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Universiteit Utrecht

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DYNAMICS AND SEDIMENTARY DEVELOPMENT OF THE DUTCH WADDEN SEA WITH EMPHASIS ON THE FRISIAN INLET

A study of the barrier islands, ebb-tidal deltas, inlets and drainage basins

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Front: Vroedschapskamer in Dokkum, D. Reynes after sketches of E.S. Burmania, 2nd half of the 18th century: *Allegoric painting. It depicts, to the left, the two nymphs Ameland and Schiermonnikoog who are standing in the water and rejoice over the pact made, to the right, between the river-god Dokkum and the Ocean.*

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DYNAMICS AND SEDIMENTARY DEVELOPMENT OF THE DUTCH WADDEN SEA WITH EMPHASIS ON THE FRISIAN INLET

A study of the barrier islands, ebb-tidal deltas, inlets and drainage basins

DYNAMIEK EN SEDIMENTAIRE ONTWIKKELINGEN IN DE NEDERLANDSE WADDENZEE MET BIJZONDERE AANDACHT VOOR HET FRIESCHE ZEEGAT

Een onderzoek van de barrière eilanden, buitendelta's, zeegaten en kombergingsgebieden

(met samenvattingen in het Nederlands en het Engels)

PROEFSCHRIFT

TER VERKRIJGING VAN DE GRAAD VAN DOCTOR AAN DE UNIVERSITEIT UTRECHT, OP GEZAG VAN DE RECTOR MAGNIFICUS PROF. DR. J.A. VAN GINKEL, INGEVOLGE HET BESLUIT VAN HET COLLEGE VAN DEKANEN IN HET OPENBAAR TE VERDEDIGEN OP DONDERDAG 23 MAART 1995 DES NAMIDDAGS TE 12.45 UUR

DOOR

ALBERT PETER OOST

Geboren op 26 augustus 1961 te Nieuw-Weerdinge

Promotor: Prof. Dr D. Eisma

Co-promotor: Dr P.L. de Boer

De druk van dit proefschrift is voor een belangrijk deel gefinancierd door Rijkswaterstaat, Rijks Instituut voor Kust en Zee (RIKZ) in het kader van het project Kustgenese. Het onderzoek is uitgevoerd aan de vakgroep Geologie, bij de afdeling Sedimentologie aan de Universiteit van Utrecht. Een deel van het onderzoek kwam tot stand met logistieke en financiële ondersteuning van de Rijkswaterstaat, Directie Noord-Nederland en Rijks Instituut voor Kust en Zee in het kader van het project Kustgenese.

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Een nieuwe lente
Op takken rust al de zweem
Van nieuwe bloesem

CONTENTS

SAMENVATTING

De nederlandse, duitse en deense Waddenzee vormen samen één van de belangrijkste natuurgebieden in West-Europa en één van de grootste intergetijde gebieden (bij eb droogvallende platen) op aarde. De sterke natuurlijke dynamiek is bepalend voor het karakter van dit gebied.

Hoofdstuk 1 - Sedimentologie en ontwikkeling

Het nederlandse Waddengebied is een barrière systeem, bestaande uit barrière eilanden en de achterliggende Waddenzee. Deze Waddenzee is deels een intergetijde gebied dat bij laag water droog valt. Het gebied wordt gedraineerd door getijdegeulen die via zeegaten uitmonden in de Noordzee. Zeewaarts van de zeegaten liggen de ebgetijde delta's, ook wel buitendelta's genoemd. De buitendelta, de aangrenzende eilandpunten, het zeegat, de geulen en de platen van één zeegat-systeem vormen één geheel. De ontwikkelingen van de verschillende onderdelen ervan zijn in sterke mate onderling gekoppeld en worden vooral bepaald door het getijdeprisma.

In het eerste hoofdstuk wordt een kort overzicht gegeven van het Waddengebied vanuit een sedimentologisch perspectief. De verschillende onderdelen, zoals eilanden, getijdemoerassen, platen, prielen en geulen, hun dynamiek en onderlinge samenhang komen hierbij aan bod. Ook wordt kort ingegaan op de flora en fauna en de historie van het Waddengebied. Een deel van de besproken onderwerpen wordt nader uitgewerkt in de navolgende hoofdstukken, waarin wordt gekeken naar enkele aspecten van de sedimentatie op een steeds kleinere tijd- en ruimteschaal.

Hoofdstuk 2 - De ontwikkeling van de oostelijke Wadden in historische tijden

Hierin wordt de ontwikkeling van het oostelijke Waddengebied en in het bijzonder van het Friesche zeegat gereconstrueerd, voornamelijk op basis van historische informatie. De oostelijke Waddeneilanden en het daarachter liggende gebied werden zo'n 6000-5000 jaar geleden, na de laatste IJstijd, gevormd. Onder invloed van de doorgaande stijging van de zeespiegel trokken de eilanden en het achterliggende gebied zich landwaarts terug over meerdere kilometers.

De reconstructies laten zien dat tijdens de laatste 2000 jaar voortdurend een waddengebied aanwezig was. Aan de zeezijde werd het beschermd door de barrière eilanden. Bodemdaling, die mede door de mens werd veroorzaakt, heeft de vorming van de Zuiderzee, de Middelzee en de Lauwerszee in de hand gewerkt. Deels hangen deze overstromingen mogelijk ook samen met een geringe stijging van de zeespiegel. Daarnaast is sprake van een zelfversterkend effect: uitbreiding van de inbraken leidde tot grotere getijdeverschillen en

daarmee tot een sterkere erosie van met name de veengebieden. Deze inbraak-gebieden zijn in de loop van het laatste millennium weer geleidelijk ingedijkt. Ondanks deze grote veranderingen is het Wadden systeem als zodanig niet teloor gegaan.

Uit de reconstructie blijkt verder dat aangrenzende zeegat-systemen elkaar onderling beïnvloeden, vooral van west naar oost (domino-effect). Onder invloed van de overheersende oostwaartse windrichting en de van west naar oost lopende getijdestroom langs de kust hebben de zeegat-systemen en daarmee samenhangend de tussenliggende eilanden de neiging om zich oostwaarts te verplaatsen. Voor het gebied oost-Terschelling tot en met Engelsmanplaat werd het domino-effect waarschijnlijk veroorzaakt door de indijking van de Middelzee waarbij het dubbele geulsysteem plaats maakte voor een enkel zeegat systeem. Het overgebleven zeegat verplaatste zich oostwaarts en tastte Ameland aan. Daarbij verschoof ook het wantij ten zuiden van Ameland oostwaarts, evenals het Pinkegat. Deze tastte op haar beurt de Engelsmanplaat aan. Indien deze aantasting door gaat mag worden verwacht dat de Engelsmanplaat in de volgende eeuw verdwijnt, waardoor Pinkegat en Zoutkamperlaag zullen versmelten.

Ten oosten van de Engelsmanplaat ligt de Zoutkamperlaag. Rond 1300 had dit zeegatsysteem waarschijnlijk nog geen of een slechte verbinding met de Lauwerszee, en werd deze nog gedraineerd via het Lauwers Zeegat (ten oosten van Schiermonnikoog). Rond 1550 had de Zoutkamperlaag de drainage van het Lauwers Zeegat geheel overgenomen. Aanvankelijk verzandde het Lauwers Zeegat grotendeels (1640 A.D.) en kon het eiland Bosch ten oosten ervan niet langer worden onderhouden, waardoor het in een zandplaat veranderde. Door oostwaarts op te schuiven verkreeg het Lauwers Zeegat weer een groter drainagegebied en vormden zich weer nieuwe platen. Dit oostwaarts opschuiven ging ten koste van het drainagegebied van het Schild Zeegat, zodat deze op zijn beurt niet langer in staat was om het onderliggende eiland Rottumeroog te onderhouden. Mede door de zandbehoefte van het Eems-Dollard estuarium werd dit eiland snel kleiner. Het verlies aan drainagegebied van het Schild kon en kan niet goed worden gecompenseerd door naar het oosten te verschuiven, omdat het gebied aan de oostkant wordt begrensd door het grote Wester Eems Zeegat.

Hoofdstuk 3 - De cyclische ontwikkeling van het Pinkegat

De ontwikkeling van het Pinkegat is sterk cyclisch: van een enkelvoudig zeegat (één hoofdgeul) tot een meervoudig zeegat (twee of meer hoofdgeulen) en weer terug. Tijdens een enkelvoudige fase is het zeegat diep en schuift het op in oostelijke richting. Hierbij wordt de oostpunt van Ameland langer. Daardoor moet de vloed een steeds langere weg afleggen naar de vloedkom van het Pinkegat. Na verloop van tijd worden in de zandplaat aan de oostpunt van Ameland nieuwe zeegaten uitgeschuurd en ontstaat een situatie met een meervoudig zeegat. In zo'n meervoudige zeegat-situatie zijn de geulen ondieper. Een meervoudig zeegat is evenwel instabiel. De westelijke geulen schuiven het snelst naar het oosten, waarbij de oostpunt van Ameland weer aangroeit. Na verloop van enkele decennia versmelten de geulen met elkaar of verzanden, zodat slechts één geul overblijft. De veranderingen in het zeegat leiden ook tot veranderingen in het achterliggende Waddengebied voor wat betreft de ligging van de hoofdgeulen en de positie van het wantij ten zuiden van Ameland.

In samenhang met dit cyclische gedrag en door de afslag aan de westzijde van Ameland is de oostpunt van Ameland én ook de hele vloedkom van het Pinkegat sinds 1800 zo'n 4,5 kilometer naar het oosten geschoven. De verkleining van het oppervlak van de vloedkom ten gevolge van het opschuiven van het wantij werd gecompenseerd doordat de westzijde van de Engelsmanplaat (oostelijk van het Pinkegat) steeds verder werd geërodeerd.

De ontwikkeling van de Engelsmanplaat verloopt ook cyclisch. De cyclus begint met erosie door geulen aan de noordzijde van de Engelsmanplaat. Daarna bouwen golven en stroming aan de Noordzee-zijde van de geul een nieuwe plaat op. Hierbij worden grote hoeveelheden sediment vastgelegd, waardoor de sedimentaanvoer naar de Engelsmanplaat afneemt. Deze wordt lager. De geul ingeklemd tussen de Engelsmanplaat en de nieuwe plaat raakt buiten gebruik. Na de opvulling ervan versmelten de twee platen. Zodoende wordt de Engelsmanplaat weer hoger en groter.

Hoofdstuk 4 - De cyclische ontwikkeling van de Zoutkamperlaag

Ook in de Zoutkamperlaag is een cyclische ontwikkeling waar te nemen. Onder invloed van getij en wind verplaatsen de geulen aan de westzijde van de ebgetijde delta zich naar het noorden. Daar aangekomen worden ze verplaatst naar het oosten. Van tijd tot tijd worden aan de westzijde nieuwe buitengeulen gevormd. Als gevolg van deze verplaatsing van de getijdestromen treedt ook in het drainagegebied van de Zoutkamperlaag een cyclische ontwikkeling op. Daarbij ontstaan in de hoofdgeul twee aparte geulen, die steeds meer van elkaar gescheiden raken. Daarna wordt de meest westelijke na verloop van tijd verlaten.

Uit vergelijking met het Pinkegat blijkt dat de beste informatie voor geologen over dynamiek en dimensies van geulsystemen vaak te verkrijgen is uit afzettingen gevormd in de zeegaten zelf.

Hoofdstuk 5 - De afsluiting van de Lauwerszee

Ten gevolge van de afsluiting van de Lauwerszee in 1969 nam het getijdevolume van de Zoutkamperlaag met éénderde af (van $305*10^6$ m^3 tot $200*10^6$ m^3). Daardoor was de buitendelta plotseling te groot waardoor daar erosie optrad. Daarnaast werden, door de verkleining van het getijdevolume, de hals van het zeegat en de hoofdgeul relatief te diep en hadden daarmee een te groot stroomvoerend oppervlak voor de afgenomen komberging.

Onder invloed van de relatief (in verhouding tot de getijdebewegingen) toegenomen golfwerking werd een deel van het sediment van de buitendelta geconcentreerd in een zandhaak vóór de kust van Schiermonnikoog. Het grootste deel van het sediment dat werd geërodeerd in de buitendelta ($26*10^6$ m^3 in de periode 1970-1987) werd echter getransporteerd naar de vloedkom. Daar kwam het voor een groot deel tot afzetting in de hals van het zeegat en in de hoofdgeul. Daardoor werd het stroomvoerend oppervlak weer verkleind en raakte het beter afgestemd op het kleiner geworden getijdevolume dat na de afsluiting van de Lauwerszee door het zeegat stroomde.

De Lauwerszee is blijvend afgesloten en daarmee is de verstoring permanent. Grote hoeveelheden sediment binnen het Zand Delend Systeem van de Zoutkamperlaag zijn blijvend verplaatst. Aldus bereikt het systeem geleidelijk een nieuw evenwicht.

Hoofdstuk 6 - De invloed van de mossel op de sedimentatie in de Waddenzee

Een belangrijke bijdrage aan het vastleggen van slib wordt geleverd door mosselen. Deze filtreren zeewater (*filter feeders*) en leggen daarbij grote hoeveelheden slib vast in de vorm van uitwerpselen. Experimenten lieten zien dat deze uitwerpselen over enige kilometers kunnen worden vervoerd voordat zij uit elkaar vallen. De geresuspendeerde uitwerpselen blijken tevens een grotere korrelgrootte te hebben dan het oorspronkelijk gefiltreerde slib.

Door intensieve bevissing, in combinatie met een aantal stormwinters, is de mossel (en daarmee haar slibvangende werking) momenteel nagenoeg geheel verdwenen uit de oostelijke Waddenzee. Een herstel van een natuurlijke *filter feeder* populatie, waarbij mosselen tot enkele tientallen procenten van de benthische biomassa vormen, zou het invangen van slib zeer ten goede komen.

Vergelijking met fossiele en recente, elders gevormde afzettingen laat zien dat in een groot deel van de geologische geschiedenis *filter feeders* actief zijn geweest en de accumulatie van fijnkorrelig sediment actief hebben bevorderd.

Hoofdstuk 7 en 8 - De vorming van kleinschalige ribbelstructuren in getijde-milieus

Experimenten (Hoofdstuk 7) laten zien dat, in tegenstelling tot wat algemeen werd gedacht, kleinschalige stroomribbels zich altijd ontwikkelen tot tongvormige ribbels. Daarbij zijn de gemiddelde dimensies van deze evenwichts-ribbels en van de opeenvolgende ontwikkelingsstadia (met als uitgangssituatie een vlakke bodem) gelijk voor alle stroomsnelheden, binnen het ribbel-stabiliteitsveld. Dit maakt het mogelijk om de ontwikkeling te voorspellen van stroomribbels bij een onregelmatige stroming, zoals in getijdegebieden. De uiteindelijke ribbelvorm blijkt afhankelijk te zijn van het traject van het stroomsnelheidsprofiel. Met deze kennis kan recente ribbelvorming beter begrepen worden en kunnen over ribbelstructuren in fossiele afzettingen conclusies worden getrokken over het regime waaronder ze vormden.

Hoofdstuk 9 - Epiloog

Hierin worden de in de eerdere hoofdstukken besproken processen en relaties beschouwd in hun onderlinge verband. Er wordt aangegeven hoe morfologische elementen en fysische processen elkaar op verschillende schalen van tijd en ruimte beïnvloeden en hoe het Waddensysteem, ook wanneer drempels worden overschreden, streeft naar dynamisch evenwichten op vele tijd- en ruimteschalen.

INTRODUCTION AND SUMMARY

The Wadden Sea, along the coasts of The Netherlands, Germany and Denmark, forms one of the major nature reserves in western Europe and one of the major intertidal areas on Earth. Numerous observations are available from this highly dynamical area, collected over the many centuries that Man has been wrestling with the sea. The growing concern about the fate of the Dutch coast in the next century, when an increase of the rate of sea-level rise is expected, has recently triggered extensive interdisciplinary research on coastal dynamics. This offered the opportunity to study in detail the dynamics and development of the Dutch Wadden Sea system, especially of the Frisian Inlet. The central theme in this study is the detailed understanding, from a geologists' point of view, of the dynamics of the morphological and sedimentological development of the Dutch Wadden Sea on different temporal and spatial scales. Special attention was given to interactions and feedbacks between sedimentary processes, sedimentary deposits and morphological elements, in order to promote a more dynamic approach of fossil analogues.

Chapter 1 - Sedimentology and development of the area

The Dutch Wadden Sea area is a barrier complex consisting of the Wadden Barrier Islands and the Wadden Sea. Large parts of the Wadden Sea are emerged at low water. Drainage occurs through tidal channels and the major tidal inlets. Seaward of the inlets, ebb-tidal deltas occur. The ebb-tidal delta, the tidal inlet and the adjacent parts of the islands, and the channels and shoals of an inlet system represent one dynamic entity, and can exchange sediment in order to compensate disturbances (Sand Sharing System). Developments in different parts are, moreover, strongly related, and are largely defined by the tidal prism.

The first chapter describes the Wadden Sea area from a sedimentological perspective. The different parts, islands, tidal marshes, channels and shoals and their interrelations and feedbacks are discussed, as well as the flora and fauna, and the historical development of the area. A part of the topics discussed is further elaborated in the subsequent chapters, in which different aspects of sedimentary processes are discussed on an increasingly smaller spatial and temporal scale.

Chapter 2 - Development of the eastern Dutch Wadden Sea in historical times

The islands along the eastern Dutch Wadden Sea and the largely intertidal area behind them were created some 6000 to 5000 years ago after the most recent Ice Age. Due to the rise of the sea level, the islands and the backbarrier area retreated over several kilometres.

The reconstructions in Chapter 2 show that intertidal backbarrier areas have been continuously present during the last 2000 years, protected at the sea side by the barrier islands.

Natural subsidence, and especially man-induced lowering of the land surface by drainage and peat excavations, led to the flooding of low-lying land and to the creation of the major embayments Zuider Zee, Middelzee and Lauwerszee. These floodings were, moreover, probably stimulated by a slight rise of the eustatic sea level. Once the inundations started, they became reinforced due to increases of tidal amplitude and tidal prism which especially led to increased erosion of peat areas. In the last millennium these inundated areas have gradually been reclaimed. Despite these great changes the Wadden Sea has maintained its character.

The reconstructions show that adjacent inlet systems influence each other, especially in the eastward direction (knock-on effect). Due to the dominant wind to the east and the eastward residual tidal current, the tidal inlets and the islands between tend to shift to the east. In the area between East Terschelling and the shoal Engelsmanplaat such a knock-on effect was likely started by the dyking of the Middelzee. The decrease of the N-S length of the Middelzee and the related decrease of tidal prism led to the transfer from a double inlet system west of Ameland towards a single inlet. The latter migrated laterally to the east and started eroding the western part of Ameland. Consequently, the tidal watershed south of Ameland migrated to the east, as did the Pinkegat Inlet which in its turn affected Engelsmanplaat, a shoal with a core of firm early Holocene clays. A continuation of the historical developments would imply that the erosion of Engelsmanplaat will continue. It is to be expected that this shoal will disappear in the next century and that consequently the Pinkegat and Zoutkamperlaag inlets will merge.

Probably the Zoutkamperlaag Inlet east of Engelsmanplaat was a small inlet around 1300 A.D., and had no or only a poor connection with the Lauwerszee, which in those days was drained by the Lauwers Inlet passing along east of Schiermonnikoog. The Zoutkamperlaag had fully taken over this drainage from the Lauwers Inlet in about 1550 A.D. Initially the Lauwers Inlet started to be abandoned and silted up (1640 A.D.), and the island Bosch east of it declined and was transferred into a sandy shoal. The Lauwers Inlet, however, started to migrate to the east, and thus obtained a larger drainage area, at the expense of the Schild Inlet further to the east. Consequently the island Rottumeroog, east of Schild Inlet decreased in size. The loss of drainage area of the Schild Inlet can not be compensated by a shift towards the east, because there the area is bounded by the large Western Ems Inlet which drains the Ems-Dollard Estuary.

Chapter 3 - Cyclical development of the Pinkegat Inlet

The development of Pinkegat Inlet shows a strongly cyclical character: it develops from a single inlet with one main channel towards a double or multiple inlet with two or more main channels, and back. In the phase with a single main channel, the inlet is deep and migrates towards the east. At the same time Ameland expands towards the east as well. Due to the eastward shift of the channel, the tidal currents have to cover an increasingly longer pathway towards the Pinkegat drainage basin, and consequently new channels develop through the shoals at the eastern end of Ameland, leading again to a multiple inlet system with shallower channels. Such a multiple inlet situation, however, is not stable. The western channels shift faster to the east than do the eastern ones, so that, after some decades, channels merge or silt up, and only one channel persists. These changes also affect the location of the channels in the backbarrier drainage basin and the position of the watershed south of Ameland.

In relation to the cyclic development and due to the erosion of the western part of Ameland, the eastern end of Ameland has expanded towards the east over about 4.5 kilometres since 1800 A.D.. The Pinkegat drainage basin moved over about the same distance to the east as well, and kept its size because of erosion of the western part of Engelsmanplaat.

The morphodynamics of the latter shoal are also cyclic. The cycle starts with erosion by channels at the northern side of the Engelsmanplaat. Then, a shoal builds up by waves and currents at the North Sea side of the channel. Sediment is trapped there and, as a consequence, the Engelsmanplaat becomes lower. The channel between the Engelsmanplaat and the new shoal becomes abandoned and, after its fill, the shoals merge. Thus the Engelsmanplaat becomes higher and larger again.

Chapter 4 - Cyclical development of the Zoutkamperlaag Inlet

Also the Zoutkamperlaag Inlet shows a cyclical development. Due to the tides and waves the channels at the westernmost side of the ebb-tidal delta shift to the north. From there they shift to the east. From time to time new outer channels are formed at the westernmost side of the ebb-tidal delta. Due to the shift of the tidal currents also the main backbarrier channel develops cyclically, from a single channel to a double channel, and back due to abandonment of the westernmost one.

Comparison with the Pinkegat learns that the better sedimentological information about the dynamics and dimensions of the system is to be obtained from the deposits formed within the inlets.

Chapter 5 - The closure of the Lauwerszee

By the artificial closure of the Lauwerszee in 1969, the tidal prism of the Zoutkamperlaag decreased with a third to $200*10^6$ m^3. In order for a new equilibrium to develop large amounts of sand have been (and still are being) relocated. Due to the closure the ebb-tidal delta was suddenly too large and started to be eroded. The gorge of the inlet and the main channel were too deep relative to the decreased tidal prism.

Due to the decrease of tidal currents and the consequent *relative* increase of wave energy, part of the sand of the ebb-tidal delta was transferred towards a recurved bar northwest of Schiermonnikoog. A much greater amount of sand ($26*10^6$ m^3 in the period 1970-1987) was transferred towards the backbarrier drainage area where it was deposited for the larger part in the gorge of the inlet and in the main channel. Thus the cross-sectional channel area decreased and became better adapted to the decreased tidal prism.

Chapter 6 - The influence of the mussel *Mytilus edulis* on sedimentation in the Wadden Sea

Mussels (*Mytilus edulis*) are filter feeders, and concentrate large amounts of fine-grained suspended sediment into faeces and pseudo-faeces, which are hydrodynamically equivalent to fine sand. Experiments showed that the biodeposits can be transported over several kilometers before they disintegrate. Upon disintegration the resuspended particles have a larger grainsize than the original suspension, which was filtered by the mussels. In these ways they thus contribute significantly to the deposition of fine-grained sediment. Due to intensive fishery and several winters with heavy storms, the mussel has largely disappeared from the eastern Wadden Sea in the last decade. Originally mussels and cockles each represented about 15% of the total biomass in the area. Their reestablishment would increase the capture of fine-grained sediment significantly.

A comparison with other recent environments and with fossil deposits indicates that filter feeders have been active during a large part of the geological history, and have favoured the deposition of large amounts of fine-grained sediment.

Chapters 7 and 8 - Small-scale current ripples in tidal environments

In contrast to the general opinion, experiments (Chapter 7) learned that equilibrium small-scale current ripples have a linguoid form. The average dimensions of these ripples and their subsequent development stages from flat bed conditions are independent of current velocity, within the current ripple stability field. This allows the prediction of the development of small-scale current ripples in tidal environments. The resulting ripple form and dimensions appear to depend on the current velocity profile. This offers tools to draw conclusions about the conditions under which small-scale current ripples in fossil tide-influenced deposits have been formed.

Chapter 9 - Epilogue

In this chapter, the previously discussed processes and relations and their mutual dependence are discussed. It is demonstrated that the various elements which constitute a barrier system can be considered hierarchical systems. It is discussed in which way morphological elements and physical processes influence each other on different spatial and temporal scales, and that - even if thresholds are passed - the Wadden Sea system continuously endeavours to maintain dynamical equilibria at many spatial and temporal scales.

CHAPTER 1

SEDIMENTOLOGY AND DEVELOPMENT OF BARRIER ISLANDS, EBB-TIDAL DELTAS, INLETS AND BACKBARRIER AREAS OF THE DUTCH WADDEN SEA

Senckenbergiana maritima, 24, 65-115

ABSTRACT

This chapter presents an overview of the Dutch Wadden Sea from a sedimentological point of view. After the pioneering work of scientists like, amongst others, Van Straaten and Reineck in the fifties and sixties, new impulses to this kind of research are being given by the need for detailed recent analogues of fossil hydrocarbon-containing rock successions and by the great concern about the future of our coastline in relation to accelerated sea-level rise. After many studies of a descriptive nature in the past, there is now a growing tendency to a more dynamical view to the Wadden Sea system. There is a strong interdependence between various tidal sub-environments within individual inlet systems. Together these sub-environments form so-called Sand Sharing Systems, whose behaviour is largely defined by the tidal prism and the wave climate. Such a dynamical approach may greatly facilitate the research and understanding of fossil barrier-related sediments. Apart from the physical processes the abundant biota plays also an important role in the sedimentological development of the Wadden Sea. The large amount of data on the development of the Wadden Sea in pre-historical and historical times, moreover, allows to test hypotheses about the evolution of the system on the scale of centuries to millennia.

INTRODUCTION

The objective of this chapter is to give an overview of the sedimentology of the Dutch Wadden Sea barrier coast. The Dutch Wadden Sea was studied in the fifties, the sixties and the seventies in great detail. In order to assess the potential impact of sea-level rise over the next decades, extensive studies are being undertaken to gain a better understanding of the interrelations between the hydrodynamics and the morphology/sedimentology of the area. These studies lead to a new and dynamic view of the sedimentology of the Wadden Sea.

Barrier islands, tidal inlets, tidal deltas, lagoons and tidal flats commonly occur along the coasts of marine basins with a micro- and especially mesotidal regime and a limited supply of sediment (Fig. 1; Hayes, 1979). Their origin is usually related to a rise of sea level which, on the one hand, leads to a decrease of gradient on the land and a consequent decrease of sediment supply and, on the other hand, to the shoreward transport and accumulation of sediments. Although tidal flats and salt marshes also occur along macrotidal coasts, barrier islands are commonly absent there. The Wadden Sea, with a tidal range from 1.4 to 3.5 m (Fig. 2), fringes the Dutch, German and Danish coasts over a distance of nearly 500 km with a maximum width of approximately 35 km. Towards the North Sea it is bordered by some 20 large and many small barrier islands and sandy shoals. Behind these islands lies the largest tidal flat area in Europe. The notion 'wadden' indicates the intertidal flats, as they occur extensively in the Wadden Sea. The mainland coast consists of dykes, some salt marshes and a few Pleistocene cliffs.

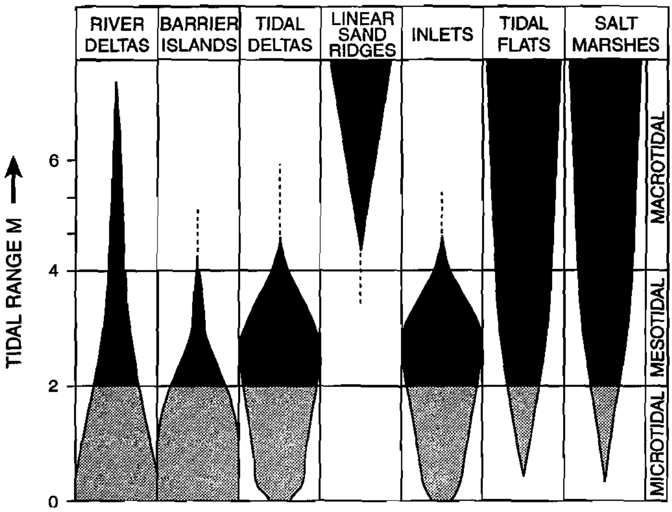


Figure 1: Relation between the morphology of the coastline and the tidal range (Hayes, 1979).

In the Dutch Wadden Sea the tidal regime is microtidal in the southwestern part, becoming mesotidal towards the east. In some of the deep channels between the larger islands the water depth may reach several tens of metres. During low tide about half of the area is exposed. Supply of fresh water and sediment by rivers is of minor importance. The bulk of the sediment is supplied from the North Sea; minor amounts of carbonates and organic matter are produced within the Wadden Sea. Generally, the salinity approaches that of the coastal North Sea, except in parts of the western Dutch Wadden Sea where fresh water from the lake 'IJsselmeer' and from small local rivers and man-made canals dilutes the coastal waters.

Shallow coastal waters such as the Wadden Sea are among the richest food supplying marine ecosystems, supporting a rich flora and fauna. A large number of marine animals lives in the Wadden Sea during their juvenile stages and thus the area serves as an important nursery ground for the North Sea fauna.

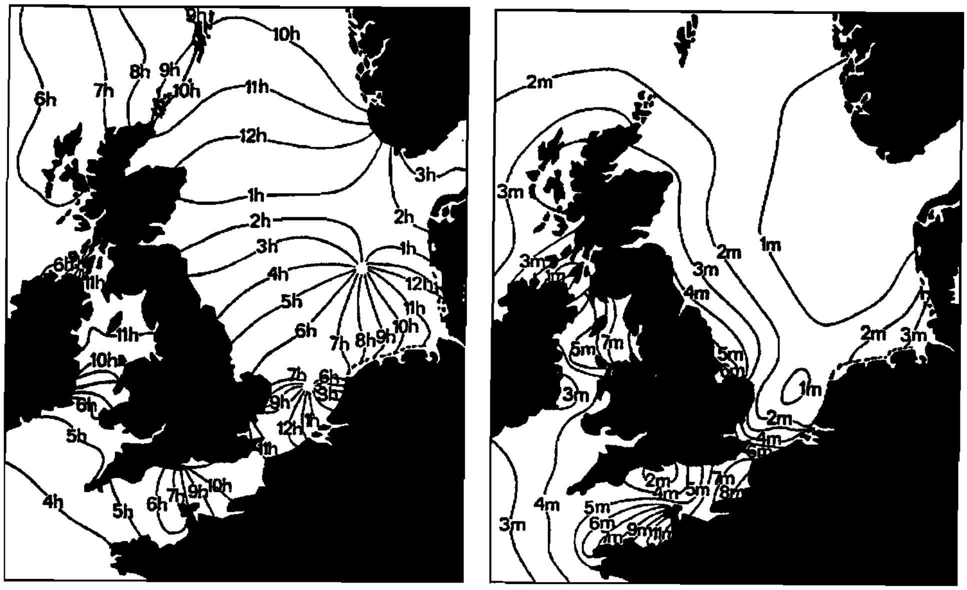


Figure 2: Tidal parameters of the North Sea and the British Isles. A. Co-tidal lines; B. Co-range lines (Reineck, 1982a).

SEDIMENTATION: SOURCES AND SINKS

Sands

In the Dutch Wadden Sea the net deposition of sand amounts to 8-21*10^6 $m^3.yr^{-1}$ (Anonymous, 1981). This sand is mostly derived from erosion of the shoreface, beaches and dunes of the North Sea coast of the barrier islands and northern Holland (cf. Winkelmolen & Veenstra, 1974; Veenstra & Winkelmolen, 1976), and to a minor extent probably also from the adjacent North Sea floor (Oost & de Haas, 1993). Deposition takes place mainly on the tidal flats, 87.5% of which comprise sandy sediments. The remaining 12.5% consist of fine-grained muddy sediments.

To compensate for the relative sea-level rise, deposition of several million $m^3.yr^{-1}$ is needed in order to maintain the same altitude with regard to the mean sea level. In addition, substantial amounts of sand are needed to compensate the extraction of sand from the Wadden Sea for building and construction purposes and to compensate the subsidence due to gas-extraction.

The recent pattern of sedimentation and erosion in the western Dutch Wadden Sea is still adapting to the change of tidal flow patterns, resulting from the closure of the Zuider Zee embayment in 1932 by the construction of the 'Afsluitdijk'. Deposition in the wake of this

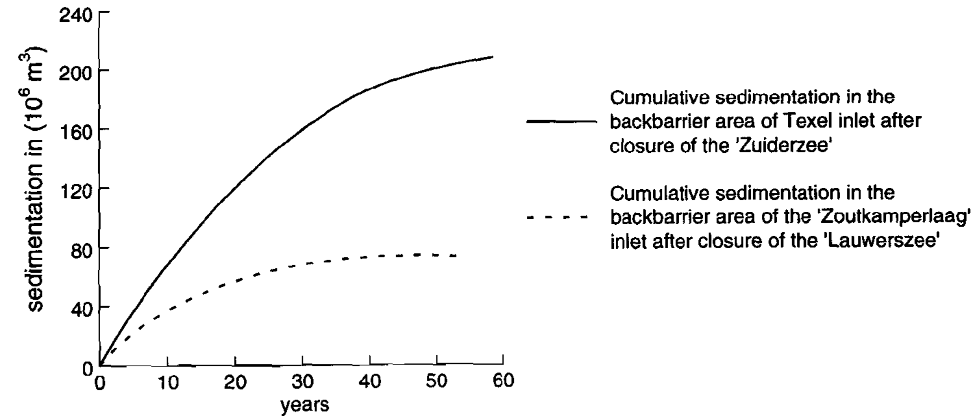


Figure 3: Effects of closures: gradual decreasing, but long-lasting sedimentation (——) after the closure of the Zuider Zee in the backbarrier area of Texel Inlet and (---) after the closure of the 'Lauwerszee' in the Zoutkamperlaag-backbarrier area (Vroom et al., 1989).

enclosure was, and still is, exceptionally large (Fig. 3): old channels changed their direction, while new channels and tidal flats were formed.

The sands in the Wadden Sea are largely composed of quartz (more than 80%) with some feldspar, $CaCO_3$ (mainly mollusc shell fragments (Cadée, in press) and mica (Fig. 4; Van Straaten, 1964). The sand deposited in the Wadden Sea is somewhat finer (Md: 170-190 μm) than the sands along the North Sea coast (usually greater than 200 μm). Due to sorting processes the coarser sand does not pass the inlets. Sorting takes place towards the mainland (Winkelmolen & Veenstra, 1974) and in the direction of the long-shore current. As a result the grain size along the coast of the barrier islands and in the backbarrier areas decreases gradually from W to E within identical environments (Veenstra & Winkelmolen, 1976; Eysink, 1979). Apart from the mineralogical similarity of the sands in the Wadden Sea and those in the adjacent North Sea, the influx from the North Sea is reflected by the occurrence of mollusc shell fragments, sea urchin spines and foraminifera (cf. Van Voorthuysen, 1960).

Clays

The fine-grained material deposited in the Wadden Sea is also brought in from the North Sea (Wiggers, 1960; Van Straaten, 1960). It is a mixture of material from different sources: the Channel, the Flemish Banks, sea-floor erosion, and the river Rhine (Van Veen, 1936). Suspended matter from the Rhine comprises 10-20% of the total supply. The fine-grained sediments consist predominantly of clay minerals, organic matter and $CaCO_3$ (mainly detrital; Fig. 4; Van Straaten, 1964).

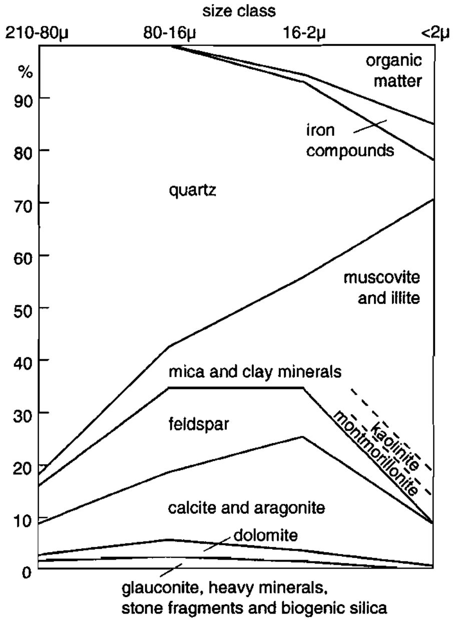


Figure 4: Average composition of Wadden Sea sediments (Van Straaten, 1964).

About 1-3*10^9 kg.yr^{-1} (dry weight) of suspended material is deposited annually in the Dutch Wadden Sea, mainly along the inner margins, near the coast of the mainland, on tidal watersheds and in the Ems-Dollard estuary (Eysink, 1979; Eisma, 1981; Dijkema et al., 1988).

