Barriers.
Figure 7: Tombolo.
Figure 8: Simple Spit.

Figure 9: Double Spit.

Figure 10: Recurved Spit.

Figure 11: Complex Spit.
Figure 12: Suggested relationship of sand supply and time to barrier growth and decay, when barrier is sufficiently large to resist washover of sand and submergence occurs at a constant rate.
Washover Transport on Sandbars

Figure 13: Microtidal Transgressive Barrier Island.

Figure 14: Microtidal Regressive Barrier Island.

Figure 15: Mesotidal Transgressive Barrier Island.

Figure 16: Mesotidal Regressive Barrier Island.
Figure 17: Cape.

Figure 18: Sea Island.

Figure 19: Chenier.
Figure 20: Schematic diagram of a barrier island system illustrating the relationship among six interactive sedimentary environments.
Figure 21: Backbarrier lagoon; variable water surface areas by flood and ebb tidal currents.

Figure 22: Backbarrier lagoon; water levels and tidal currents.
Figure 23: Development of ebb and flood tidal deltas.

Figure 24: Aerial views of the Ancão Inlet flood tidal delta (13 September 1984); Flood tidal delta components are labeled: (1) flood ramp, (2) flood channel, (3) ebb shield, and (4) ebb spit.
Figure 25: Fuzeta Inlet (B.F.) at low spring tide (27 September 1984). Incorporated are remnants of flood tidal deltas mark the former positions of the rapidly migrating inlet (Algarve Barrier Islands, Portugal).
Figure 26: Ebb-tidal deltas of tidal inlets. Arrows illustrate the relative forces of the onshore, longshore, and offshore currents.

A: The prevailing force of longshore and onshore currents is greater than the offshore force.
B: The prevailing force of the southward longshore current is greater than the three remaining component forces.
C: The prevailing force of the northward longshore current is greater than the three remaining component forces.
D: The prevailing forces of the inlet currents are greater than the forces of northward and southward longshore currents.
Figure 27: Overwash surges during storm conditions (above) and resulting washover fans (below).

Figure 28: Overwash breach in the high dune line of the Ancão Peninsula. By 1984 this 1960s pass was nearly sealed by reformation of the vegetated dune (28 August 1980).
Figure 29: Schematic illustration of internal structure of extensively overwashed spit at Atkinson Point (U.S.) showing beach cross-bedded sets and washover planar stratification.

Figure 30: The primary coastal landform types, showing the main types of sand accumulation.
Figure 31: Schematic diagram of a barrier island illustrating the relationship among major sedimentary environments and landforms.

Figure 32: Beach Terminology (from U.S.A. CERC).
Figure 33: Principal factors affecting barrier beach migration (after Leatherman, 1980).

Figure 34: A barrier island may (1) remain in place, (2) prograde seaward, or (3) retreat, depending on sediment supply, sea level, and sea energy conditions (after Swift, Hoyt).

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(a) 12,000 years ago when sea level was about 50 m lower than present;
(b) Landward migration and upward growth by waves as sea level rises, forming a lagoon behind the barrier;
(c) Continued migration to present (after Dolan & Lins, 1987).

Figure 35: Evolution of barrier islands by sea level rise and spit-extension.
Figure 36: With sea level rise a barrier must retreat up the gradually coastal plain over geologic time. Without migration the barrier can be drowned.

Figure 37: Generalized conceptual model for sand dune development along the Caminada-Moreau coast in Louisiana as it relates to the role of storms and the return period of hurricane impact.
Figure 38: Profile H example of rapid cycle of destruction (hurricane Bob) and reconstruction of high coastal dune ridge.

Figure 39: Development of extensive salt marshes on a flood tidal delta, following inlet closure.
Washover Transport on Sandbars

Figure 40: Sequential diagrams of inlet breaching, migration, closing, and development of large flood tidal delta.

