Summary Coastal Dynamics

Chapter 2: Large-scale geographical variation of coasts

1. Introduction

2. Cumulative evolution of coastal systems

2.1. Geological timescale

The timescale is divided in mane periods and subdivided in many sub-periods. The sub-periods are becoming shorter due to our knowledge and records of the later periods. The Current Geological period Quaternary can be divided into the Pleistocene and the Holocene. During this period continental separation determined the geological initial state of the boundaries. Secondly, ice-sheets retracted, causing sea level to rise significantly.

2.2. Continental drift

Around 200 million years ago a single super continent existed (Pangaea). Land masses that formed Pangaea started drifting apart. The theory of the drifting continents has been fully developed by Wegener around 1912-1915. It was until the 1960s this was proven by geological evidence. The movement of plates has governed the formation of coastlines and determines the basic features of coasts.

2.3. Pleistocene inheritance

During the Pleistocene, earth experienced several periods of cooling and heating up. The vast amount of water trapped in ice caused severe water level fluctuations (up to 100m). The inheritance is greatest in the rocks formed by ice that have been deposited in coastal systems. The so called Pleistocene layers is the layer on which most building foundations in the Netherlands have been constructed.

3. Tectonic control of coasts

3.1. Plate tectonic theory

The continents are part of the lithosphere. This sphere is divided in 12 large and tightly fitting plates and several small ones. These plates are in permanent motion moving up to 10 cm per year. Where plates meet, either mountains or oceanic trenches are formed.

3.2. Tectonic plate setting of coasts

Broad coastal characteristics are related to the position of these coasts o the moving tectonic plates. We recognize 3 types:

- *Leading edge or collision coasts:* Coasts near converging plates. Usually rugged, cliffed coastlines in tectonically unstable regions.
- *Trailing edge coasts:* Coasts located away from plate boundaries in tectonically stable areas because the continent and ocean floor are the same plate. This category of coasts can be divided into Neo/Amero/Afro. The difference between these types lays in the state of the developed ocean and the sediment that is available in the coastal area.
- *Marginal sea coasts:* Tectonically stable coasts protected from the open ocean by island arcs at converging plate boundaries.

3.3. *First-order coastal sedimentary features* – Factors important to coastal sedimentary features are:

- *Continental shelf width*
  Wide and flat shelves facilitate the development of extensive sedimentary features whereas narrow and steep shelves facilitate the opposite. This causes large deltas and barrier islands systems to easier develop on coasts with wide shelves. Hence, leading-edge coasts are dominated by sea cliffs and rocky headlands whereas trailing-edge coasts are dominated by depositional features like barrier islands.

  The shelf width also has hydrodynamic effects. Narrow shelves lack the inability to dampen wave energy due to lack of a long gentile slope inducing bottom friction. This causes the waves to be more powerful near narrow shelf coasts. Wider shelves do have this frictional dampening mechanism and thus wave energy is lower. However, tidal amplitudes are generally higher at wider shelves because of resonance of the (by friction) slowed tidal waves.

  Marginal sea coasts also have the potential to develop large sedimentary features due to the presence of a gentle slope and shallow waters of the continental shelves in these areas, the reduction of wave energy due to the sheltering by nearby objects as island arcs or land masses and the restricted size of these seas limiting fetch. This all results in low wave energy and therefore little erosion and a relative high deposit of sediment.

- *Sediment availability*
  As said before, sediment supply determines the type of trailing edge coast. American trailing edge coasts receive large supplies of sediment due to the mountainous origin of the rivers. These trailing edge coasts are therefor called Amero-trailing edge coasts. In Africa, both sides of the continent have developed trailing edge coasts. Due to the lack of mountainous regions and a much less erosive climate, these trailing edge coasts typically lack sedimentary features as Deltaic areas along the coast. The coasts are therefore called Afro-trailing edge coasts.

  In Marginal sea coasts, sedimentary features can easily develop due to the low-energy coastal conditions and the large amount sediment. Some of the world largest delta's have developed in these regions (for e.a. the Mississippi Delta)

  Leading edge coasts are usually fed by short, steep and straight rivers. These may transport large amounts of sediment. However, large sedimentary features rarely develop due to a narrow and steep shelf. Most of this sediment is therefore lost to deep sea.

4. Pleistocene inheritance of cliffed coasts

Roughly 75% of continental and island margins is lined with cliffs. These cliffs are commonly related to tectonics. There are however cliffs, that where formed under different conditions. Glaciers from the Pleistocene are known to have created the fjords by moving ice carving out these steep valleys.

Cliffs can also be formed by the deposition of sediment by the ice. This has formed a thick layer called a till. The layer usually forms near the edge or beneath the ice sheet and contains a wide variety sediment sizes. Large parts of the British & Danish coast where formed in this manner.

A third variety of rocky cliffed coast is associated with areas where carbonate sediments from skeletal remains and coral debris accumulated. Lithification of the sediment caused the it to cement together.

5. (Holocene) transgression versus regression

5.1. Geological sea level changes

Local mean sea level → height with respect to land. Local changes to the perceived mean sea level can be either relative (for ex. Land subsidence) or absolute sea level rise. The absolute sea level rise can also be called eustatic and can be caused by:

- *Changes in volume of the ocean water*
- *Changes in the volume of the ocean basis due to tectonic plate divergence*
- *Changes in the distribution of water due to differences in earth's gravity field*

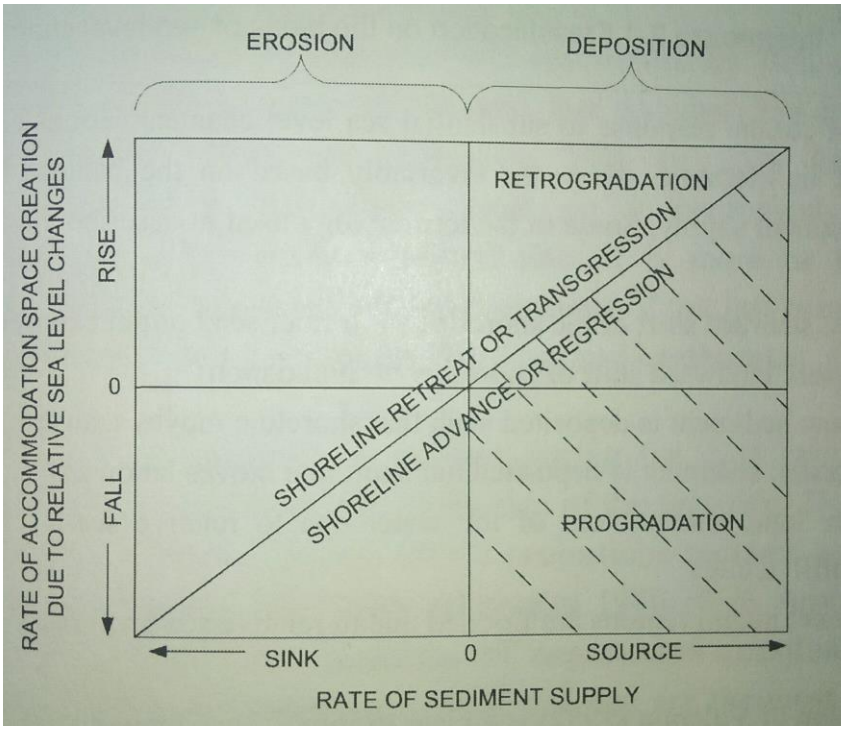
Of the eustatic sea level rise, volume change of the ocean water has been the most important in earth's history. The alternating glacial and interglacial periods caused the sea level to vary significantly (more than 100m). The eustatic sea level changes cause deformations in earth's crust. These deformations are caused by loading and unloading of ice during a Glacial period (Glacio-isostasy), and loading and unloading of water during an interglacial period (Hydro-isotasy)

In addition to eustatic & isostatic effects, regional effects are also of importance:
Regional subsidence: subsidence caused by compaction of sediments or withdrawal of resources.
Tectonic activity: movement related due to tectonic movement.