MORPHOLOGICAL ELEMENTS AND SEDIMENTATION PATTERNS

The Wadden Sea is characterized by a series of adjacent tidal inlet systems. Each inlet system consists of an ebb-tidal delta with the inlet proper, parts of the barrier islands adjacent to the inlet, and the backbarrier area. These elements, i.e. the barrier islands, the inlets with their ebb-tidal deltas and the backbarrier areas form the most important morphological units of the Wadden Sea.

Barrier Islands

On its seaward side the Dutch Wadden Sea is bordered by barrier islands (Fig. 5). These are formed by the combined transport action of wind, waves and tides. The wind strength varies seasonally. Average wind velocities reach 15 m.s^{-1} in winter, and 7 m.s^{-1} in summer.

The tidal wave in the North Sea moves from the S(W) to the N(E), that is from Den Helder in Holland to Esbjerg in Denmark (Fig. 2). The vertical range is largest (over 3 m) in the German Bight, because there the distance to the amphidromic point is greater than else-

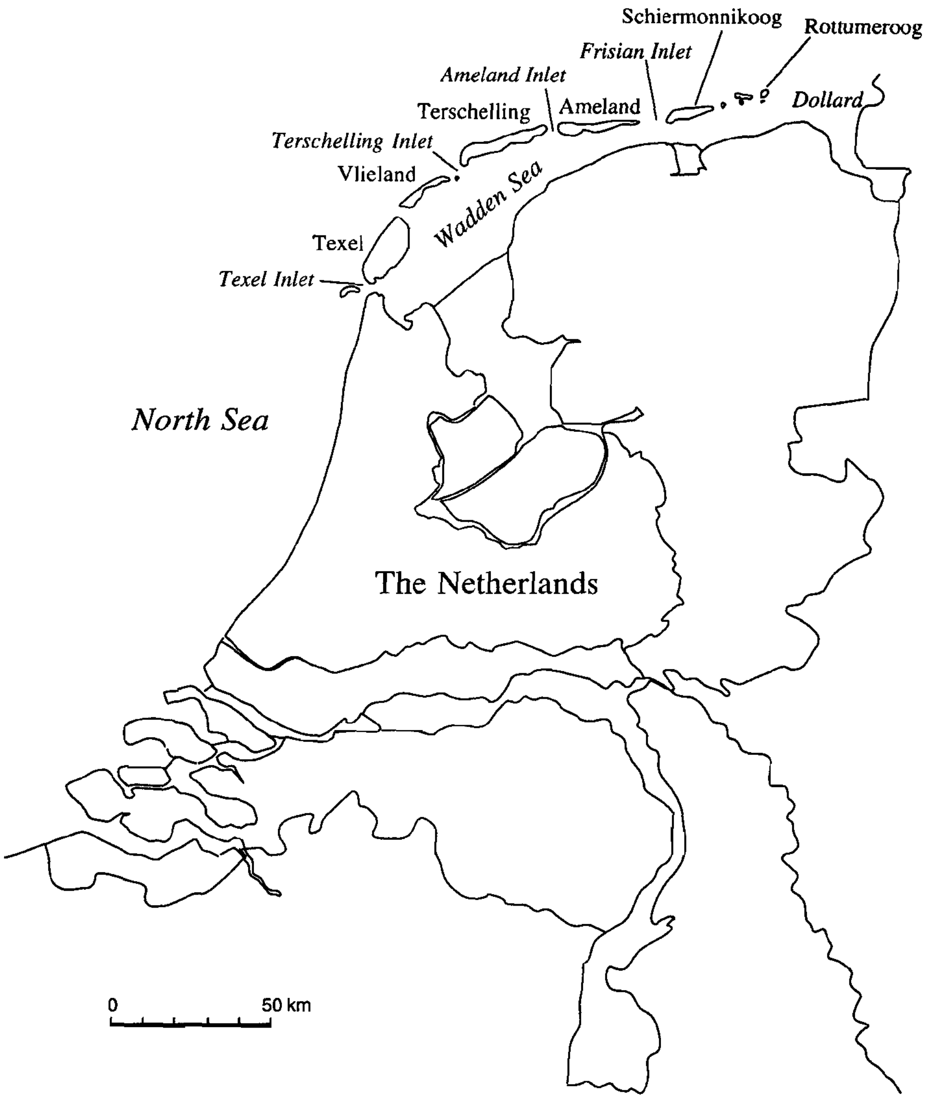


Figure 5: Physiography of the Dutch Wadden Sea (Steijn, 1991).

where along the Wadden Sea coast. Also, the increased friction due to the funnel shape of German Bight increases the tidal range. Smallest ranges are found near Den Helder (1.3 m) and Esbjerg (Dijkema et al., 1980). Hayes (1975) stated that the length of barrier islands depends on the mean tidal range: the larger the tidal range, the smaller the island length.

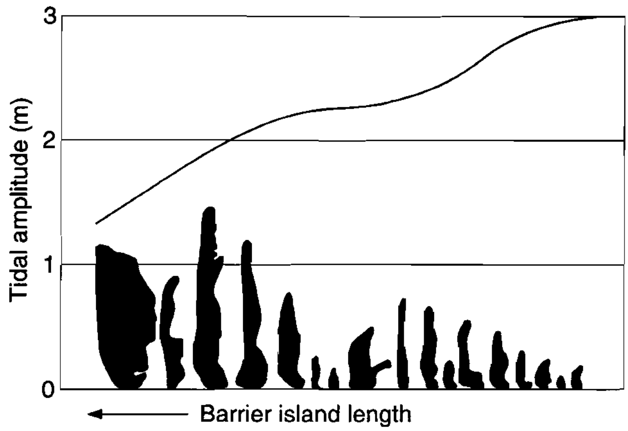


Figure 6: The relation between tidal amplitude and the length of the Wadden Sea barrier islands (Wolff, 1986).

Indeed, for the barrier islands of the Wadden Sea such a relationship exists (Fig. 6; Wolff, 1986). At present the islands do not migrate strongly. This is mainly the result of artificial coastal protection. Moreover, some of the islands and shoals are fixed by erosion-resistant Pleistocene or early Holocene cores.

The barrier islands consist of the following sedimentary subenvironments: the lower and upper shoreface, the intertidal beach along the North Sea coast, the dunes, and the tidal marshes which merge with the back-barrier tidal flats. Under natural conditions the islands are cut by washover channels. Part of the tidal marshes were moulded into polders by the construction of dykes, and washovers were blocked by artificially stimulating the growth of dunes.

Schiermonnikoog provides a good example of the original, natural situation on the barriers, as it has been left relatively undisturbed by man (Fig. 7). To the west the island is bordered by the large Zoutkamperlaag Inlet and to the east by the small Eilanderbalg Inlet. The tidal amplitude near Schiermonnikoog is 2.3 m, and the regime thus is mesotidal (Postma & Dijkema, 1982). In winter waves are highest, with a mean wave height of 2 m at 20 m water depth. The coast of this barrier island is thus classified as a mixed energy, tide-dominated shoreline, influenced by both tides and waves (cf. Hayes, 1975, 1979). The island has the typical 'drumstick' form, which is characteristic for such mesotidal coasts (Hayes, 1979).

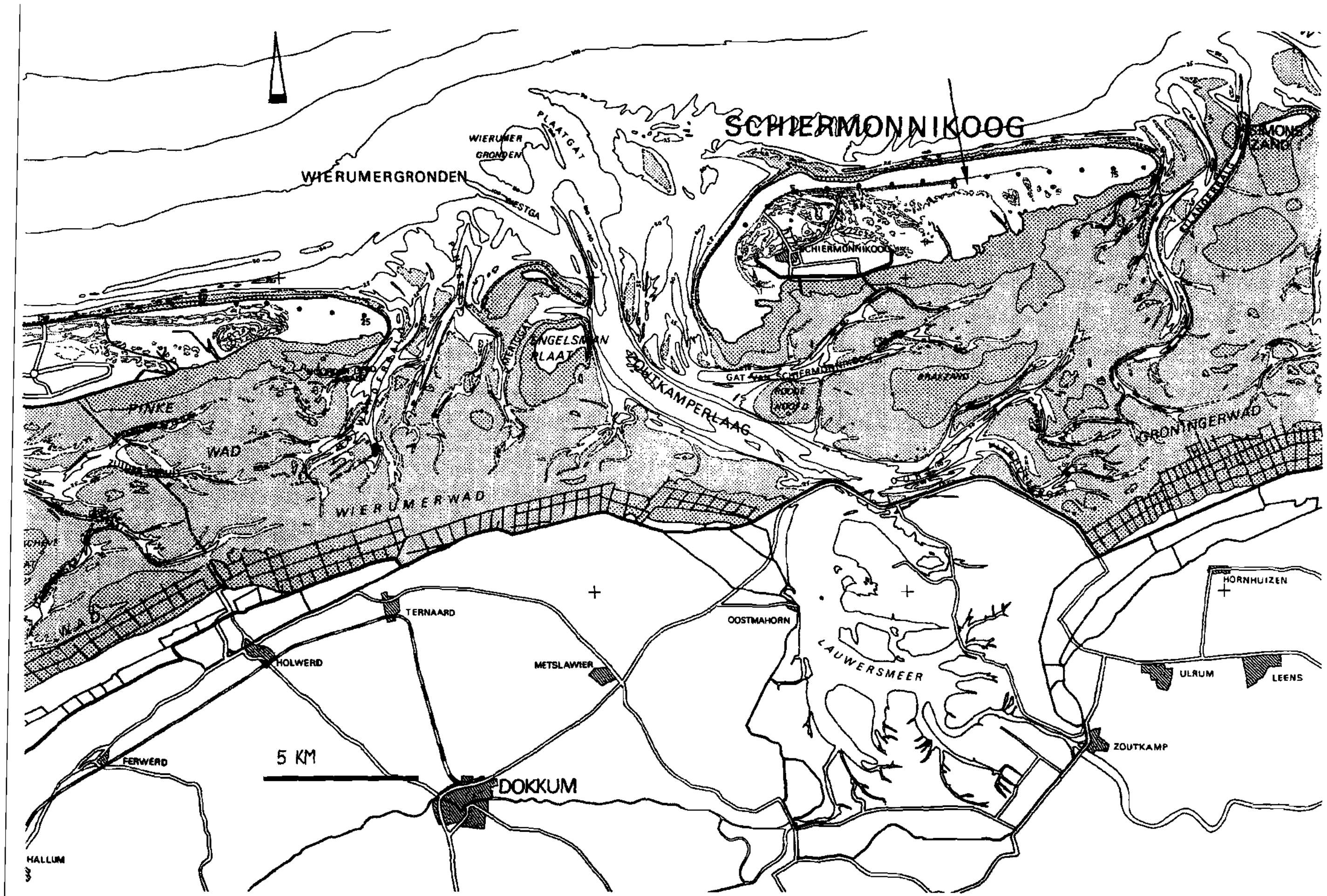


Figure 7: The Zoutkamperlaag and the barrier island of Schiermonnikoog. Arrow shows position of washover channel.

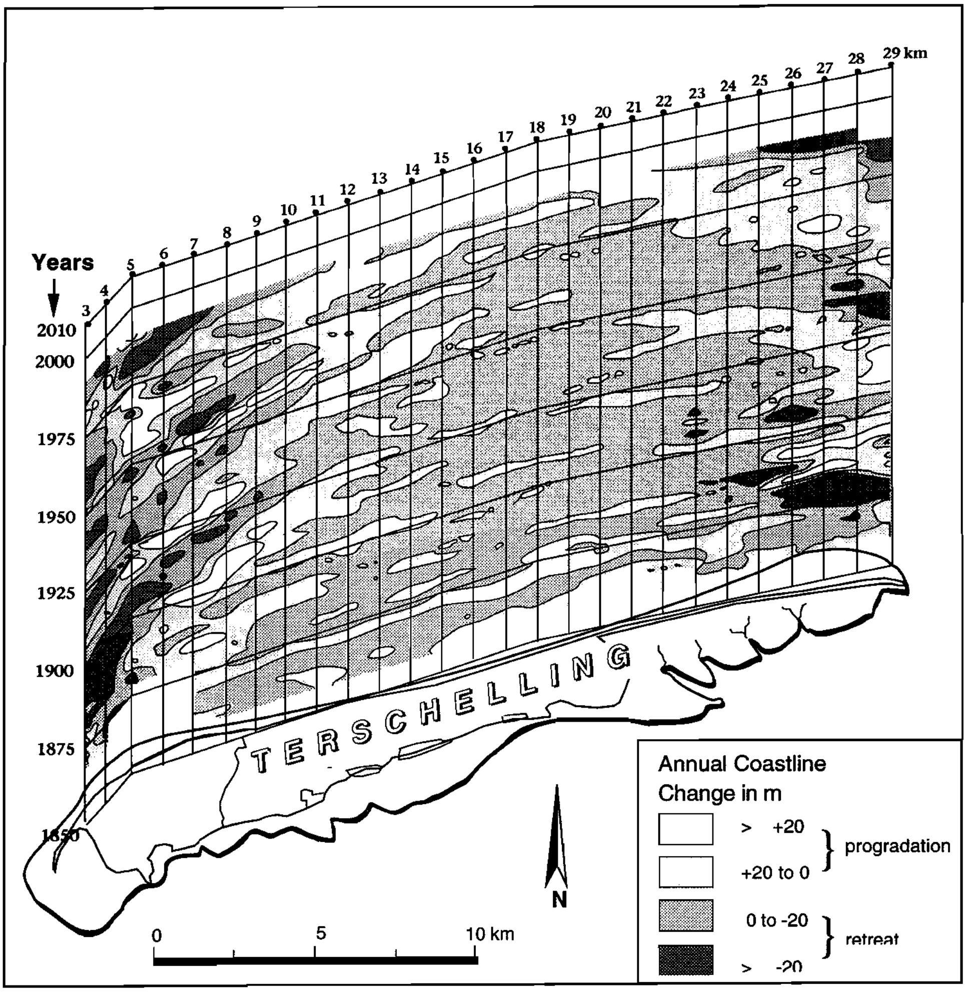


Figure 8: Patterns of erosion and sedimentation as reflected by coastal retreat and progradation in $m.yr^{-1}$ along the North Sea coast of Terschelling (Oost et al., 1993).

The North Sea barrier island coast

For a full discussion of the lower and upper shoreface the reader is referred to Flemming & Davis (1992, 1994). Sedimentation and erosion on the intertidal beaches of the islands is mainly brought about by wave action, coast-parallel, eastward directed residual currents (tidally and wind-induced) and wind-driven aeolian processes (the latter occurring especially above the high-water mark).

Over a longer time-span, sites of erosion and sedimentation shift in position along the North Sea barrier coast from the adjacent inlets towards the centre of the island in a wave-like mode (Fig. 8). Loci of erosion and sedimentation move slowly and migration of an erosion/sedimentation wave from the inlets towards the centre of the islands covers periods of 30 to 70 years. Such patterns largely dominate sedimentation and erosion along the island coasts and show significant periods of 18 to 19 years (Oost et al., 1993).

On an annual scale, higher frequency sedimentation patterns can be observed. During the summer a coastal profile is built up, which is partly destructed during the stormy winter season (Fig. 9). The development and position of beach ridges and the locations where rip-current channels develop is largely determined by wave action on the sandy coast. The beach ridges are formed by breaking waves during the high tide and the swash during low tide.

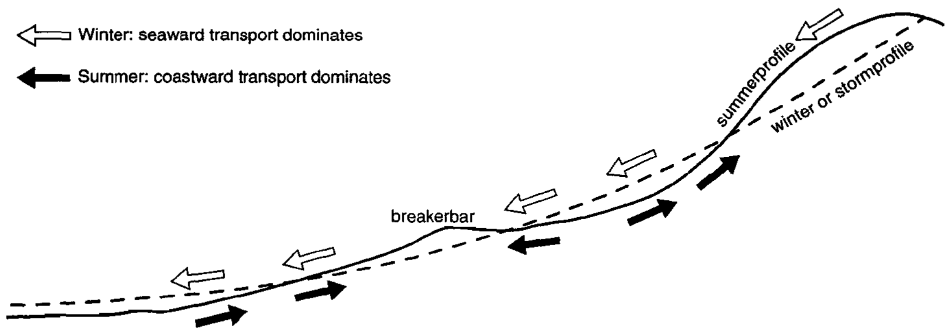


Figure 9: Schematic beach profile: build-up in summer (---), destruction in winter (—).

Sediment transport along the beach takes place in three zones (Fig. 10):

1) The swash zone on the berm during high water,

2) The zone on the ridge, where sand brought in by waves is transported longshore, via the runnel by the long-shore current and subsequently brought out by the rip current. Not much sediment is exchanged between zones 1 and 2, except during the winter when the beach profile is partly eroded (Bleuten, 1971),

3) The aeolian zone where, depending on the wind force and direction, large amounts of sediment can be transported, can extend to the low-water line. In the intertidal zone a sediment layer several cm thick can be eroded or deposited by the wind during the period between two successive high waters. Calculations show that aeolian transport is probably as important as coast parallel transport (Steijn, 1991).

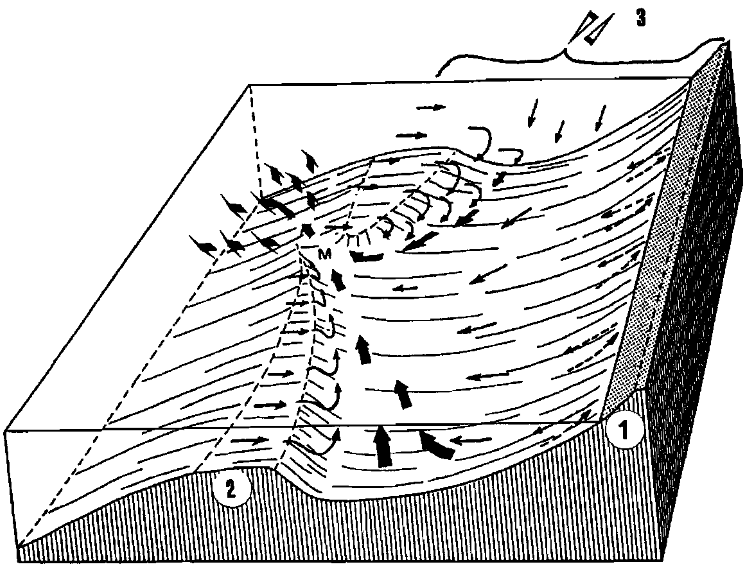


Figure 10: Sediment transport pattern on and along a barred beach. The triangular arrows indicate aeolian transport in various directions, depending on the wind direction (zone of aeolian transport at low tide indicated). All other arrows indicate directions of sediment transport by water at high tide. For explanation of numbers see text. After Bleuten (1971).

Sand transport rates within these zones are high. Over a period of only 8 days with one weak storm (26-8-1970 to 3-9-1970):

1) a landward shift of the berm in the period from neap- to springtide,

2) a net eastward sand transport of 70 m^3 on the beach ridge during one storm period (i.e., a net sedimentation of an average surface layer of 2 mm thickness during one storm),

3) a flattening of the berm profile and an eastward shift of the rip current channel during the storm were observed within an area of 150*200 m^2 on the barrier island Schiermonnikoog (Fig. 11) by Bleuten (1971).

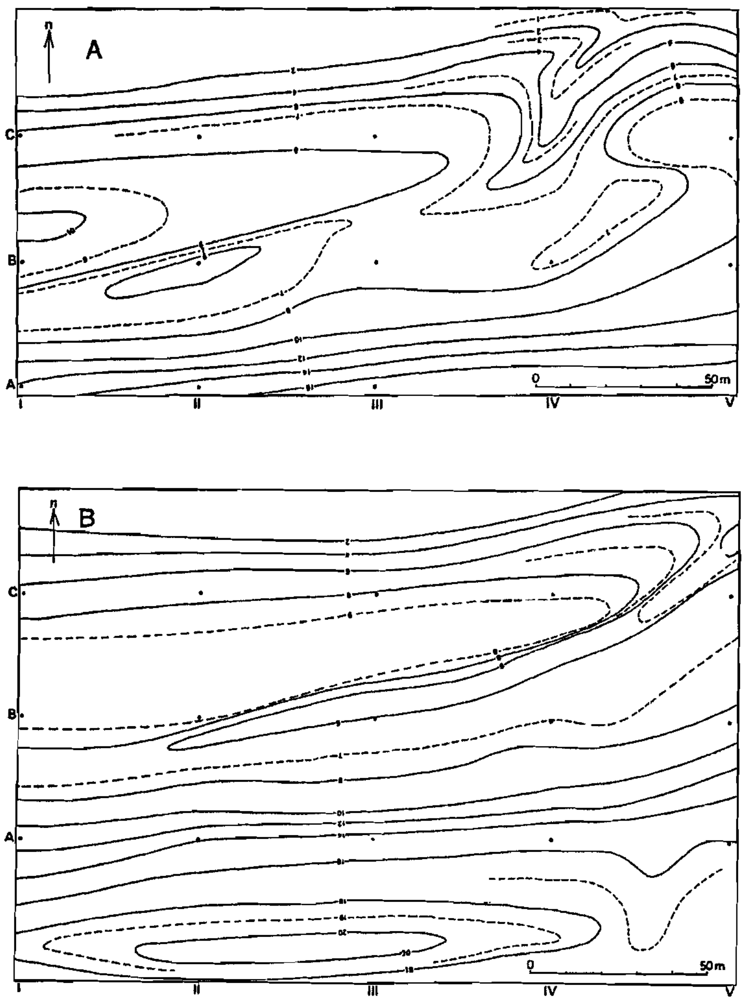


Figure 11: Development of a beach section on Schiermonnikoog. Elevation in dm. A. Situation on 26 August 1970; B. Situation on 3 September 1970 (Bleuten, 1971).

The great variety of sedimentary structures on a ridge and runnel dominated beach plain is discussed in detail by several authors (e.g. Davis et al., 1972). It should be realized, however, that sedimentary structures in intertidal sandy deposits will not always be preserved. Air filled cavities, up to 1 cm in diameter, are formed within fine-grained sand; at rising tide the ground water level does not rise quickly enough to replace the interstitial air before flood water covers the sediment surface. Air is trapped in the sand between the ground water level and the sediment-water interface. Because of the weight of the overlying water column and the slow downward movement of water due to capillary action, the entrapped air is compressed and eventually may attain a pressure which enables it to lift a thin layer of overlying sediment. Then air cavities (cavernous sand) are formed, often to a depth of 20 cm (P.L. de Boer, 1979), by reorientation of grains with flat surfaces (Wunderlich, 1979). The cavities are sustained by the friction of the grain fabric and the surface tension of the sand-water-air contacts. Observation and experiments show that, in layers of fine sand, bubbles develop preferentially in better sorted and coarser zones. This is probably because capillary forces are greater in finer-grained and less sorted sand. Thus, water will penetrate by preference into the latter, pressing the interstitial air into the better sorted and coarser sand. The high content of air cavities in certain layers can result in a complete homogenisation of the structures in such layers. It can also provide a density instability responsible for deformational processes which can lead to the formation of convolute laminations. In the intertidal zone this appears to be a slow process that covers a number of ebb and flood cycles. Such structures can be formed in all intertidal areas where some relief is present and may contribute to the instability of, amongst others, channel and gully walls.

Dunes

Along the North Sea side of most islands an almost continuous dune belt separates the beach from the rest of the island. The dunes are partly natural and partly man-made. Part of the

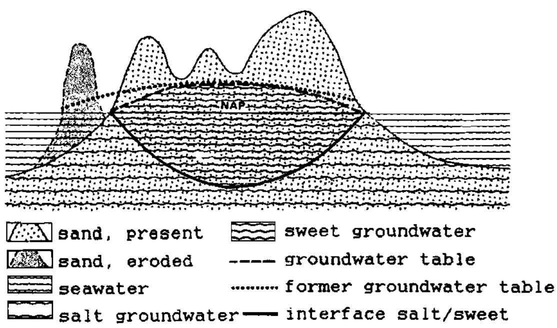


Figure 12: Freshwater lens in the dunes (Van Oosten, 1986).

dunes on Schiermonnikoog, for example, have been formed as an artificial aeolian sand dyke, which was constructed by trapping sand blown inland from the beach between and behind sticks and branches stuck upright into the ground. Other parts of the dunes are natural, forming large parabolas which have developed in different periods.

On the island Schiermonnikoog, which is a National Park, the vegetation of one dune area to the NW of the village is absent and a natural development is allowed. This dune area consists of actively moving dune complexes, with areas locally eroded down to the ground water table. The ground water table depends, amongst others, on the size of the dune complex (Fig. 12) and on the sea level (Van Oosten, 1986). Under natural conditions, aeolian deposits below the ground water table are preserved from erosion. A rise of sea level and an increase of the surface area of dune complexes therefore favour the preservation potential of dune deposits.

Tidal marshes and washover channels

On most Dutch Wadden Sea islands washover channels transect the dunes where these are not protected by man, i.e., especially at the east side of the islands. During storms, water and sand are transported through the washover channels from the North Sea into the backbarrier area or into small embayments on the islands. In the natural parts the higher intertidal zones behind the dunes consist of extensive tidal marshes, whereas the protected parts mainly comprise polders. For a discussion of the salt marshes, see below.

Again, Schiermonnikoog is one of the most natural examples. At present, the eastern part of the barrier island is largely undisturbed by man (Fig. 7). In the period 1950-1978 a sand dyke was constructed by stimulating aeolian deposition in the rear of the northern part of Schiermonnikoog. This artificial dune blocked the southern part of a broad beach plain with some isolated dunes. As a result salt marsh vegetation developed on top of the sandy beach deposits. Towards the North Sea, the salt marsh sediments wedge out. Originally the artificial dune extended further to the east, but part of it was destroyed by storm floods. The government repaired the dunes several times, but stopped doing so after 1984, so that washovers became active again. The channel at the eastern side of the island, near beach marking point No. 10 (Fig. 7), is an active washover channel during storms, especially in the winter season. Through such washover channels the whole outer-dyke tidal marsh area of Schiermonnikoog is flooded several times a year. The channels are flanked by sandy, vegetated levees which are locally more than 0.5 m high. They consist of North Sea beach sands brought in during the floods. These sandy deposits pinch out laterally and interfinger with fine-grained silty and clayey overbank deposits, which have settled from the seawater that stands above the tidal marsh for a number of days after floods. During the more quiet summer season the washover channel is partly filled and blocked at the North Sea side by aeolian deposits such as sand sheets, aeolian ripples and small dunes.

Since repairs of the artificial dune have been stopped, the washover channel has become deeper and has extended gradually towards the North (Fig. 7). This has led to a slight panic among the island inhabitants, who became afraid that the washover might become an active channel, splitting the island into two parts. However, the washover channel enters the Wadden Sea close to the tidal watershed, thus greatly diminishing the risk of it evolving into a tidal inlet channel (cf. Ehlers & Kunz, 1993).

The washover channel near beach marking point 10 ends in the Wadden Sea on the tidal watershed south of the island (see arrow in Fig. 7). The washover-lobe deposited there consists mainly of sand. Biota flourish in this quiet part of the backbarrier area and bioturbation is intensive. Lugworms and burrowing bivalves rework the bottom sediment continuously. Especially at the turn of the high tide millions of shrimps, crabs and flat fish destroy the small-scale ripples, which form during the ebb and the flood. At low tide birds visit their feeding grounds to prey for invertebrates, thereby strongly bioturbating the tidal flats (Hertweck, 1982; Ehlers, 1988; Cadée, 1990). In places where the diatom/algal mats are destroyed, erosion is stronger and may result in deep erosion scours. Such bioturbation, in combination with intensive reworking by waves and currents results in a complete destruction of the washover delta-lobe structures.

Ebb-tidal deltas and inlets

Ebb-tidal deltas

Ebb-tidal deltas are situated on the seaward side of the inlet throats between neighbouring barrier islands. They are the result of sedimentation by the tidal currents flowing through the inlets, wave action and the tidal wave along the coast (Oertel, 1975; Hayes, 1975, 1979, 1980; Hubbard et al., 1977; Nummedal et al., 1977; Nummedal & Fischer, 1978; FitzGerald et al., 1984; Sha, 1989a-d, 1990a, b; Sha & de Boer, 1991). Mesotidal regimes in particular are considered to favour their evolution and maintenance (Hayes, 1979, 1980), although it should be noted that tidal prism rather than tidal amplitude defines the strength of the tidal currents within and adjacent to tidal inlets (Sha, 1989c). The tide-induced currents and the morphology of ebb-tidal deltas are strongly interrelated (Sha, 1989a, 1990a). A general empirical relation quantifying the sand volume of ebb-tidal deltas (above the profile along the barrier island coast) was proposed by Dean & Walton (1975), Walton & Adams (1976) and Steijn (1991) as:

$$V=65.6*10^{-4}P^{1.23}$$

where V is the sand volume of the ebb-tidal delta and P is the mean tidal prism. The ebb-tidal delta volume decreases with increasing wave influence (Fig. 13; Dean, 1988). The Dutch ebb-tidal deltas plot on the line of strong wave action (Eysink & Biegel, 1992). They consist of large volumes of sediment, which are of the same order of magnitude as the adjacent barrier islands and thus play an important morphological role in the Wadden Sea system.

Studies of tidal deltas have revealed that the morphologies of ebb-tidal deltas in different geographic places are quite similar and generally fit well to models of the kind as, for example, given by Oertel (1972) and Hayes (1980). Typical components of an ebb-tidal delta

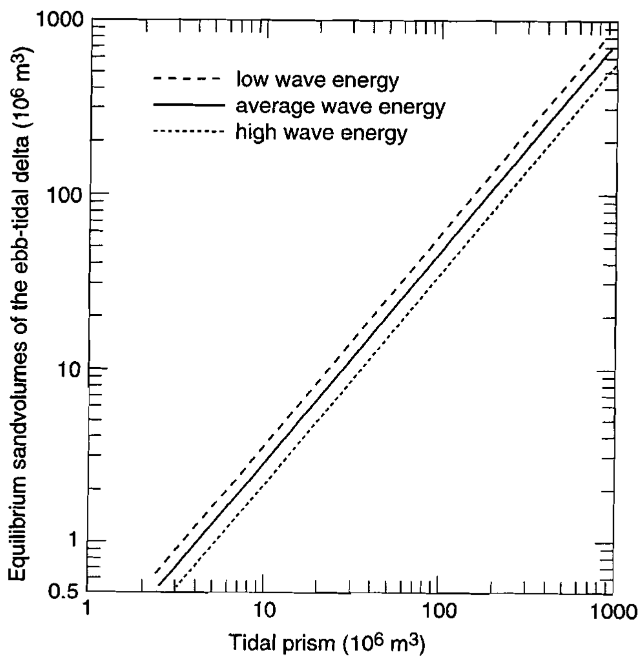


Figure 13: Relationship between the sand volume of the ebb-tidal delta (with reference to the normal barrier coast), the tidal prism and the influence of wave climate. After Dean (1988).

are the main ebb channel, marginal flood channels, channel margin linear bars, swash platform and swash bars (Fig. 14).

Inlets

The barrier islands are separated by tidal inlets, through which the tidal water passes to and from the backbarrier area in the course of a tidal cycle. Inlets consist of one or more main channels separated by sandy shoals. The strongest currents in barrier systems generally occur in the inlets (and in the larger channels of the backbarrier area). Tidal current velocities increase from the open North Sea towards the inlet throats, where they reach values of 1-2 $m.s^{-1}$ at maximum. Only during storms, higher velocities have been measured. Empirical relations have been proposed for the cross-sectional area in the throat of the inlet and the tidal prism (e.g. O'Brien, 1969; Jarrett, 1976; Dieckmann et al., 1988; Gerritsen, 1990; Hume & Herdendorf, 1990; Niemeyer, 1990; Sha, 1990a; Steijn, 1991; Biegel, 1991b; Eysink & Biegel, 1992).

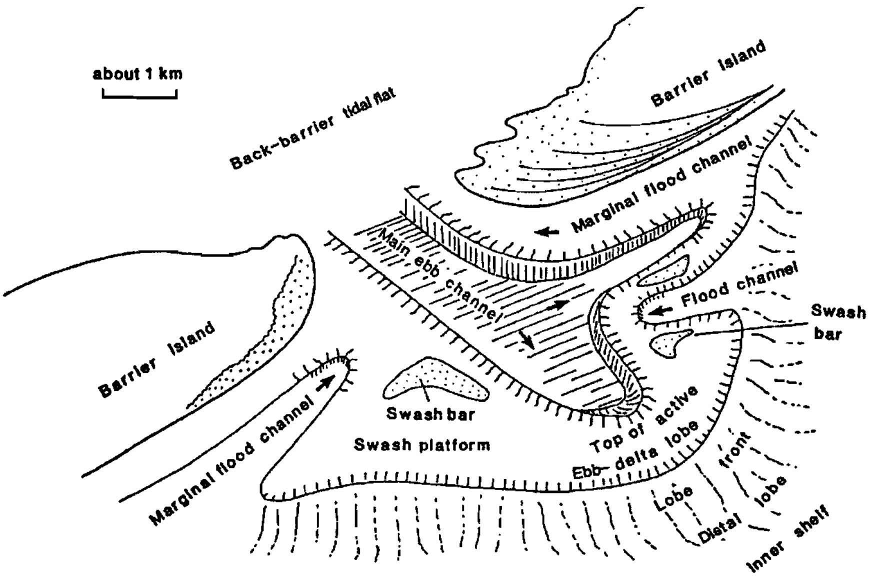


Figure 14: Morphology of a typical ebb-tidal delta. Arrows indicate direction of dominant tidal currents (Sha, 1990a).

For 162 inlets along the American coast the general empirical relation in metric units, with or without jetties is (Jarrett, 1976):

$$A_c=158*10^{-6}P_s^{0.95}$$

where A_c is the cross-sectional area and P_s is the tidal prism at spring tide. Dieckmann et al. (1988) showed a similar relation based on 26 tidal inlets along the Wadden Sea:

$$A_c=372*10^{-6}P^{0.915}$$

where A_c is the cross-sectional area and P the mean tidal prism.

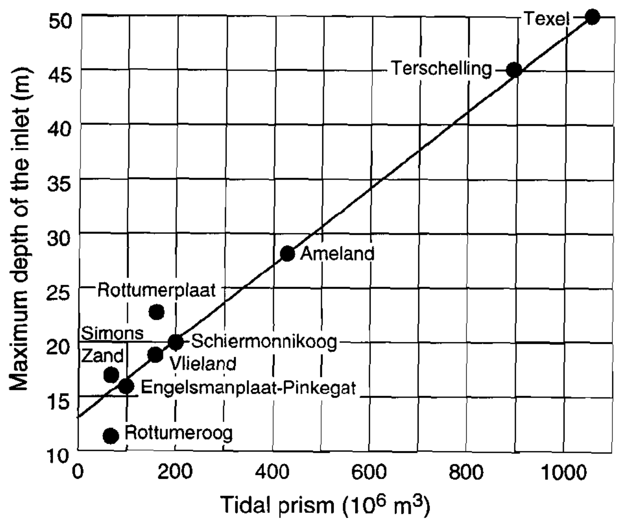


Figure 15: Relationship between the maximum depth of an inlet channel and the tidal prism (Sha, 1990a)

The main ebb channel is mostly the deepest element in the ebb-tidal delta because the maximum velocities during the ebb commonly occur at lower water levels than during the flood (Van Straaten, 1964). Sha (1990a) demonstrated that, for the Dutch Wadden Sea, the maximum depth of the main inlet channel is linearly related to the tidal prism (Fig. 15), with the relation:

$$h=13.3 + P/30*10^6$$

where h is the maximum depth (in m) and P the mean tidal prism (in m^3). Flemming (1991) found a relation between the local catchment area and the width of the inlets for the East Frisian Wadden Sea:

$$A=37.67\ L^{0.8}$$

where A is the catchment area (in km^2) and L is the inlet width (in km). Relations between tidal prism and maximum depth of inlets have also been found in other tidal areas (Dieckmann et al., 1988; Hume & Herdendorf, 1990; Eysink & Biegel, 1992). The rather large scatter in the data from different areas suggests that universally applicable relationships between channel widths and mean or maximum depths as a function of tidal prism cannot be given. The quantities probably depend strongly on local flow and bed conditions (Eysink & Biegel, 1992), as well as on the intensity of wave action.

The ebb current is concentrated in the main ebb channel, draining the backbarrier in form of an ebb-jet. The ebb channel is flanked on either side by levee-like marginal linear bars, formed by the interaction of ebb and flood currents in combination with waves and wave-generated currents (Hayes, 1976; Sha, 1989b, 1990a).

The flood waters enter the inlet over a large front, and together with wave-induced currents (wave-pump-concept; Bruun & Viggosson, 1977) they transport sediment into the backbarrier area. The asymmetry of tidal current velocities influences the development of tidal deltas. When the water level starts rising after low water, the ebb current in the main ebb channel is still flowing strongly, forcing the flood currents into the marginal flood channels. The main channel is thus ebb-dominated and the marginal channels are flood-dominated, the latter being located laterally of the main ebb channel, adjacent to the barrier islands and between the swash platforms (cf. Hayes, 1976).