A. Initiation of storm:
   ▶ Wave overwashing of barrier island at low places in dune lines.
B. Flow of superelevated water in the bay, by strong offshore winds:
   ▶ Creation of inlet at low, narrow points along the island.
C. Normal tidal currents through inlet throat:
   ▶ Creation of flood tidal delta in the bay.
D. Sediment accumulation in bay:
   ▶ Continuous growth of flood tidal delta;  
   ▶ Small ebb tidal delta due to disturbance of waves.
E. Net longshore current to the south:
   ▶ Southward migration of inlet;
   ▶ Increased size and growth flood tidal delta.
F. Water path trough inlet throat area becomes very long and tortuous:
   ▶ Inlet’s efficiency greatly reduced.
G. Inlet sediment transport by tidal current overpowered by longshore sediment transport:
   ▶ Inlet closes.
1. Cross-section of a low barrier island at any starting time.

2. A severe storm will push high tides over the beach berm and as the water flows back across the island it carries beach sand into grasslands.

3. A second, less severe storm carries a smaller quantity of sand onto the earlier deposit. Abnormal high tides have pushed a small quantity of sand over the edge of the berm. Grasses have recolonized the earlier layer by pushing up through the deposit. The front of the island has retreated a small amount.
4. A very severe storm completely inundates the island and carries a great deal of sand across which adds to the earlier layers onto the backbarrier, elevating the subtidal and tidal environment. Old dunes have been knocked down and the sand spread out. The beach front has retreated substantially. In some places, the berm front may be lowered as sand moves away from this source.

5. A few years later (less than 10 usually), and after a period of relative storm quiet, the island surface has completely recovered, but the vegetation zones have been displaced slightly. Low salt marsh has developed on the new base in the backbarrier lagoon, the former low marsh has become high marsh, and the barrier flat has extended into the old high marsh border (after Godfrey, 1969).

Figure 42: Spit formation along a coastline by oblique wave approach and resulting littoral drift.
Figure 43: Formation of the island of Sakalin in front of one of the mouths of the Danube (from J. Petrescu, 1957)
The south western tip of the Wai San Ting sandbar (on old delta Peikang River) and Heifeng sandbar (on delta from Choshui River) and other submarine formations (F) show a sharp turn to north. This reflects the influence of the strong tidal flow on the submerged remains of old deltas.
Washover Transport on Sandbars

Figure 45: Depositional model for a transgressive barrier shoreline development in the Mississippi River delta plain; erosional headland with flanking barrier islands (Penland and Boyd, 1981)

Figure 46: A historical map time series showing coastal changes associated with the Bayou Lafourche barrier shoreline (1887-1978)
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Figure 47: Generalized flow diagram for sand movements from barrier sources to dune and washover depositional landforms.
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Figure 48: Schematic of factors affecting sediment budgets on washover fans.
Figure 49: The continuum of overtopping and overwashing modes by which gravel barrier crest migration may occur; t1 and t2 are crestal profiles before and after storm generated overtopping and/or overwashing run-up.

R=run-up elevation; B=Barrier crest height; Qw=run-up volume
Washover Transport on Sandbars

Figure 50: Schematic view of barrier crestal stability domains as a function of seaward and back-barrier shoreline migration.

Figure 51: Winter and summer beach profiles (above) and envelope of profiles from January-August 1984 (below), Duck, North Carolina, U.S.A. (after USACE CRRC, 1984).
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Figure 52: Schematic diagrams of beach profiles showing the volume of beach sand associated with (a) net recession of the shoreline $\Delta x$, and (b) with sea level rise $\Delta z$. In both cases it is assumed that the beach profile remains constant.
Figure 53: Conceptual geomorphic model of spit and barrier island migration in the southeastern Canadian Beaufort Sea, showing idealized plane view and cross-section with major sediment transport paths.

(1) spit; (2) barrier island. Arrow dimensions are proportional to the volume of sediment transport. (Héquette and Hill, 1989).
Figure 54: Study area; Taiwan
Figure 55: Study area; Yunlin coast
Figure 56: Bathymetry of the Wai-San-Ting
Figure 57: Locations of the Wai San Ting sandbar in different years
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Figure 59: Location of profiles
Figure 60: Spring tide and neap tide (below): Surface elevations north and south side of WST
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Figure 62: Cross-sectional definitions for "erosion susceptibility index".

Figure 63: Profile locations Trabucador Bar, Spain.