In the past century, global sea level rise has accelerated. Global warming has been blamed for this.

5.2. Role of sea-level rise in Holocene coastal evolution

When sea-level rises, the local equilibrium of the coast is distorted and the accommodation space enlarges. If there is no source of sediment the land will retreat (transgression). If there is a source of sediment the coast can advance (regression). During the Holocene coasts in the Mediterranean and the Caribbean have transgressed due to sea relative sea level rise while parts of Asia, Southern parts of Africa, South America and Oceania have seen regression of the coastal areas.



Figuur 1; Curray's diagram : shoreline migrational factors

5.3. More recent coastal development.

Around 70% of the world's sandy coasts has shown retreat in the last century; about 10% had propagation, while the remaining 20% has been stable. The sea level rise is seen as the cause of this. Sea level is expected to rise even further the coming century. Especially low lying countries with poor coastal defenses are threatened.

6. Nature and abundance of coastal material

6.1. Sources of sediments deposits

There are roughly 2 sources of sediments:

Continental sediments: sediments formed from weathered continental rock as Granite, basalt and feldspar.

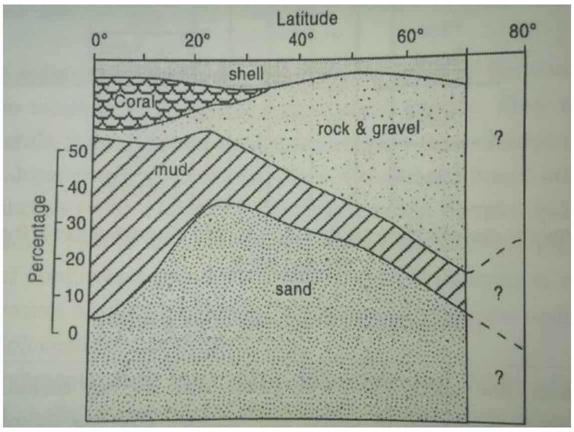
Carbonate sediments: fragments of shells or remains of marine life. These sediments consist mostly of calcium carbonate

6.2. Sediment sizes

The size of sediment can vary greatly; the Wentworth classification can be used to classify the sediments into clay, silt, fine sand, medium sand, coarse sand, coarse gravel and cobble.

6.3. Geographical variation

The material present on beaches and shore faces shows a latitudinal zonality. The average distribution can be seen in the following diagram:



Figuur 2; Latitudanal distribution of sediment types

Availability of the material to the coast: the availability of material to the coast is determined by: (1) the presence of rivers, (2) the type of weathering (mechanical for cold areas and chemical for warm areas), (3) Pleistocene glaciation and (4) water temperatures.

Coastal processes

These processes play an important role in determining the type of material found at a certain location with processes like: Erosion of rock, winnowing out of muddy material by wave and tidal processes, and enhanced deposition through flocculation.

6.4. Muddy coasts

These can be found at all latitides and continents, but mostly in tropical areas. They usually settle near quit areas an near river mouths.

6.5. Sandy coasts

Locations where you can find sandy coasts.

6.6. Vegetations

Mangroves and salt marches

Salt resistant vegetation requiring calm conditions and a salty substratum. The salt marches can be found in moderate climate zones. Mangroves favor tropical and subtropical climates.

A Marsh environment is similar to that of river and delta floodplains, with channels cutting through the marshy plain. Sediment enters the marsh by storm tides that push it in or by sluggish currents. Marsh grasses are very important sediment stabilizers and prevent currents and waves from removing sediment.

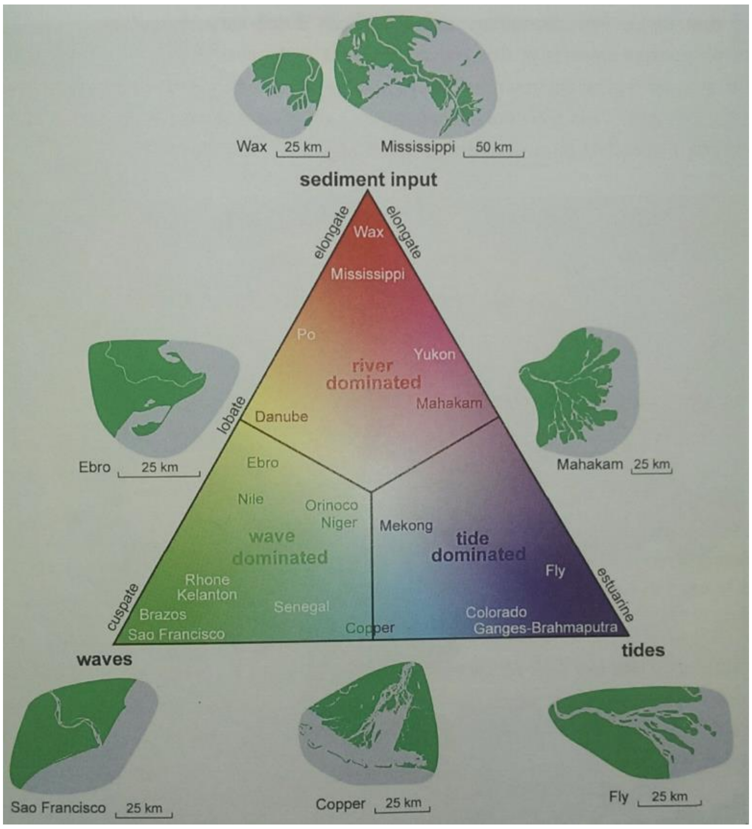
Mangroves are woody trees. The root systems they form are dense but also diverse. Mangroves accumulate and stabilize sediment and protect the coastline from erosion. They also from a habitat for a wide variety of organisms. Worldwide the mangrove vegetation is disappearing rapidly because of human influence. Reforestation is extremely difficult because the younger plants are very vulnerable.

<u>Dune vegetation.</u>
Non-marine plant life that lives on the dry beach. The plants have to be drought resistant. The vegetation prevents the wandering of dunes by Aeolian(wind-driven) transport.

7. Process-based classification

7.1. Dominance of fluvial, wave or tidal processes

Fluvial, wave or tidal processes can influence the shape of the coastal features. How they influence this depends strongly on local factors.