The swash platforms are broad sheets of sand topped by isolated swash bars which are built by wave action. If such bars grow high enough, as in case of the shoal 'Noorderhaaks/-Razende Bol' in the Texel Inlet, aeolian processes may contribute to the accumulation of sediment. As a result a new barrier island may develop, or, more commonly, the platform may migrate and connect to the adjacent downdrift barrier island (e.g., Texel: Sha, 1989a, 1990a). Oertel (1972) pointed out that swash platforms are generally located along the axis of, and on the highest parts of onshore migrating shoals, parallel to the direction of the wave-bore approach. The orientation of swash bars on the platform is generally perpendicular to the incident wave-bore.

Geological and historical developments show that part of the Dutch and German inlets tend to move in the direction of the resulting sand transport along the coast, i.e., towards the east (Fig. 16; Luck, 1975; Ligtendag, 1990; Sha, 1992; Oost & Dijkema, 1993; Chapter 2). This migration is not continuous through time and space, but is influenced by the underlying palaeotopography (Sha, 1992) as well as the morphological developments in the backbarrier area (Flemming & Davis, 1992; Oost & de Haas, 1992; Van der Spek, 1995), by the variations in tidal amplitude (Sha, 1990; Flemming & Davis, 1992; Oost et al., 1993) and by variations in sea-level rise. Eastward directed sediment transport occurs in the ebb-tidal delta mainly as a result of the tide and waves, by migration of channels and shoals and by sand transport through the channels (Oost & de Haas, 1992). Many inlets have today been artificially stabilized.

Morphological patterns: asymmetry and orientation

The larger ebb-tidal deltas of the Dutch barrier coast have an updrift asymmetry, which is greatest in the ebb-tidal delta of Texel Inlet (tidal prism about $1050*10^6$ m^3) and the one of Terschelling Inlet (about $850*10^6$ m^3). Tidal prisms through inlets with downdrift oriented ebb-tidal deltas are considerably smaller.

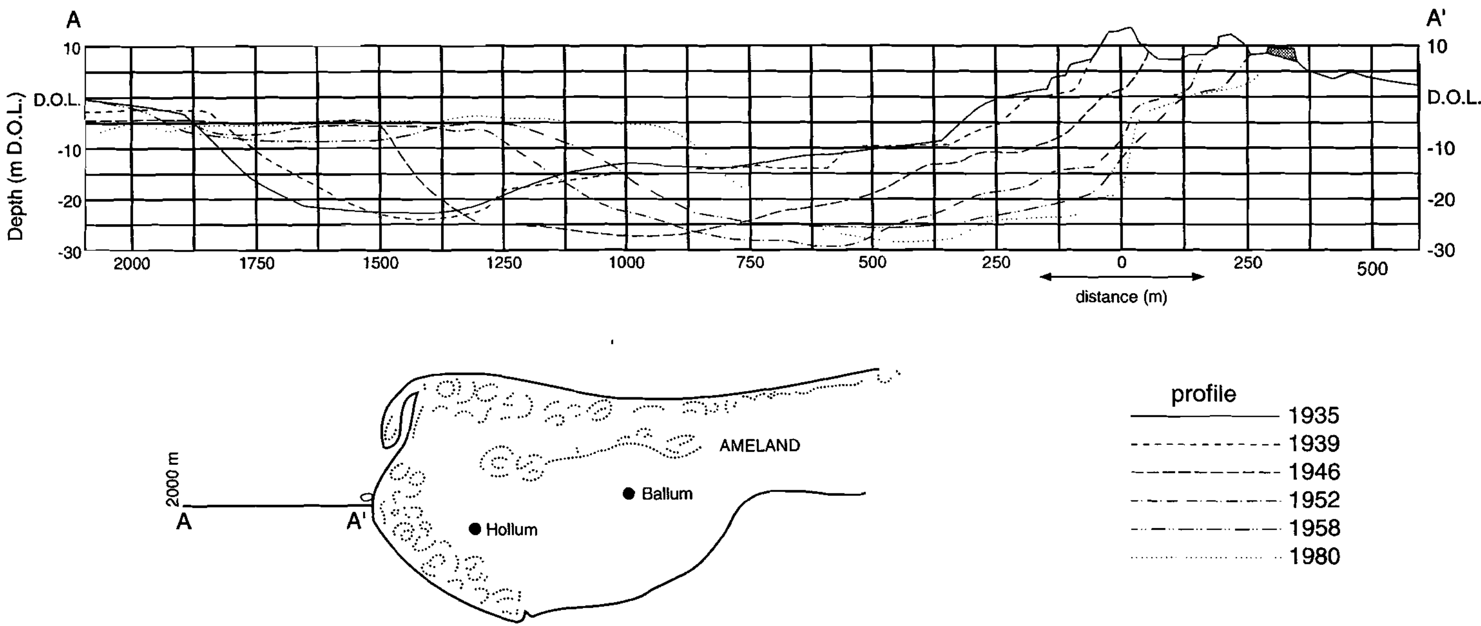


Figure 16: Lateral shift of the Ameland inlet, situated to the west of Ameland. After De Boer et al. (1991).

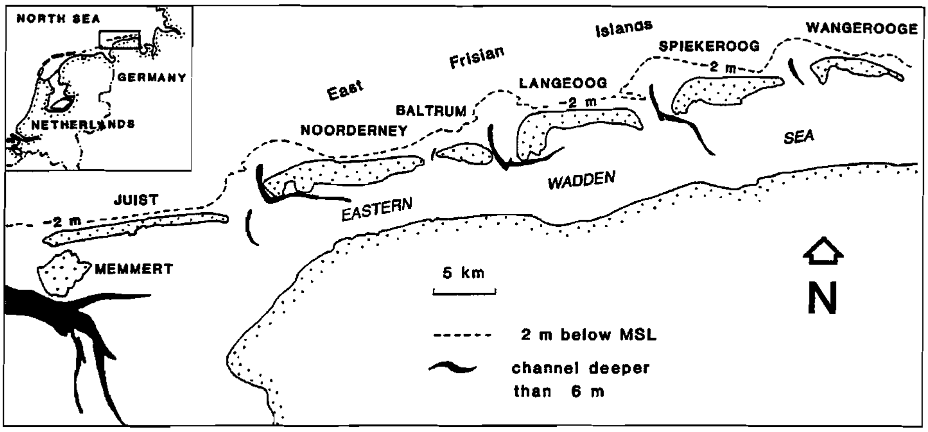


Figure 17: Location and shape of the ebb-tidal deltas along the East Frisian Islands. After FitzGerald et al. (1984). Note that the ebb-tidal deltas and their main ebb channels are asymmetrical to the east (Sha, 1989c).

The asymmetrical orientation of smaller ebb-tidal deltas along the Wadden Sea coast and the downdrift (to the east) orientation of their main ebb channels (Fig. 17) is commonly ascribed to the fact that wind-driven waves dominantly attack the ebb delta from the west. Together with the residual tidal current they produce an easterly longshore drift which forces a downdrift, i.e. eastward directed asymmetry on the ebb-tidal deltas (Luck, 1976; Nummedal & Penland, 1981; Sha, 1990a).

The updrift orientation of the larger ebb-tidal deltas is due to the influence of tidal currents which pass through the inlet, as is illustrated in Fig. 18. The morphodynamic character and orientation of ebb-tidal deltas is thus defined by the relative influence of waves from the open sea and tidal currents through the inlets. Waves and longshore drift tend to force the ebb-tidal deltas into a down-drift asymmetry, whereas the tidal wave tends to force the ebb-tidal delta into an updrift asymmetry. Because of the relatively large influence of waves and longshore drift, most of the ebb-tidal deltas of the Wadden Sea face to the N and NE, i.e., they are downdrift-asymmetrical. Only inlets with a relatively large tidal prism, such as the one of Texel, are oriented updrift (Sha, 1990a; Fig. 18).

Texel Inlet

A large-scale study of depositional character and sediment transport in the ebb-tidal delta of Texel Inlet was made by Sha (1989a-d, 1990a, b). He found a flood-dominated net sediment transport through the minor tidal channels and over the shoals (Fig. 19A), and an ebb-dominated net sand transport in the larger and deeper channels. Sand, which is transported to the

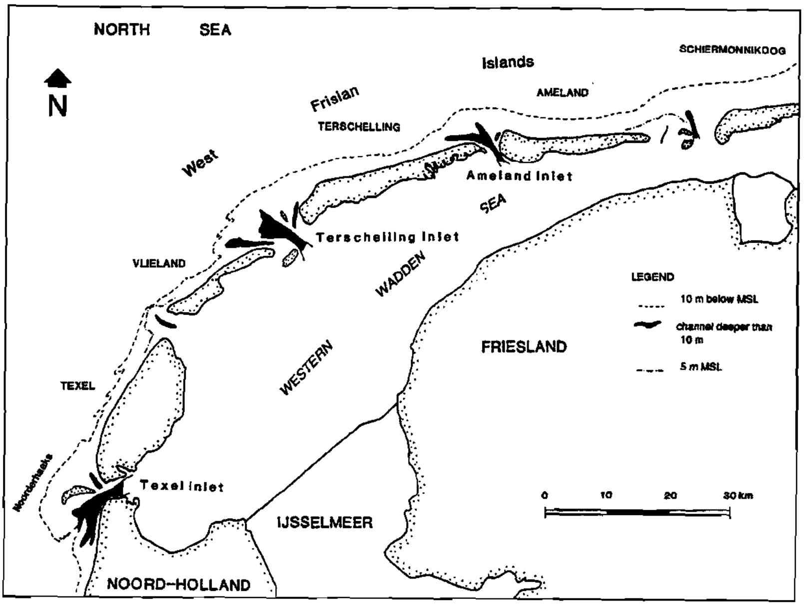


Figure 18: Location and shape of the ebb-tidal deltas of Texel Inlet (Marsdiep), Terschelling Inlet (Vliestroom) and Ameland Inlet (Borndiep). The -10 m depth contour outlines the position and shape of the ebb-deltas; black areas are channels within the ebb-tidal deltas deeper than 10 m below MSL (Sha, 1989c). The IJsselmeer is the former Zuider Zee which was blocked by the 'Afsluitdijk' in 1932.

seaward margin of the ebb-tidal delta through the southwest-directed main ebb channels, is carried north by flood-dominated residual currents along the ebb delta front. Landward sediment transport dominates the northern part of the ebb-tidal delta. Sand is also deposited north of the ebb-delta shoal 'Noorderhaaks' owing to weak, rotational tidal currents (Fig. 19B). From here the sand is transported shoreward by waves in the form of swash bars. Part of the sand returns to the inlet through the flood-dominated channel in the north (Fig. 19A). The lower and higher intertidal parts of shoals between the major channels are dominated by (storm) waves and wind. The larger subtidal (parts of the) shoals are dominated by waves and have swash bars superimposed. Flood-dominated channels to the NE of the shoals tend to silt up and are abandoned upon attachment of the shoals to the island in the NE (Sha, 1990a).

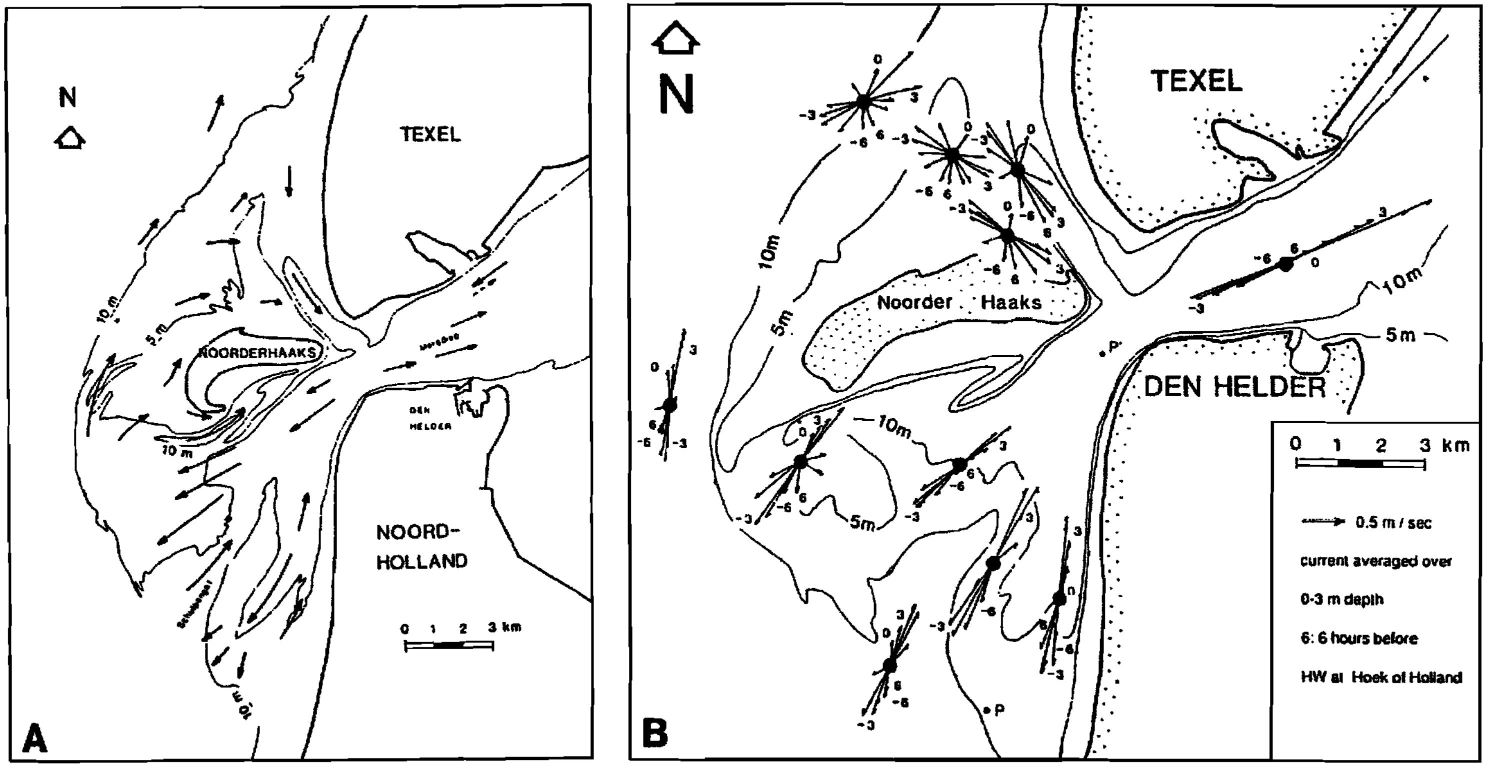


Figure 19: A. Schematic current and sand transport pattern in the ebb-tidal delta of Texel Inlet. Arrows indicate the net sand transport directions. B. Tidal current roses for a tidal cycle in the ebb-tidal delta of Texel Inlet integrated from the "Stroomatlas van Nederland" (M.M.A.H., 1963). Time near current roses indicates hours before or after high water at the Hook of Holland. Arrow length represents current velocities, averaged from the water surface to 3 m below the surface (Sha, 1989a).

Cyclic morphological changes of positions and orientations of shoals and channels

An important feature of ebb-tidal deltas along the Wadden Sea coast is a clockwise rotation and translation of channels, together with the onshore movement of lower and higher intertidal shoals (Johnson, 1919; Sha, 1989a, 1990a; Oost & de Haas, 1992, 1993). Each cycle, covering a variable number of decades, begins with the development of a new main (ebb)channel, or by the transformation of a flood-dominated channel into an ebb-dominated one, which then gradually translates and rotates clockwise until it degenerates into a marginal flood channel, while a new main ebb channel develops in the updrift direction.

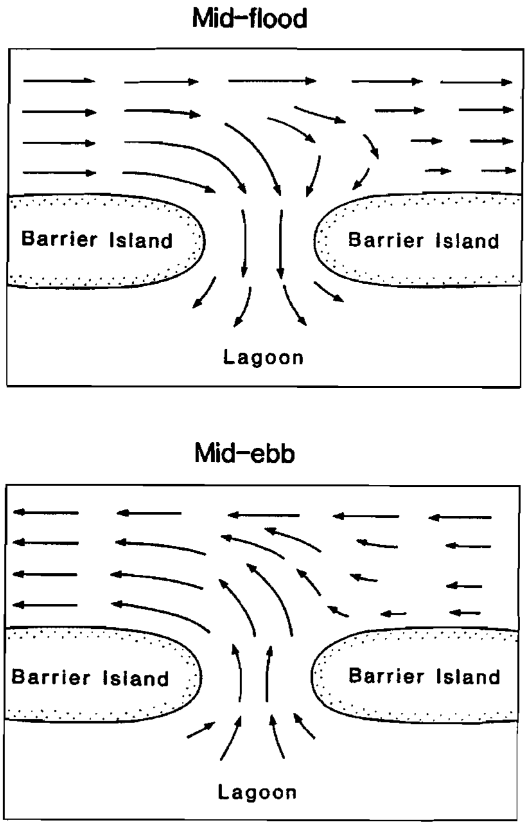


Figure 20: Schematic current patterns at mid-flood and mid-ebb showing the effect of the interaction of tidal currents parallel to the coast with the tidal currents through the inlet. Strong, bidirectional currents are produced on the left (SW) side of the ebb-tidal deltas and a weak, rotational tidal current pattern occurs on the right (N/NE) side. See also Fig. 19B (Sha, 1989c).

This process can result in fast migration of the channels: on the island Schiermonnikoog two light houses built in 1853/54 were positioned in such way that, if kept in one line, one could safely enter into the Zoutkamperlaag from the North Sea through the main outer channel. However, the constructors did not anticipate the fast migration of the outer ebb-delta channels, so that the light-line became obsolete soon after it had been built. The southern light house is today used as a water tower.

Upon rotation and northward migration of tidal channels, shoals tend to attach to the downstream island. The repetitive attachment of shoals contributes to the drumstick-form of the islands, i.e., being thin at the eastern head and much broader at the western head (cf. Fig. 18).

Such cyclic morphological changes are explained by the tendency of the system to be gradually pushed to a more downdrift orientation (Sha & de Boer, 1991) by wave action and longshore currents. In this way the length of the main (ebb) channel and the pathway of the tidal currents entering and leaving through the main inlet channel increases (cf. Fig. 20). As a result, the tidal currents through this channel become weaker, and when the length of ebb (and flood) currents (of decreasing strength) through the main ebb channel reach a certain limit, the channel is transformed into a marginal flood-dominated channel, an updrift channel taking over the role of the main ebb channel. As a consequence, other flood-dominated channels further downdrift diminish in importance, wave and storm attack pushing the ebb delta shoals towards the island across the abandoning downdrift marginal flood channel. In this process the channels rotate clockwise, thereby producing characteristic, one-directional lateral migration patterns in the depositional record.

Sedimentary facies and sequences

The character of surface sediments in ebb-tidal deltas is largely defined by the local tidal and wave energy, water depth, and location with respect to ebb and flood channels. Large amounts of sand are transported through the ebb- and flood-dominated channels. Coarsest sediments are found in ebb-dominated channels which are deeper than flood channels because they are fewer in number and are fed by fast-flowing water draining from the back-barrier area through the relatively narrow inlet throat during the ebb phase. Into the backbarrier, in the throat of the inlet, flood-dominated dunes have been observed in Texel Inlet (Fig. 19a; Sha, 1990a). Such flood dominance is enhanced by the asymmetry of the tidal wave in the North Sea and the low percentage of intertidal shoals in the drainage basin (Sha, 1989b; Steijn, 1991). Typical channel sequences are mostly characterized by ebb-dominated channel deposits (large-scale dunes) in the deeper parts and increasingly smaller and more flood-dominated structures higher up in the sequence. Upon abandonment of channels, as in the case of shoal attachment to the downdrift island, the development of the (fining-upward) channel sequence will show a break, with bioturbated organic-carbon-rich mud and intercalated fine-grained sandy ripples forming the top of the sequence. Between the channels, swash bars form relatively shallow zones where waves are dominant and tidal currents are weak. Due to the strong wave action, the sands are clean and well sorted. If such shoals build up to the intertidal level, wind may cause further vertical accretion, thus producing supratidal shoals that are dominated by beach processes and aeolian transport.

The ebb-delta lobes represent the terminal lobes of the ebb-tidal channels, i.e., accumulations of sand in the area where the ebb-tidal channels loose their transport capacity due to increasing depth and width. The active shallower parts of the lobes are affected by waves and tidal currents, while the more distal parts are finer grained and influenced by longshore tidal currents, rather than by waves and tides which pass through the inlet. In these distal parts of the ebb-delta lobes, storm deposits and bioturbation structures are preserved. Due to lateral channel migration, parts of the ebb delta may become abandoned and subject to wave reworking and erosion. Hummocky cross-bedding and shell lags may develop in the abandoned parts, bioturbation structures being abundant because of the low rate of sedimentation and erosion.

For a more detailed description of the distribution of surface sediments in the ebb-tidal delta of Texel Inlet see Sha (1989b, 1990a,b) and Sha & de Boer (1991).

Sequence models

Lateral migration of channels and ebb-delta lobes, and decreases and increases of tidal prism and of wave attack may produce different sedimentary sequences. The three most characteristic sequences are (Sha & de Boer, 1991):

1) *Progradational ebb-delta-lobe sequences* (fig. 21A), deposited from reworked inner-shelf to distal ebb-delta-lobe sediments, coarsening upward into active ebb-tidal delta-lobe sediments with diminishing bioturbation towards the top.

2) *Abandoned ebb-tidal delta sequences* (Fig. 21B), formed under conditions of a relative sea-level rise. Along the Wadden Sea islands such successions have formed during the Holocene due to the glacio-eustatic rise of sea level (cf. Sha, 1990a). Such sequences are truncated at the top by a ravinement surface formed during shoreface retreat. A reworked, inner-shelf tidal sand sheet covers the sequence. If the rise of sea level continues, the series may ultimately be covered by shelf mud deposits. Sha (1990a) demonstrated that the preservation potential of ebb-tidal delta deposits depends mainly on the maximum depth of shoreface erosion and the depth of the seaward limit of the ebb-tidal deltas.

3) *Inlet and channel sequences* (Fig. 21C), formed especially in the proximal parts of the ebb-tidal delta, where migrating channels produce successions with ebb-dominated channel deposits at the base, topped by flood-dominated channel deposits. Depending on the local conditions, the top of the sequence may be formed by abandoned channel deposits or shoal deposits.

In all cases, preservation in the fossil record depends on the location with respect to the depth of shoreface erosion which, in turn, depends on the ratio between lateral and vertical migration of the shoreline in relation to the depth to which waves rework and erode the shoreface sediments.

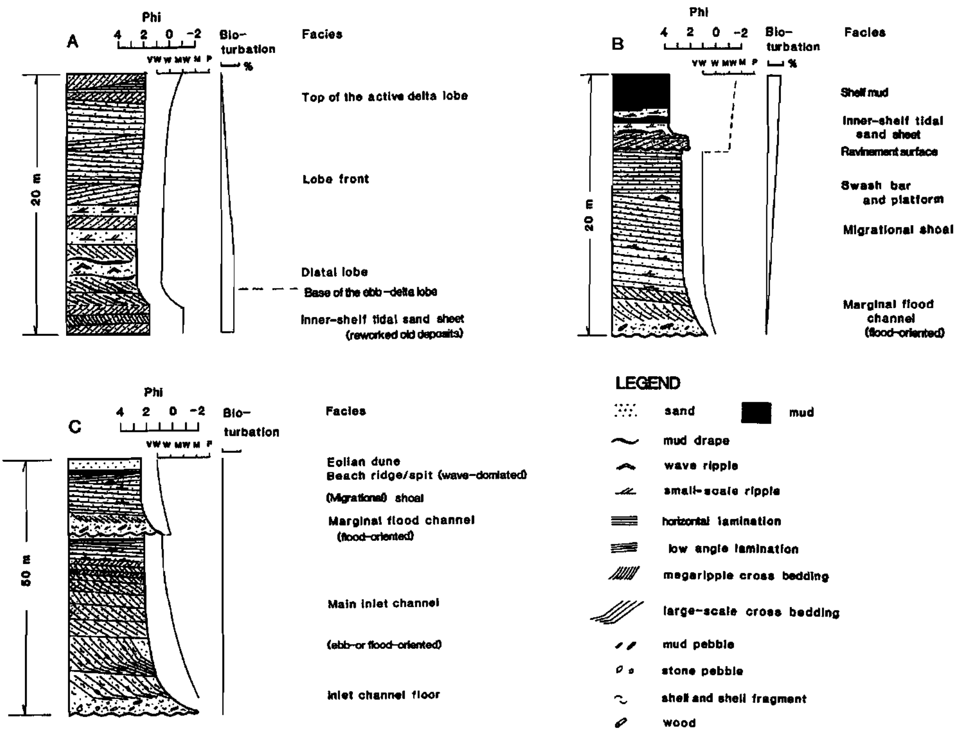


Figure 21: Hypothetical vertical sedimentary successions in inlet and ebb-delta deposits. A. Active progradational ebb-tidal delta lobe. B. Abandoned ebb-tidal delta lobe facies. C. Migrating inlet. Paleocurrent directions in the lower part are commonly ebb-dominated, whereas higher in the sequence flood-oriented structures are dominant (Sha, 1990a).

Preservation of ebb-delta deposits

Ebb-delta deposits can be preserved in the course of a transgression. On seismic profiles off Terschelling and Ameland, the remnants of ebb-delta/inlet deposits can be recognized (Sha, 1990, 1992). They show low-angle, seaward-dipping foresets of the ebb-delta lobe and low-angle inclined beds formed by the inlet channel, migrating laterally, parallel to the coast.

As indicated above, a prerequisite for preservation is, of course, that the maximum erosional depth of the shoreface is less than the depth of part of the ebb-tidal delta deposits. In other words, the ravinement surface formed during shoreface retreat should be positioned above the lower bounding surface of the ebb-tidal delta. The lower bounding surface is generally produced by migrating channels or is formed by the pre-existing sea bed in the more distal parts. The relative positions of these lower and upper (erosional) bounding surfaces depend on the rate of relative sea-level rise, the rate of sediment supply as a function of the tidal prism draining a particular inlet, and the wave erosion base. Early Holocene ebb-tidal-

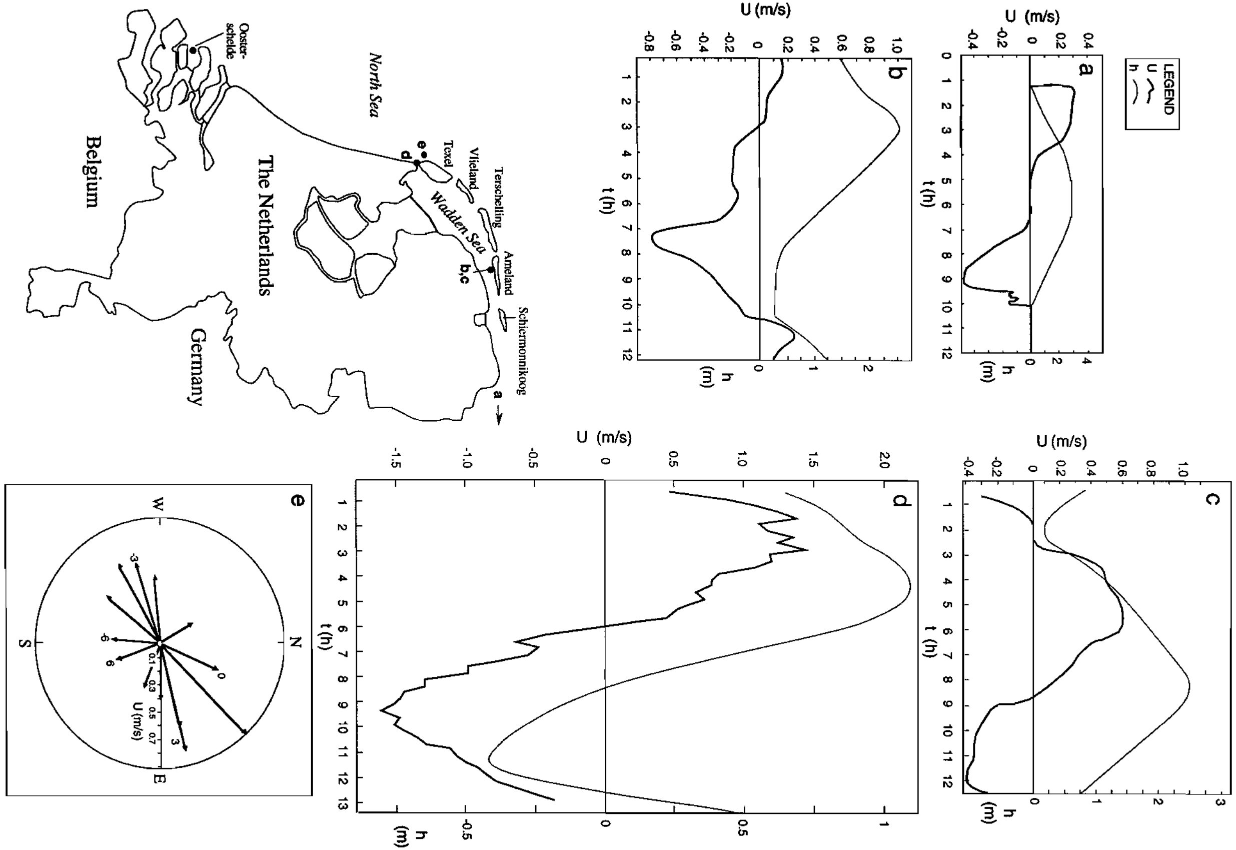


Figure 22: Typical tidal flow velocity curves (thick line) along the mesotidal Dutch/German coast. Thin line indicates water level. Map shows locations. (a) Flow velocity curves for the intertidal flats of Germany (modified after Reineck, 1982a); (b) Flow velocity curve for an intertidal gully south of Ameland (modified after Van Straaten, 1954); (c) Flow velocity curve for a shallow subtidal channel (modified after Van Straaten, 1954); (d) Flow velocity curve for an inlet channel (modified after Postma, 1954); (e) Ellipsoidal flow velocity curve in the Dutch offshore region (modified after Sha, 1990a).

delta deposits in the subsurface offshore of the West Frisian Islands show that preservation of ebb-tidal delta deposits during a rise of the relative sea level does occur indeed (Sha & de Boer, 1991). In the offshore of the Dutch Wadden Sea the deposits of tidal inlets and the related tidal deltas constitute a relatively large part of the offshore sub-fossil barrier-related sediments (Fig. 21; Sha, 1990a, 1992).

Backbarrier area

Waves from the North Sea (which may reach a height of 10 m during severe storms) enter the Wadden Sea mainly around high tides. The influence of waves entering from the North Sea into the backbarrier area (the Wadden Sea) decreases rapidly due to refraction and strong decrease of water depth towards the backbarrier area. Indeed, direct observations show that the wave energy is strongly reduced right after the passage of the inlet throat (Niemeyer, 1986). Most waves in the Wadden Sea are locally generated, reaching heights of up to 2 m (max. approx. 4 m). These waves influence sedimentation on the intertidal shoals and in the salt marshes. Since the tide is asymmetrical due to the morphology of the Wadden Sea, especially that of the tidal flats, the ebb phase lasts longer than the flood phase. As the latter thus is stronger, it can transport larger quantities of sediment (Van Straaten, 1964; Postma, 1967; Dronkers, 1986). The resulting net sediment transport is thus directed towards the backbarrier area. In this process the Wadden Sea acts as a 'sorting machine'. The large differences in current velocity (Fig. 22) combined with the ebb and flood movements and the influence of waves on the various parts of the backbarrier area efficiently sort the sediment with respect to its transportability (mainly grain size and shape).

In the backbarrier area of the Wadden Sea clearly defined flood-tidal deltas, as known from other barrier island systems, e.g. those along the east coast of North America, do not occur. This may be due to the relatively strong currents which supply and redistribute so much sediment that the larger part of the Wadden Sea is filled up to the intertidal level. As a result, the term flood-tidal delta could be applied to the Wadden Sea as a whole.

The backbarrier region is commonly subdivided into three zones, often referred to as sub-, inter-, and supratidal. Actually, the terms lower intertidal and upper intertidal are to be preferred over 'inter-' and 'supratidal' because during (very) high tides the latter zone is also subject to flooding. The backbarrier environment, which constitutes the Wadden Sea proper, can thus be divided into:

1) a zone below low water springs: channels;

2) a zone between low water springs and mean high water: shoals and gullies;

3) a zone above mean high water: high shoals (mainly at the North Sea side) and salt marshes.

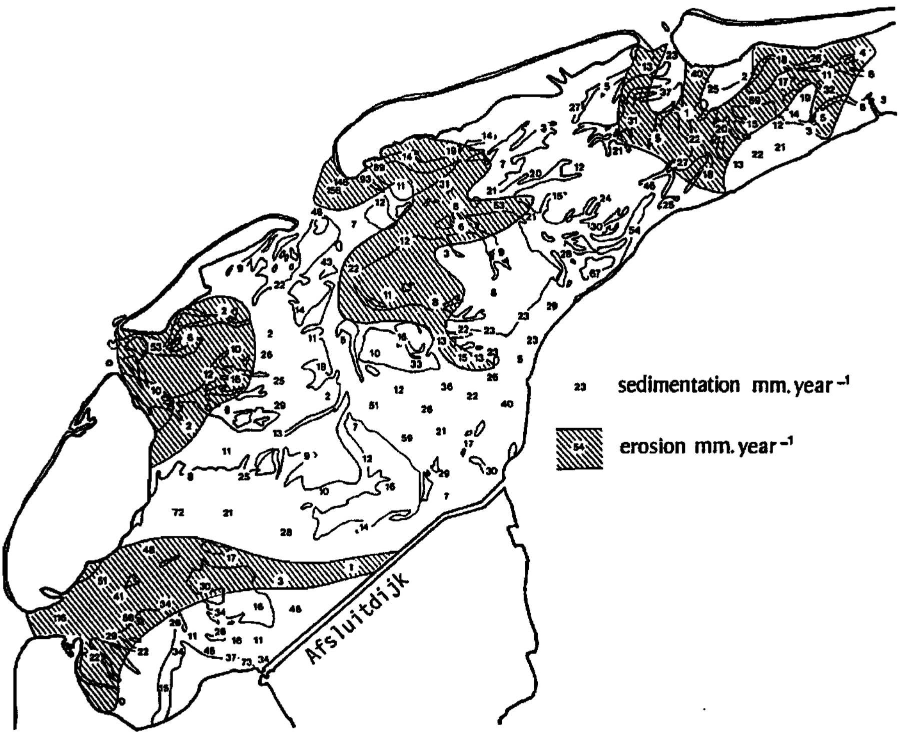


Figure 23: Erosion and sedimentation in the southwestern part of the Wadden Sea over the last decades, caused mainly by the shifting of channels close to the inlets (Eisma & Wolff, 1980).

Channels

In the Wadden Sea the inlet channel splits up into an extensive drainage network (Fig. 5). Through these channels the tidal waters flow to and from the intertidal shoals, thus forming the main pathway for water and sediment (Winkelmolen & Veenstra, 1974) transfer. The current velocity in the major channels is almost everywhere similar (maxima around 1 $m.s^{-1}$, Fig. 22), the cross-section decreasing with decreasing tidal volume. For the larger channels of one of the tidal systems of the Dutch Wadden Sea (the Vlie system) it was shown that an empirical relation exists between the tidal prism and the cross-sectional area of the channels (Gerritsen & de Jong, 1985; Gerritsen, 1990), i.e.,:

$$A_c=7.16*10^{-5}P+135$$

where A_c is the cross-sectional area of a channel and P is the mean tidal prism. Van der Spek (1995) furthermore proposes an empirical relationship between the maximum depth of the backbarrier channels and the tidal prism:

$$h=1.48*P^{0.46}$$

where h is the maximum depth and P is the mean tidal prism.

These and earlier mentioned numerical relationships have all been defined empirically; the exact nature of the generating physical mechanisms is still subject to debate (cf. Van den Berg, 1986; Dieckmann et al., 1988; Stive & Eysink, 1989; Gerritsen, 1990; Hume & Herdendorf, 1990; Misdorp et al., 1990; Niemeyer, 1990; Sha, 1990; Biegel, 1991b; Flemming, 1991; Steijn, 1991; Van Kleef, 1991; Eysink, 1992; Eysink & Biegel, 1992; Bilse, 1993; De Vriend & Bakker, 1993; Eysink, 1993; Van der Spek, 1995).

In smaller channels the maximum current velocity is lower. In such cases the clay content of the channel sediments is inversely related to the size of the channel. Local sediment transport depends on the local dominance of the ebb or the flood current. In contrast to estuaries, where ebb and flood currents often have separate channels, tidal channels in the Wadden Sea mostly have both flood and ebb dominated sections.

Lateral channel migration is one of the most prominent mechanisms controlling erosion and deposition (Fig. 23). The rate of lateral accretion can be quite fast, reaching several 100 $m.yr^{-1}$. The sediments in the channels are generally not or only poorly bioturbated.