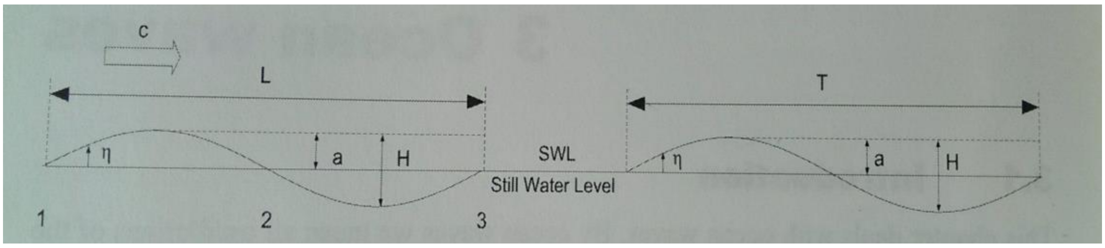
Summary Coastal Dynamics

Chapter 3: Ocean Waves

1. Introduction

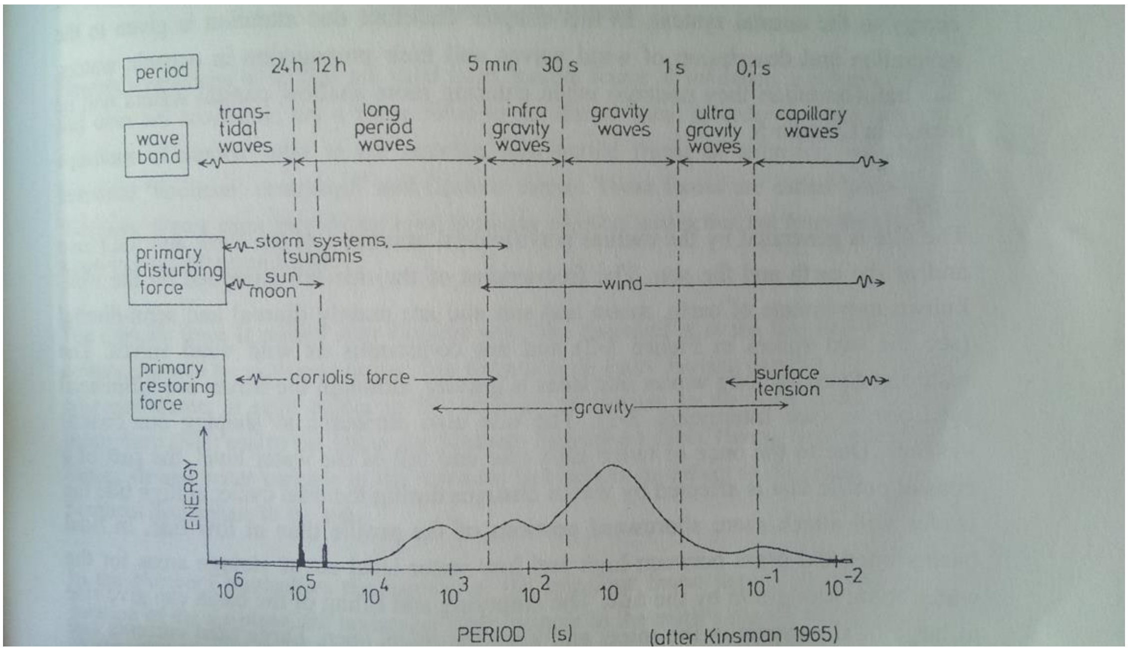
2. Oscillation of the Ocean water surface

MSL→ Sea level when all fluctuations average out.
Waves can be described by a sine or cosine.



With:
H→ Wave Height (height betweeen crest & though)
L → Wave length (Distance a wave will travel within wave period **T**)
T → Wave period (Period a wave requires to pass a location)
η → Surface elevation (asin(**W**t-**K**x)
K → Wave number (2π/**L**)
W→ Angular frequency (2π/**T**)

Waves can be Classified to generating force and period. The classification can be derived from the next picture:



Summary Coastal Dynamics

Wind Waves:
Wind generated waves have periods from 0.25 to 30 second. The waves are called gravity waves and, when generated by local wind fields, are usually short, random & irregular. Dispersion can cause the fields to transform in to regular swell due to frequency dispersion. The waves will have to travel large distances for this effect to become visible. The wind generated gravity waves are the primary supplier of energy to coastal systems

Tides:
Waves generated by the gravitational attraction of the sun and the moon. The restoring force of these waves is gravity; they are however also influenced by the Coriolis force. Tides contribute considerably to the coastal profile.

The Coriolis force can be considered a fictitious force, since it does not arise from physical interaction but from a non-inertial reference frame. In the Northern hemisphere, all motion is deflected to the right while on the southern hemisphere the deflection is to the left. At the equator the Coriolis force is equal to zero. Near the poles the force is strongest. The Coriolis acceleration can be described as:

$$a_c = 2w_e sin\varphi V$$

With :

$w_e = 72{,}9 * 10^6 \left[\frac{rad}{s}\right]$ (Angular velocity of earth)

V = current velocity
φ = latitude (positive NH and negative in SH)

The significance of the Coriolis deflection is given by the Rossby number:

$$R = \frac{V}{|2w_e sin\varphi| * L}$$

In which L is the length scale. If the rossby number is smaller than 1 Coriolis deflection is important.

Tsunamis:
Tsunamis are waves caused by a displacement of the sea level. Tsunamis are not visible in deep water but grow when they reach shallow water. A devastating Tsunami develops roughly once every 15 years.

Storm surges:
These elevations of the water surface are on a scale of the storm fields that generate them. The wavelength and period are slightly shorter than those of tides. Storm surges will pile up water against the coast when they approach it and can be very dangerous.

Summary Coastal Dynamics

3. Measuring Ocean Surface Elevations

There are in situ ways like buoys and poles but also remote-sensing techniques, like radar, available to measure the vertical elevation of the sea level. Most techniques have an accuracy of ±10% or better

4. Short-term wave statistics

4.1 Description of wave characteristics

a seemingly unpredictable wave field can be described in a statistical way by taking average parameters. The two basic ways to characterize a wave record are:

1) Based on direct analysis of the time series and regarding it as a sequence of individual waves each with their own wave height and wave period (see 4.2).

2)Through a spectral analysis using the fact that the surface ca be seen as a summation of an infinite number or sine waves with different heights, periods and directions (see 4.3).

4.2 Analysis of time series

From a time series one can derive a mean and a standard deviation (σ). The variance of the demeaned surface elevation is σ^2. The energy can be described by: $E = \rho g \sigma^2$ (half kinetic, half potential).

When dealing with a short time record one can derive other parameters like the significant wave height H_s or $H_{1/3}$. This is the average height of the highest one third of the waves. H_{rms} is also often used and obtained by taking the square root of the mean of the wave heights squared. The significant wave period can be obtained similarly to the significant wave height.

4.3 Spectral analysis

An alternative approach to the Time series analysis. Waves are unravelled into many harmonic components that can be written in terms of sine or cosine.

4.4 Short-term wave height distribution

If waves can be considered as the sum of a large number of components, they can also be described by a Rayleigh distribution. Observations have shown that for both narrow and broader spectra the wave heights obey this distribution.

5. Wind wave generation and dispersion

5.1 Locally generated sea

Wave characteristics and their duration depend on the characteristics of the wind field, fetch and depth. Wind conditions are irregular but can be described statistically. The JONSWAP spectrum is often used to estimate characteristics for (developing) wind sea in oceanic waters.

5.2 Wave dispersion

For a linear sine wave the relation between frequency ω and k is given by:

$\omega = \sqrt{gk \tanh kh}$ (dispersion relation)

The phase velocity is the rate at which any phase of the wave (for example the crest) propagates, is then given by:

$$c = \frac{gT}{2\pi} * \tanh kh$$

5.3 Wave groups

In deep water wave crest move faster than the wave energy. This is visible when a wave is within a group of waves. It will originate at the end of the group and move through it until it reaches the leading edge, where it disappears. In shallow waters this characteristic does not occur.

The ratio between the group velocity and the phase velocity is called n and can be depicted by:

$$n = \frac{c_g}{c} = 0{,}5(1 + \frac{2kh}{\sinh 2kh})$$

5.4 Sea versus swell waves

From section 5.2 we learned that the phase speed of waves differs due to a difference in their frequency. This means that on large distances away from the point of generation shorter waves will have been filtered out because dissipation more strongly affects shorter waves. The waves that eventually reach the coastal areas will be fairly regular & unidirectional. These waves are called swell.

6. Long term statistics and extreme values

There are several ways to represent long term data. There are both Histograms and scatter plots for the significant wave height Hs. Tables can include information on wave period and wave angle. Wave roses give the directional distribution of waves.

7. Generation of the tide

7.1 Equilibrium theory of the tide

This theory was developed by newton, assuming the earth to be completely covered by water and assuming water responds instantly to the tide generating forces.

7.2 Gravitational pull

Both the sun and the moon influence the tide on earth. This influence is proportional to the mass and distance of the object. The earth and moon revolve around a common center of mass causing centripetal forces who also influence the tide.

From calculation we can determine that the gravitational pull of the sun = $6{,}0*10^{-4}$ g while the gravitational pull of the moon = $3{,}4*10^{-6}$ g. This may however vary for different locations on earth due to variation in angle and distances to the attracting mass.

7.3 Differential pull or the tide-generating force

Although the gravitational pull of the sun is far the largest, it only contributes to roughly 31% of the tidal amplitude. The moon is responsible for the other 69%. The actual force causing the tidal amplitudes is the difference between the gravitational pull between the surface and the center of the earth for the attracting object. Because the moon is far closer to earth than the sun, the effect is much stronger

The rapid diurnal rotation of earth causes the earth to rotate underneath the tidal bulges and thus produces a semi-diurnal tide (2 highs, 2 lows).

7.4 Spring and neap tide

When the sun earth and moon are in one line the tides reinforce each other (spring tide) When the solar and lunar tides are 90^0 out of phase the effects cancel each other out (neap tide).

7.5 Daily inequality

The daily inequality between the tides can be explained by the tilted axis of the earth and the time-varying declination angle between the equatorial plane and the earth-sun and earth-moon connection lines.

7.6 Tidal constituents

There are numerous tidal components. The most important ones are the M2 (influence moon) and the S2 (influence sun). All the components can be seen in the following picture.

Tidal constituents	Name	Equilibrium Amplitude (m)	Period (hr)
Semidiurnal			
Principal lunar	M2	0.24	12.42
Principal solar	S2	0.11	12.00
Lunar elliptical	N2	0.046	12.66
Lunar-solar declinational	K2	0.031	11.97
Diurnal			
Lunar-solar declinational	K1	0.14	23.93
Principal lunar	O1	0.10	25.82
Principal solar	P1	0.047	24.07
Lunar elliptical	Q1	0.019	26.87
Long Period			
Fortnightly	Mf	0.042	327.9
Monthly	Mm	0.022	661.3
Semiannual	Ssa	0.019	4383

Figuur 3; Tidal Constituents

8. Propagation of the tide

8.1 Dynamic theory of tides

The earth's seas are too shallow for waves to move fast enough to allow the equilibrium tide to exist. Also the continents prevent the equilibrium tide from forming. Only near Antarctica are both conditions met.

When the tidal wave enters for example the much shallower North-Sea, celerity (speed) goes down but amplitude increases. Due to its decreased speed, tidal waves are slowed significantly and occur roughly two days after the corresponding moon configuration.

8.2 Amphidromic systems

Because the movement of the tides is deflected by Coriolis and blocked by land masses, rotary moments are formed in ocean basins, bays and seas. This means the tide will propagate around a certain point in the water mass. Due to the slow propagation speed in the North sea, two nodes have developed. One is between England and the Netherlands, the other one is between Norway and the Netherlands.

8.3 Kelvin waves

Is a wave that is coastally trapped. It rotates around an Amphidromic point while Coriolis is also affecting it. The wave travels parallel to the coast.

9. Tidal analysis and prediction

Because the tides are being caused by regular astronomical phenomena, they can predicted accurately a long time ahead. The method used for this is called harmonic analysis.

Summary Coastal Dynamics

Chapter 4: Global wave and tidal environments

1. Introduction

2. Zonal wind systems and ocean circulation

2.1 Solar radiation and temperature distribution

Winds and ocean currents develop due to an uneven distribution of heat over the earth's surface. The main source of thermal energy is electro-magnetic radiation from the sun. The absorption of radiation leads to heating. This may be transmitted by conduction or convection (fluids). A balance must exist between incoming solar radiation and outgoing terrestrial radiation. Incoming radiation is strongly dependent on the latitude. Uneven distribution forces a transfer of this heat. This generates winds (60% of heat transfer) and ocean currents (40%).

The amount of incoming solar radiation is determined by average distance between the sun and earth, transparency of the atmosphere and the angle at which the sun's rays strike earth. These factors vary with latitude and seasons.Oceans respond slower to changes in incoming radiation because radiation penetrates further in water than in land. Water has a greater specific heat capacity and a big storage possibility.

2.2 Atmospheric circulation and wind patterns

Near the equator air is warmed, allowing it to attain a maximum vertical altitude of about 14 km, when it starts flowing horizontally towards the poles. This creates a low pressure area known as the Intertropical Convergence zone. Due to rotation of the earth, the created convection cell breaks up in 3 smaller ones.

The air in the upper atmosphere is deflected when moving away from the equator. Around 30 degrees of latitude the air begins to flow from west to east causing an accumulation of air in the upper atmosphere. Some of the air I the upper atmosphere sinks, creating the subtropical high pressure zone. From this zone, the surface air travels in two directions. For the NH there are Northeast trades and a flow towards the poles, for the SH there are Southeast trades and flow towards the poles.

At 60^0 the subtropical westerlies collide with the polar easterlies. This results in the uplift of air and the creation of sub polar low pressure zones. Westerly winds (30^0-$70^{0)}$ are strong and variable. Trade winds (10^0-30^0) are moderate but persistent throughout the year. Near the continents they are generally overruled by tropical and subtropical seasonal winds called monsoons. Polar easterlies (>70^0) are moderate and blow over land or ice. Near the ITCZ wind climate is predominantly calm (10^0N-10^0S, called doldrums). In the same region tropical storms can develop (hurricanes, cyclones, typhoons)

The non-uniformity of earth's surface makes the situation far more complex. The large scale pressure belts are broken up caused by the local topography and differential warming of the oceans and land. Westerlies are the strongest winds and their seasonality is largest in the NH. Some tropical areas are dominated by trade winds while other regions are dominated by monsoons (seasonally reversing wind).

2.3 Ocean circulation

ocean water circulation also contributes to re-distribution of excess heat of the equatorial zone by a density driven flow. Warm, salty surface water is chilled in the North Atlantic and sinks to flow southwards. After upwelling in the Pacific and Indian Oceans it returns as surface flow to the North Atlantic. This surface flow is primarily wind-driven.

3. Large-scale variation in wave environments

3.1 Wave height variation

Global wind systems determine the global wave environments. Wave heights are highest at mid-latitudes as a result of Westerlies. The mid-latitude wave climate in the North Pacific and North Atlantic has much larger wave heights in the winter than in the summer. The southern Ocean has an almost unlimited fetch resulting in high waves but a much smaller seasonality. Wave heights in the tropics are generated by the gentle trade winds making them moderate. In the tropics and subtropics, larger waves are generated by swell or seasonal winds. The wave heights of monsoons strongly vary per region, while Tropical and east coast cyclones are too seasonal to greatly impact longer-term wave climates.

3.2 Wave environments

1) Storm wave climate: Is the most important and energetic wave environment. Locally generated by westerlies and associated mid-latitude cyclones between 40^0 and 60^0. Operates during the winter in the NH and all year long in the SH. Generally a combination of sea and swell is present. Waves are steep, short-crested, irregular and multi-directional.

2) West coast swell climate: swell waves generated in the storm wave belt. Located between 0^0-40^0. Year-round in SH while seasonal in NH. Waves are uniform in direction, shape and size and wave heights are moderate to high.

3) East coast swell: Same characteristics of west coast swell but only reach east facing coasts where west coast swell only reaches west facing coasts. Waves tend to be a little lower and arrive less frequently.

4) Protected wave environments: areas protected from swell. Shielded by ice, reefs, island archipelagos or land masses.

3.3 Coastal impact of different wave conditions

1) On open coasts, a storm wave climate is characterized by waves which are highly variable in height, period and direction. This results in a dynamic coastal profile. Larger waves break at relatively deep water, so the littoral zone extends to relatively larger water depths. Breaking waves tend to be of the spilling type.