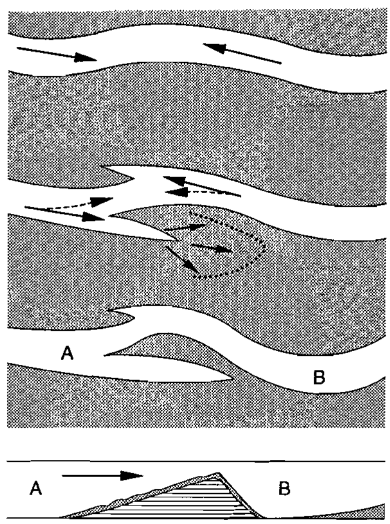


Figure 24: Development of ebb and flood chutes. Bottom: section from A to B (Van Straaten, 1964).

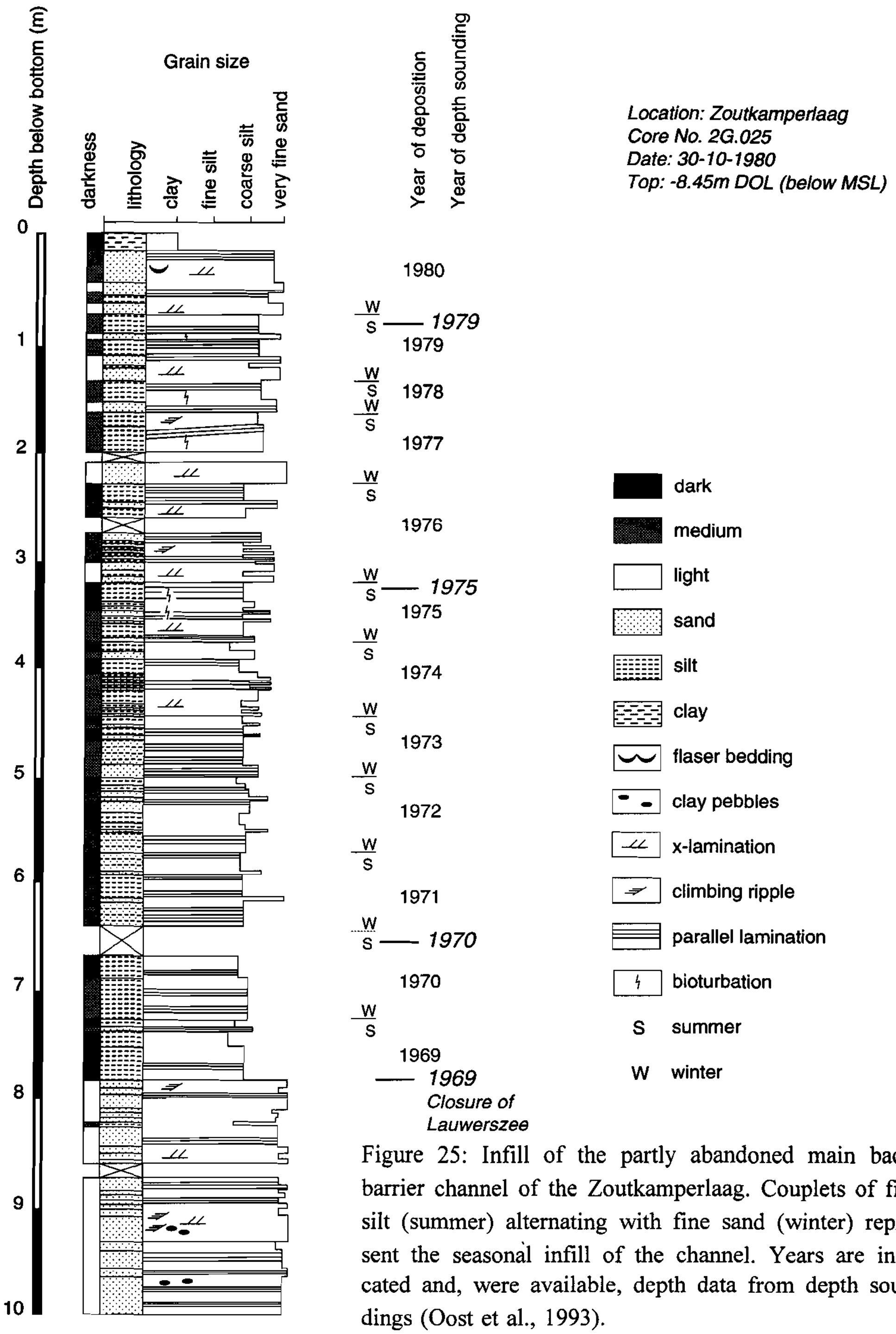


Figure 25: Infill of the partly abandoned main back-barrier channel of the Zoutkamperlaag. Couplets of fine silt (summer) alternating with fine sand (winter) represent the seasonal infill of the channel. Years are indicated and, were available, depth data from depth soundings (Oost et al., 1993).

Depending on the rate of lateral migration relative to the general rate of subsidence and sediment accumulation, a large part of the inter-channel sediments deposited in the inter- and subtidal zone may be eroded. Consequently, the ratio of channel to inter-channel deposits in the rock record may exceed the ratio in recent environments by several orders of magnitude.

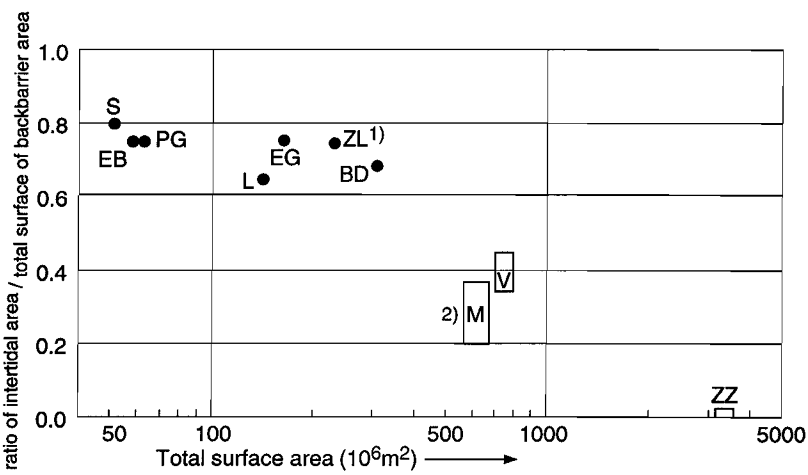
Lateral migration of channels is partly restricted by ebb- and flood-dominated tidal chutes (Fig. 24; Van Straaten, 1954). Due to inertia effects, flowing water has a tendency to flow in a straight line if possible. In the bends of channels the water flow is pushed against the outer bend. During the flood, water eventually rises above the channel banks. The flooding water tends to maintain its straight course and thus may create a secondary channel: a flood chute. In this way bends in tidal channels can be cut off. Often the current is not strong enough to create a new cut-off channel, so that these flood chutes rapidly become shallower and narrower in the direction of the flow and vanish completely downstream, often developing a lobe of sediment at the end. Similar processes can be observed in the ebb currents (Fig. 24).

Because of meander cut-offs and of take over of drainage of one channel system by another, (parts of) channels can become abandoned. Abandoned channels are filled with sediment (clay, sand or a combination of both) in relatively short periods, i.e., in months to years (Fig. 25).

Channel lags concentrated along the channel axis consist of (rare) stones, clay pebbles, relatively coarse sands and shells (for a discussion see Flemming et al., 1992). Above the channel lag, especially in the inner bends of the channel wall, sand and fines are deposited. When channels are sufficiently deep and wide, dunes may develop within them. In general, the dimensions of such dunes increase with the depth of the channel (De Boer et al., 1991; Oost & de Haas, 1992). Heights vary from several decimetres to several metres and spacings may reach several hundreds of metres in extreme cases. Most of the dunes are ebb-dominated because the channels are commonly ebb-dominated, the highest velocities being reached when the water level has dropped below the channel banks. On the other hand, the flood current is strongest at higher water levels. The dunes may register the tidal influence in the form of double mud drapes, thick-thin alternations at foreset laminae, due to the diurnal inequality of the tide, and bundle sequences formed by neap-spring cycles (Visser, 1980; De Boer et al., 1989).

Shoals and gullies

Shoals are fully exposed during low water springs and flooded during mean high water. They form a substantial part of the Wadden Sea. As much as 80% of the backbarrier parts of the inlet systems of the Dutch Wadden Sea may consist of intertidal shoals (Fig. 26). This value is lower for Texel Inlet and Vlie Inlet. According to Eysink & Biegel (1992) there is a relation between relative size of the intertidal shoals and total size of the backbarrier area. An alternative possibility is that the Texel and Vlie systems have not yet reached an equilibrium, because of the relatively young age of these basins (around 1000 years) and also because of the closure of the Zuider Zee in 1932 (see historical section).



1) before closure of the 'Lauwerszee'
2) after closure of the 'Zuiderzee' ('IJsselmeer')

● Backbarrier area

ZZ = Zuiderzee	PG = Pinkegat
M = Marsdiep / Zeegat van Texel	ZL = Zoutkamperlaag
EG = Eijerlandse Gat	EB = Eilanderbalg
V = Vlie	L = Lauwers
BD = Borndiep	S = Schild

Figure 26: Relationship between percentage of intertidal area and total surface of catchment areas of different inlets of the Dutch Wadden Sea (Stive & Eysink, 1989). Note that the Zuider Zee (ZZ) represents a special case, whereas before its closure by the 'Afsluitdijk' in 1932, interference between entering and leaving tidal waves led to a damping of tidal current velocities and a consequent decrease of sediment transport capacity. With respect to the Texel (Marsdiep) and Vlie inlets it is questionable if these catchment areas have reached equilibrium after their origin about 1000 years ago and the interference caused by the closure of the 'Afsluitdijk' in 1932.

In the channels the maximum current velocity during the flood occurs just before or around the time that the shoals are being submerged (Van Parreeren, 1980; Postma & Dijkema, 1982). As soon as the flood water spreads over the lower intertidal shoals its velocity decreases, and accordingly its sand transport capacity decreases strongly. As a result, sandy levees with flood-oriented dunes often form along the channel margins. Such levees can be several dm higher than the surrounding shoals. Due to the relatively high resistance (a thin layer of water over a large area) tidal current velocities over these shoals are generally low (Fig. 22). Sediment transport on the shoals in the lee of the levees is also somewhat flood-dominated, but especially influenced by waves.

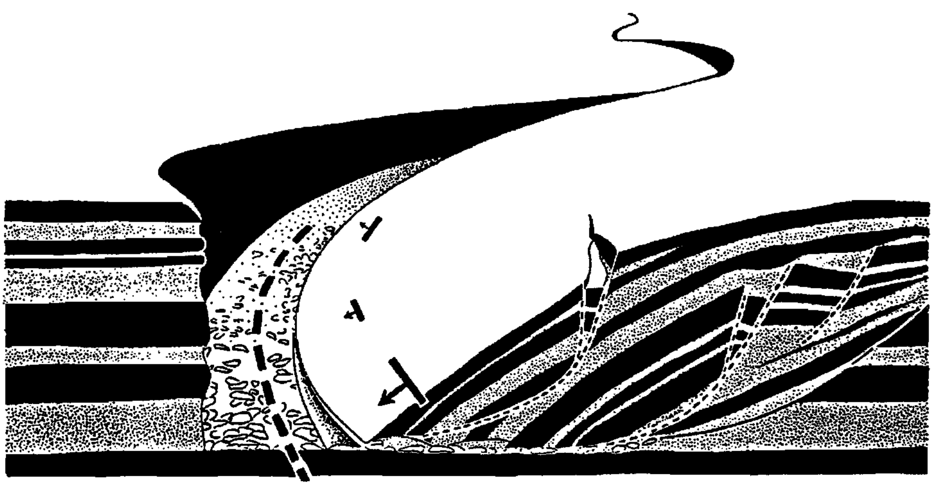


Figure 27: Soft sediment deformation in relation to lateral migration of channels (Reineck, 1982b).

The ebb current reaches its maximum velocity after the water level has fallen below the level of the shoals. The water flowing from and (as groundwater) out of the flats is concentrated in shallow, meandering gullies that drain into the subtidal channels (Fig. 27). Current velocities in the gullies may reach 1 $m.s^{-1}$ around low water. Sedimentation in the gullies is thus dominated by falling stage currents (Van Straaten, 1964). The relation between water level and current velocities in the gullies is shown in Fig. 22.

Lateral shift of the shoals occurs mainly by lateral migration of channels and gullies. Due to the weakness of muddy deposits, channel and gully erosion may be accompanied by various types of soft sediment deformation along the channel walls in the intertidal zone (Fig. 27). For sandy layers, the instability of the channel walls is likely enhanced by repeated submergence/emergence and the intrusion of air (cf. Wunderlich, 1967; P.L. de Boer, 1979).

Slow net vertical accretion, of the order of $mm.yr^{-1}$, occurs over large areas on the shoals. The hydrodynamic conditions and the high-water level, however, restrict vertical accretion. With increasing height of the sandy shoals the influence of waves on the bottom increases. When the shoals become sufficiently high, the major part of sediment transport is accomplished by waves, resulting in a strongly diffuse spreading of the sediment (Eysink, 1979). Moreover, tidal currents can transport the sediment suspended by waves, whereas wave action prevents re-settling. Finally, storms and drift ice, occurring mainly in autumn and winter, may level the shoals. Depending on the phase of the tide, the eroded sediment will be transported towards the sub-tidal areas, to the open sea or to the tidal marshes.

Sand transport into and within the Wadden Sea has been estimated from volume changes based on repeated depth soundings (De Boer et al., 1991; Oost & de Haas, 1992, 1993). It appears that the net sediment transport for individual inlet systems of the Wadden Sea is of the order of millions of cubic metres per year. Based on direct measurements, extrapolations have been made (continuous measurements over a whole year have not been carried out), which show that the total annual sediment volume imported and exported through the inlets must amount to many millions of cubic metres per year (Eysink, 1993).

Mean sedimentation rates in the Wadden Sea are of the order of 1 mm to 2 mm.yr^{-1}. This value has been determined on the basis of repeated soundings, dating by pollen, ^{210}Pb and by ^{14}C determinations. Over shorter spans of time (years) vertical sedimentation rates vary from several cm to 1 m.yr^{-1} over larger areas.

By the above processes of erosion and sedimentation, a dynamic equilibrium is established (Heynis et al., 1987; Nichols, 1989; Eisma et al., 1989; Eysink & Biegel, 1992; Oost & Dijkema, 1993; Oost & de Haas, 1993). For the Dutch Wadden Sea the maximum elevation of shoals is 0.3 m below mean high water (except for the Dollard), whereas the mean height is 1.3 m below mean high water. The mean shoal height in particular seems to keep pace with the mean high water level (Eysink, 1979; Eysink & Biegel, 1992). Together with the more stable part of the intertidal areas this suggests that sedimentation keeps up with relative sea-level rise. The same can also be concluded from historical data (Oost & Dijkema, 1993; chapters 2 to 4).

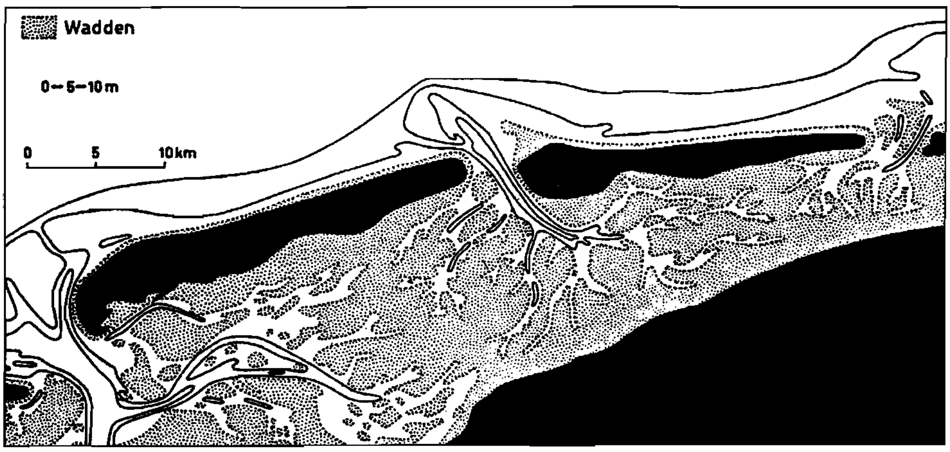


Figure 28: Physiography of the Wadden Sea south of Terschelling and Ameland. Depth contours: 0 m, 5 m, and 10 m below the local low-water line (Van Straaten, 1964). Note the abundance of tidal channels close to the inlets and their near-absence along the tidal watersheds.

The slow vertical accretion of intertidal areas gives ample opportunity for bioturbation. Layers of shells and shell fragments (Cadée, in press) buried at 15-20 cm below the tidal flat surface are formed by the burrowing activity of the lugworm (*Arenicola marina*; Van Straaten, 1954). Layers of shells can also form where dense populations of shells die in situ (for instance *Mya arenaria*). Such mass mortality layers are formed during extremely cold winters (cf. Van den Berg, 1981).

Important morphological elements of the backbarrier flats are the tidal watersheds separating the tidal basins of adjacent inlet systems. Morphologically a watershed forms a slightly elevated ridge comprising lower intertidal and upper subtidal sediments connecting the mainland with the barrier island. The watersheds can commonly be crossed by foot during low tide. In the Wadden Sea a morphological watershed is formed where the tidal wave entering through a westerly inlet meets the wave entering through the next, i.e., more easterly inlet. At the watersheds low-velocity rotational currents (whirls) develop so that sediments (sands and fines) are easily deposited. Because the tidal wave approaches from the west, the tidal wave enters each consecutive inlet with a certain time lag. As a result the watersheds occur asymmetrically displaced towards the east in the rear of the islands and the mainland (Fig. 28). Since the positions of these watersheds continually shift due to inlet migration, changes in tidal currents and wind effects (M. de Boer, 1979; FitzGerald, 1988; de Boer et al., 1991; Oost & de Haas, 1992, 1993), the formation of higher intertidal areas along the watersheds is prevented (Eysink, 1987).

As stated before, the grain size of the sediment decreases from the open North Sea towards the mainland and the higher parts of the shoals. The deeper parts of the Wadden Sea are subject to high energy conditions and thus consist mainly of sand. On the higher parts of the shoals, on the other hand, energy levels are lower due to dissipation of wave energy (Ehlers & Kunz, 1993), shorter duration of water cover and a general decrease in current velocity. This favours the sedimentation of silts and clays. There is a great variety of sedimentary structures on sandy and mixed tidal flats (cf. Van Straaten, 1954; Reineck, 1982b, c; Wunderlich, 1982a, b; Ehlers, 1988; Oost & Baas, 1994). Towards the higher, muddier parts not only population densities of burrowing organisms increase, but algal mats and higher plants (*Zostera*) are also common.

These grain size trends are brought about mainly because current velocities are reduced towards the higher shoals. Sand settles mostly in the early stages, being preferentially deposited on the lower parts of the shoals. Moreover, there is an increase in the concentration of suspended material from the inlets towards the more sheltered and shallower parts of the backbarrier area which results in higher sedimentation rates of fine sediments. This increase in concentration is brought about by the following mechanisms:

1) Higher current velocities are needed to erode particles from the bottom once they have settled, than the velocity at which these particles settle from suspension (scour lag or Hjulström effect (Hjulström, 1935; Van Straaten & Kuenen, 1958). This effect is enhanced if the

deposited particles have had some time to consolidate, a process significantly affecting the erodability after as little as an hour (Creutzberg & Postma, 1979).

2) In the inner parts of the Wadden Sea the tidal wave becomes asymmetrical in such a way that around low water the tide turns much more quickly than around high tide (cf. Fig. 22). The time span during which low current velocities prevail, allowing suspended matter to settle, is therefore much greater at high tide than at low tide (Postma, 1961).

3) The average water depth (in the areas which are covered by water) is, paradoxically, much less at high tide, when the water covers the entire flats, than at low tide when the water only fills the channels (Fig. 29). The probability of suspended matter settling to the bottom over the tidal flats at high tide is therefore much greater than over the channel beds at low tide (Van Straaten & Kuenen, 1958).

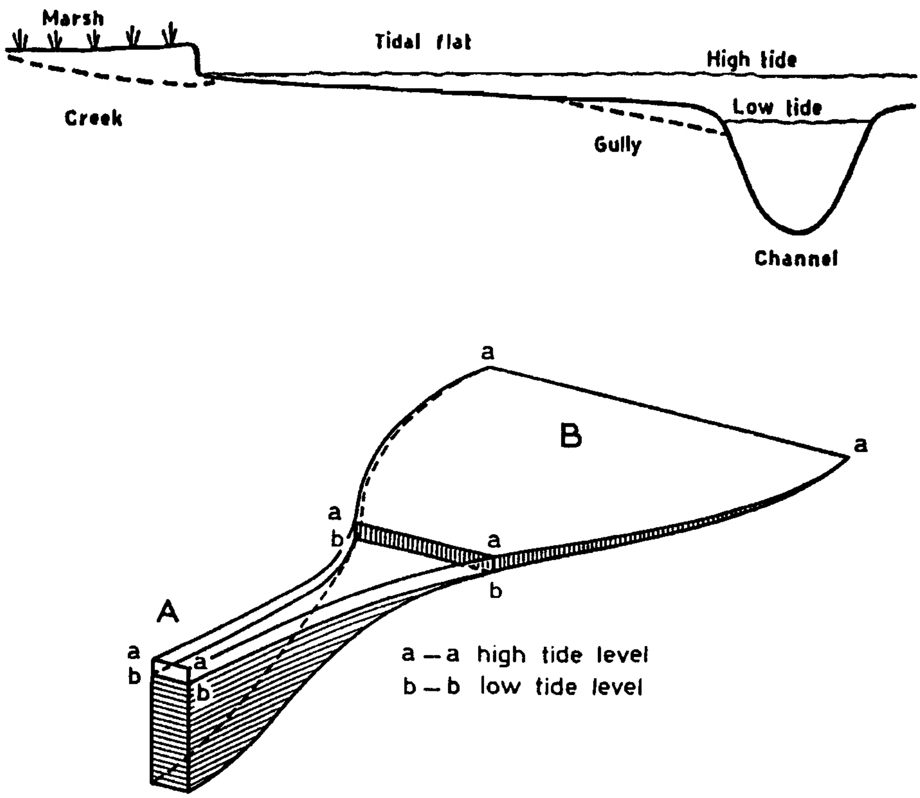


Figure 29: Difference in average depth between high and low-water level (Van Straaten & Kuenen, 1958).

4) Particles will start to settle as soon as the current velocity drops below the settling velocity, the particles being carried along some distance before they reach the bottom (settling lag effect; Fig. 30; Van Straaten & Kuenen, 1958). The flood current will transport a particle from its original starting position (1) in the direction of the land and it will settle somewhat landward (5) from the point where the current velocity becomes too low for transport (3). As a consequence the particle can only be eroded during the subsequent ebb by a water mass (B), which was landward from the water mass (A), which brought the particle in during the flood. Because of this, and because of the effects mentioned above at points b and c, the particle will settle out landward (9) from its original starting position (1) during the second slack water after ebb (Van Straaten & Kuenen, 1958).

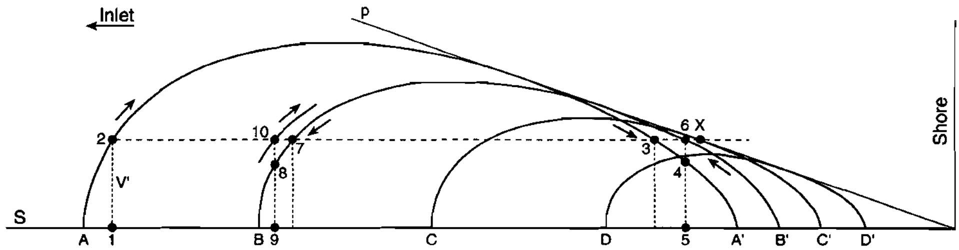


Figure 30: Settling lag effect. Numbers indicate the successive pathways of suspended particles during a number of successive tidal cycles; letters indicate the pathway of the successive water masses capable of (eroding and) transporting the particles (Van Straaten & Kuenen, 1958). See text for further explanation.

5) Suspended matter is transported by waves in the direction of wave propagation which is mostly directed inward into the Wadden Sea (Ehlers & Kunz, 1993). Suspended matter deposited during calm periods can be resuspended and transported by waves. Since this is less common in the more sheltered areas, fine-grained sediments preferentially accumulate there. Indeed, the western Wadden Sea, which is more exposed to wave action than the eastern part, is poor in mud deposits.

6) Organisms also play an important role in the deposition of suspended matter. They filter particles from the water and aggregate them into faeces and pseudo-faeces (e.g. *Mytilus*, *Cerastoderma* sp.; Kamps, 1962). Other animals eat freshly deposited mud and aggregate it into faeces, pseudo-faeces and bind sediments by the secretion of mucus (e.g. *Macoma baltica*, worms). Furthermore diatoms bind particles below and in algal mats. On the salt marshes suspended matter (and during storms also sand) is retained between the vegetation.

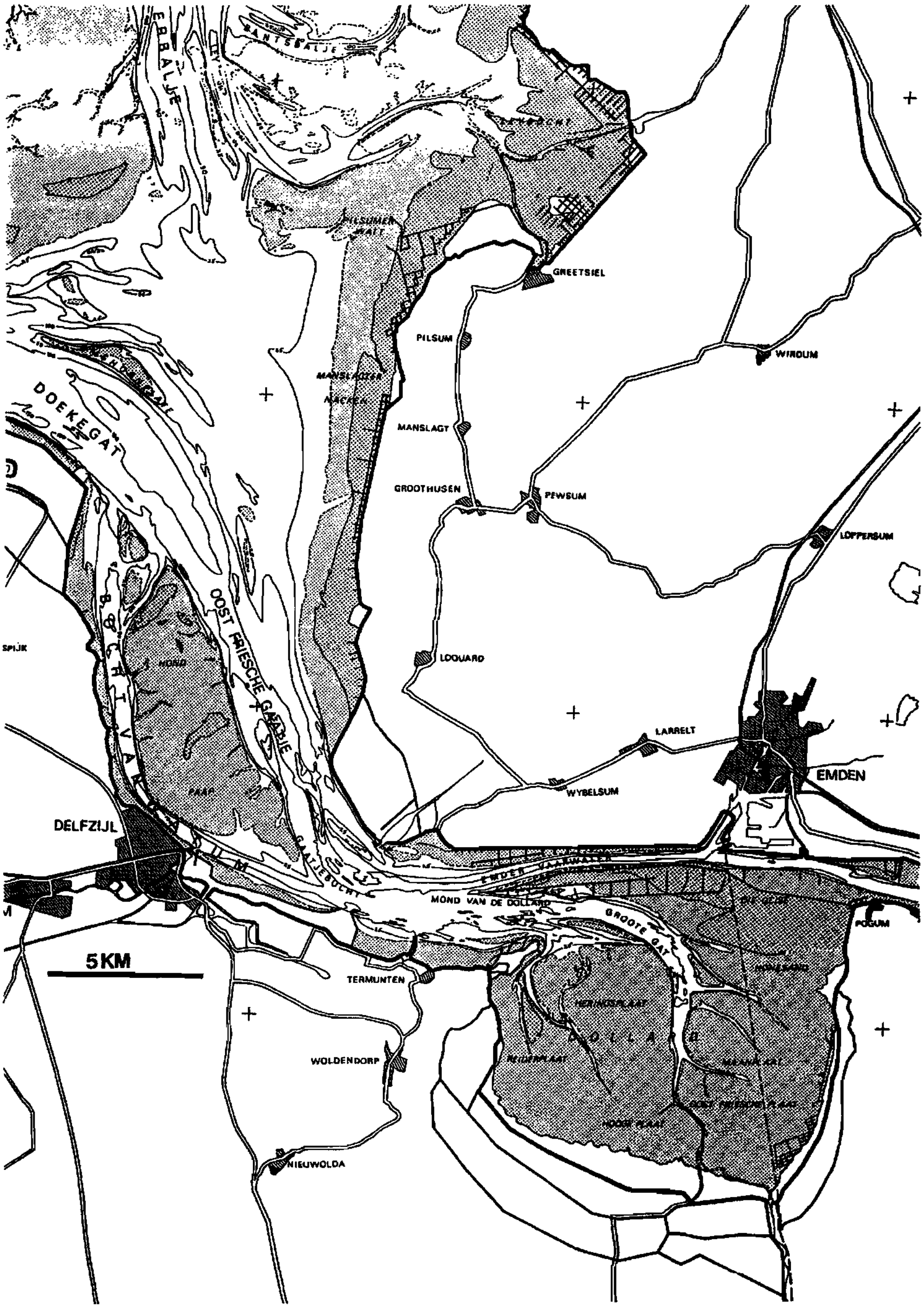


Figure 31: Geography and physiography of the Dollard Estuary.

7) Furthermore, river water influx causes locally a seaward directed residual current at the surface and a residual salt-water current along the bottom. Suspended matter from both currents is trapped by this estuarine-type circulation and concentrated at the convergence zone in so-called 'turbidity maxima'. This effect is especially important in estuaries (Postma, 1967), and contributes to e.g. the large-scale deposition of mud in the Ems-Dollard Estuary.

The Dollard Embayment

The grain-size distribution in the Dutch Wadden Sea illustrates the effects of the concentration of fines on sedimentation. One of the best examples can be observed in the Ems-Dollard region, which is a tidal embayment in the northeastern part of the Dutch Wadden Sea (Fig. 31). It consists of muddy tidal flats intersected by one main channel and some smaller branches. The area is bordered by dykes. In the southwest, in the south and in the east salt marshes are present. These receive some fresh water from a small (canalised) river in the south, the 'Westerwoldse A'. In the north the Dollard Estuary is separated from the river Ems by a dam (Fig. 31).

The sediments in the Dollard are very fine-grained. The median grain size is mostly less than 150-200 µm; even medians of 50 µm occur over large parts of the estuary (Wiggers, 1960; Van Heuvel, 1991) as a consequence of the sorting processes described above (Van Straaten, 1960). Moreover, part of the clays is derived from the river Ems (Favejee, 1960). Measurements show that when the flood waters start to cover the intertidal flats, suspended sediment concentrations become very high as a result of erosion of the surface layer (De Haas & Eisma, 1992). As a result, fine-grained sediments are transported to and concentrated on the intertidal flats.

Sedimentation occurs mainly during summer and erosion during winter, especially during storms. ^{210}Pb and pollen dating have revealed that sedimentation rates in the Dollard lie between 1.4 and 2.8 $mm.yr^{-1}$ (Eisma et al., 1989; Heynis et al., 1987). The relative rise of the sea level in the Dollard is about 1-2 $mm.yr^{-1}$. Since this is less then the average rate of deposition, it confirms that the Dollard is an area of net sedimentation, a feature also evident from the historical records of land reclamation and dyke construction during the last centuries.

The clays, which form the bulk of the fine-grained sediments, are fairly resistant to erosion. Lateral channel migration and the formation of tidal chutes is thus strongly restricted. This is one plausible explanation why channel patterns in this muddy embayment are relatively stable, as opposed to sandy intertidal areas (in a way comparable to anastomosing rivers versus meandering rivers).

Sedimentary sequence of backbarrier sediments

Independent of the relative vertical sea-level movements, tidal flat deposits locally tend to show regressive successions due to the vertical accretion of channel deposits and also to the strong sediment trapping capacity of tidal flat systems. As stated above, many of the deeper channels are dominated by the ebb current, whereas flood dominance prevails in the shallower parts. This is due to the asymmetry of the tide, with the maximum flood-current velocities during the second half of the flood, and the maximum ebb-current velocity during the later stage of the ebb period. Indeed, fossil tidal deposits frequently show this pattern, with ebb directions in the deeper parts and flood directions higher in the sequence.

High shoals and salt marshes

High shoals, like Noorderhaaks (Fig. 19) and Richel, are mostly situated in, or in front of inlets on the North Sea side. They are mainly formed as a result of wave transport and aeolian processes. Their development depends directly on the development of the adjacent inlet channels (see Chapter "Ebb-tidal Deltas and Inlets"; Sha, 1990a; Sha & de Boer, 1991). Exceptions are Griend and Engelsmanplaat, which top older fine-grained deposits acting as erosion resistant nuclei. The higher intertidal shoals are flooded only during spring tide and/or storms.

Natural tidal marshes are today largely restricted to the barrier islands. Along the mainland coast most tidal marshes are artificial, their growth being stimulated and protected by shields of twigs. This is largely the result of dyke construction over the past centuries. Land reclamation obstructs the natural extension of tidal environments in the direction of the mainland (Dijkema, 1987a).

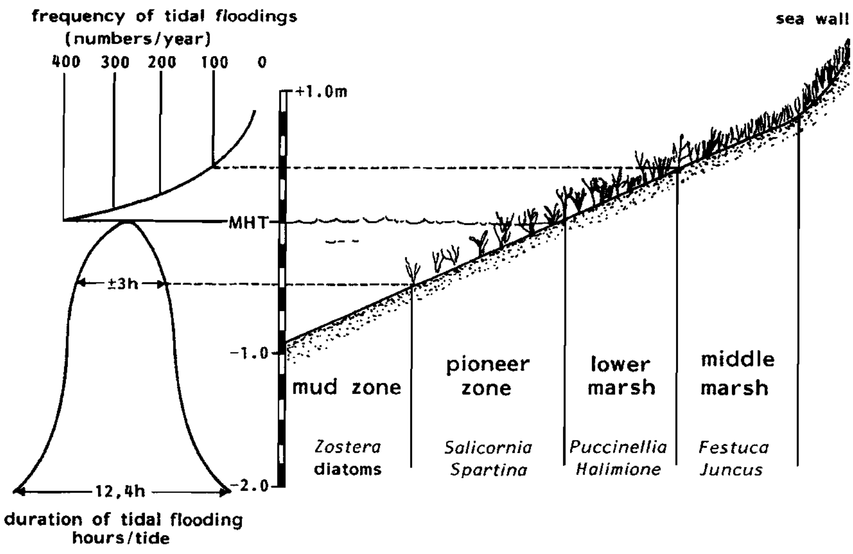


Figure 32: Zonation of the tidal marshes in relation to the duration and frequency of flooding (Dijkema et al., 1990).

Salt marshes are characterized by dense vegetation, comprising halophytes and algae. The vegetation reduces the currents, thus allowing sands, silts and clays to settle out. This sediment is then stabilized by the root systems of the marsh plants. The tidal marshes are drained by a system of small intertidal channels (creeks). When the height of the marsh increases relative to the low-water line, this upper intertidal zone is flooded less frequently than the lower intertidal zone. Sedimentation and erosion, and the wet, salty and poorly ventilated bottom produce rather hostile conditions for higher plants (Dijkema et al., 1990).

On the lower intertidal flats only algae and sea grasses (*Zostera*) can exist. Where sedimentation increases the height of the area to several decimetres below mean high water, pioneer plants (*Salicornia dolichostachya, Spartina anglica*) settle, followed with increasing height by *Suaeda maritima* and *Aster tripolium* (Dijkema et al., 1990). Apart from the elevation, successful settlement also depends on the wave energy and the firmness of the sediment (König, 1948; Van Eerdt, 1985; Groenendijk, 1986; Dijkema, 1987b). The pioneer zone is flooded almost every day (Fig. 32). *Salicornia* does not significantly enhance sedimentation, but enables other plants to settle (Kamps, 1962). *Spartina*, on the other hand, is known to enhance sedimentation (Christiansen & Miller, 1983). When the height of the pioneer zone increases, the vegetation cover of pioneers becomes denser and other plants start to settle (Van Oosten, 1986).

The border between the pioneer zone and the lower marsh, i.e. the area around or above the mean high-water level is characterized by the appearance of marsh grasses (e.g. *Puccinella maritima*).The lower marsh vegetation can only exist if the bottom is well ventilated (oxygenated; Dijkema et al., 1990). The vegetation of, amongst others, *Limonium vulgare* (prevalent in relatively sandy areas), *Puccinella maritima, Halimione portulacoides, Spergularia maritima, Triglochin maritima, Plantago maritima, Atriplex prostrata, Cochlearia anglica* and *Aster tripolium* enhance sedimentation rates to maximum values (Dijkema et al., 1988, 1990). In addition, erosion is strongly reduced (Kamps, 1962; von Weihe, 1979).

The denser vegetation, especially *Puccinella maritima*, also triggers the formation of creek systems (Van Straaten, 1964; Dijkema et al., 1990). The water flow is concentrated in the lower areas between the patches of plants, where it restricts sedimentation or even scours the bottom. When vegetation becomes denser, the originally randomly distributed lower-lying areas develop into strongly meandering creek systems. Lateral migration of these creeks is slow or absent (Van Straaten, 1964; pers. obs.). The levees of the creeks consist of an alternation of clay and sand layers, several mm to cm thick, which accumulate through settling from suspension. Deformation processes comparable to those in lower intertidal gullies may occur in the creeks. The lower marsh is flooded several hundred times per year (Fig. 32). The creeks strongly enhance the drainage of the area and improve the ventilation of the sediment, thus promoting further the colonisation by plants (Kamps, 1962; Van Diggelen, 1988; Dijkema et al., 1990).