2) Swell waves are relatively low and long. A swell climate gives a relatively narrow sandy littoral zone. Breaking waves tend to be of the plunging type

3) The monsoon climate of SE Asia has its highest waves in the summer. This results in a narrow sandy inner littoral zone with a gently sloping outer part dominated by finer sediments.

4) Cyclones generated very high waves and storm surge causing erosion and storm bars. Coastal morphology will generally be decided by the normal wave climate though.

4. Large-scale variations in tidal characteristics

4.1 Global tidal environments

Tidal characteristics like the magnitude and the character also vary globally. The tidal wave is distorted by local differences in water depth, location and shape of land masses, and embayment's. There are 3 categories based on magnitude.

1) Micro–tidal (<2m) (open coasts and fully enclosed seas)
2) Meso–tidal (2-4m)
3) Macro-tidal (>4m)

The tidal character is defined by the form factor: $F = (K_1 + O_1)/(M_2 + S_2)$

Category	Value of F
Semidiurnal	<0,25
Mixed, mainly semidiurnal	0,25-1,5
Mixed, mainly diurnal	1,5-3
Diurnal	>3

The diurnal components are introduced due to the declination of the earth axis. Daily inequality increases with latitude and is further influenced by the presence of land masses which can locally magnify the larger tide. Most of the world's coastlines are semi-diurnal or mainly semi-diurnal mixed tides. Areas with diurnal and mainly diurnal mixed tides tend to have smaller tidal ranges compared to semi-diurnal systems.

4.2 Coastal impact of tide and classification

Semi-diurnal or diurnal rise and fall creates a intertidal zone that is exposed during low water. In absence of waves coasts develop wide low-gradient tidal flats. Tidal currents in combination with the horizontal translation of the water line determine the morphology of these tidal flats. The coasts are generally dominated by fine sediments. In tropical and sub-tropical regions mangroves occupy the intertidal zone.

Tide-dominated flats occur for large tidal ranges and small wave heights. A useful parameter to delineate between wave and tide influence is the relative tidal range: $RTR = MSTR/H_b$. Where MSTR is the mean spring tidal range and H_b is the wave height just before breaking. For RTR<3 beaches are wave-dominated. For RTR>15 beaches approach the pure tidal flat situation. Between both beaches are shaped by waves with some distinct tidal characteristics.

Summary Coastal Dynamics

Chapter 5: Coastal hydrodynamics

1. **Introduction**
2. **Wave transformation**

2.1 Energy Balance

Waves can be transformed by refraction, shoaling, bottom friction and wave-breaking, which we will get to later. Models like HISWA ans SWAN have been developed to simulate these effects when waves enter the shallower coastal area. When you integrate all frequencies and directions in an irregular wave field you get the following Energy balance

change of energy + import of energy in x-direction + import of energy in y direction = gain of energy

$$\frac{\partial E}{\partial t} + \frac{\partial}{\partial x}\left(Ec_g cos\theta\right) + \frac{\partial}{\partial y}\left(Ec_g sin\theta\right) = S - D$$

In this balance D represents Dissipation. The easiest way for waves to dissipate energy is break. This mainly occurs in the surf zone.

2.2 Shoaling

When waves enter shallow waters, the bed friction starts to kick in. Wave speed will slow down significantly, yet energy is not dissipated that quickly. As a result, wave heights increase to accommodate the loss in speed.

2.3 Refraction

As waves travel faster in deeper water, a wave approaching under an angle will bend towards the coast because the parts further away from the coast will propagate faster and allow the wave to turn.

2.4 Diffraction

When waves are obstructed by an object, waves can bent around it decreasing the wave height of the wave that bends around it significantly.

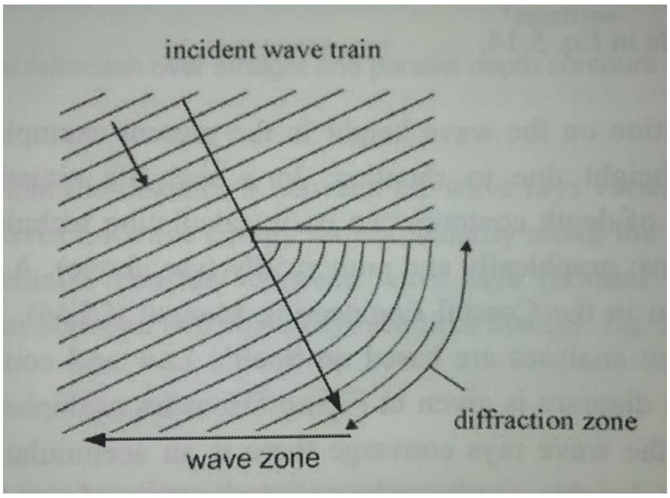


Figure 4; Diffraction

2.5 Wave breaking

Effect of bed slope on breaking process

the manner in which waves break can be described by the Iribarren parameter ξ.

$$\xi = \frac{\tan \propto}{\sqrt{\frac{H_0}{L_0}}}$$

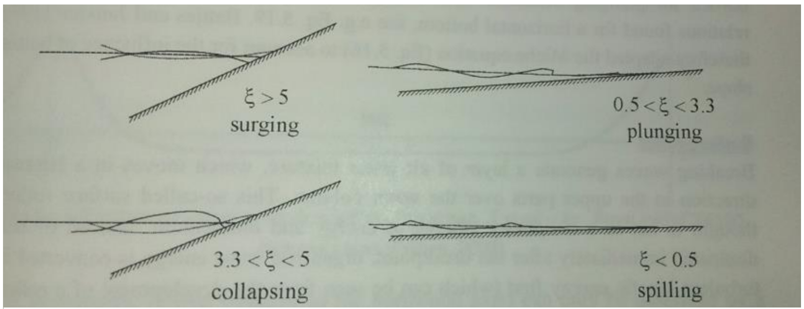


Figure 5; Wave breaking for different Iribarren parameters

Surging – Occur along steep coasts. More than half of the energy is reflected back into deeper water. Waves tend to reflect more than actual breaking.

Collapsing – Between Surging (non-breaking) and Plunging (breaking).

Plunging – Curling top breaking over the lower part of the wave. Wave energy is dissipated into turbulence.

Spilling – Found along flat beaches, the waves begin breaking a a relatively great distance from shore and break gradually.

Surface roller

A layer of air-water mixture, which moves in a landward direction in the upper parts over the water column.

3. Wave asymmetry and skewness

Waves propagating to the shore become more asymmetric until the point of wave breaking. There are roughly two forms of this:

Skewness

Asymmetry to the horizontal axis. Waves crest gradually peak while the through flattens out. The phenomena can be described by a sum of sinusoidal waves with higher harmonics.

Asymmetry

In shallow water the wave through tends to move slower than the crest. Which causes the wave to pitch forward (lean forward).

4. Wave orbital velocities, pressure and bed shear stress

4.1 Wave orbital velocities

Water particles beneath the wave surface move in an elliptical path. The further the particles are away from the surface, the flatter there elliptical orbit becomes while the horizontal movement remains the same. The ellipse flattens out and reduces to zero

4.2 Dynamic pressure

This is the wave-induces pressure that varies harmonically due to the surface elevation caused by the wave. It can written as:

$$\hat{p} = \frac{\rho g H}{2} * \frac{\cosh k(h+z)}{\cosh kh}$$

4.3 Wave boundary layer

5. Wave-induced set-up and currents

5.1 Wave-induced mass flux or momentum

Summary Coastal Dynamics

Chapter 6: Sediment

1. Introduction

2. Sediment properties

2.1 General

We can distinguish many different sizes of particles. They can roughly be divided into silt, clay, sand, gravel, and cobbles. The chemical properties vary greatly causing them to possible be very cohesive (clay), or even not cohesive at all (Carbonate sands)

2.2 Grain size, density and bulk properties

For sediment transport D_{50} (Median grain size) and D_{90}/D_{10} are important. If D_{90}/D_{10} is small (<1.5) sediment can be called well sorted. If it is large (>3) we can call it well-graded sediment.

Other factors of relevance to sediment transport are:

- Grain density (ρ_s) – Depends on mineral composition
- Relative density (s) – Ratio sediment/water density
- Fall velocity (w_s) – Depends on grain &fluid characteristics
- Angle of repose (φ) – the slope that can be sustained by sediment
- Porosity (p) – ratio of pore space of the sediment
- Sediment concentration (c) – defined by either volume or mass concentration.

2.3 Fall velocity

Defined as the maximum velocity a particle can obtain while settling. Is the balance between buoyance and gravity.

Basic equation

$$F_G = (\rho_s - \rho) g \left(\frac{\pi}{6} * D^3\right) \; (downward\ force) \qquad F_D = \frac{1}{2} C_d \rho w_s^2 \left(\frac{\pi}{4} D^2\right) (upward\ force)$$

$$w_s = \sqrt{\frac{4(s-1)gD}{3C_d}}$$

Dependence on Reynolds number

The drag coefficient can be described by:

$$C_D = 24/Re$$

Hindered settling

Other particles can obstruct the fall velocity of a particle. We can derive an effective fall velocity with the following formula: $w_e = (1 - concentration)^{\propto} w_s$

3. Initiation of the motion

3.1 Forces on a single grain

Sediment can be transported if the shear stress becomes large enough. The lift and drag forces will tend to move the grain (F_L& F_D), while the gravity force (F_G) will try to hinder this.

The driving force = $\rho u^2 D^2$ and the resisting force = $(\rho_s - \rho)g(D^3)$

From which the so-called shields parameter can be deduced: $\theta_{cr} = \frac{\tau_{b,cr}}{(\rho_s - \rho)gD} = C$

Experiments for sand on a smooth bed let C become 0,05

3.2 Shields Curve

Shields found that C was a weak function of the grain Reynolds number. He visualized the data in a curve we call the shields curve. The curve shows us the value for C for each Reynolds number. In reality the situation is more complex do to uniformity of the bed, Graduation of the bed, Slope of the bed and cohesive forces between the grains. The shields curve has been represented as a function of a non-dimensional grain size $D_* = D_{50}\left(\frac{g(s-1)}{v^2}\right)^{1/3}$

4. Basic principles of transport modelling

4.1 Definitions

4.2 Bed load versus suspended load

Bed – Load transport at low shear stresses

Sediment particles start rolling or sliding over the bed. If the shear stress increases further, the particles can start making small jumps (saltations).

Bed – Load transport at high shear stresses (sheet flow)

Sediment particles start moving in layers. The top layer moves back and both as a sheet of sand over a flat immobile bed.

Suspended load transport

Sediment particles which are suspended under relative low concentrations in the water and move horizontally without any interaction with the bed.

4.3 Practical modelling of sediment transport

Bed load transport formulations are often expressed in terms of bed-shear stress due to currents and waves, often supplemented with a criterion that describes initiation of motion. This approach can be called quasi-steady. Suspended load transport is often modelled as the product of the sediment concentration and the horizontal velocity. The suspended sediment transport can be computed by integrating the suspended sediment flux from the top of the bed load layer to the water level.

5. Bed load based on the Shields parameter

5.1 Importance of the Shields parameter

In the bed load layer the turbulent mixing is often assumed to be still small, so that it only slightly influences the motion of sediment particles. We can also assume that the bed load transport responds instantaneously to the bed shear stress. Many approaches for bed load transport are based on this reasoning. In such formulas the dimensionless sediment transport is invariably a function of the Shields parameter

5.2 Including waves

Near shore the influence of waves has to be included into the Shields parameter. This can be done by using the time-averaged bed shear stress or the instantaneous bed shear stress. The instantaneous bed load transport vector can be written as: $\Phi_b(t) = \frac{S_b}{\sqrt{(s-1)gD_{50}{}^3}}$

It responds quasi-steadily to the instantaneous bed shear stress. The instantaneous dimensionless effective shear stress is given by: $\theta' = \frac{\tau_b'}{(\rho_s-\rho)gD_{50}}$.
The shields parameter $\theta'$is a measure of the forcing on the sediment grains. The effective bed-shear stress $\tau_b'$is that part of the total bed shear stress which is transferred directly to the grains in the bed as skin friction. We can compute the skin friction through the roughness height related to grain size.

The non-dimensional critical shear stress $\theta_{cr}$is the threshold of motion of sand grains. It can be computed using the classical Shields curve.

5.3 Instantaneous bed load transport

Approach and explanation (read entire section)

5.4 Bed load transport based on time-averaged shear stress

Approach and explanation (read entire section)

5.5 Summary and concluding remarks

Many formulas where written for rivers and later applied to coastal environmenst. These formulations can normally be written in the form: $\langle\Phi_b(t)\rangle = f(\langle\theta'(t)\rangle, \theta_{cr})$

6. Diffusion approach for suspended transport

6.1 General formulation

If upward forces are large enough particles can be lifted from the bed and go into suspension. For not too high sediment concentrations and not too heavy particles, we can assume the particles move through a vertical plane with the horizontal water velocity.

6.2 Sediment continuity

To obtain a sediment concentration, a mass balance equation must be solved.

$$\frac{\partial c}{\partial t} + \frac{\partial uc}{\partial x} + \frac{\partial vc}{\partial y} + \frac{\partial wc}{z} - \frac{\partial w_s c}{\partial z} = 0$$

Horizontal advective terms are often samler than the vertical advective terms. We n

Summary Coastal Dynamics

Chapter 9: Coastal inlets and tidal basins

1. **Introduction**
2. **Basin and inlet types**

2.1 Bays, lagoons and estuaries

- *Tidal lagoons:*
  Basins that are enclosed by wave-shaped coastal barrier islands or spits. The Wadden Sea is an example of this. Penetration of waves into the lagoons is limited. Water flows into the lagoon with the flood and out during ebb. Inlets are called throats or gorges and sometimes occur sub-surface.
- *Tidal bays:*
  Basins that are more open to the deep sea water**.** The bays have a limited fresh water run-off. Waves can enter unhindered but usually lose their energy close to the entrance.
- *Estuaries:*
  These bays experience a (strong fresh-water run-off. Seawater in these basins is measurably diluted. Estuaries tend to be tide-dominated. Sediment is primarily imported from the adjacent coastal region. The entrance of the estuary can be constricted by spits, shoals or barriers. Estuaries can also be classified in the degree of mixing of the salt and fresh water: stratified, partially mixed/stratified and mixed or homogeneous.

2.2 Hydrodynamical classification

A slightly different distinction between the different basins types. Tidal basins are tide dominated and show influences by the tide in the form of tidal current ridges, extensive salt marshes and tidal flats. We can than distinguish between tidal basins with and without a strong river influence.

2.3 Hydraulic boundary conditions and geometric controls

Hydraulic boundary conditions are important for determining the morphology of tidal basins and inlets. A few geometric and hydraulic controls are:

- Tidal prism is the volume of water that has to flow in and out through the inlet during one tidal cycle and is determined by the surface area of the basin and the tidal range.
- The morphologically active part of basins consist of deeper areas where the flow is concentrated during lower tidal water levels and of tidal flats that are covered during higher water and are exposed during lower water
- Tidal waves propagating into basins may be progressive or standing or a mixture of the two. This depends partially on the length of basin. Resonance can occur when the basin length is approximately a quarter of the tidal wavelength.
- In very short basins the combination of tidal range, channel depth and intertidal storage areas or flats results in a different type of asymmetry than asymmetry between ebb and flood duration.