When the marsh becomes even higher, sedimentation rates decrease strongly due to the decreasing number of floods and also due to the smaller sediment supply with each flood.

As a result, marsh deposits typically show a fining and thinning upward of the sediment layers (Van Straaten, 1964). Due to sedimentation the lower marsh zone evolves into a middle marsh zone: *Puccinella maritima* disappears and plants like *Festuca rubra, Juncus gerardii, Agrostis stolonifera, Armeria maritima, Artemisia aritimae, Elymus pychanthus* and *Glaux maritima* appear (Dijkema & Bossinade, 1990). Above this zone the upper marsh commences with normal grassland plants, such as *Elymus repens, Leontodon autumnalis, Lolium perenne* and *Potentilla anserina* (Van Oosten, 1986; Dijkema & Bossinade, 1990).

Normally deposition of fine silt and clay only occurs in the middle and higher marshes during (very) high water. Beds of sand and shells, deposited during storms, may be continuous over kilometres. Shells of over a dm in length and other animal remains can also be transported over small distances by strong winds. During this process left/right sorting can take place (Cadée, 1992). More substantial shell hash deposits, situated further inland, are accumulated in the upper intertidal zone by birds such as Eiderduck, Oystercatchers, Gulls and Crows. Up to several percent of the annual shell production is probably transported in this way (Goethe, 1937, 1958; Remane, 1951; Leopold et al., 1984; Oost & Leopold, 1988; Cadée, 1989, 1992, in press). For a more extensive discussion of Dutch tidal marshes the reader is referred to Van Straaten (1954), Van Oosten (1986), Dijkema (1987a, b), Dijkema et al. (1988) and Dijkema & Bossinade (1990).

Sand sharing system

The different morphological elements constituting the Wadden Sea show a strong mutual interaction. All elements influence the local tidal currents and the wave regime and thus also the local sedimentation patterns. During the last decades many empirical relationships have been recognised for the different parts of the system. Especially the tidal prism greatly influences the character of the different morphological elements, as well as the interaction between them. For this reason each individual tidal inlet system can be considered to form a separate sand-sharing system (cf. Dean, 1988). As a first approximation each sand-sharing system can be considered to comprise an inlet, the related ebb-tidal delta, the island points at either side of the inlet and the related backbarrier drainage area between the watersheds behind neighbouring islands. However, it should also be realized that adjacent drainage areas can influence each other across the watersheds (Oost & Dijkema, 1993).

By definition, all parts of a sand-sharing system are coupled and are in a dynamic equilibrium with each other. Changes in any part of the system will primarily be compensated by sediment transport to or from other parts of the same system. When changes are temporary, the old dynamic equilibrium will eventually be restored. If changes are more permanent (e.g. by loss of drainage area), a new equilibrium will be established. In both situations sediments can be imported or exported from or to areas outside the normal sand-sharing system.

Sedimentologists studying barrier-related tidal deposits should thus realize that observations of any part of the system may provide important clues to other parts of the system. For instance, observed changes in channel size and depth may be related to changes in the (backbarrier) drainage area (when tidal range is constant) which, in turn, will also be expressed in the ebb-tidal delta volume (e.g. Flemming & Davis, 1992; Chapter 5).

RELATIONSHIPS BETWEEN FLORA AND FAUNA AND SEDIMENTS

The influence of biota on siliciclastic sedimentation is commonly ignored. This is certainly not justified for the Wadden Sea. As is the case in many backbarrier areas, the biomass of the Wadden Sea is very large compared to other marine environments. Biota influence erosion, transport and sedimentation of all sediments and, moreover, they form part of it.

The high nutrient supply enables a high primary production. Important primary producers are floating microscopic algae (phytoplankton) and, most important, microscopic benthic algae (micro-phytobenthos). Besides the local primary production, the influx of plankton and detritus from the North Sea is also important. This food supply is utilized by the primary consumers (or secondary producers). The majority of the fauna (molluscs, polychaetes) lives in the sediment as deposit feeders or suspension feeders. This fauna is predated upon by crustaceans, molluscs, echinoderms, fishes and birds (Creutzberg & Kuipers, 1984).

As in most backbarrier areas, the faunal diversity is low in the Wadden Sea. Normally, four species (*Arenicola marina* (lugworm), *Mytilus edulis* (blue mussel), *Cerastoderma edule* (cockle) and *Mya arenaria* (soft-shell clam)) each make up more then 15% of the total benthic fauna (together even 75%!). Some 40 species constitute the other 25% (Beukema, 1977). This is partly due to the low variability of the substrate, allowing only a small range of ecological niches. More important, however, are the extremely large variations of the environmental conditions in time (Creutzberg & Kuipers, 1984). Particularly on the tidal flats, large fluctuations in temperature, humidity, current velocity, salinity and sediment supply occur in the course of a year. Only a few species can cope with these extreme variations. Some authors argue that estuaries have not existed long enough in geological time, or occur too isolated in space and time, to permit the evolution of a complete euryhaline estuarine fauna (Nybakken, 1982). This view is conjectural: other researchers state that the animals which live on the tidal flats, and which also occur in fully marine environments, are not particularly adapted to the extreme conditions on the tidal flats, but also occur in other environments (Hertweck, 1992).

The marine benthic community in the Wadden Sea consists primarily of species that have a free-swimming larval stage. In general, as is also the case in the Wadden Sea (e.g. Beukema et al., 1977; Buhr & Winter, 1977), benthic communities are relatively stable over large areas through time. This stability is partly maintained because the larvae are able to detect if the sediment is suitable for settlement or not (Nybakken, 1982). Larvae often preferentially settle where the adults are already living (Seed, 1976) and many larvae are able to detect the presence of adults of their own species by certain pheromones released into the water (Crisp & Meadows, 1962). Larvae may also respond to physico-chemical factors such as light, pressure and salinity (Thorson, 1966). In the early stages of their larval life a large part of the free-floating larvae reside in the upper, faster-moving water where dispersal is the strongest. Later, when time for settlement approaches, they migrate to the bottom. Other larvae are especially sensitive to pressure and light, hence being restricted to particular levels of the water column. These larvae can only settle if the water in which they reside comes into contact with the seabed, an important mechanism in the establishment of non-randomly distributed community patterns (Nybakken, 1982).

Furthermore, in the Wadden Sea it has been noted that mass mortality during severe cold winters is compensated for by greater amounts of larvae in the following year, ensuring a quick re-settlement in places which have become devoid of life (Seed, 1976; Beukema, 1982). The influence of some specific elements of the biota on the sediment is briefly discussed below. For a more extensive discussion on faunal influence the reader is referred to Dörjes (1982) and Hertweck (1982, 1990, 1992).

Bacteria

Sections through Wadden Sea sediments commonly show a light-coloured to rusty brown layer some mm to cm thick at the sediment surface. This layer is aerobic and contains algae, bacteria and ferric ions. Aerobic bacteria oxidize the organic compounds and produce hydroxides. Where erosion has occurred recently, or where the concentration of organic matter is extreme, this layer may be missing locally.

This upper aerobic layer is followed by a darker, often black-coloured layer. This is the anoxic zone where anaerobic bacteria flourish. Some of these bacteria can reduce sulphate ions to form sulphur-hydrogens. These react with iron-hydroxides to form colloidal iron-sulphides which have an intense black colour. For a fuller treatment of this subject the reader is referred to Gerdes et al. (1985) and Van Gemerden (1993). It is still not clear what effect these bacterial processes have on sedimentation, especially on diagenetic processes. The strong diagenetic alterations around burrows, which have been attributed to the influence of anoxic-aerobic transitions (e.g. Aller, 1982; Zijlstra, 1994), indicate that the influence may be considerable.

Diatoms

Diatoms live on aeolian dunes, sandy shoals, in the wet troughs of ripples, and on mudflats. These siliceous algae are rarely fossilized because the slightly alkaline waters tend to dissolve the silica skeletons. Although benthic diatoms hardly produce any sediment, they nevertheless have an important effect on sedimentation. Epipelic diatoms move over the sediment surface and leave a trail of sticky slime. Where diatoms are abundant (e.g. at the sediment-water interface) they bind the sediment which has been deposited after a period of high current velocities. Moreover, sediment grains are glued together by epipsammic diatoms or by the mucus produced by diatoms. In case of very dense algal mats, grains may even become part of the organic carpet. Moreover, abundant mucus attached to the sand grains may decrease the bottom roughness. This mechanism is most effective in the case of poorly rounded, angular grains.

Field experiments in an intertidal area of the Oosterschelde (SW Netherlands) have shown that in higher energy parts of the intertidal flats, where the activity of algae cannot be recognized on a macroscopic scale, strong erosion may occur on places where the algal cover is chemically destroyed (De Boer, 1981). Both epipsammic and epipelic diatom populations may increase the threshold velocity for the erosion of sands in temperate climatic zones, as in the case of the Wadden Sea. Laboratory experiments have revealed that a diatom population that was allowed to re-establish itself for 24 hours after stirring, increased the threshold velocity by several tens of percents, as compared to freshly stirred and redeposited sand; sandy sediment surfaces with a rich epipelic diatom population may even show an increase of more than 100% (Vos et al., 1988). Stabilization by diatoms can therefore be an important factor in the stabilization of sediment in intertidal and shallow subtidal areas.

Arenicola marina (L.) (Lugworm)

On the higher parts of the tidal flats, where migrating gullies are rare, organisms have sufficient time to completely modify the original depositional structures. The combined action of burrowing organisms works like a powerful conveyor-belt in all directions. One of the most important bioturbators is the lugworm *Arenicola marina*, which forms U-burrows. The animal lives at depths of 15 to 20 cm, almost continuously ingesting sand (deposit feeder). There is no need for *Arenicola* to migrate because sand is sliding down to the mouth by gravitational forces, creating a small pit at the sediment surface in the process. From time to time the animal excretes the indigestible sand in spaghetti-like strings at the surface (Van Straaten, 1954, 1964; Beukema, 1977; Hertweck, 1982). It has been calculated that the top 30 cm of the tidal flat sediments can be reworked completely by *Arenicola* spp. each year (Cadée, 1976). The grains that are too large to be swallowed are concentrated at the deepest point of the U-burrow. In this way extensive layers of shells, especially of *Hydrobia* spp., are formed (Van Straaten, 1954, 1964).

U-burrows of the type produced by *Arenicola* spp. are found throughout the stratigraphic column, being considered typical for intertidal sediments. This does not, of course, imply that all U-burrows in the rock record relate to the same polychaete species. The strategy of making U-burrows allows animals to continuously ingest sediment without exposing themselves at the sediment surface to predators or hostile physical and chemical conditions. This favourable way of life in the intertidal domain has been adapted by a large variety of different animal species (see e.g., Bromley, 1990). It should be noted, however, that such burrows are not exclusively restricted to intertidal environments, but can also be found in subtidal deposits (Hertweck, 1992).

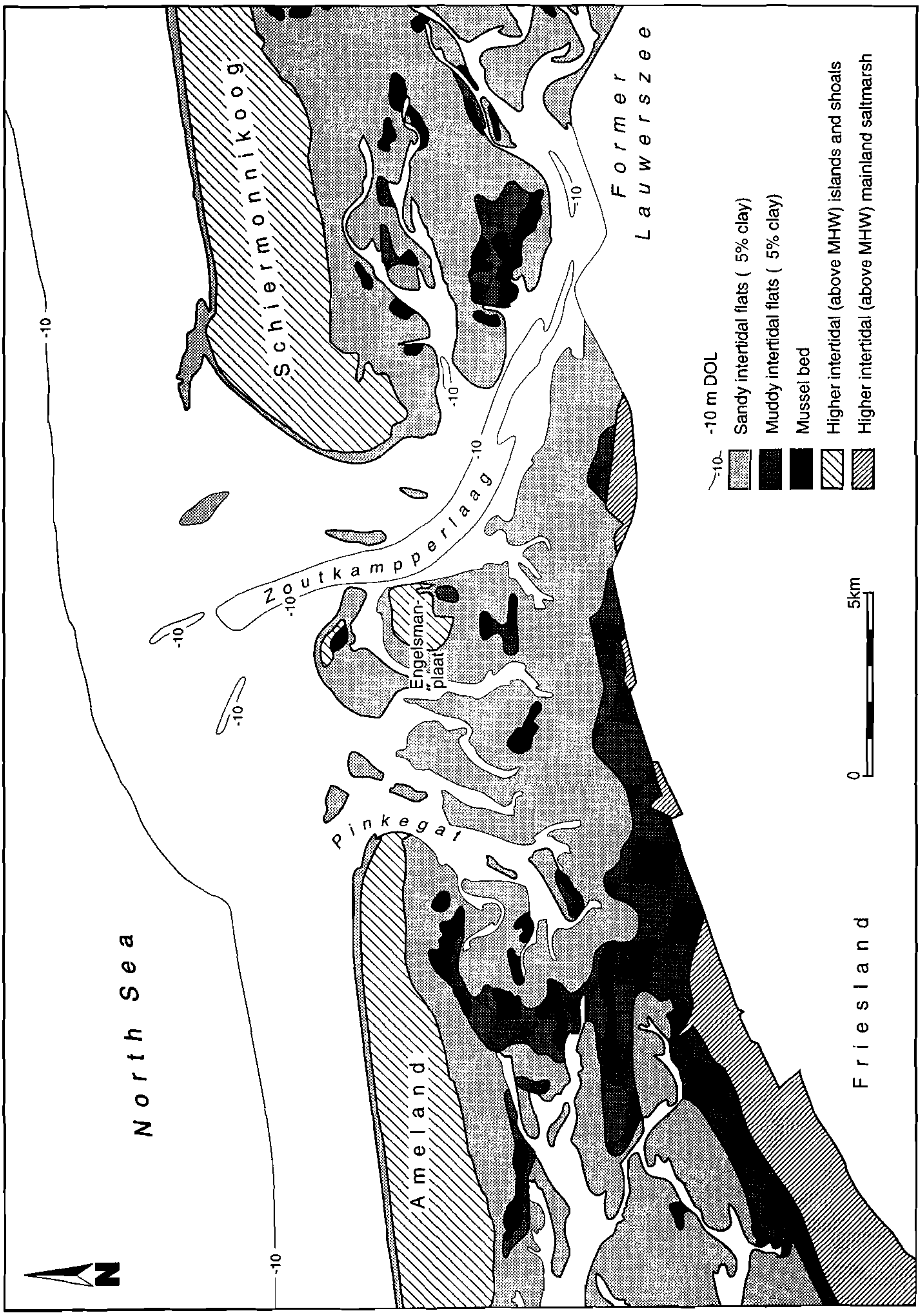


Figure 33: Mud flats in the surroundings of mussel colonies consist partly of faeces and pseudo-faeces produced by the mussels. After Dijkema (1989).

Mytilus edulis L. (Blue Mussel)

The blue mussel lives on the surface and filters seawater for organic particles (suspension feeder). The part of the filtrate which is indigestible is compressed into mud faeces and pseudo-faeces. The animal excretes faeces strings which break up into individual pellets approximately 1 mm long. These pellets are hydraulically equivalent to sand particles of about 60 μm in diameter. They are fairly resistant to erosion and can be transported over several tides before complete disintegration occurs (Oost, in prep a, Chapter 6). As observed on the tidal flats, small-scale ripples may locally consist predominantly of mud in the form of faecal pellets. Large amounts of mud are also deposited in mussel colonies, forcing them to move upward in order to avoid suffocation by their own faeces. Mussels attach themselves with byssus threads to avoid being swept away by waves and currents. In doing so, the colonies form a semi-rigid framework which tends to bind the sediment, thereby protecting it from erosion (cf. Flemming et al., 1993).

In the more protected parts of the Wadden Sea, faecal pellets may form an important element in the deposition of fine-grained sediments at some distance from the mussel colonies (Fig. 33). The amount of sediment filtered by the mussels in the Dutch part of the Wadden Sea is estimated at several 10^9 kg.yr^{-1}, depending on the size of the population (Dankers & Koelemaij, 1989; Dankers et al., 1989; Chapter 6). Theoretically this could form a layer of dry fine-grained sediment of about 0.5 mm.yr^{-1} covering the entire surface of the Dutch Wadden Sea. Most of the pelletal sediment that is not resuspended, accumulates in the more sheltered parts of the Wadden Sea. The blue mussel is thus of major importance in the deposition of fine-grained sediments in the Wadden Sea.

Plants

In upper intertidal flats and coastal dunes plants are the principal biotic component influencing sedimentation next to man. In tidally influenced areas a clear zonation can be observed from the lower intertidal to above the higher intertidal zone (see Fig. 32). The different plants are adapted to the extreme conditions of the salt marsh environment. Some influence sedimentation by decelerating the flow of the current during flooding. In this way large amounts of sediment are retained and the marsh accretes vertically (see also the part 'High shoals and salt marshes').

In the coastal dunes plants form a succession which mainly depends on the availability of water. Of these plants *Ammophila arenaria* plays an important role in the fixation of aeolian sediments with its strong vertical straws and long horizontal root system. The hardy character of this plant enables it to survive and even to propagate in this barren sand environment.

SEDIMENTARY HISTORY OF THE DUTCH WADDEN SEA

Pleistocene

From a geological point of view the Wadden Sea is relatively young. The Dutch Wadden Sea is underlain by an irregular surface of glacial till and fluvial sediments dating from the Riss (Saalian) glacial stage (approx. 180,000-130,000 B.P.). Most of these glacial deposits have been covered by younger sediments (Eemian interglacial, Würm/Weichselian glacial and post-glacial). Locally, e.g. on the island of Texel, the Riss deposits crop out as small (ice-pushed) moraine hills that reach a height of 10 m to 20 m above present mean sea level (Fig. 34). The area between Den Helder (south of Texel) and the 'Afsluitdijk' also contains Pleistocene outcrops, mainly comprising glacial moraines and fluvial sediments. The deposits form a highly irregular micro-relief in the otherwise flat Dutch landscape (Fig. 34). These Pleistocene deposits form the backbone of the western Wadden Sea and have strongly influenced the morphological development of the area. The development of Texel Inlet, i.e. the tidal inlet between the island of Texel and the mainland, for example, was strongly hampered by Pleistocene deposits in the subsurface. Also, the position of this island is stabilized by its Pleistocene core.

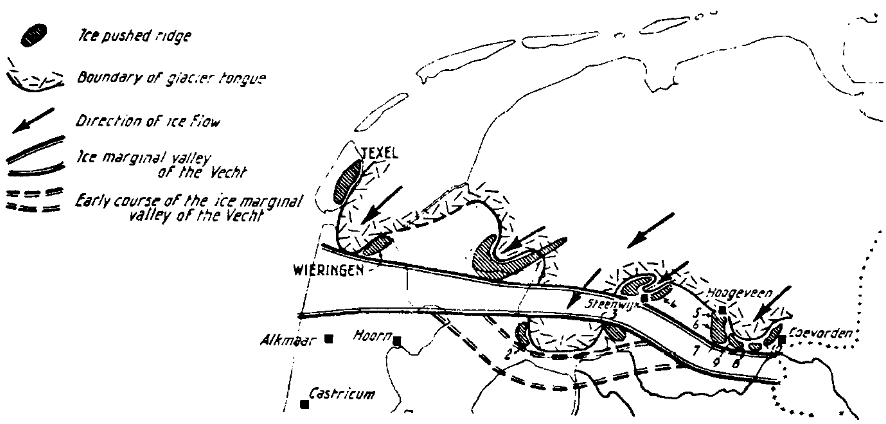


Figure 34: Ice-pushed ridges dating from the Riss glacial.

During the Eemian (Riss-Würm) interglacial sea level was relatively high (e.g. approximately 75,000 years B.P.), low-lying areas being submerged. Shallow marine and tidal flat sediments with a characteristic fauna were deposited in the embayments.

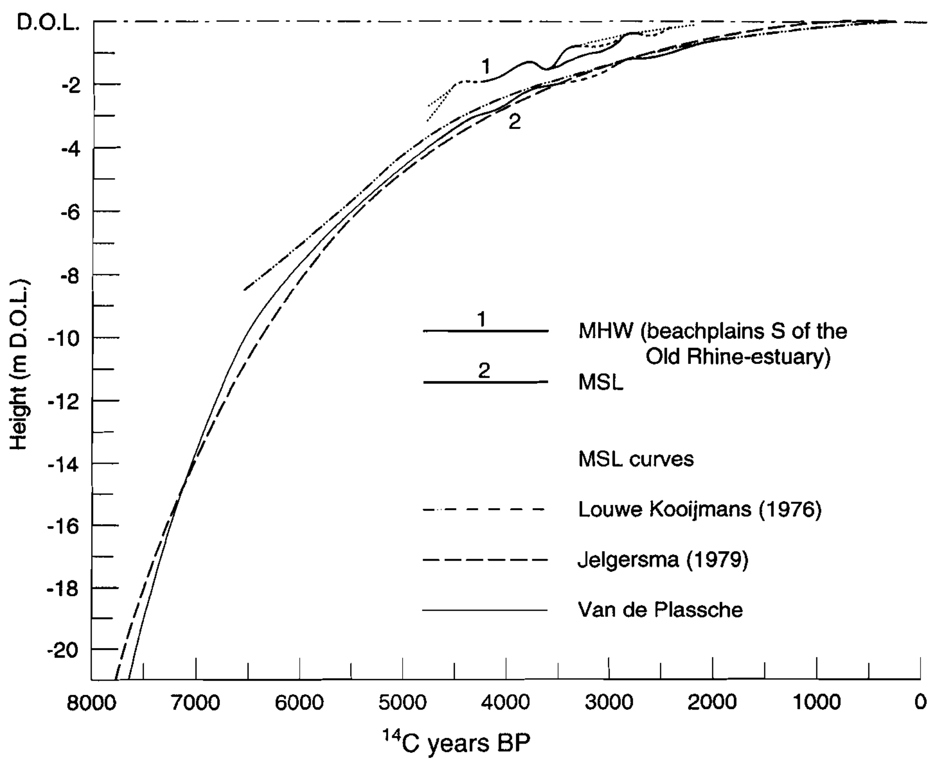


Figure 35: The relative rise of sea level in The Netherlands according to different authors. After Van de Plassche (1982).

During the last glacial stage (Würm-Weichsel, which ended about 10,000 years B.P.) the icecap did not reach the Dutch region. Sea level was about 100 m lower than at present and extensive aeolian deposits ('cover sands' and loess) were formed on the emerged land surface. In many places these deposits form the substratum of the Holocene sediments. In general, the Pleistocene subsurface dips gently in the seaward direction.

Holocene: Pre-historic sea-level rise and sedimentation

During the early parts of the Holocene transgression, mean sea level rose at a rate of approx. 1 $cm.yr^{-1}$, mostly because of melting of the icecaps, but partly also by subsidence of the North Sea basin (Fig. 35). After 6,000 B.P. subsidence (at a rate of 0.10-0.15 $cm.yr^{-1}$) began to dominate, eustatic sea level now rising at a rate of only 0-0.1 $cm.yr^{-1}$. Tidal deposits buried below the Recent marine sands in the southern North Sea and the widespread occur-

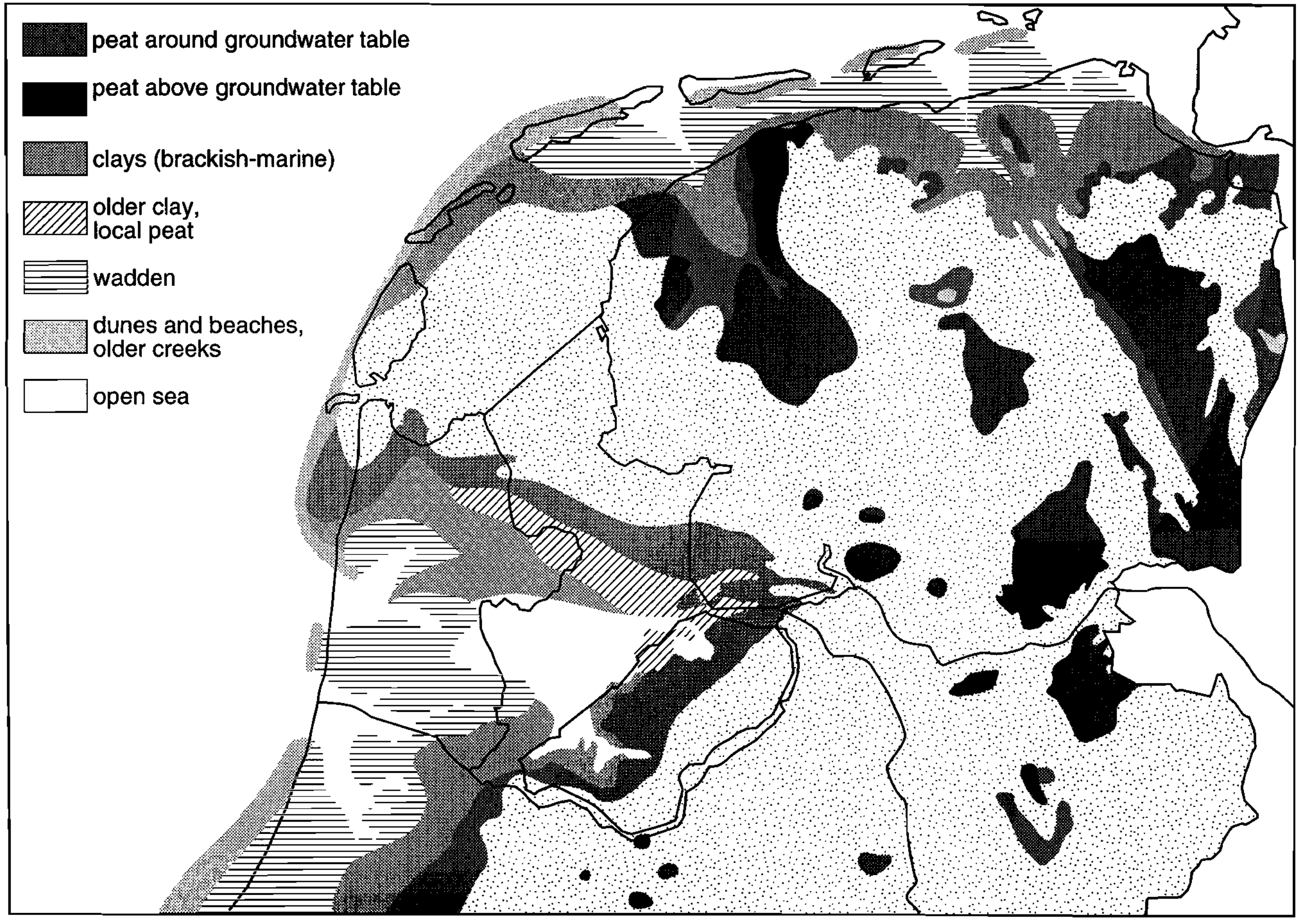


Figure 36: Reconstruction of the Wadden Sea region around 5,300 B.P. (Zagwijn et al., 1985). Position of the barrier islands was several kilometres further to the north than indicated in the figure.

rence of tidal flat mollusc shells suggest that much of the southern North Sea basin comprised tidal flats for some time in the period of 10,000 to 5,000 B.P.. During the Holocene large amounts of sediment were transported to the modern Wadden Sea area, derived from the rivers and from the North Sea. It is assumed that a large part of the Wadden Sea sediments was derived from a high area west of Texel and, during the last millennia, from the coast of North Holland (Schoorl, 1973; Zagwijn et al., 1985; Zagwijn, 1986).

During the early sea-level rise the coast retreated and the ground water table rose. The coastal areas became marshy and a peat layer was formed on the Pleistocene subsurface. From land to sea the following zones succeeded each other: a peat layer (Basal Peat), a fine-grained zone of wadden deposits, brackish lagoons and salt marshes and a sandy coastal zone consisting of beach plains and dunes. During the landward retreat of the coast, the sediments formed in these zones were deposited on top of each other. The transgression continued until approx. 5,000 B.P., after which the Dutch coastline became more or less stable. The topography defined by the Pleistocene subsurface dominated the evolution of the area (Fig. 36). Around 5,000 B.P. a tidal basin was formed in the west (Zeeland-Northern Holland), being separated by a Pleistocene high (Texel-Vlieland) from the precursor of the present Wadden Sea (Zagwijn et al., 1985). Each area experienced a different evolution:

Western Area: coast of South Holland and North Holland

In the western area (coast of Holland) sediment supply by various large rivers overruled the effect of the sea-level rise, causing the coast to prograde. Sediment was also derived from erosion of headlands, old ebb-tidal deltas and from the North Sea (Beets et al., 1992). The dominance of sediment supply by rivers may further have been enhanced by changes in wave climate and tidal amplitude (Zagwijn, 1985). The intertidal areas behind the coastal zone became shallower and tidal channel systems were eventually abandoned (5,500-3,300 B.P.; Beets et al., 1992; Van der Spek & Beets, 1992). By that time the coastline had been closed and large surfaces behind it were logged by fresh water. As a result, a second period of large-scale peat formation (Holland Peat) commenced, becoming especially important between the Middle Subboreal to Lower Subatlantic (Roeleveld, 1974; Zagwijn, 1985). This was followed by a period characterized by a shortage of sediment, so that large parts of the coastal zone were once again submerged.

Western Dutch Wadden Sea

In the western Wadden Sea (Texel-Terschelling) the Pleistocene surface was at a relatively high elevation, thus strongly reducing the influence of the sea (Fig. 36). Until 3,700 B.P. this area developed in the same way as the western one (see above), except for Texel High, where no submergence and formation of peat took place. After 3,700 B.P. part of the area east of the Vlie channel (Edelman, 1964; Eisma & Wolff, 1980) or east of the line central Texel-Bolsward (Zagwijn et al., 1985) gradually changed into a Wadden area. West of this line the surface was covered with peat until the Early Middle Ages. In the east the small

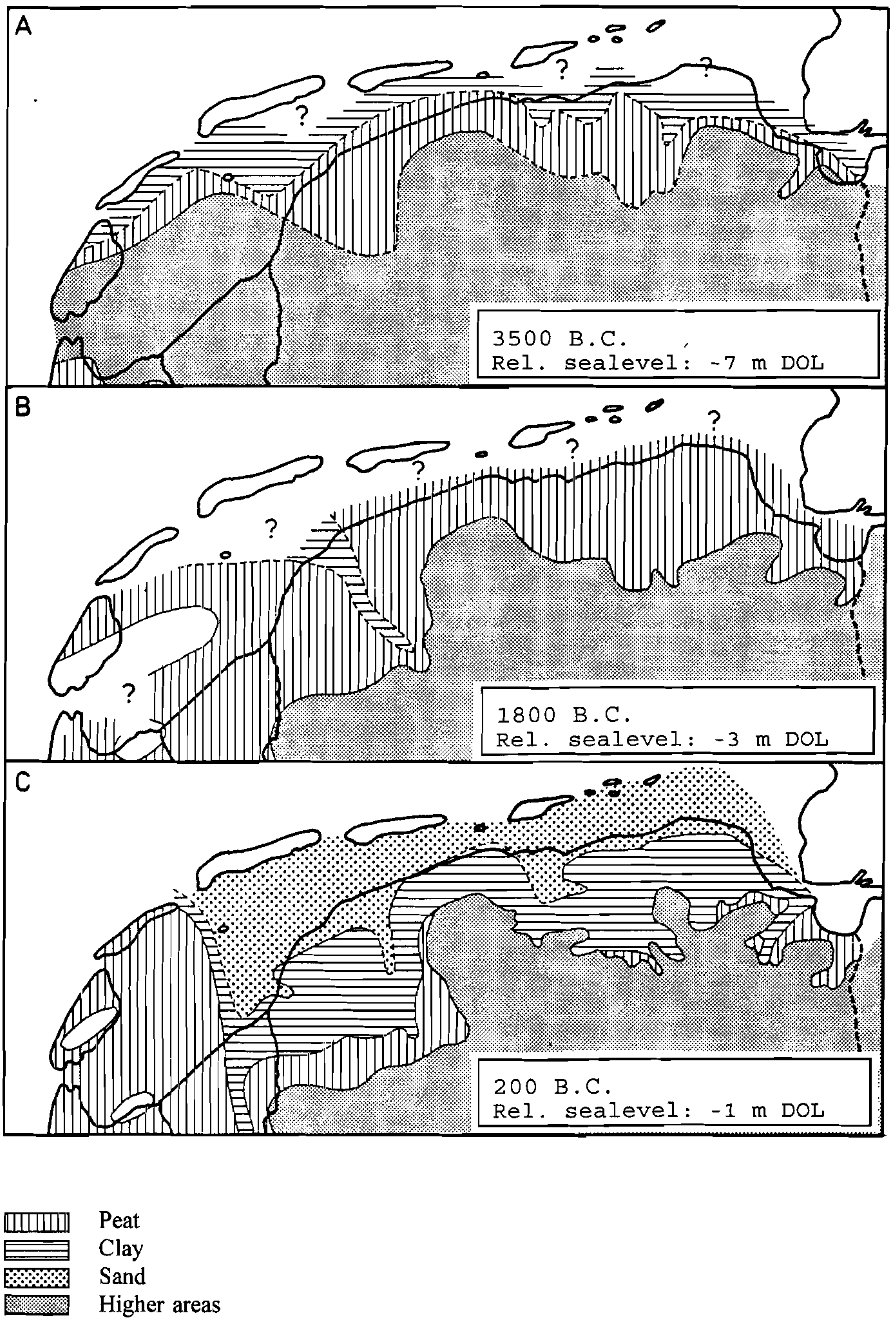


Figure 37: Sedimentation in the period 3500-200 BC (Mazure et al., 1974).

Boorne river debouched, whereas in the west a river was situated at the site of the present Texel Inlet channel (Schoorl, 1973). The sea entered this part relatively late (approx. 1,000 B.P.), and after the peat layer was eroded this area too became tidally influenced (Edelman, 1964; Schoorl, 1973; Eisma & Wolff, 1980).

Eastern Dutch Wadden Sea

In the eastern Dutch Wadden Sea the Pleistocene surface was lower, with a number of small rivers incised (amongst others Boorne and Hunze Valleys). The sea entered the river valleys between 8,000 B.P. and 5,000 B.P., especially after 6,500 B.P. (Griede, 1978; Roeleveld, 1974; Griede & Roeleveld, 1982). Barrier islands were probably generated between 6,000 and 5,000 B.P. (Sha, 1992). Peat formation and seaward extension of the land were important between 4,200-3,000 B.P.; afterwards the mainland coast retreated (Griede & Roeleveld, 1982). Low sediment supply in combination with continued sea-level rise forced the islands to shift landward. Sediment was needed to compensate the relative rise of the sea, i.e., to maintain the barrier islands and the Wadden Sea and to replace eroded Pleistocene and early Holocene fine-grained and peaty deposits. This sediment was largely derived from the island coasts. As a result, the islands shifted coastward (11 km for Ameland and 15 km for Schiermonnikoog in approximately 5,000 years; Sha, 1992). Although shifts of the islands towards the east may have been partly caused by human interference (Flemming & Davis, 1992), geological and historical evidence in the Dutch Wadden Sea clearly demonstrates that the tips of the islands also tend to shift eastward due to the coast-parallel eastward directed drift, caused by residual tidal and wave-driven currents.

In the area of the present eastern Dutch Wadden Sea, lower rates of sea-level rise often did not result in peat formation. On the contrary, clayey deposits were formed, suggesting a larger marine influence than in the areas where peat was formed (Fig. 37). This is probably due to the presence of large tidal inlets at relatively short distance and the rather small width of the tidal area, as compared to the tidal area in the western area (coast of Holland). There, the inner flats were more sheltered behind long coastal barriers and peat formation was far more extensive. An important additional effect has probably been the much greater terrestrial run-off into the western part of The Netherlands as compared to the rather small fluvial input especially in the eastern part of the Wadden Sea.

Historical development: man and flooding

The development of the western and eastern Dutch Wadden Sea in historical times was also different and will therefore be discussed separately.

Western Dutch Wadden Sea

In Roman times and during the early Middle Ages the coastline of northern Holland, up to the present Vlie Inlet (between Texel and Vlieland), was rather continuous, interrupted by

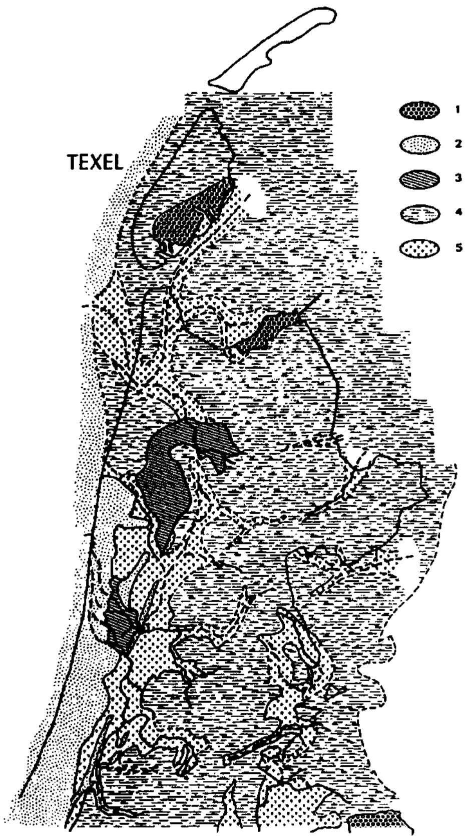


Figure 38: Reconstruction of North Holland and the southwestern part of the Wadden Sea region during the early Middle Ages (Eisma & Wolff, 1980). 1. Pleistocene outcrops; 2. Dunes and beaches; 3. Clay deposits; 4. Oligotrophic peat; 5. Eu- and mesotrophic peat.