3. The main morphological elements

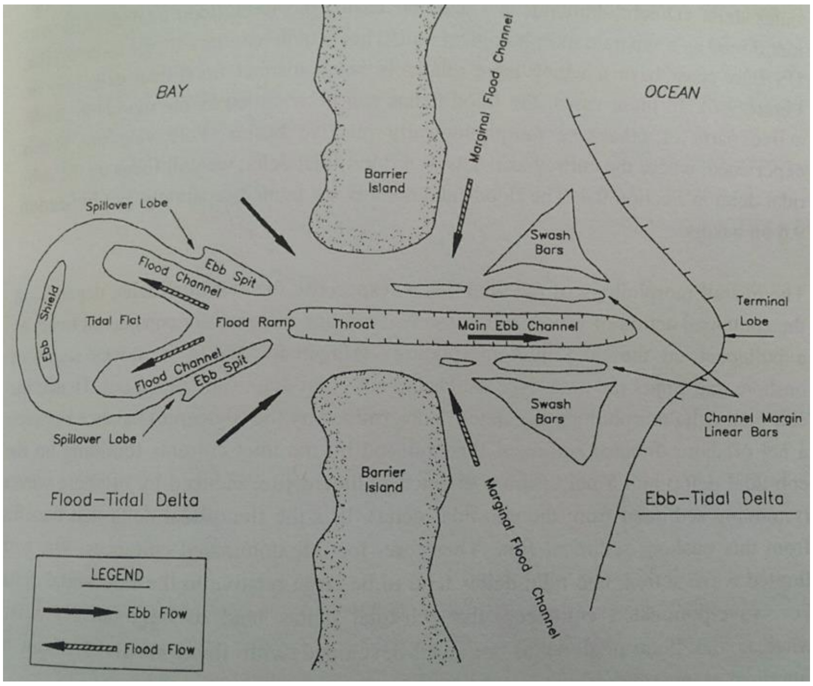
3.1 Introduction

3.2 Tidal deltas

Large tidal discharges cause strong sediment exchange between the basin and the outside area. This leads to the formation of sand deposits at either side of the entrance, called tidal deltas. Basins with large tidal ranges tend to have well-developed ebb- and flood-tidal deltas and very deep inlet gorges.

Ebb- and flood-tidal deltas are dissected by channels. The flats and channels exhibit a highly dynamic behavior. A typical ebb-tidal delta includes:

- *Marginal flood channels:* Flood channels to the sides of the inlet
- *Main ebb channel:* Due to inertia, the water continues flowing towards the sea in this channel even when the flood cycle starts.
- *Channel margin linear bars:* Bars that flank the ebb tidal channel. Built from deposits resulting of the interaction of flood and ebb tidal currents.
- *Terminal lobe: S*teep seaward-sloping body of sand
- *Swash platforms and bars:* Broad sheets of sand, isolated swash bars can develop on these sheets.



The flood-tidal delta may be fan ore horse-shoe shaped with a flood ramp that slopes upward ending in the ebb shield, the most elevated outer edge of the flood-tidal delta. For relative small basins with abundant sediment supply flood-tidal deposits are found everywhere in the basin.

The overall morphology of the tidal deltas depends on the combined action of waves and tides. Wave action acts as a bulldozer on the tidal inlet morphology. Ebb-tidal delta morphology is generally determined by the balance between a net offshore directed sediment flux induced by the inlet currents and a net onshore directed sediment flux induced by offshore waves. The flood tidal delta benefits from this onshore sediment flux. For tide-dominated entrances ebb-tidal deltas are well-developed with flood shoals that can be emergent at low tide

3.3 Basin characteristics

Tides fill and empty the basins via channels while cutting through tidal sand and mud flats. Higher parts remain dry at high water and are called supra-tidal flats or salt marshes. Ebb-dominant channels follow a meandering course, whereas flood-dominant channels shoal landwards. Along barrier islands coasts the basins are often rectangular or near square, and the channel structure is often more branched than braided. In the case of larger rivers discharging in tidal basins, the tidal basin is often funnel shaped, and the channel structure is not branched but more braided

4. The ebb-tidal delta or outer delta

4.1 Waves and currents at the outer delta

The morphology of the outer delta is highly complicated and variable. Meaning waves and currents encounter a very complex bed topography. Via refraction, diffraction and reflection, this can lead to complex wave patterns with a strong spatial variability. Currents in the vicinity of a tidal inlet are partly tidal, partly wave-driven, and partly wind-driven. Tidal currents are primarily concentrated in the main channels. Wave-driven currents in areas where waves are breaking and wind-driven currents are occur rather episodic

4.1.1 Wave patterns: In case of a typical inlet between two barrier islands. Due to refraction, waves turn towards depth contours. Wave energy is therefore concentrated in the central front edge of the ebb-tidal delta. Due to breaking the gorge will also be less exposed.

Another important aspect is wave penetration into the inlet. The barrier islands provide considerable shelter. Still wave energy penetrates through the gorge. The waves quickly radiate into the basins and wave heights rapidly decay.

The complexity of the bed topography and the mixture of sea/swell waves make modeling for tidal inlets very difficult.

4.1.2 Tidal residual currents: Is a very complicated picture. There usually is a distinct ebb-dominated current and there are often flood channels near the tips of the island. Another type of residual current is the Stokes' drift, which is dependent on the phase-coupling between the horizontal and the vertical tide. If in phase, there can be a considerable residual current.

4.1.3 Wave-induced currents: In tidal inlets, the wave-driven current around shoals on the outer delta can be sos strong that they dominate the tidal residual currents.

4.1.4 Wind-induced currents: Induced by shear stress on the water surface in the inlet, or indirectly, via set-up of the water level against the coast. Wind tends to be more effective in driving a current when it acts on shallower water. Due to the large variations in water depth which are inherent to a tidal inlet, wind-driven current field will therefor strongly vary in space.

4.1.5 3D combined current field: Current field around an inlet is more complex than on a uniform straight coast. Therefor 3-D modelling is essential.

4.1.6 Wave-current interaction

The tidal and wave-driven current pattern on the outer delta is largely concentrated in the deeper channels. There can be strong currents which affect the wave propagation via current refraction. This can even result in wave blocking.

Another form of wave-current interaction is the effect of waves on the bottom shear stress experienced by the current. The bed shear stress enhancement induced by waves may have major effects on the current pattern: tidal flow will tend to avoid shallow areas, where wave action and shear stress enhancement are strongest

4.2 Sediment transport patterns

Summary Coastal Dynamics

Summary Coastal Dynamics

Chapter 10: Coastal protection

1. **Introduction**
2. **Coastal protection strategies and methods** I

   n general, there are two possible solutions to coastal morphological problems, that is 'hard' measures (coastal structures) and 'soft' (natural) measures. The principle of soft measures like beach or foreshore nourishment is to compensate for the eroded sand by nourishing sand. Examples of hard measures are series of groynes, series of offshore breakwaters, submerged breakwaters and revetments or seawalls. They avoid that sediment is eroded by interfering in the sediment transports in alongshore and cross-shore directions of the coasts. The use of beach nourishment is becoming increasingly popular, but the use of hard measures cannot be disregarded.

3. **Coastal erosion**

   We speak about structural erosion in the case of a clear long-term erosional trend. Hence, structural erosion is a permanent erosion phenomenon. For example on the lee side of port entrances. Structural erosion means that the volume of sand gradually reduces as a function of time. The reduction is in the order of 10 to 50 m^3/m per year and a few meters of coastal retreat per year.

   A (severe) storm surge is not necessarily a structural erosion problem. The total volume of sand in a cross-shore profile has not changed during the storm surge. Only a redistribution of sand in the cross-shore profile. Dune erosion after a severe storm surge is thus a temporary instead of a permanent erosion. Generally a recovery towards the original situation occurs under the combined action of moderate waves and wind.