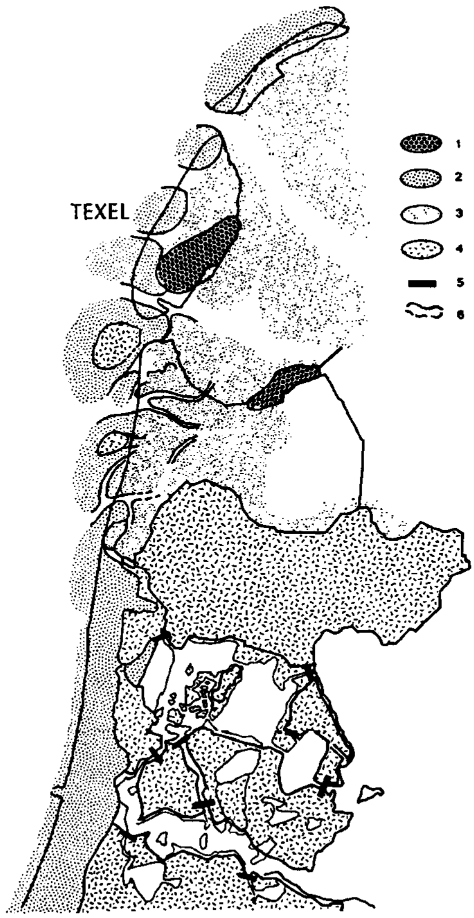


Figure 39: Reconstruction of the situation in the 14th century (Eisma & Wolff, 1980). 1. Pleistocene outcrops; 2. Dunes and beaches; 3. Tidal flats and salt marshes; 4. Low-lying cultivated and inhabited polder area; 5. Dams across streams and inlets.

relatively few river mouths and tidal inlets, peat marshes and woodlands covering large parts of the western Dutch Wadden Sea (Fig. 38). During the following centuries a major part of the coast in the western part was destroyed and the land behind it was flooded, while the coast as a whole retreated (Fig. 39). Also the Zuider Zee (the embayment which was closed

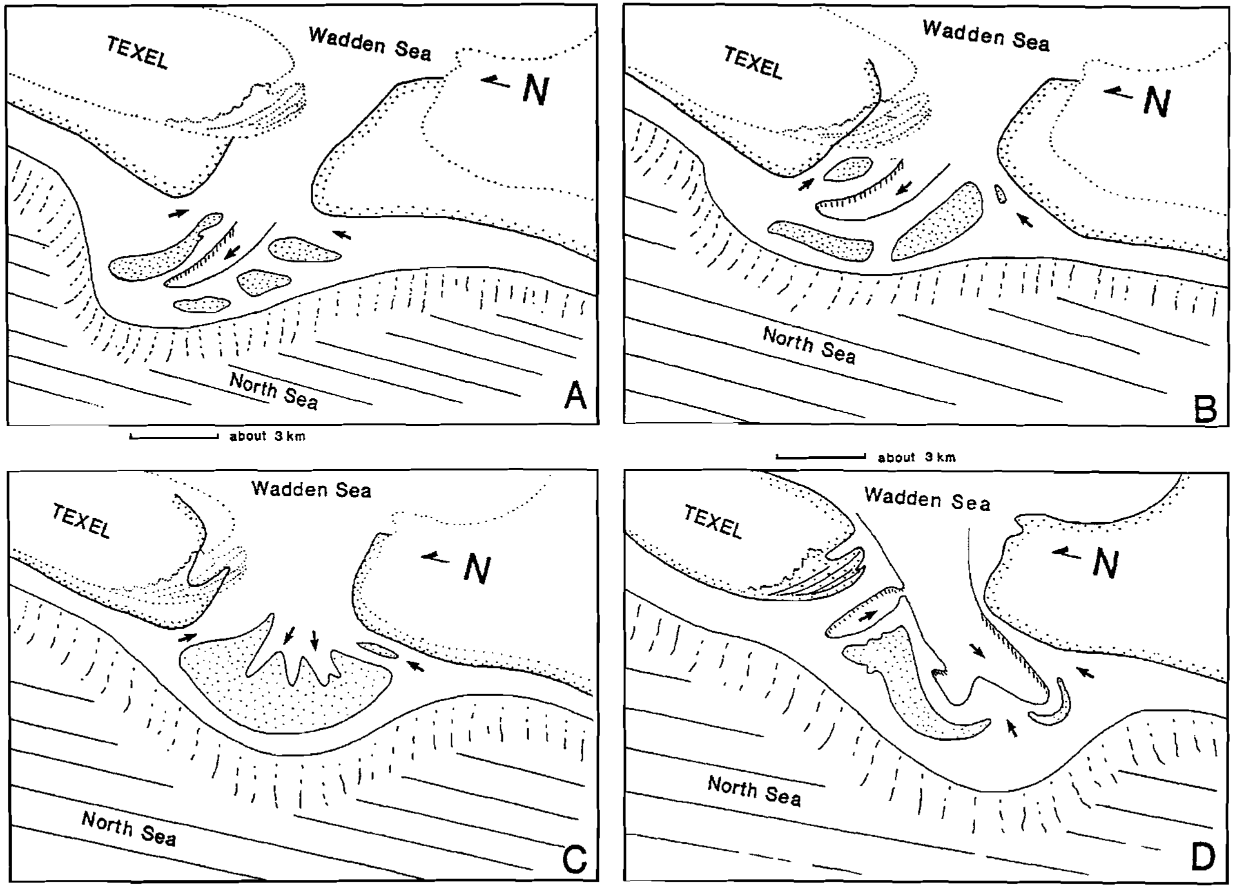


Figure 40: Evolution of Texel Inlet from a downdrift orientation in the 16th century (A) via a more symmetrical orientation in the 17/18th century (B) and 18/19th century (C) to the present-day situation with an updrift orientation of the ebb-tidal delta (D) (Sha, 1990a).

in 1932 by the 'Afsluitdijk') came into being during the Middle Ages, the Wadden Sea thus reaching its greatest extension in historical times. New inlets were formed.

The evolution of the morphology of the ebb-tidal delta of Texel Inlet in relation to the evolution of the hydrodynamic regime can be explained by the mechanisms discussed in Chapter "Ebb-tidal Deltas and Inlets". Texel Inlet probably formed during a severe storm surge in 1170 A.D. It could not be navigated until about 1300 A.D. (Schoorl, 1973). Historical maps show that the maximum inlet depth increased from 25 m in about 1583 (earliest reliable map) to about 50 m at present (Sha, 1990a). Based on the relationship between the maximum inlet depth and the tidal prism, the historical development of the tidal prism was reconstructed by Sha (1990a). He concluded that the tidal prism through Texel Inlet increased from about $240*10^6$ m^3 in 1583 to $1050*10^6$ m^3 in 1970. The increase of the tidal prism in historical times is related to the rise of sea level and the resulting increase of tidal amplitude together with an increase of drainage area. Reconstructions by Sha (1990a) on the basis of historical navigation charts since the 16th century, show that the ebb-tidal delta of Texel Inlet was asymmetrically oriented north to northwest before the middle of the 18th century (Fig. 40). It became symmetrical between the late 18th century and the early 19th century, and since the early 19th century it has been directed towards the southwest (Fig. 40). This is a clear development from a down-drift, relatively more wave-dominated to an updrift tide-dominated ebb delta, which is in close accordance with the observed increase in tidal prism.

Much of the flooding was enhanced by the exploitation of the land behind the coast: ditches and small canals were dug to drain the lowlands and peat was dug for fuel and salt extraction (Edelman, 1964; Eisma & Wolff, 1980). The marshland subsided due to levelling, dewatering and oxidation upon aeration, and became vulnerable to floodings.

Counter measures against further flooding started around 1000 A.D., when dykes were constructed to enclose sheltered areas and to reclaim lost land. In about 1250 A.D., a large dyke protecting a larger part of the inhabited area at the mainland which bordered the western part of the Wadden Sea was established (Schoorl, 1973). The former dune coast or barrier beach along the western coast had been breached during the Middle Ages and former land had been flooded, leaving only a few small islands. From the 16th century onward, a number of such islands were connected by sand dykes which were constructed by trapping aeolian sand between branches and mats of reed. This technique is still in use. Along the mainland, coastal protection works were intensified from the early 19th century onward, and at present the mainland coast of the Wadden Sea is almost everywhere bordered by dykes. Despite the dykes and other coastal protection works, the Dutch Wadden Sea is still a relatively undisturbed area where natural sedimentary processes proceed freely.

During the Middle Ages loss of land generally dominated over reclamation. This process was gradually reversed during the early 17th century when larger areas were reclaimed, culminating in the large reclamations of the 19th and 20th centuries. The present extension of the Dutch Wadden Sea was reached in this century after the enclosure of the Zuider Zee by the 'Afsluitdijk' and the enclosure of the much smaller 'Lauwerszee'. The history of land reclamation in the northwestern Netherlands is shown in Fig. 41.

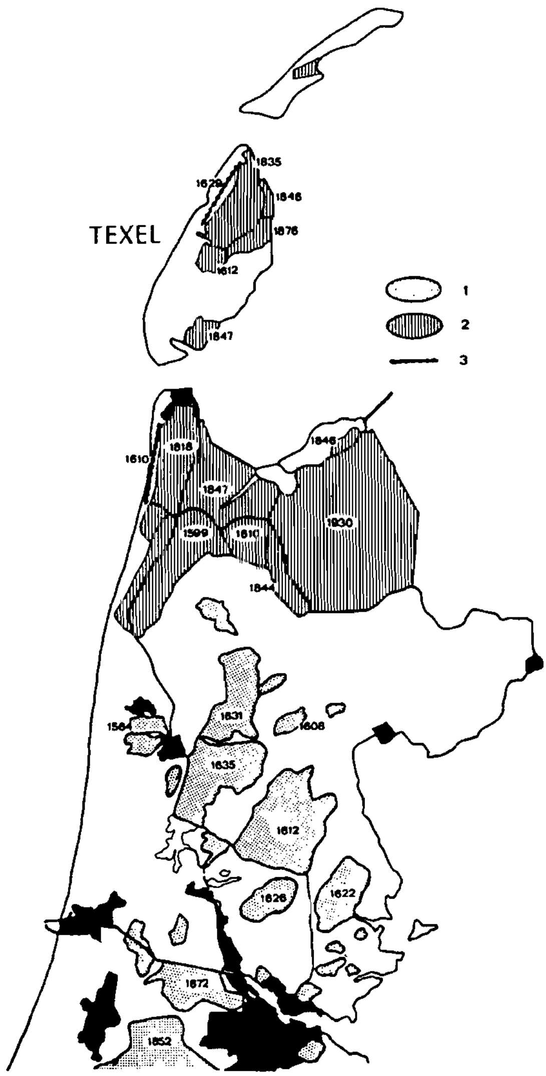


Figure 41: Land reclamation in the northwestern part of Holland and the SW Wadden Sea. Numbers refer to year of reclamation (Eisma & Wolff, 1980). 1. reclamation of lakes; 2. reclamation of tidal areas; 3. sand dykes.

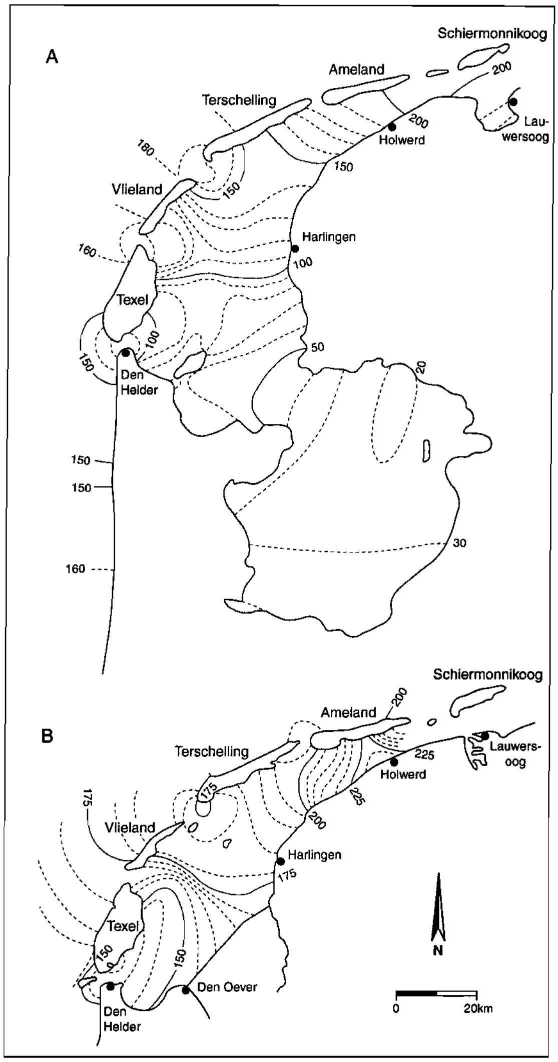


Figure 42: Mean tidal amplitude in cm before (A) and after (B) the closure of the Zuider Zee. Data: A: Klok & Schalkers (1980); B: Vroom et al. (1989).

The 'Afsluitdijk'

The 'Afsluitdijk' has a length of 32 km. It was built in 1932 and is one of the world's longest causeways across a backbarrier area (Fig. 18). South of it the lake 'IJsselmeer' (before the closure: Zuider Zee) is situated. Before the dyke was built, the Zuider Zee was a

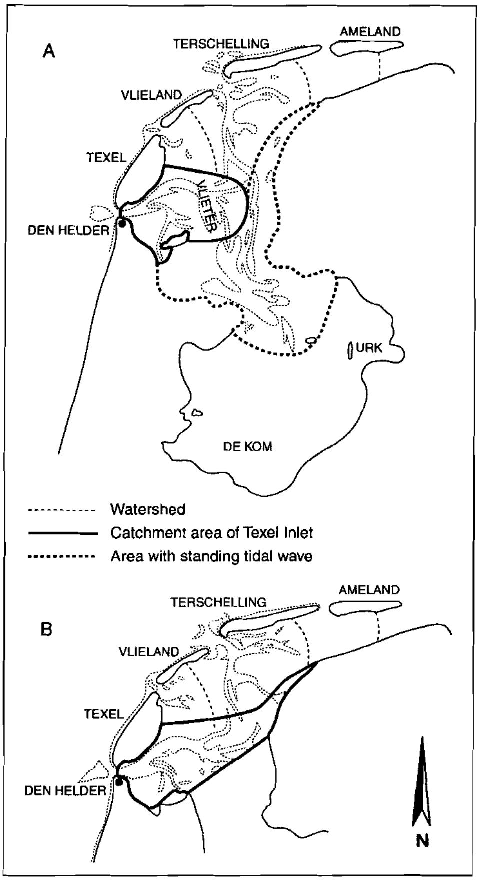


Figure 43: Catchment area of Texel inlet before (A) and after (B) the closure of the Zuider Zee (Klok & Schalkers, 1980).

marine to brackish water backwater with an open connection to the Wadden Sea. It was influenced by the tidal wave entering it. The size of the Zuider Zee basin was such, that the reflected outgoing tidal wave interfered with the incoming tidal wave, resulting in a standing wave of low tidal amplitude (Klok & Schalkers, 1980). As a consequence, the tidal prism through the nearby Texel Inlet was also low (Fig. 42A). After closure, the tidal amplitude increased considerably (approx. 20% near Texel Inlet; Fig. 42B) and the catchment area of Texel Inlet within the remaining part of the Wadden Sea became larger (Fig. 43). In this way the tidal prism of Texel Inlet and, as a consequence, also its depth increased (Sha, 1990a).

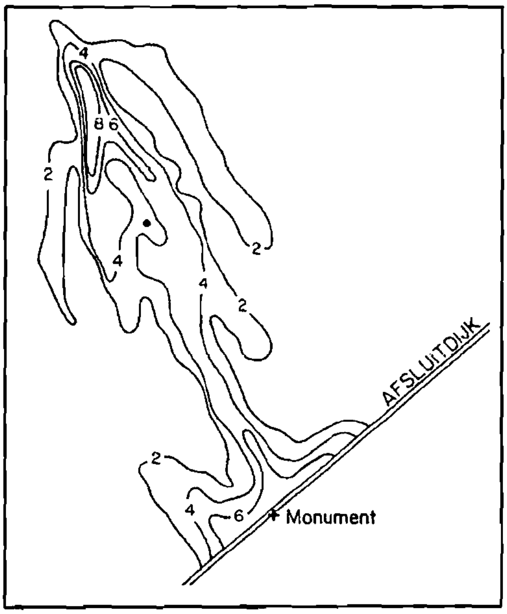


Figure 44: Sedimentation (in m) in the former Vlieter channel based on a comparison of soundings in 1928/34 and 1975. The dot indicates the place of the corings (Berger et al., 1987).

The closure of the large Zuider Zee tidal basin also blocked the tidal channels draining it. In these channels the tidal currents were reduced to almost zero. The infill of one of these channels, the Vlieter (Fig. 44), was studied by Berger et al. (1987). Above the coarse sandy/shelly channel bed of 1932, a sequence of 1-3 cm thick mainly undisturbed layers formed, consisting of fine sand/silt, alternating with silt/clay and occasionally intercalated medium sand layers. Based on ^{210}Pb measurements, average sedimentation rates were calculated to have been 6.5 cm.yr^{-1} (Berger et al., 1987; Eisma et al., 1989). When compared with partially abandoned tidal channels, such as in the Oosterschelde (see Fig. 22) and the Zoutkamperlaag, where the rate of sedimentation is of the order of dm.yr^{-1}, the Vlieter is being filled quite slowly (cm.yr^{-1}). This must have been due to the drastic reduction of tidal currents through the Vlieter channel, the ability of the system to transport sediment being a major constraint on the infill of such channels (Oost et al., 1993).

Eastern Dutch Wadden Sea

From Greek(?), Roman and Early Medieval sources it can be concluded that at least since O A.D. the eastern parts of the coast consisted of barrier islands and intertidal flats. The geological reconstructions are thus in accordance with the early historical sources (De Jong, 1984; Sha, 1992; Bosch & Vos, 1992). The landscape of the Province of Friesland, east of the 'Afsluitdijk', is quite flat because it largely consists of old tidal flat and tidal marsh deposits. Prehistoric man entered the area in several waves: the first around 500 B.C., the second around 50 B.C., the third 600-700 A.D., and the last between 900-1100 A.D. Man started to influence the morphology of the area by constructing dwelling mounds on which their settlements were protected against storm floods. These so-called 'terpen' or 'wierden' were built between 300 B.C. and 1200 A.D. The total amount of earth and dung piled up in these artificial mounds is estimated at about 30 times the amount of building material used in the Great Pyramid (Bruun, 1986). The 'terpen' were often situated on the levees of the larger creeks in the higher intertidal marshes. In later times, much of the fertile earth of these dwelling mounds was removed for agricultural purposes. Many of the churches in the centre of the old villages are still standing on the top of such mounds, the remnants of larger 'terpen'.

Since at least 12 B.C. peat has been dug for heating and later also for salt production. Because of this and also for agriculture, the area was drained. As in the western Wadden Sea, large parts of the mainland were flooded, being at least partly caused by the mining of peat (Edelman, 1974; Griede, 1978). In the period of c. 800-1300 A.D. the 'Middelzee' (Van der Spek, 1995), the 'Lauwerszee' (Griede, 1978) and the Dollard estuary were formed. Loss of land in the Dollard continued up to the 16th century.

Dyke construction began in the 11th or 12th century A.D., initially to protect land from the sea, but later also to reclaim land. Sedimentation in the higher intertidal marshes was actively encouraged in order promote land reclamation. Remnants of the old dykes can be observed in various places of the landscape. The total area of the dyked land is considerable (Tab. 1). In this way the tidal catchment area decreased strongly. Simultaneously the inlet channels, the channels in the backbarrier area and the ebb-tidal deltas became smaller, due to an almost linear relationship between these features and the tidal prism (see above). As the tidal prism decreased, the ebb-tidal deltas were strongly eroded by wave action. Part of the eroded sand served as infill for the inlet channels and the channels in the backbarrier areas (cf. Oost & de Haas, 1992; Van der Spek, 1995), the process vividly illustrating the concept of the sand-sharing system.

The formation of new washover channels on the islands was strongly reduced by the protection of the vegetation cover (since at least 1354). This led to the stabilisation of dunes on the barrier islands and probably helped to limit the coastward retreat of the islands (Ehlers & Kunz, 1993; Oost & Dijkema, 1993).

Table I: Dyke construction in the Dutch Wadden Sea, Zuider Zee (now lake 'IJsselmeer') and 'Dollard' in km². Data Mazure et al. (1974).

Period (century)	Original character of the dyked area			
	Intertidal/ marshes	Subtidal flats	Water-covered	Total
11th-12th	52			52
13th-16th	421			421
17th-18th	242	159		401
19th	313	56	19	388
20th	115 *	1650	2106	3871
Total	1143	1865	2125	5133

*) In the 20th century general land reclamation was reduced in favour of the closure of the Zuider Zee and 'Lauwerszee' (Anonymous, 1981).

Around 1300 A.D. the following islands and inlets were probably present from west to east: island Terschelling, Ameland Inlet (Bordena, 1297), island Ameland (Ammeland, 1307), Zoutkamperlaag Inlet (Scudbalwe, 1309), island Schiermonnikoog (1323), Lauwers Inlet (Lavicam, 1287), island Bosch (mentioned 1535), Schild Inlet (?), island Rottummeroog (1354) and the Ems River. In the eastern Wadden Sea several small islands were situated (Cornasant, 1323; Bant and Heffesant; Chapter 2).

Take-over of the drainage of the 'Lauwerszee'

One of the most dramatic changes in the backbarrier region of the eastern Dutch Wadden Sea has been the take-over of drainage of the 'Lauwerszee' embayment. Around 1300 A.D. the watershed of the barrier island Schiermonnikoog was situated on the coast of the province of Friesland (Fig. 45). The precursor of the Zoutkamperlaag Inlet (west of Schiermonnikoog) was probably small and had no or only a small connection with the 'Lauwerszee' embayment, which was entirely drained by the, at that time, large Lauwers Inlet situated east of Schiermonnikoog (cf. Bosch & Vos, 1992). The watershed of Schiermonnikoog was probably breached in the period 1350-1450. Around 1500 A.D. both the Lauwers Inlet and the Zoutkamperlaag Inlet drained the 'Lauwerszee'. Later the drainage was rather rapidly taken over fully by the Zoutkamperlaag which, as a result, increased in cross-sectional area. Judicial information shows that by 1556 the Lauwers Inlet had lost its connection with the 'Lauwerszee' (Formsma, 1954, 1958). After the breaching, the western inlet (Zoutkamperlaag) took over the drainage of the 'Lauwerszee'. Schiermonnikoog became approx. 6 km (1300-1850) longer at its eastern head, the Lauwers Inlet also shifting eastward, while gradually growing smaller. At the same time the small inlet at the western side of Schiermonnikoog and its ebb-tidal delta grew. Erosion of the western part of Schiermonnikoog, amounting to approx. 3.5 km over the period 1300-1850, further facilitated the connection of the Zoutkamperlaag with the 'Lauwerszee' (Oost, in prep b, chapter 2).

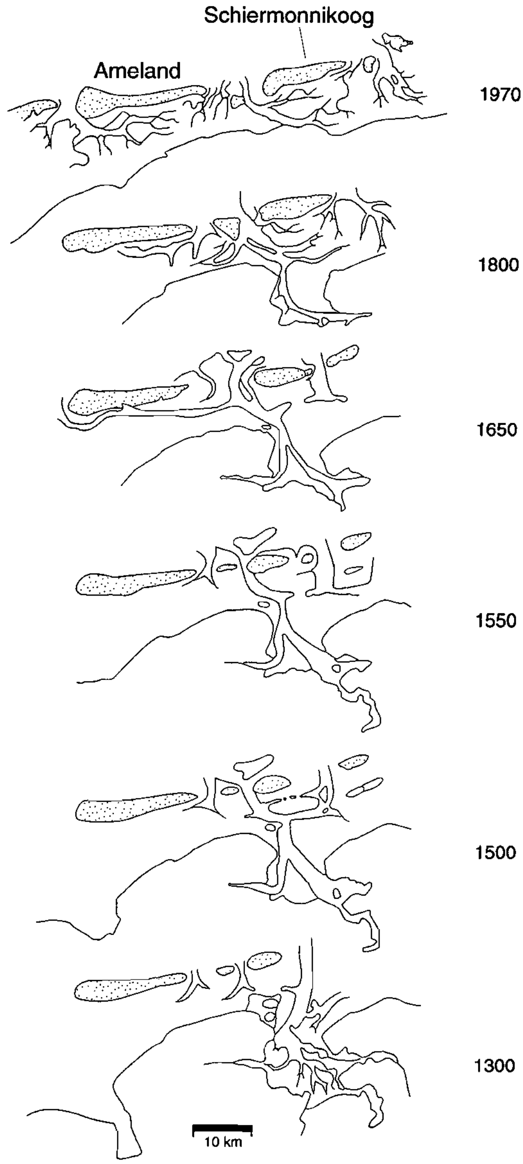


Figure 45: Sketch of the physiographic evolution of the eastern Dutch Wadden Sea.

Around 1700 A.D., the hamlets 'Westerburen', 'de Dompen' and 'Oosterburen' were situated west of the present village of Schiermonnikoog. Due to the migration of dunes towards the east the village of 'Oosterburen' eventually had to be abandoned. Strong erosion occurred at the western side of the island, especially during the storms of 1717, 1720, 1756 and 1760, and the hamlets 'Westerburen' and 'de Dompen', as well as the manor house of the island ruler had to be given up. The present village was built in the period 1721-1761. Erosion at the SW side continued until 1843, when the low-water line reached the present dune area. A new tidal flat on the SW side developed after 1843, becoming quite high (mainly higher intertidal) after the closure of the 'Lauwerszee' in 1969 (Oost & de Haas, 1992). Around 1538/45 the backbarrier island Cornasant merged with the barrier island Bosch. The All Saints Flood of 1570 changed the barrier island Bosch and the backbarrier island Heffesant into sandy shoals, the latter disappearing shortly afterwards. Although man tried to restore the dunes by planting grass, the island Bosch did not recover from this major blow. The shoal started to migrate to the south at about 1650 and disappeared in the 19th century (Lang, 1958). Later a new shoal developed at the place where Bosch had been situated in 1600 (Chapter 2).

The changes which resulted from the take-over of drainage of the 'Lauwerszee' are a good illustration of how a dynamic equilibrium (Lauwers Inlet drains 'Lauwerszee') can change in a rather short time to a new dynamic equilibrium (Zoutkamperlaag Inlet drains 'Lauwerszee').

In contrast to Schiermonnikoog, the more western barrier island of Ameland migrated less strongly to the east. Some erosion occurred at the western head of the island (1.5 km in the period 1650-1970) and sedimentation at the eastern head (3.5 km in the period 1650--1991, especially in the period 1800-1991). The difference to Schiermonnikoog can be explained by the fact that the inlet of Ameland was not located as asymmetrically to the drainage area as was the case for the Zoutkamperlaag. Also, the inlets were originally situated more then 10 km to the west: nowadays the eastern end of Terschelling.

Sedimentation in a sheltered embayment: The infill of the Dollard

Along the Wadden Sea coast several semi-enclosed embayments have existed and still exist (in the Dutch Wadden Sea: Zuider Zee, 'Middelzee', 'Lauwerszee' and 'Dollard'). Such embayments, which are enclosed by the mainland on three sides, strongly influence sedimentation. A good example is the Dollard embayment (Fig. 31). The present morphology of the Dollard is the result of: 1. the Holocene rise of sea level; 2. tectonic subsidence; 3. compaction of peat and clay; 4. erosion of peat; 5. human activities (land reclamation).

As a result of (1) and (2) relative sea level rose and the coastline shifted landward until about 1000 A.D. At this time the people who lived in this area began to build dykes. The marshy land behind the dykes was drained in the interest of agriculture, which resulted in an increased compaction of clay and peat. Man thus stimulated an increase of subsidence in a period of rising sea level. The dykes were breached from time to time by severe storms and

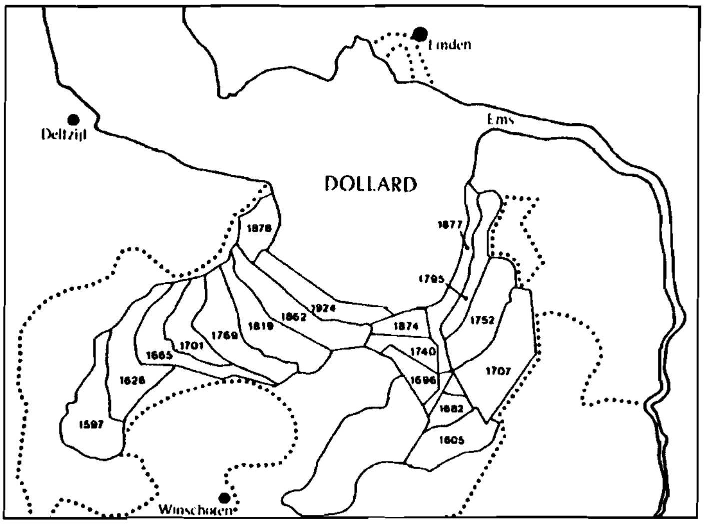


Figure 46: Land reclamation in the Dollard Estuary. Numbers show the year of reclamation.

huge areas were flooded. As a result, large amounts of clay and peat were eroded from the areas behind the dykes. In this way the formation of the Dollard estuary began in the 13th century. The Dollard had its largest extension around 1520 A.D.

The relatively sheltered embayment allows the settling of large amounts of sediments (clay to very fine sand). Sedimentation rates are high and cause a gradual silting up of the embayment. This was encouraged by man by the construction of small sedimentation fields, protected by rows of sticks, by which wave action was damped. Subsequently more and more of the land was reclaimed, until the Dollard reached its present shape (Fig. 46; Van Voorthuysen & Kuenen, 1960; Streif, 1982a, b). The decreasing size of the embayment resulted in a reduction of the tidal prism and hence strong infilling of the main channel (reduction of cross-sectional area).

Lateral migration of channels in sheltered embayments is strongly restricted by the clayey sediments (e.g. Irion, 1992), but also because the drainage area itself can not shift (much), in contrast to other parts of the Wadden Sea. Gradual infilling of the main channel(s) will thus often result in nested sandy channel-fill complexes, as was the case in the 'Lauwerszee' (Chapter 2).

Sedimentary effects of the closure of the 'Lauwerszee'

Tidal systems are able to react rapidly to changes in the hydrodynamic regime, whereas the strong tidal and wave forces, which dominate the system, allow the rapid transfer of massive amounts of sediments so that old equilibria can be restored or new ones can be reached (sand-sharing system). This is illustrated in great detail by the effects of the closure of part of the backbarrier system of the Zoutkamperlaag (Fig. 7; Oost, 1995, Chapter 5).

Due to the closure of the 'Lauwerszee' in 1969, the tidal prism of the Zoutkamperlaag was reduced from $305*10^6$ m^3 to $200*10^6$ m^3 (Van Sijp, 1989a, b). As a consequence of this permanent change, the system was no longer in equilibrium with the hydrodynamic conditions and both the ebb-tidal delta and the backbarrier area started to change towards a new morphodynamic equilibrium. The ebb-tidal current diminished, as a result of which the ebb-tidal delta of the Zoutkamperlaag was affected by wave erosion. Erosion was dominant between -4 m DOL (Dutch Ordnance Level = about Mean Sea Level) and -12 to -13 m DOL, which is the local storm-erosion base (Figs. C5 to C10, and D1). From the total amount of sediment eroded in the period 1970-1987 ($26*10^6$ m^3) a small part was transported longshore and perhaps offshore by storms. The larger part of this sand was transferred into the backbarrier area in the period 1970-1987. A relatively fast rotation of the outer channels and inlet was brought about by an increase in sedimentation at the western side and an increase of erosion at the eastern side of the channels, caused by the decrease in tidal volume and the relative increase of wave influence (Oost & de Haas, 1992).

Within the ebb-tidal delta, sediment was concentrated by waves into a large intertidal recurved bar with a length of 4 km in the E-W direction and 4 km in the N-S direction. This bar already started to form before closure of the 'Lauwerszee', after abandonment of an outer tidal channel directly west of Schiermonnikoog. The large size of the bar (deposition of $6*10^6$ m^3 above the -2 m DOL line in the period 1970-1987) is due to the large amount of sand which became available by the partial erosion of the ebb-tidal delta (Biegel, 1991a; Oost & de Haas, 1992; Biegel & Hoekstra, 1995). The large dimensions of the recurved bar affected the direction of wave approach over the period 1970-1987, aligning them more or less perpendicular to it. The waves, in their turn, contributed to the build-up of the recurved bar by accumulating large amounts of sand. Originally the recurved bar surrounded a small embayment. After the bar was breached during a series of storms (1990), strong erosion by tidal currents occurred and gullies of up to -5 m DOL were formed. Moreover, storms and aeolian erosion tended to erode the bar, as a result of which it has no significant preservation potential.

As stated above, a large part of the sediment (mainly sand) was transferred to the backbarrier area in the period 1970-1987 (Oost & de Haas, 1992). Strong sedimentation took place in the main gorge due to the deceleration of the currents. In the backbarrier area the change in tidal volume caused rapid vertical sedimentation in the main channel. The sediments consist of fine sand, clay or an alternation of both. The alternations have a rhythmic appearance (winter/summer couplets), with climbing ripple structures, linzen, loadcasts and bioturbation (Fig. 25). In the first period (1970-1975) the sediments were partly derived from the surrounding tidal flats (Winkelmolen & Veenstra, 1974). After 1979 sediment was mainly derived from the ebb-tidal delta. In total, some $30*10^6$ m^3 of sediment (fine sand and clay) was deposited in the backbarrier area during the period 1970-1987 (Figs. C5 to C10, and D1; Oost & de Haas, 1992, 1993).

The deposits in the channels largely represent a vertical build-up. This results in nested channel deposits and vertical infill with clays and sands, features commonly observed in the rock record. Under the conditions of a decreased tidal prism, any newly established channels will be shallower than the previous ones. The preservation potential of these older channel deposits is therefore rather good (Chapter 5).

Another result of the closure of the 'Lauwerszee' was that the position of the watershed south of Schiermonnikoog was shifted. In the period 1970-1979 the watershed slowly shifted towards the east; after 1979 the eastward migration accelerated. To the west of the watershed new, small channels were formed (Figs. C5 to C10, and D1). After closure of the 'Lauwerszee' the large width of the Zoutkamperlaag main backbarrier channel resulted in a faster transport of water towards the watershed (Van Parreeren, 1980; Postma & van Parreeren, 1982). No significant changes have been observed in the inlet system east of the watershed. The lateral shift of the watershed must therefore have been generated mainly by the closure of the 'Lauwerszee' (Oost & de Haas, 1992). Due to the migration of the watershed, the (abandoned) channel deposits of the inlet system east of Schiermonnikoog were covered, thereby improving their preservation potential.

Future developments: Greenhouse effect and sea-level rise

In the Wadden Sea the relative rise of sea level amounted on average to 17 cm over the last 100 years. About half of this was due to the eustatic rise of sea level and the other half to subsidence and compaction.

Due to rising concentrations of carbon dioxide and some trace gases (CH_4, N_2O, CFK-11, CFK-12) in the atmosphere, the radiation of heat from the atmosphere to outer space is reduced (Houghton et al., 1990). As a result, a global warming of the lower atmosphere is anticipated in the near future: the Greenhouse effect. It is expected that this warming will cause thermal expansion of ocean water, melting of small icecaps, the retreat of glaciers on Greenland and in high mountains, and increased accumulation of snow on Antarctica (Warrick & Oerlemans, 1990). The net effect will be an increase in the volume of sea water, resulting in an estimated global sea-level rise of 0.31 to 1.10 m by the year 2100 (Warrick & Oerlemans, 1990). To date it has not been possible to conclusively demonstrate an acceleration in global sea-level rise (Ekman, 1988; Woodworth, 1990; Warrick & Oerlemans, 1990). Moreover, it has been suggested that the Greenhouse effect will result in an increased number of storms. Thus the mixed energy coast may shift from tide-dominated to relatively more wave-dominated, resulting in less inlets and more washover channels (cf. Hayes, 1979).

The rates of future sea-level rise predicted above are comparable to early Holocene rates. The important difference is that during the early Holocene the complete coastal system was mobile, being freely able to retreat landwards in response to the rise of sea level. At present, however, dykes block any landward migration of the coastal system. Extensive research is presently in progress to assess the effects of a relative sea-level rise along the Dutch Wadden Sea coast (Bruun, 1986; Nichols, 1989; Dijkema et al., 1990; Misdorp et al., 1990; Eysink, 1992, 1993; Eysink & Biegel, 1992; Van der Spek & Beets, 1992; Anonymous, 1993; Oost & Dijkema, 1993). Over a short time period, i.e. in the coming centuries, it is expected that sea-level rise will be compensated by increasing sedimentation on the backbarrier tidal flats (Anonymous, 1993; Eysink, 1993; Oost & Dijkema, 1993). Since sedimentation is commonly lagging behind a sea-level rise, it is expected that under conditions of accelerated sea-level rise a small percentage of the intertidal flats will change into subtidal areas (Eysink, 1993).

The material needed to compensate sea-level rise will be derived from the coastal zone (to approximately -20 m DOL) of the barrier islands (Anonymous, 1993) and probably to a small extent also from deeper offshore parts (Oost & Dijkema, 1993). Thus, under the conditions of an artificially fixed mainland coastline, the anticipated response of the barrier and backbarrier systems to a further rise of sea level would be a retreat of the barrier islands and a narrowing of the backbarrier tidal basins. Artificial fixation of the Wadden Sea barrier islands in their present position might, in the long run, lead to a sediment deficiency (Ehlers & Kunz, 1993; Oost & Dijkema, 1993) and thus to a permanent drowning of large parts of the intertidal backbarrier region.

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REFERENCES

Aller, R.C., 1982: The effects of macrobenthos on chemical properties of marine sediment and overlying water. In: McCall, P.L. & Tevesz, M.J.S. (Eds.): Animal-Sediment Relations: 53-102; New York, London, Plenum Press.

Anonymous, 1981: Zandwinning in de Waddenzee, resultaten van een hydrografisch-sedimentologisch onderzoek. Stuurgroep, Rijkswaterstaat, Directie Friesland, 40 pp.

Anonymous, 1993: Invloed van gaswinning op de Waddenzee. Publ. of Rijkswaterstaat, The Hague, 33 pp.

Beets, D.J. & van der Valk, L. & Stive, M.J.F., 1992: Holocene evolution of the coast of Holland. Mar. Geol., 103: 423-443.

Berger, G.W., Eisma, D. & van Bennekom, A.J., 1987: 210Pb derived sedimentation rate in the Vlieter, a recently filled-in tidal channel in the Wadden Sea. Netherl. J. Sea Res., 21: 287-294.

Beukema, J.J., 1977: Dierlijk leven in en op de bodem. In: Abrahamse, J. & Joenje, W. & van Leeuwen-Seelt, N. (Eds.): Waddenzee, 2nd. pr.: 125-131, Harlingen

Beukema, J.J., 1982: Annual variation in reproductive success and biomass of the major macrozoobentic species living in a tidal flat area of the Wadden Sea. Netherl. J. Sea Res., 16: 37-45.

Beukema, J.J., Cadée, G.C. & Jansen, J.J.M., 1977: Variability of growth rate of *Macoma balthica* (L.) in the Wadden Sea in relation to availability of food. In: Keegan, B.F. & Ceidigh, P.O. & Boaden, P.J.S. (Eds.): Biology of Benthic Organisms, 11th Eur. Symp. on Mar. Biology, Galway, Oct. 1976: 69-77; London, Pergamon.

Biegel, E.J., 1991a: De ontwikkelingen van de ebgetijde delta en het kombergingsgebied van het Friesche Zeegat in relatie tot de sluiting van de Lauwerszee. Univ. of Utrecht, Fac. of Geogr. Sci., GEOPRO, 1991.07: 57 pp.

Biegel, E.J., 1991b: Equilibrium relations in the ebb tidal delta, inlet and backbarrier area of the Frisian inlet system. Univ. of Utrecht, Fac. of Geogr. Sci., GEOPRO, 1991.028/GWAAO-91.016: 79 pp.

Biegel, E.J. & Hoekstra, P., 1995: Morphological response characteristics of the Zoutkamperlaag, Frisian Inlet, The Netherlands to a sudden reduction in basin area. In: Flemming, B.W. & Bartholomä, A. (Eds.): Tidal signatures in modern and ancient sediments. Int. Assoc. Sedimentol., Spec. Publ., 24: 85-99.

Bilse, D.P., 1993: A large scale model of the morphological behaviour of the outer delta of tidal inlets. Delft Hydraulics, Report H 1215: 42 pp.

Bleuten, W., 1971: Longshore en ripcurrents op het strand van Schiermonnikoog. Int. Rep. Fac. of Geogr. Sci., Univ. of Utrecht: 16 pp; Utrecht.

Bosch, A. & Vos P.C., 1992: Paleogeografische reconstructie van het Lauwersmeergebied, concept. State Geol. Survey-Project 40009: 22 pp.

Bromley, R.G., 1990: Trace Fossils; Biology and Taphonomy. 280 pp.; London, Unwin Hyman.

Bruun, P., 1986: Worldwide impact of sea-level rise on shore lines. Rijkswaterstaat, Dienst Getijdewateren, The Hague, Publ. 86-8-713: 81-111.

Bruun, P. & Viggosson, G., 1977: The wave pump: conversion of wave energy to current energy. Proc. ASCE, J. waterway, port, coast. ocean div., 103: WW4.

Buhr, K.J. & Winter, J.E., 1977: Distribution and maintenance of a *Lanice conchilega* association in the Weser Estuary (FRG), with special reference to the suspension-feeding behaviour of *Lanice conchilega*. In: Keegan, B.F. & Ceidigh, P.O. & Boaden, P.J.S. (Eds.): Biology of Benthic Organisms, 11th Eur. Symp. on Mar. Biology, Galway, Oct. 1976: 101-113; London, Pergamon.

Cadée, G.C., 1976: Sediment reworking by *Arenicola marina* on tidal flats in the Dutch Wadden Sea. Neth. J. Sea Res., 10: 440-460.

Cadée, G.C., 1989: Size-slective transport of shells by birds and its palaeoecological implications. Palaeontol., 32: 429-437.

Cadée, G.C., 1990: Feeding traces and bioturbation by birds on a tidal flat, Dutch Wadden Sea. Ichnos, 1: 23-30.

Cadée, G.C., 1992: Aeolian transport and left/right sorting of *Mya* shells (mollusca, bivalvia). Palaios, 7: 198-202.

Cadée, G.C., in press: Eider, Shelduck, and other predators, the main producers of shell fragments in the Wadden Sea, palaeoecological implications. Paleontology.

Christiansen, C. & Miller, P.F, 1983: *Spartina* in Mariager Fjord, Denmark: the effect on sediment parameters. Earth Surface Processes and Landforms, 8: 55-62.

Creutzberg, F. & Kuipers, B.R., 1984: Cursus Mariene Ecosystemen (praktische gedeelte). Int. Rep. Neth. Inst. Sea Res.: 40 pp; Texel.

Creutzberg, F. & Postma, H., 1979: An experimental approach to the distribution of mud in the southern North Sea. Netherl, J. Sea Res., 13: 99-116.

Crisp, D.J. & Meadows, P.S., 1962: The chemical bases of gregariousness in cirripedes. Roy. Soc. London Proc., B 150: 500-520.

Dankers, N. & Koelemaij, K., 1989: Variations in the mussel population of the Dutch Wadden Sea in relation to monitoring. Helogoländer Meeresunters., 43: 529-535.

Dankers, N., Koelemaij, K. & Zegers, J., 1989: De rol van de mossel en de mosselcultuur in het ecosysteem van de Waddenzee. RIN-Rep., 89/9: 66 pp.

Davis, R.A., Fox, W.T., Hayes, M.O. & Boothroyd, J.C., 1972: Comparison of ridge and runnel systems in tidal and non-tidal environments. J. sedim. Petrol., 32: 413-421.

de Boer, M., 1979: Morfologisch onderzoek Ameland. Verslag van het onderzoek op het Amelander Wantij in 1973. Rijkswaterstaat, Directie Waterhuishouding en Waterbeweging, District kust en zee, Studiedienst Hoorn, Nota WWKZ-79.H005.

de Boer, M., Kool, G. & Lieshout, M.F., 1991: Erosie en sedimentatie in de binnendelta van het Zeegat van Ameland, 1926-1984, Deelonderzoek no.4. Rijkswaterstaat, Directie Noord-Holland, Rep. ANVX-91.H202: 42 pp.

de Boer, P.L., 1979: Convolute lamination in modern sands of the estuary of the Oosterschelde, the Netherlands, formed as the result of entrapped air. Sedimentol., 26: 283-294.

de Boer, P.L., 1981: Mechanical effects of micro-organisms on intertidal bedform migration. Sedimentol., 28: 129-132.

de Boer, P.L., Oost, A.P. & Visser, M.J., 1989: The diurnal inequality of the tide as a parameter for recognizing tidal influences. J. sediment. Petrol., 59: 912-921.

de Haas, H. & Eisma, D., 1992: Suspended sediment transport in the Ems-Dollard estuary. Netherl. J. Sea Res., 31: 37-42.

de Jong, J., 1984: Age and vegetational history of the coastal dunes in the Frisian Islands, The Netherlands. Geol. en Mijnb., 63: 269-276.

de Vriend, H.J. & Bakker, W.T., 1993: Sedimentary processes and morphological behaviour models for mixed-energy tidal inlets. Netherl. Centre coast. Res. Delft, Rep. H 1887: 25 pp.

Dean, R.G., 1988: Sediment interaction at modified coastal inlets: processes and policies. In: Aubrey, D.G. & Weishar, L. (Eds.): Hydrodynamics and Sediment Dynamics of Tidal Inlets. Lecture Notes on Coastal and Estuarine Studies, 29: 412-439; New York, Springer.

Dean, R.G. & Walton, T.L., 1975: Sediment transport processes in the vicinity of inlets with special reference to sand trapping. In: Cronin, L.E. (Ed.): Estuarine Research, Vol. 2: Geology and Engineering: 129-150; New York, Academic Press.

Dieckmann, R., Osterthun, M. & Partenscky, H.W., 1988: A comparison between German and North American tidal inlets. Proc. 21st Coast. Engng. Conf., ASCE, 3, 199: 2681-2691.

Dijkema, K.S., 1987a: Changes in salt-marsh area in the Netherlands Wadden Sea after 1600. In: Huiskes, A.H.L., Blom, C.W.P.M. & Rozema, J. (Eds.): Vegetation between Land and Sea: 42-49; Dordrecht, Junk.

Dijkema, K.S., 1987b: Geography of salt marshes in Europe. Z. Geomorphol., N.F., 31: 489-499.

Dijkema, K.S., 1989: Habitats of The Netherlands, German and Danish Wadden Sea, 1:100.000. 30 pp.; Texel, Leiden (RIN, Veth Foundn.).

Dijkema, K.S. & Bossinade, J.H., 1990: Vegetatieclassificatie van Waddenzeekwelders volgens een vast typenstelsel. RIN, Int. Rep., 90/15: 37 pp.

Dijkema, K.S., Bossinade, J.H., Bouwsema, P. & de Glopper, R.J., 1990: Salt marshes in the Netherlands Wadden Sea: rising high-tide levels and accretion enhancement. In: Beukema, J.J., Wolff, W.J. & Brouns, J.J.W.M. (Eds.): Expected Effects of Climatic Change on Marine Coastal Ecosystems: 173-188; Dordrecht, Kluwer.

Dijkema, K.S., Reineck, H.E. & Wolff, W.J., 1980: Geomorphology of the Wadden Sea Area. Rep. 1 of the Wadden Sea Working Group.: 135 pp; Rotterdam, Balkema.

Dijkema, K.S., van den Bergs, J., Bossinade, J.H., Bouwsema, P., de Glopper, R.J. & van Meegen, J.W.Th.M., 1988: Effecten van rijzendammen op opslibbing en omvang van de vegetatiezones in de Friese en Groninger landaanwinningswerken. Rijkswaterstaat, Dir. Groningen, Nota GRAN 1988-2010, RIN-Rep. 88/66, Texel & RIJP-Report 1988-33 Cbw, Lelystad: 119 pp.

Dörjes, J., 1982: Das Watt als Lebensraum. In: Reineck, H.-E. (Ed.): Das Watt, Ablagerungs- und Lebensraum, 3rd Edn.: 107-143; Frankfurt a. M., Kramer.

Dronkers, J., 1986: Tidal asymmetry and estuarine morphology. Netherl. J. Sea Res., 20: 117-131.

Edelman, T., 1964: De historische veranderingen in de natuurlijke gesteldheid van het Nederlandse waddengebied. In: Anderson, W.F., Abrahamse, J., Buwalda, J.D. & van Straaten, L.M.J.U. (Eds.): Het Waddenboek: 37-65; Zutphen, Thieme.

Edelman, T., 1974: Bijdrage tot de historische gegrafie van de Nederlandse kuststrook. Rijkswaterstaat, Dir. Waterhuishouding en waterbeweging, The Hague: 84 pp.

Ehlers, J., 1988: Morphodynamics of the Wadden Sea. 397 pp.; Rotterdam, Balkema.

Ehlers, J. & Kunz, H., 1993: Morphology of the Wadden Sea natural processes and human interference. Coastlines of the southern North Sea, Proc. 8th Symp. on Coastal and Ocean Management: 65-84.

Eisma, D., 1981: Supply and deposition of suspended matter in the North Sea. In: Nio, S.D., Shüttenhelm, R.T.E. & van Weering, T.C.E. (Eds.): Holocene Marine Sedimentation in the North Sea Basin. Int. Assoc. Sedimentol., Spec. Publ., 5: 415-428.

Eisma, D., Berger, G.W., Chen Wei-Yue & Shen Jian, 1989: Pb-210 as a tracer for sediment transport and deposition in the Dutch-German Wadden Sea. Proc. KNGMG Symp. Coastal Lowlands, Geol. and Geotechnol.: 237-253.

Eisma, D. & Wolff, W.J., 1980: The development of the westernmost part of the Wadden Sea in historical time. In: Dijkema, K.S., Reineck, H.-E. & Wolff, W.J. (Eds.): Geomorphology of the Wadden Sea Area. Rep. 1 of the Wadden Sea Working Group: 95-103; Rotterdam, Balkema.

Ekman, M., 1988: The world's longest continued series of sea level observations. Pure and Appl. Geophys., 127: 73-77.

Eysink, W.D., 1979: Morfologie van de Waddenzee, gevolgen van zand- en schelpenwinning, verslag literatuuronderzoek. Waterl. Lab., Rep. R-1336: 92 pp.

Eysink, W.D., 1987: Gaswinning op Ameland-oost, effekten van de bodemdaling, verslag onderzoek. Waterl. Lab. & RIN, Rep. H114: 53 pp.

Eysink, W.D., 1992: Impact of sea-level rise on the morphology of the Wadden Sea in the scope of its ecological function. Proposed set-up of a dynamic morphological model for Wadden Sea basins and estuaries based on empirical relations. Vol. 1. Rijkswaterstaat, Dienst Getijdewateren, ISOS*2 Proj., Phase 3: 38 pp.

Eysink, W.D., 1993: Impact of sea-level rise on the morphology of the Wadden Sea within the scope of its ecological function. General considerations on hydraulic conditions, sediment transports, sand balance, bed composition and impact of sea-level rise on tidal flats. Rijkswaterstaat Dienst Getijdewateren, ISOS*2 Proj., Phase 4: 35 pp.

Eysink, W.D. & Biegel, E.J., 1992: Impact of sea-level rise on the morphology of the Wadden Sea in the scope of its ecological function. Investigations on empirical morphological relations. Rijkswaterstaat Dienst Getijdewateren, ISOS*2 Proj., Phase 2: 73 pp.

Favejee, J.Ch.L., 1960: On the origin of the mud deposits in the Ems-estuary. In: van Voorthuysen, J.H. & Kuenen, Ph.H. (Eds.): Das Ems-Estuarium (Nordsee), ein sedimentologisches Symposium. Verh. kon. Nederl. geol. mijnbouwk. Genoot., 19: 147-151.

FitzGerald, D.M., 1988: Shoreline erosional-depositional processes associated with tidal inlets. In: Aubrey, D.G. & Weishar, L. (Eds.): Hydrodynamics and sediment dynamics of tidal inlets. Lecture Notes on Coastal and Estuarine Studies, 29: 186-225; New York, Springer.

FitzGerald, D.M., Penland, S. & Nummedal, D., 1984: Control of barrier island shape by inlet sediment bypassing: East Frisian Islands, West Germany. Mar. Geol., 60: 355-376.

Flemming, B.W., 1991: Holozäne Entwicklung, Morphologie und fazielle Gliederung der Ostfriesischen Insel Spiekeroog (südliche Nordsee). Senckenberg-am-Meer, Ber. 91/3: 51 pp.

Flemming, B.W. & Davis, R.A., Jr., 1992: Holocene evolution, morphodynamics and sedimentology of the Spiekeroog barrier island system (southern North Sea). Tidal Clastics 92, Wilhelmshaven, Germany, Field Guide-Book: 32 pp.

Flemming, B.W. & Davis, R.A., Jr., 1994: Holocene evolution, morphodynamics and sedimentology of the Spiekeroog barrier island system (southern North Sea). Senckenbergiana marit, 24: 117-155.

Flemming, B.W., Delafontaine, M.T. & Bartholomä, A., 1993: Morphodynamics and mass balance of biodepostion in a backbarrier mussel bank (Spiekeroog, Southern North Sea). In: Amler, M.R. W. & Tietze, K.W. (Eds.): Geologica et Palaeontologica, Sediment 93, Marburg, Kurzfassungen: 30-31.

Flemming, B.W., Schubert, H., Hertweck, G. & Müller, K., 1992: Bioclastic tidal-channel lag deposits: a genetic model. Senckenbergiana marit., 22: (3/6): 109-129.

Formsma, W.J., 1954: Een reis naar Schiermonnikoog in 1556. Groningse Volksalmanak 1954: 65-71.

Formsma, W.J., 1958: De grens tussen de provincies Groningen en Friesland in de Wadden. Groningse Volksalmanak 1958: 27-42.

Gerdes, G., Krumbein, W.E. & Reineck, H.E., 1985: Distribution and geological importance of marine microbial mats in the North Sea tidal flats. Facies, 12: 75-96.

Gerritsen, F., 1990: Morphological stability of inlets and channels of the western Wadden Sea. Rijkswaterstaat, Dienst Getijdewateren, Nota GWAO-90.019: 86 pp.

Gerritsen, F. & de Jong, H., 1985: Stabiliteit van doorstroomprofielen in het Waddengebied. Rijkswaterstaat Adviesdienst Vlissingen, Nota WWKZ-84.V016: 53 pp.

Goethe, F., 1937: Beobachtungen und Untersuchungen zur Biologie der Silbermöwe (*Larus argentatus* Pontopp.) auf der Vogelinsel Memmertsand. J. Ornithol., 85: 1-119.

Goethe, F., 1958: Anhäufungen unversehrter Muscheln durch Silbermöwen. Natur u. Volk, 88: 181-187.

Griede, J.W., 1978: Het ontstaan van Frieslands Noordhoek, een fysisch-geografisch onderzoek naar de holocene ontwikkeling van een zeekleigebied. Thesis Free Univ. Amsterdam: 186 pp.

Griede, J.W. & Roeleveld, W., 1982) De geologische en paleogeografische ontwikkeling van het Noordelijk Zeekleigebied. K.N.A.G. Geogr. Tijdschrift 16, 439-543.

Groenendijk, A.M., 1986: Establishment of *Spartina anglica* population on a tidal mudflat: a field experiment. J. Environmental Management, 22: 1-12.

Hayes, M.O., 1975: Morphology of sand accumulations in estuaries. In: Cronin, L.E. (Ed.): Estuarine Research, Vol. 2: Geology and Engineering: 3-22; New York, Academic Press.

Hayes, M.O., 1979: Barrier island morfology as a function of tidal and wave regime. In: Leatherman, S. (Ed.): Proceedings of the coastal symposium on barrier islands: 1-27; New York, Academic Press.

Hayes, M.O., 1980: General morphology and sediment patterns in tidal inlets. Sed. Geol., 26: 139-156.

Hertweck, G., 1982: Die Bewohner des Wattenmeeres in ihren Auswirkungen auf das Sediment. In: Reineck, H.-E. (Ed.): Das Watt, Ablagerungs- und Lebensraum, 3rd Edn.: 145-172; Frankfurt a. M., Kramer.

Hertweck, G., 1990: Die Zonierung der Lebensspuren im Watt der Jade, Südliche Nordsee. Beiträge zur Geologie und Paläontologie Norddeutschlands. Ber. Fachbereich Geowiss. Univ. Bremen, 10: 118-159.

Hertweck, G., 1992: Zonation of benthos and lebensspuren in the tidal flats of Jade Bay (Southern North Sea). Forschungsinstitut Senckenberg, Abt. Meeresforschung, Wilhelmshaven, Ber. 92/3: 13 pp.

Heynis, H., Berger, G.W. & Eisma, D., 1987: Accumulation rates of estuarine sediment in the Dollard area: comparison of Pb-210 and pollen influx methods. Netherl. J. Sea Res., 21: 295-301.

Hjulström, F., 1935: Studies of the morphological activity of rivers as illustrated by the River Fyris. Bull. geol. Inst. Univ. Uppsala, 25: 221-527.

Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (Eds.), 1990: Climate Change, the IPCC scientific assessment. 328 pp.; Cambridge, Cambridge Univ. Press.

Hubbard, D.K., Barwis, J.H. & Nummedal, D., 1977: Sediment transport in four South Carolina inlets. ASCE, Proc. Coastal Sediments '77: 582-601.

Hume, T.M. & Herdendorf, C.E., 1990: Morphologic and hydraulic characteristics of tidal inlets on a headland dominated low littoral drift coast, Northeastern New Zealand. J. coast. Res., Spec. Iss: Proc. Skagen Symp.: 527-563.

Irion. G., 1992: Morphological and sedimentological evolution of the Jade Bay (Southern North Sea). Tidal Clastics 92, Wilhelmshaven, Germany, Field Guide-Book: 15 pp.

Jarrett, J.T., 1976: Tidal prism-inlet area relationships. GITI, Rep. 3, US Army Eng. Waterways Exp. Stat. Vicksburg.

Johnson, D.W., 1919: Shore Processes and Shoreline Development. 584 pp.; New York, Wiley.

Kamps, L.F., 1962: Mud distribution and land reclamation in the eastern wadden shallows. Rijkswaterstaat Commun., 4: 73 pp.; The Hague.

Klok, B. & Schalkers, K.M., 1980: De veranderingen in de Waddenzee ten gevolge van de afsluiting van de Zuiderzee. Rijkswaterstaat, Rep. 78.H238: 13 pp.

König, D., 1948: *Spartina townsendii* an der Westküste von Schleswig-Holstein. Planta, 36: 34-70.

Lang, A.W., 1958: Gestaltungswandel des Emsmündungstrichters, Untersuchungen zur Entwickling der Emsmündung von der Mitte des 16. bis zum Beginn des 20. Jahrhunderts. Schriften der wirtschaftswissenschaftlichen Gesellschaft zum Studium Niedersachsens E.V., Reihe A, N.F., 58: 153 pp.

Leopold, M.F., Marteijn, E.C.L. & Swennen, C., 1984: Long distance transport of prey from the intertidal to high tide roosts by the Oystercatcher. Ardea, 73: 76-82.

Ligtendag, W.A., 1990: Van IJzer tot Jade, een reconstructie van de zuidelijke Noordzeekust in de jaren 1600 en 1750. Rijkswaterstaat, Meetkundige Dienst: 55 pp.

Luck, G., 1975: Der Einfluss der Schutzwerke der Ostfriesischen Inseln auf die morphologische Vorgänge im bereich der Seegaten und ihrer Einzugsgebiete. Mitt. Leichtweiss-Inst. Wasserb. Techn. Univ. Braunschweig, 47: 122 pp.

Luck, G., 1976: Inlet changes of the East Frisian Islands. Proc. 15th Conf. Coastal Eng., Am. Soc. Civ. Eng., Honolulu, 1938-1957.

Mazure, J.P. & many others, 1974: Rapport van de Waddenzeecommissie; Advies inzake de principiele mogelijkheden en de voor- en nadelen van inpolderingen in de Waddenzee. Rijkswaterstaat, Rep. Wadden Sea Comm., The Hague: 328 pp.

Misdorp, R., Steyaert, F., Hallie, F. & de Ronde, J., 1990: Climate change, sea-level rise and morphological developments in the Dutch Wadden Sea, a marine wetland. In: Beukema, J.J., Wolff, W.J. & Brouns, J.J.W.M. (Eds.): Expected Effects of Climatic Change on Marine Coastal Ecosystems: 123-131; Dordrecht, Kluwer.

Nichols, M.J., 1989: Sediment accumulation rates and relative sea-level rise in lagoons. Mar. Geol., 88: 201-220.

Niemeyer, H.D., 1986: Ausbreitung und Dämpfung des Seegangs im See- und Wattengebiet von Norderney. Jber. Niedersächs. Landesamt Wasserwirtschaft, Forschungsstelle Küste, Norderney, 1985, 37: 49-95.

Niemeyer, H.D., 1990: Morphodynamics of tidal inlets. In: Coastal Morphology Stichting Postacad. Ond. Civ. Techn. en Bouwtechn., Delft.

Nummedal, D. & Fischer, I.A., 1978: Process-response models for depositional shorelines: The German and Georgia bights. ASCE, Proc. 16th Coastal Engineering Conf.: 1215-1231.

Nummedal, D., Oertel, G.F., Hubbard, D.K. & Hine, A.C., 1977: Tidal inlet variability Cape Hatteras to Cape Canaveral. ASCE, Proc. Coastal Sediment '77: 543-562.

Nummedal, D. & Penland, S., 1981: Sediment dispersal in Norderneyer Seegate, West Germany. In: Nio, S.D., Shüttenhelm, R.T.E. & van Weering, T.C.E. (Eds.): Holocene Marine Sedimentation in the North Sea Basin. Internat. Assoc. Sedimentologists, Spec. Publ., 5: 187—210.

Nybakken, J.W., 1982: Marine Biology, an ecological approach. 446 pp.; New York, Harper and Row.

O'Brien, M.P., 1969: Equilibrium flow areas ofinlets on sandy coasts. J. Waterways and Harbors Div., ASCE, WWI: 43-52.

Oertel, G.F., 1972: Sediment transport on estuary entrance shoals and the formation of swash platforms. J. sediment. Petrol., 42: 857-863.

Oertel, G.F., 1975: Ebb-tidal deltas of Georgia estuaries. In: Cronin, L.E. (Ed.): Estuarine Research, 2: 267-276; New York, Academic Press.

Oost, A.P., in prep a: The influence of *Mytilus edulis* on fine grained sedimentation in the Dutch Wadden Sea. (chapter 6).

Oost, A.P., in prep b: Development of the eastern Dutch Wadden Sea with emphasis on the Frisian Inlet in historical times: morphodynamic behaviour of islands and inlets (chapter 2).

Oost, A.P., 1995: Sedimentological implications of changes in the ebb-tidal delta, the inlet, and the drainage basin of the Zoutkamperlaag tidal inlet (Dutch Wadden Sea), induced by a sudden decrease of tidal prism. In: Flemming, B.W. & Bartholomä, A. (Eds.): Tidal signatures in modern and ancient sediments. Intern. Assoc. Sedimentol., Spec. Publ., 24: 101-119 (chapter 5).

Oost A.P. & Baas J.H., 1994: The development of small-scale bedforms in tidal environments: an empirical model for unsteady flow and its applications. Sedimentol., 41, 883-903.

Oost, A.P. & Dijkema, K.S., 1993: Effecten van bodemdaling door gaswinning in de Waddenzee. IBN-Rep. 025: 133 pp.

Oost, A.P. & de Haas, H., 1992: Het Friesche Zeegat, morfologisch-sedimentologische veranderingen in de periode 1970-1987, een getijde inlet systeem uit evenwicht, Deel 1 & 2. Rep. Fac. Earth Sci., Univ. Utrecht, Coastal Genesis: 68 pp.

Oost, A.P. & de Haas, H., 1993: Het Friesche Zeegat. Morfologisch-Sedimentologische veranderingen in de periode 1927-1970, Deel 1 & 2. Rep. Fac. Earth Sci., Univ. Utrecht, Coastal Genesis: 83 pp.

Oost, A.P., de Haas, H., IJnsen, F., van den Boogert, J.M. & de Boer, P.L., 1993: The 18.6 year nodal cycle and its impact on tidal sedimentation. Sediment. Geol., 87: 1-11.

Oost, A.P. & Leopold, M.F., 1988: Shell transport by birds. Internat. Symp. on "Theoretical and Applied Aspects of Coastal and Shelf Evolution, Past and Future", jointly with the inaugural meeting of IGCP Proj. 274 "Quaternary Coastal Evolution; Case Studies, Models and Regional Patterns", Amsterdam, The Netherlands, 19-24 sept. 1988, Suppl. Abstracts: 29-30.

Postma, H., 1954: Hydrography of the Dutch Wadden Sea. Arch. Neerl. Zool., 10: 405-511.

Postma, H., 1961: Transport and accumulation of suspended matter in the Dutch Wadden Sea. Netherl. J. Sea Res., 1: 148-190.

Postma, H., 1967: Sediment transport and sedimentation in the estuarine environment. In: Lauff, G.A. (Ed.): Estuaries: 158-179; Washington, D.C. (Amer. Assoc. Advancem. Sci.).

Postma, H. & Dijkema, K.S., 1982: Hydrography of the Dutch Wadden Sea: Movements and properties of water and particulate matter. Final report on 'Hydrography' of the Wadden Sea Working Group, Rep. 2: 75 pp.; Rotterdam, Balkema.

Postma, J.T. & van Parreeren, D., 1982: Onderzoek naar de bevaarbaarheid van de wantijen Engelsmanplaat, Lutjewad en Hornhuizerwad. Rijkswaterstaat, Dir. Groningen, Nota 82-:, 26 pp.

Reineck, H.-E., 1982a: Allgemeines, Hydrographie. In: Reineck, H.-E. (Ed.): Das Watt, Ablagerungs- und Lebensraum, 3rd Edn.: 12-17; Frankfurt a. M., Waldemar.

Reineck, H.-E., 1982b: Rinnen. In: Reineck, H.-E. (Ed.): Das Watt, Ablagerungs- und Lebensraum, 3rd Edn.: 89-93; Frankfurt a. M., Kramer.

Reineck, H.-E., 1982c: Schichtungsarten und Wühlgefüge. In: Reineck, H.-E. (Ed.): Das Watt, Ablagerungs- und Lebensraum, 3rd Edn.: 69-80; Frankfurt a. M., Kramer.

Remane, A., 1951: Marine Schillablagerungen im Süßwasser und aeolischer Hydrobienschill. Kieler Meeresforsch., 8: 98-110.

Roeleveld, W., 1974) The Groningen coastal area: a study in Holocene geology and low-land physical geography. Berichten van de Rijksdienst voor Oudheidkundig Bodemonderzoek, 20-21, 1970-71 & Vol. 24, 1974, supplement.

Schoorl, H., 1973: Zeshonderd jaar water en land, bijdrage tot de historische Geo- en Hydrografie van de Kop van Noord-Holland in de periode 1150-1750. Verh. kon. Nederl. Aardrijkskdg. Genoot., 2: 534 pp.; Groningen, Wolters-Noordhoff.

Seed, R., 1976: Ecology. In: Bayne, B. L. (Ed.): Marine Mussels: Their Ecology and Physiology. Internat. biol. Progr., 10: 13-65; Cambridge, Cambridge Univ. Press.

Sha, L.P., 1989a: Cyclic morphologic changes of the ebb-tidal delta, Texel inlet, The Netherlands. Geol. en Mijnb., 68: 35-48.

Sha, L.P., 1989b: Sand transport patterns in the ebb-tidal delta off Texel Inlet, Wadden Sea, The Netherlands. Mar. Geol., 86: 137-154.

Sha, L.P., 1989c: Variation in ebb-delta morphologies along the West and East Frisian Islands, The Netherlands and Germany. Mar. Geol., 89: 11-28.

Sha, L.P., 1989d: Holocene-Pleistocene interface and three-dimensional geometry of the ebb-delta complex, Texel Inlet, The Netherlands. Mar. Geol., 89: 207-228.

Sha, L.P., 1990a: Sedimentological studies of the ebb-tidal deltas along the West Frisian Islands, the Netherlands. Ph.D.-Thesis. Geologica Ultraiectina. Mededelingen van het Instituut voor Aardwetenschappen der Rijksuniversiteit te Utrecht, 64: 159 pp.

Sha, L.P., 1990b: Surface sediment characteristics in the ebb-tidal delta of Texel Inlet, The Netherlands. Sediment. Geol., 68: 125-141.

Sha, L.P., 1992: Geological Research in the Ebb-Tidal Delta of 'Het Friesche Zeegat', Wadden Sea, The Netherlands. State Geol. Survey-Proj. 40010: 20 pp.

Sha, L.P. & de Boer, P.L., 1991: Ebb-tidal delta deposits along the West Frisian Islands (The Netherlands): processes, facies architecture and preservation. In: Smith, D.G., Reinson, G.E., Zaitlin, B.A. & Rahmani, R.A. (Eds.): Clastic Tidal Sedimentology. Canad. Soc. Petrol. Geol. Mem., 16: 199-218.

Steijn, R.C., 1991: Some Considerations on Tidal inlets. A literature survey on hydrodynamic and morfodynamic characteristics of tidal inlets with special attention to "Het Friesche Zeegat". Delft Hydraulics, Coastal Genesis, Rep. H840.45.: 109 pp.

Stive, M.J.F. & Eysink, W.D., 1989: Voorspelling ontwikkeling kustlijn 1990-2020, fase 3, deelrapport 3.1: dynamisch model van het Nederlandse kustsysteem. Rapport H825, Waterloopkundig Laboratorium, Delft: 66 pp.

Streif, H., 1982a: Geologie des Küstenraumes. In: Reineck, H.-E. (Ed.): Das Watt, Ablagerungs- und Lebensraum, 3rd Edn.: 24-30; Frankfurt a. M., Kramer.

Streif, H., 1982b: Der holozäne Meeresspiegelanstieg und die heutige Küstengestalt. In: Reineck, H.E. (Ed.): Das Watt, Ablagerungs- und Lebensraum, 3rd Edn.: 31-38; Frankfurt a. M., Kramer.

Thorson, G., 1966: Some factors influencing the recruitment and establishment of marine benthic communities. Netherl. J. Sea Res., 3: 267-293.

van de Plassche, O., 1982: Sea-level change and water-level movements in the Netherlands during the Holocene. PhD Thesis Free Univ. Amsterdam: 138 pp.

van den Berg, J.H., 1981: Rhythmic seasonal layering in a mesotidal channel fill sequence, Oosterschelde Mouth, the Netherlands. In: Nio, S.D., Schüttenhelm, R.T.E. & van Weering, T.C.E. (Eds.): Holocene Marine Sedimentation in the North Sea Basin. Internat. Assoc. Sedimentologists, spec. Publ. 5: 147-159.

van den Berg, J.H., 1986: Aspects of sediment- and morphodynamics of subtidal deposits of the Oosterschelde (The Netherlands). Rijkswaterstaat Commun., 43; The Hague.

van der Spek, A.J.F., 1995: Reconstruction of tidal inlet and tidal channel dimensions in the Frisian Middelzee, a former tidal basin in the Netherlands Wadden Sea. In: Flemming, B.W. & Bartholomä, A. (Eds.): Tidal signatures in modern and ancient sediments. Int. Assoc. Sedimentol., Spec. Publ., 24: 239-258.

van der Spek, A.J.F. & Beets, D.J., 1992: Mid-Holocene evolution of a tidal basin in the western netherlands: a model for future changes in the northern Netherlands under conditions of accelerated sea-level rise? Sediment. Geol., 80: 185-197.

van Diggelen, J., 1988: A comparative study on the ecophysiology of salt marsh halophytes. PhD Thesis Free Univ. Amsterdam: 208 pp.

van Eerdt, M.M., 1985: The influence of vegetation on erosion and accretion in salt marshes of the Oosterschelde, the Netherlands. Vegetatio, 62: 367-373.

van Gemerden, H., 1993: Microbial mats: A joint venture. Mar. Geol., 113: 3-25.

van Heuvel, Tj., 1991: Sedimenttransport in het Eems-Dollard estuarium, volgens de methode McLaren. Rijkswaterstaat, Dienst Getijdewateren, Rep. GWWS-91.002: 25 pp.

van Kleef, A.W., 1991: Empirical relationships for tidal inlets, basins and deltas. Fac. Geogr. Sci., Univ. Utrecht, Rep. GEOPRO 1991.19: 53 pp.

van Oosten, M.F., 1986: Bodemkaart van Nederland, Schaal 1:50,000, Toelichting op de kaarten van Vlieland, Terschelling, Ameland en Schiermonnikoog. Stichting Bodemkartering Wageningen: 129 pp.; Arnhem, van der Wiel.

van Parreeren, D., 1980: Waterloopkundige aspecten van de doorbaggering wantij Schiermonnikoog. Rijkswaterstaat, directie Groningen, Nota 80-29: 40 pp.

van Sijp, D., 1989a: Correcties op gemeten eb- en vloedvolume bij de omrekening naar gemiddeld getij, in het Friese Zeegat. Rijkswaterstaat, directie Friesland, Notitie ANW 89-02: 4 pp.

van Sijp, D., 1989b: Erosie vooroever Lauwersmeerdijk in relatie met de afsluiting van de Lauwerszee. Rijkswaterstaat, directie Friesland, Notitie ANW 89.19: 12 pp.

van Straaten, L.M.J.U., 1954: Composition and structure of Recent marine sediments in the Netherlands. Leidse geol. Mededel., 19: 1-110.

van Straaten, L.M.J.U., 1960: Transport and composition of sediments. In: van Voorthuysen, J.H. & Kuenen, Ph.H. (Eds.): Das Ems-Estuarium (Nordsee), ein sedimentologisches Symposium. Verh. kon. Nederl. geol. mijnbouwkdg. Genoot., 19: 279-292.

van Straaten, L.M.J.U., 1964: De bodem der Waddenzee. In: Anderson, W.F., Abrahamse, J., Buwalda, J.D. & van Straaten, L.M.J.U. (Eds.): Het Waddenboek: 75-151; Zutphen, Thieme.

van Straaten, L.M.J.U. & Kuenen Ph.H., 1958: Tidal action as a cause of clay accumulation. J. sediment. Petrol., 28: 406- 413.

van Veen, J., 1936: Onderzoekingen in de Hoofden. 252 pp.; The Hague, Landsdrukkerij.

van Voorthuysen, J.H., 1960: Die Foramineren des Dollart-Ems-Estuarium. In: van Voorthuysen, J.H. & Kuenen, Ph.H. (Eds.): Das Ems-Estuarium (Nordsee), ein sedimentologisches Symposium. Verh. kon. Nederl. geol. mijnbouwkdg. Genoot., 19: 237-269.

van Voorthuysen, J. & Kuenen, Ph.H., 1960: Das Ems-Estuarium (Nordsee), ein sedimentologisches Symposium. Verhandelingen van het K.N.G.M.G., geol. Ser. 2, 29: 300 pp.