   Often a gradient in the longshore sediment transport is the main reason, this is because the eroded sediments from the upper part of the cross-section will not fully return under normal conditions, but will be removed in an alongshore direction.

4. **Modification of longshore transport processes**

   n most cases, the use of structures for coastal protection relies on the ability of such structures to interfere with the existing sediment transport processes. Erosion can be stopped by ensuring that the existing sediment transport is changed to dS/dx=0 on the specific section. In the upstream coastline higher rate erosion is expected, resulting in lee-side erosion.

5. **Structure influencing longshore transport rates**

   5.1 Introduction

   Structures of which the primary aim is to change the longshore transport rates under both normal and extreme conditions are: jetties, shore-normal breakwaters, series of groynes, and detached parallel offshore breakwaters.

5.2 Jetties on shore-normal breakwaters

Function 1: blocking the longshore transport of sand, which would otherwise settle into a dredged approach channel. The impact of long structures on the morphodynamics of the adjacent coasts can be very large.

Function 2: Stabilizing a natural river mouth or coastal inlet. A sand bypass is created to ensure stable sediment situation. If onle the growth of a spit is taken into account, the rates of accretion and erosion after stabilization and the necessary capacity of a bypass system may be largely underestimated.

Function 3: Flushing the entrance channel to a harbor or river by constriction of the entrance. In order to push the shoal formation to deeper water, two jetties can be built at the river mouth, in such a way that the entrance is kept narrow until deeper water is reached.

Function 4: Preventing structural erosion of a sandy coast near a tidal inlet. By building a long pier (dam) near the inlet at the end of the eroding coast one prevents that sediments disappear into the tidal basin.

5.3 Groynes

A field of groynes is a series of smaller jetties extending into the surf zone and spaced at relatively short intervals along a beach. They can be very effective in reducing the existing longshore sediment transport rate along a coast.
Two types:
(1) Impermeable, high-crested structures. Sawtooth appearance of overall shoreline.
(2) Permeable, low-crested structures. Slightly reduce the littoral drift in the inner surf zone and create a more regular shoreline (without saw-tooth effect). They are for instance made from sheet piles.

5.4 Detached shore parallel offshore breakwaters

Detached breakwaters are breakwaters parallel to the coast a certain distance from the coastline. The breakwaters are either emerged or submerged. In the shadow zone behind the structures, tombolo and salient development is stimulated.

<u>Emerged detached breakwaters</u>

Since emerged breakwaters are effective in reducing the longshore sediment transport capacity in their shadow, structural coastal erosion can be solved with a series of emerged shore-parallel offshore breakwaters. From the above, it follows that the optimal emerged breakwaters in terms of coastal protection are built close to the shore with a high crest level and small gap lengths, but such a structure will largely block the horizon and is not attractive in terms of beach recreation.

Submerged detached breakwaters

Examples are known from field and laboratory tests in which severe erosion is generated landward of submerged breakwaters. Wave breaking on the structure drives a current pattern with diverging shore-parallel currents in the shadow zone of the breakwater. Along the end sections of breakwaters, the currents escape in the seaward direction. This current pattern can transport a lot of sediment outside the area and counteract the accretional tendency.

5.5 Piers and trestles

these are rather long structures with a horizontal deck on a series of piles extending perpendicular to the coast into the sea. It might impact the adjacent coast and sediment transports may be reduces. Some accretion may occur.

5.6 Concluding

Long breakwaters, series of groynes or detached offshore breakwaters can be used to solve structural erosion problems. The principle of these measures is that they locally affect (reduce) the existing longshore sediment transports and hence change the longshore transport gradients.

6. Structures providing protection against storm-induced erosion

6.1 Introduction

If the rate of erosion due to a severe storm surge is unacceptably large during design conditions, the use of structures may be helpful in reducing the rate of erosion → seawalls, revetments or sea-dikes.

6.2 Seawalls

A seawall is a shore parallel and (nearly) vertical structure at the transition between the low-lying (sandy) beach and the (higher) mainland or dune. Unfortunately, a rigid massive seawall tends to reflect the incoming waves. The increased turbulence from this reflection may erode a deep trough along the toe of the sea wall. This can be prevented by maintaining a beach in front of the seawall.

Neither seawalls nor revetments can protect a coast from structural erosion due to longshore transport gradients.

6.3 Revetments

A revetment is similar to a seawall in the sense that it is a shore-parallel structure that can prevent storm erosion but is ineffective against structural erosion. Compared to a seawall, a revetments is more gently sloping. During tests is turned out that with a slope of 1:3.6 a deeper scour hole was found than with a slope of 1:1.8. For a rough slope a smaller scour depth may be expected than for a smooth slope.

The higher the level to which the revetment is applied the smaller the dune erosion is, but by implication also the deeper the scour trough. Revetments (and sea walls) physically prevent the loss of material from the dunes or land and thus reduce the dune or mainland retreat. The potential scouring in front of the structures has to be taken into account.

6.4 Sea-dikes

Sea-dikes differ from revetments in the sense that a beach in front of the structure is absent. Just like dikes along e.g. a river, sea-dikes are often meant to prevent flooding.

7. Nourishments

7.1 Introduction

The basic idea of nourishments (a 'soft' measure) is to supplement sand by artificial means (dredge, truck) in places where the loss or lack of sand is causing problems.

They can be applied for various reasons:

(1) To compensate for losses as a result of structural erosion
(2) To enhance the safety of the hinterland against flooding
(3) To broaden a beach, create new beaches or reclaim large areas of new land such as artificial islands

In the first case, with ongoing structural erosion, the nourishment will have to be repeated from time to time. An interval of 5 years between nourishments is generally acceptable.

7.2 Design aspects

Origin of the sand

The required volumes of sand are in the order of magnitude up to several hundreds of millions m^3 per project for large scale reclamation projects and up to 10 million m^3 per project for smaller-scale nourishments. The marine sources can be estuaries or the seabed. It is often attractive to try to combine excavation works with nourishment activities ('werk met werk maken')

In the Netherlands, some sand is obtained by maintenance dredging in the access channels of ports or from dredging thin layers at a distance of at least 20km from the shore.

Quality requirements for the sand

Major changes in the slopes and other coastal features are to be expected when the grain size of the supplied sand differs from the original material. Usually, this is not acceptable. It must also be ascertained that the beach nourishment material contains little or no fines (silt), this may have a negative impact on the marine environment.

Location of the nourishment

(1) On the inner slope of the dunes
(2) On the outer slope of the dunes
(3) On the dry beach
(4) On the shoreface

If compatible sand is available in nearby borrow areas, shoreface nourishments are more economical and recreation-friendly than beach nourishments because the sediment can be placed at the seaward edge of the surf zone where the navigational depth is sufficient for hopper dredgers.

7.3 Structural erosion of coasts

When a beach suffers from structural erosion, artificial nourishments can be applied as a 'soft' remedy. Generally lifetimes of 5-10 years are strived for since the initial costs of a nourishment operation are often rather high due to mobilization costs. For shoreface nourishment a relatively large nourishment volume is required as only part of the nourishment volume (approx. 20% to 30%) will reach the beach zone after 5 years. But the costs of a shoreface nourishment against a beach nourishment are much lower. Artificial nourishments can also be applied to counteract the structural coastal retreat due to sea-level rise.

7.4 Dune reinforcement

Nourishment of the dune area is usually done if the volume of material in the dune ridge is insufficient to cope with dune erosion during the design storm. The protection level of the land behind the dunes is increased more effectively by making a row of dunes wider than by making them higher. However, the widening of the dunes in the seaward direction disturbs the dynamic equilibrium of the cross-shore profile.