Veenstra, H.J. & Winkelmolen, A.M., 1976: Size, shape and density sorting around two barrier islands along the north coast of Holland. Geol. en Mijnb., 55: 87-104.

Visser, M.J., 1980: Neap-spring cycles reflected in Holocene subtidal large-scale bedform deposits: a preliminary note. Geol., 8: 543-546.

von Weihe, K., 1979: Morphologische und ökologische Grundlagen der Vorlandsicherung durch *Puccinellia maritima* (*Gramineae*). Helgol. wiss. Meeresunters, 32: 239-254.

Vos, P.C., de Boer, P.L. & Misdorp, R., 1988: Sediment stabilisation by benthic diatoms. In: de Boer, P.L., van Gelder, A. & Nio, S.D. (Eds.): Tide-influenced sedimentary environments and facies: 511-526; Dordrecht, Reidel.

Vroom, M.G., Steyaert, F.H.I.M., Misdorp, R., Venema, H., Elgershuizen, J.H.B.W., Rakhorst, H.D. & Damoiseaux, M.A., 1989: Wadatlas Ministerie Verkeer en Waterstaat: 95 pp.; Amsterdam, Stadsdrukkerij.

Walton, T.L. & Adams, W.D., 1976: Capacity of inlet bars to store sand. Proc. 15th Conf. coast. Engng., Am. Soc. Civ. Eng., Honolulu, 1919-1938.

Warrick, R. & Oerlemans, J., 1990: Sea level rise. In: Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (Eds.): Climate Change, the IPCC scientific assessment: 259-261; Camb., Univ. Press.

Wiggers, A.J., 1960: Die Korngrössenverteilung der holozänen Sedimente im Dollart-Ems-estuary. In: van Voorthuysen, J.H. & Kuenen, Ph.H. (Eds.): Das Ems-Estuarium (Nordsee), ein sedimentologisches Symposium, Verh. kon. Nederl. geol. mijnbouwkdg. Genoot., 19: 75-81.

Winkelmolen, A.M. & Veenstra, H.J., 1974: Size and shape sorting in a Dutch tidal inlet. Sedimentol., 21: 107-126.

Wolff, W.J., 1986: De Waddenzee: eigenschappen van een dynamisch kustgebied. Flevobericht 252, RIJP, 11-22.

Woodworth, P.L., 1990: A search for accelerations in records of European mean sea level. Internat. J. Climatol., 10: 129-143.

Wunderlich, F., 1967: Die Entstehung von "convolute bedding" an Platenrändern. Senckenbergiana lethaea, 48: 345-349.

Wunderlich, F., 1979: Die Insel Mellum (südliche Nordsee) Dynamische Prozesse und Sedimentgefüge. I. Südwatt, Übergangszone und Hochfläche. Senckenbergiana marit., 11: 59-113.

Wunderlich, F., 1982a: Schichtbänke. In: Reineck, H.-E. (Ed.): Das Watt, Ablagerungs- und Lebensraum, 3rd Edn.: 81-88; Frankfurt a. M., Kramer.

Wunderlich, F., 1982b: Marken. In: Reineck, H.-E. (Ed.): Das Watt, Ablagerungs- und Lebensraum, 3rd Edn.: 95-105; Frankfurt a. M., Kramer.

Zagwijn, W.H., 1985: De paleogeografische ontwikkeling van Nederland in de laatste drie miljoen jaar. Geogr. Tijdschrift 9, 181-201.

Zagwijn, W.H., 1986: Nederland in het Holoceen, Geologie van Nederland. Deel 1. Rijks Geol. Dienst, Staatsdrukkerij, Den Haag, 46 pp.

Zagwijn, W.H., Beets, D.J., van den Berg, M., van Montfrans, H.M. & van Rooijen, P., 1985: Atlas van Nederland in 20 delen, deel 13, Geologie: 23 pp.; The Hague, Staatsuitgeverij.

Zijlstra, J.J.P., 1994: Sedimentology of the type section of the Late Cretaceous and Early Tertiary (Tuffaceous) Chalk of Northwest Europe. Geol. Ultraeictina, 119: 192 pp.

CHAPTER 2

DEVELOPMENT OF THE EASTERN DUTCH WADDEN SEA WITH EMPHASIS ON THE FRISIAN INLET SYSTEM IN HISTORICAL TIMES: MORPHODYNAMIC BEHAVIOUR OF ISLANDS AND INLETS

INTRODUCTION

The development of the Frisian Inlet system and adjacent areas is reconstructed on the basis of historical information. The reconstructions give insight into the processes involved in barrier island migration and inlet and channel migration on a large spatial/temporal scale. The interdependency of the tidal drainage basins, and the influence of sea-level rise, climate, storm frequency, and human interference is discussed.

PRESENT SETTING

The area studied is the eastern part of the Dutch barrier chain. It is part a system of barrier islands and inlets (Fig. 1) forming the North Sea coast of The Netherlands, Germany, and Denmark. The North Sea is a broad and shallow shelf sea on a passive continental margin. The tidal regime is semi-diurnal. The tidal wave propagates from the S(W) to the N(E), i.e., from Den Helder in Holland to Esbjerg in Denmark (Fig. 1), with a residual current towards the E. Along the Dutch Wadden Sea the tidal amplitude increases from 1.3 m near Den Helder to 2.8 m in Delfzijl. In the Frisian Inlet area, the centre of the area, the tidal range is 2.2 m and the tidal regime thus is mesotidal (cf. Hayes, 1979; Postma & Dijkema, 1982). Dominant wave height is 0.5-1.5 m at 20 m water depth (Data Directorate General for Public Works and Water Management = Rijkswaterstaat, further referred to as RWS). Following the classification of Hayes (1975, 1979) the coast thus is a tide-dominated mixed energy shoreline, influenced by both tides and waves.

The westernmost inlet of the Dutch barrier chain is Texel Inlet, with N of it the island Texel, followed to the E by the Inlet Eijerlandse Gat, the barrier island Vlieland, and the Vlie Inlet. East of that inlet (west of the area studied) lies the barrier island Terschelling. East of it the Ameland Inlet system and the barrier island Ameland are situated. East of Ameland, the Frisian Inlet system drains the backbarrier area. The Frisian Inlet consists of two separate inlet systems: the western Pinkegat and the eastern Zoutkamperlaag (Fig. 1). The mainland coast of the Pinkegat drainage area is part of the province Friesland. The mainland coast of the Zoutkamperlaag drainage area is largely part of the province Groningen (Fig. 1). A supratidal shoal (Engelsmanplaat) and the adjacent watershed separate

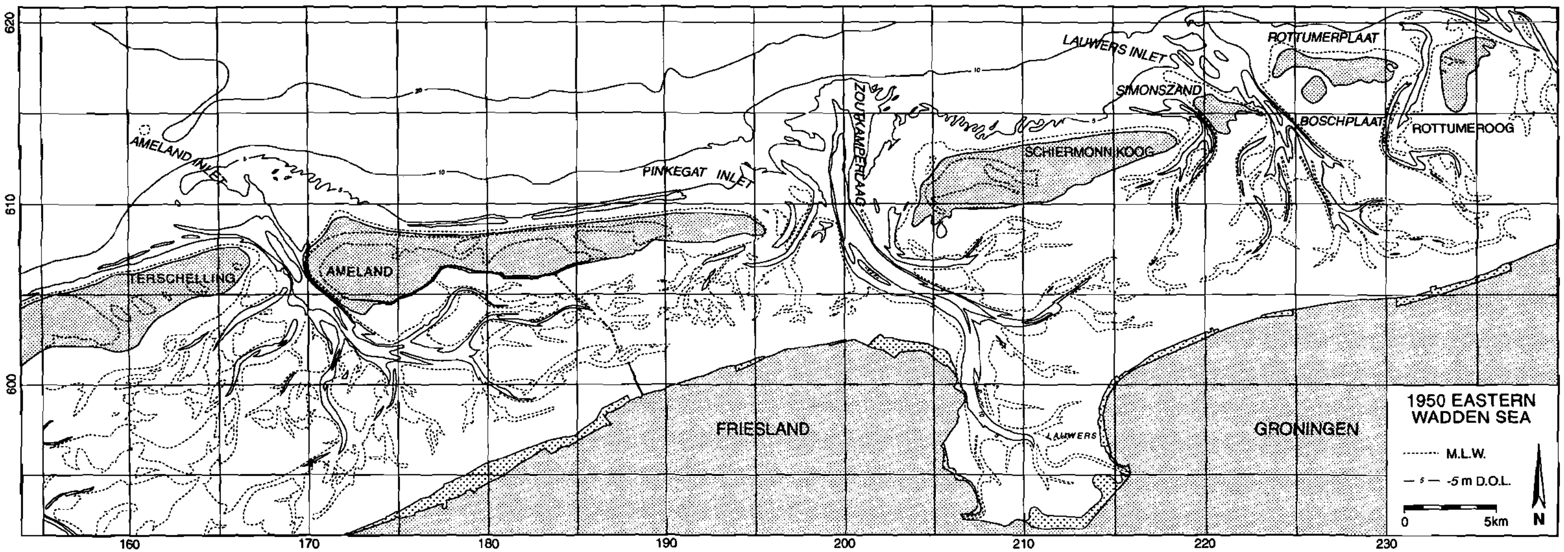

Figure 1: Overview of the eastern part of the Dutch Wadden Sea in 1950. Embayment below is the Lauwerszee embayment. East of Rottumeroog enters the Western Ems-Dollard Estuary.

the inlets and their largely intertidal backbarrier flood basins. The shoal is situated on top of a massive, relatively stable clay body at -5 m with reference to Dutch Ordnance Level (DOL ≈ mean sea level; Sha, 1992; Oost & De Haas, 1992). East of the Zoutkamperlaag Inlet system the barrier island Schiermonnikoog is situated. From there, from west to east, the following barrier inlets and barrier islands form the easternmost part of the Dutch Wadden system: the Eilanderbalg Inlet, the shoal Simonszand, the Lauwers Inlet, the shoal Rottumerplaat, the Schild Inlet, the shoal Rottumeroog and the Western Ems Estuary.

METHODS USED FOR THE RECONSTRUCTIONS

Data after 1300 A.D. allow detailed reconstructions of the easter Dutch Wadden Sea, with special reference to the Frisian Inlet, mainly based on historical data and maps (Appendices 1 and 2). For the period 400 B.C.-1300 A.D. only a general reconstruction of the Dutch Wadden Sea can be given, based on the geological reconstructions and the scarce historical data. Topographic reconstructions were made for 1300 and between 1500 and 1800 A.D. with time steps of 50 years. From 1800 to 1975 several composites were made of hydrographical charts. The reconstructions are based on the following sources:

1) copies of written sources or accurate transcriptions,
2) copies of old maps and earlier reconstructions,
3) reconstructions of the dykes,
4) geological data, and
5) other information, such as names and local history.

Ad 2 and 3: In general, for each specific part of the area those maps were chosen which were especially made for that area (cf. Ligtendag, 1990, 1991). For example, nautical charts were used to reconstruct the channel configuration, whereas island maps were used to reconstruct the barrier islands.

Throughout the text Roman numerals in superscript refer to the list of maps and sailing directions of appendix 1. At the moment there is no systematic and complete overview of the available maps of the Wadden Sea. The choice of maps, charts and sailing-directions used in the study is for the large part based on the overviews given by Isbary (1936), Vredenberg-Alink (1974), Lang (1958, 1968), Donkersloot-de Vrij (1981), Koeman (1985), Koenders (1986), Ligtendag (1990), and Van der Top (1992). These overviews were supplemented by other studies; among others Bodel Nyenhuis & Eekhoff (1846), Koeman (1967-1971), De la Ronciere & Mollat du Jourdin (1984), and Lang (1986). Furthermore, maps and charts of RWS Friesland were used for the reconstructions after 1800. Attention was given to avoid later copies of older maps, in particular for the islands.

Each reconstruction is compiled from several original maps and thus is a composite. All reconstructions started with the reconstruction of the mainland coastline. The development of it is rather well known (Kooper, 1939; Rienks & Walther, 1955; RWS, 1948, 1959; Stiboka, 1973, 1976, 1981; Griede, 1978; Knol, 1991).

Subsequently, the barrier islands were reconstructed on the basis of most accurate maps available, mostly following the method of Ligtendag (1990). When such maps were not available the coastline was reconstructed by interpolation between the preceding and the subsequent period. For the islands (especially for Schiermonnikoog) the reconstruction of the dune development by Isbary (1936) was used to reconstruct the migration direction of the island. Furthermore, the reconstructions of the morphological development of the islands, in particular W of Ameland and E of Schiermonnikoog by Isbary (1936), Lang (1958), Ligtendag (1990) and Schoorl (in prep.) were taken into account. For the reconstructions, especially for the interpolations, care was taken to select the smallest changes needed to obtain a logical series of reconstructions.

Thereafter the configuration of the channels and intertidal areas was reconstructed using information from nautical charts and from archives. Care was taken that the channel reconstructions are in agreement with the local sedimentological data (Stiboka, 1973, 1976, 1981; Roeleveld, 1974; Van Staalduinen, 1977; Griede, 1978; Van Oosten, 1986; Bosch & Vos, 1992; Sha, 1992, cores of RWS/RGD). Because the historical maps of the Wadden Sea were mostly rather crude, the channel configuration, in particular of the smaller channels, is not very accurate; errors of >1 km can occur. In the descriptions, the position of the HW lines of the islands is given relative to their position in circa 1975. Always the distance between the most extreme points was measured. Furthermore the position of the axis of the inlets, measured between the islands, is given with reference to the position in circa 1975.

The configuration of tidal marshes and channels near the mainland was reconstructed from contemporaneous province maps and other detailed maps, or by assuming that parts which were enclosed shortly after the year of the map, were already supratidal.

GEOLOGICAL PRE-HISTORICAL DEVELOPMENT

Tectonics

The area is tectonically not very active. Possible earthquakes at 27 October 1225, 28 January 1262 (Wittewierum, Rozenhof, Bloemhof and Fiskmare; Emo & Menko, 1225 and 1262), and 3/4 February 1825 (Schiermonnikoog; Reitsma, 1988) and 2 earthquakes, around 55^0 N in the North Sea (between 1980-1986; Burton & Marrow, 1988) indicate that there is, however, some seismic activity in the north of The Netherlands and the adjacent North Sea. It was suggested, however, that the 'earthquakes' of 1225 and 1262 actually have been storms (cf. Houtgast, 1991). Also, during the 'earthquake' of 1825 a severe storm surge was active. It cannot be excluded that all these 'earthquakes' were the effect of storms.

It has been noticed that the orientation of lineaments in the field and part of the Holocene and Pleistocene sedimentation patterns coincide with the tectonic pattern in the deeper subsurface (Sesören, 1976). The major fault zone is the Hantum fault, which has been active since the Carboniferous (Brolsma, pers. comm.) until at least the beginning of the Pleistocene. Where the Hantum fault pattern crosses Ameland a deep former Holocene inlet fill is present in the subsurface. The faults W of the Lauwerszee Trough are parallel to the Early Holocene sandy valley fills. These valley fills were zones, where during the late Holocene some of the major tidal channel systems were formed. The area of major subsidence (Lauwerszee Trough) is located parallel to the Lauwerszee embayment. The centre of the area coincides with the position of the Early Holocene Hunze valley. It is therefore likely that the ancient tectonic pattern influences sedimentation patterns up to the present day. The influence is for the larger part, if not totally, indirect, by means of the relief of the Pleistocene surface.

Pleistocene

From a geologist's point of view, the Wadden Sea is relatively young. The Dutch Wadden Sea is underlain by an irregular surface of glacial till and fluvial sediments dating from the Riss (Saalian) glacial stage (circa 180,000-130,000 year B.P.). Most of these glacial sediments are or have been covered by younger deposits (Eemian interglacial, Würm/Weichsel glacial and Holocene post glacial). Locally, for instance on the island of Texel, and in the area between Den Helder (south of Texel) and the 'Afsluitdijk', Riss deposits crop out as small (ice-pushed) moraine hills that reach a height of 10 m to 20 m above present mean sea level. These Pleistocene deposits form the backbone of the Western Wadden Sea and have strongly influenced the morphological development of probably the whole Dutch Wadden Sea area. During the Eemian (Riss-Würm) interglacial (circa 75,000 years B.P.) sea level was relatively high. The low-lying areas (at present below MSL) were submerged and shallow marine and tidal flat sediments with a characteristic fauna were deposited.

During the last glacial stage (Würm-Weichsel, which ended about 10,000 years B.P.) the ice cap did not reach the Dutch region. During the glaciation, sea level was about 100 m lower than at present. Extensive aeolian deposits were laid down on the emerged land surface. At many places these deposits form the substratum of the Holocene sediments. The Pleistocene surface dips mainly in a seaward direction.

Holocene: pre-historical development; sea-level rise and sedimentation

During the early part of the Holocene transgression, mean relative sea level rose at a rate of approximately 1 $cm.yr^{-1}$, mostly because of melting of the icecaps and partly also by subsidence of the North Sea basin (Fig. 35 in Chapter 1; Jelgersma, 1979; Van de Plassche, 1982). After 6,000 B.P. sea-level rise slowed down. Early Holocene tidal deposits in the southern North Sea, and the widespread occurrence of tidal-flat mollusc shells indicate that most of the southern North Sea was covered by a tidal flat for some time between 10,000 and 5,000 B.P. (Eisma et al., 1981). During the Holocene large amounts of sediment, derived from the rivers and from the North Sea, were transported to the Wadden Sea area. It is assumed that a large part of the present Wadden Sea sediments was derived from an area west of Texel and, during the last millennia, from the coast of North Holland (Schoorl, 1973; Zagwijn et al., 1985). During the Early Holocene the sea transgressed.

An irregular topography, largely defined by the Pleistocene subsurface, dominated the development of the area, in particular in the early Holocene. Around 7,000 B.P. a brackish to marine lagoon was present in the west (Northern Holland). Some authors suggested that the basin was largely closed off from open sea by a long spit (cf. Zagwijn, 1985), whereas others found indications that the basin may have been partly open (Beets et al., 1994). The lagoon was separated from the precursor of the present Wadden Sea by an extensive Pleistocene high (Texel-Vlieland; Zagwijn et al., 1985). The various areas experienced a different evolution:

1) Western area (coast of South Holland and North Holland)

Due to the rising sea level the coastal areas became marshy and a peat layer was formed on top of the Pleistocene subsurface. The whole area from Bergen (North Holland) to Zeeland (Southern Netherlands) became influenced by tides. From land to sea the following zones occurred: a peat area (Basal Peat), a zone of fine-grained wadden deposits, brackish lagoons and salt marshes, and a sandy coastal zone consisting of beach plains and dunes. During the landward retreat of the coast, the sediments, formed in these zones, were deposited on top of each other. The sea also invaded the area of the Zuider Zee from the west and entered the valleys of the Vecht and IJssel (Koopstra et al., 1993). Landward retreat continued until approximately 5,500 B.P. (in the south, where the tidal basin was small) to 4,500 B.P. (in the north).

After that time the supply of sediment, derived from erosion of headlands (Rhine-Meuse delta area, Texel High), old ebb-tidal deltas and sediment supply from the North Sea and the large rivers (Eijndhoven, 1992; Beets et al., 1992), more than compensated sea-level rise of 20-40 cm per century. This was possibly enhanced by changes in wave climate and/or tidal amplitude (Zagwijn, 1986). As a result of the sediment surplus, the tidal basins were filled up; first in the south, and later also in the north. The intertidal areas behind the coastal zone became shallower and tidal channel systems were gradually abandoned (largely between 5,500-3,300 B.P.; Beets et al., 1992; Van der Spek & Beets, 1992). Because of the decreasing need for sediments with the increasing infill of the tidal basins, a sediment surplus became available and the coast started to prograde; in the south of Holland between 6,000 and 5,000 B.P. (northern side of Rhine/Meuse area) and in the N around 4,200 B.P. (Alkmaar; Beets et al., 1994). The inlets in the coastline closed and large surfaces behind it became logged with fresh water, and a second period of large-scale peat growth (Holland

Peat) occurred (mainly in the Middle Subboreal to Lower Subatlantic; Roeleveld, 1974; Zagwijn, 1985). Erosion continued in the area N of Bergen-Texel High (Beets et al., 1994).

Coastal progradation continued, but sediment supply gradually decreased. As a result the coastal profile became steeper from circa 3,600 B.P. to at least 2,000 B.P., and, finally, gave way to coastal erosion (Beets et al., 1994). Extrapolation of profiles suggests that this happened somewhere around 9th century A.D. (Wassenaar) to 11th century A.D. (Haarlem; Pool, 1992).

2) Western Dutch Wadden Sea

In the western Dutch Wadden Sea (S of Texel-Terschelling) the Pleistocene surface was relatively high, thus strongly reducing the influence of the sea (Fig. 36; Chapter 1). In the period of especially 7,000-5,300 B.P. the western, seaward part of the area was gradually eroded. Afterwards the remaining high area SE of Texel was gradually flooded, except for a small part on Texel. Until 3,700 B.P. the area developed much in the same way as the western area, although a tidal inlet remained open at the Boorne side, in the east. Texel High was an exception; it was not flooded and no formation of peat took place on it. After 3,700 B.P. part of the area east of the Vlie channel (Edelman, 1964; Eisma & Wolff, 1980) or east of the line central Texel-Bolsward (Zagwijn et al., 1985) gradually changed into a wadden area. West of the line the surface was mainly covered with peat until the early Middle Ages. In the east, the small Boorne river debouched. In the west there was a river at the site of the present Texel Inlet channel (Schoorl, 1973). The sea entered the area relatively late (approximately 1,000 B.P.) the area became influenced by tides, and the peat layer was eroded (De Waard, 1948; Edelman, 1964; Schoorl, 1973; Eisma & Wolff, 1980). After the closure of the coast of Holland, the Zuider Zee drained along the Vlie-channel (Koopstra et al., 1993). The connection of the Zuider Zee with the North Sea was enlarged around 2,000 B.P., and the Zuider Zee changed gradually in a brackish lagoon. The entrance of the Zuider Zee became fully marine around circa 1200 A.D. (Koopstra et al., 1993; Beets et al., 1994).

3) Eastern Dutch Wadden Sea

In the eastern Dutch Wadden Sea the Pleistocene surface was lower than in the west and incised by a number of small river valleys (amongst others those of the Boorne, Hunze, and Ems rivers). As the sea entered these depressions the valleys turned into drowned estuaries during the Early to Late Atlantic (8,000 to 5,000 year B.P.), especially after 6,200 year B.P. (cf. Griede, 1978; Van der Spek, 1994). From sea to land the following zones succeeded each other: tidal flats, lagoons (with deposition of clays and reed-growth), and peat marshes. The peat zones retreated landward during the transgression (Griede & Roeleveld, 1982).

Tidal inlet channel fills cutting into Early Atlantic lagoonal clay indicate that barrier islands existed seaward of Ameland and Schiermonnikoog latest in the Late Atlantic (6,000-5,000 B.P.; Sha, 1992). Sediment needed to compensate the relative rise of the sea level, i.e., to maintain the barrier islands and the Wadden Sea, and to replace eroded Pleistocene and early Holocene fine-grained and peaty deposits, was largely derived from the island coasts, probably because external sediment supply was low. As a result, the islands shifted coastward (at least 9.5 km for eastern Terschelling (Oosterend; cf. Sha, 1990); 11 km for central Ameland and 15 km for central Schiermonnikoog; Sha, 1992). Radiocarbon datings of peats on the barrier islands indicate the time that these have been at approximately their present position. Terschelling reached its present position since at least 1,620 B.P. (Staalduinen, 1977), Ameland since at least 1,260 B.P. (De Jong, 1984), and Schiermonnikoog since at least 525 B.P. (the latter partly due to strong lateral migration; Sha, 1992). How to explain the increaseing distance, from W to E, over which the coastward retreat took place, and the decreasing age of the peat deposits in the same direction? This might be explained by a rotating alignment of the barrier islands, the hinge point being some 7 km NW of mid-Vlieland. Models indicate that during the last 10 m of sea-level rise (since 7,400 B.P.) tidal amplitude decreased slightly at the western side of the Dutch Wadden Sea and increased at the eastern side (Franken, 1987). This may have led to a gradual decrease of tidal forces in the W and a gradual increase to the E, and to an increasing sediment demand towards the E, explaining the observed differences. That the sediment demand of the Wadden Sea is of major importance is also suggested by the observation that Vlieland, Texel, and Huisduinen have been relatively stable until circa 800-1,000 A.D., after which the coast became open and strong erosion occurred at the North Sea side (cf. Schoorl, 1973; Ligtendag, 1990).

During 4,200 to 4,000 B.P. (Holland IVa regression) the coastal peats and tidal marshes expanded seaward (Griede & Roeleveld, 1982). Between 3,900 to circa 3,700 B.P. (Calais IVb transgression) the sea invaded the area again, it retreated around 3,650 B.P., to invade once more between 3,500 and 3,200 B.P. (Duinkerke 0 transgression; Griede & Roeleveld, 1982). Griede & Roeleveld (1982) observed that the in NE Friesland the peats expanded seaward during the Duinkerke 0 transgression, whereas other areas were flooded. After another retreat of the sea around 3,200 B.P., the peat zones and marsh areas started to retreat after circa 3,000-2,900 B.P. (Griede & Roeleveld, 1982). Most of the time the Boorne, Hunze, and Ems estuaries remained partly open, surrounded by intertidal flats and/or tidal marshes. In the Boorne valley the sea came almost 30 km inland in the Late Subboreal. After an initial enlargement, sediments filled up the basin (Van der Spek, 1994). The river Boorne shifted its course to the low-lying area to the E.

In the area of the present eastern Dutch Wadden Sea, lower rates of sea-level rise often did not lead to peat formation. Instead, clay was deposited, suggesting a larger marine influence than in the areas where peat was formed. This is probably due to the presence of large tidal inlets at relatively short distances, and the rather small width of the tidal area, as compared to the tidal area in the west (coast of Holland). There, the inner flats were more sheltered behind long coastal barriers, and hus peat formation could be far more extensive. An important additional effect must have been the much greater fluvial run-off into the western part of The Netherlands as compared to the rather limited fluvial activity in especially the eastern part of the Wadden Sea.

REVIEW OF HISTORICAL DATA

5th century B.C. - 8th century A.D.

5th century B.C. - 0

Around 500 B.C. the mainland area must have been dissected by many estuaries, which were bordered by tidal flats. The supratidal zone of the mainland was more land inward than nowadays (Bosch & Vos, 1992). Towards the North Sea barrier islands were present.

In ancient literature, the barrier coast of the Wadden Sea has been mentioned several times, because it was on the route to the Nordic amber. The Nordic amber was highly valuated by mediterranean people, as is, for instance, indicated by many Nordic amber artifacts found in Minoan Knossos, Greece (Henning, 1944; Waterbolk & Waterbolk, 1992). Early Greek records from the 7th century B.C. may refer to the river Elbe, when they mention the '*Amber river*' (Henning, 1944; Russchen, 1974).

In 350-310 B.C. Pytheas of Massilia (Henning, 1944) may have mentioned the North Sea (Metuonis) and the bordering Wadden Sea. From the German mainland, he stated, it was a one-day sail to the amber-rich island Abalus[1] . The amber was said to be traded or used as fuel (probably confused with peat; *Pliny, XXXVII*). Timaeus (356-260 B.C.) confirms the existence of the amber island, but calls it Basilia (*Pliny, XXXVII, 11*). Late Iron Age pottery found on the island Griend shows that man migrated into the Dutch back-barrier area during the 2nd-1st century B.C. (Taayke, 1988).

Diodorus Sicilus (of Sicily, 54 B.C.) mentioned islands with large intertidal flats and inlet systems (*Diodorus Sicilus, XXII, 1-4*), situated between Britain and Europe, which may have been the Wadden islands. Also he mentioned the inhabited amber island Basileia *Diodorus Sicilus, XXIII, 1&4*). The stranding of the ships of Claudius Drusus (12 B.C.) during ebb on intertidal flats west of the Ems was mentioned by *Dio Cassius (LIV, 32*).

0 - 2th century A.D.

In the period 500 B.C.-0 A.D. the supratidal zone of the mainland expanded slightly seaward. Large estuaries surrounded by tidal flats dissected the area (Bosch & Vos, 1992). Tacitus' writings (*Ann. I, 70*) indicate that the Romans sailed the Wadden Sea in 15 A.D.. The same description says that the supratidal mainland between the Ems Estuary and the Hunze Estuary was dissected by tidal creeks and flats, and was easily flooded during storms. In July 16 A.D. the Roman fleet made a journey through the Wadden Sea from the Ems towards the winter quarters in the centre of The Netherlands (*Tacitus, Ann. II, 23&24*). A storm from the south in combination with the ebb current threw part of the (shallow) ships on the barrier islands and (intertidal) shoals, while others drifted into the North Sea (even to England).

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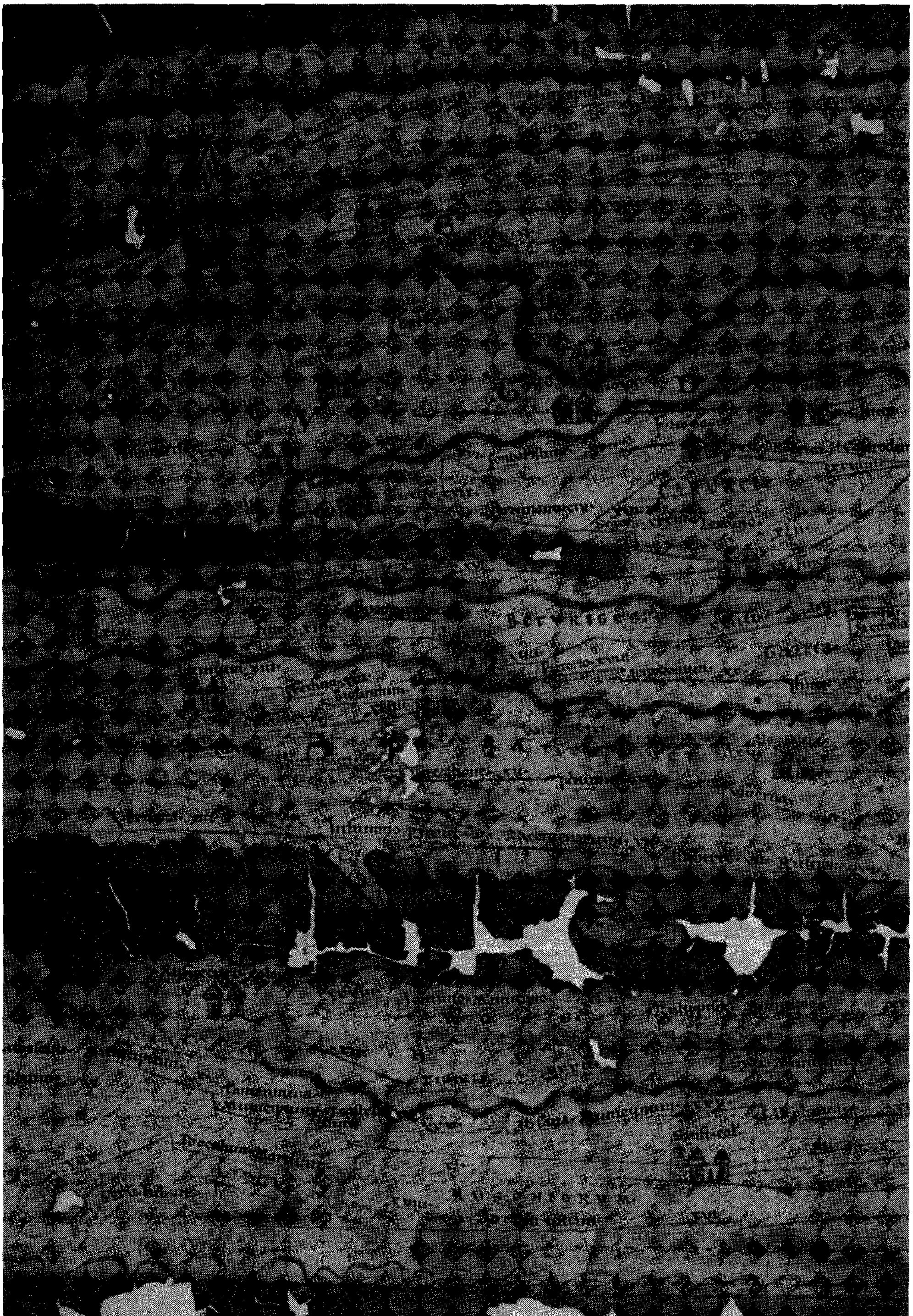
[1] This may also have been situated in the Balthic Sea.



The descriptions of the journeys of the Romans between the German coast and their winter quarters in the centre of The Netherlands furthermore state that the Zuider Zee embayment still consisted of several (probably two) fresh water lakes (cf. Koopstra et al., 1993). These gradually became larger by erosion (*Tacitus, Ann. I, 60, II, 8; Pliny, Hist. Nat., XVI, 2*; Van Es, 1981) and had a connection with the North Sea (cf. Koopstra et al., 1993). These lakes were fed by a northern branch (IJssel; Van Es, 1981) of the Rhine-system (*Pliny, Hist. Nat., IV, 15*) and by the northern Vecht river (Van Es, 1981).

Around 47 A.D. 23 (barrier) islands were situated along the North Sea from the river Rhine to approximately Esbjerg, Denmark (*Pliny, Hist. Nat., IV, 14*; Henning, 1944). The bean island Burcana (Borkum?; cf. Hettema, 1951; Lang, 1958) and the amber island Austeravia/Actania/Glaesaria (Ameland/Oostergo?) were mentioned by name (*Pliny, Hist. Nat., IV, 14; XXXVII, 11*). Also the island Flevum (Vlieland or Velzen?; *Pliny, Hist. Nat., IV, 15*; Beelaerts van Blokland, 1943; Van Es, 1981) was mentioned. Pliny described the prevailing northern winds and the semidiurnal tide of the area. Furthermore he told about the presence of extensive intertidal flats, artificial dwelling hills situated above the highest flood level (terps), and the exploitation of peat for firing (*Pliny, Hist. Nat., XVI, 1*) in the area E of the

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Figure 2 (opposite page): Tabula Peutingeriana (12th to early 13th century): Copy of a Late Roman map (4th century) which was based on the world map made during the rule of Emperor Augustus (27 B.C.-14 A.D.; Stuart, 1991). The map gives an overview of the travelling distances of the Roman Empire and the surrounding world. For the Dutch area only some details are given for the area S of the Rhine. Note the large estuaries drawn at the end of the Rhine and Meuse. N is to the top. The part depicted shows a part of the Dutch area. To the west is the North Sea and the English area.

Explanation of the Dutch names from left to right and from top to bottom: *(C)Haci* = Chaucken; *Crepstini* = unknown tribe; *Chamavi qui el Pranci* = Chamavi, also named Francs; *Fl. Rhenus* = river Rhine; *Lugduno* = probably Brittenburg; *Pretoriu Agrippine* = Valkenburg (S.H.), *Matilone* = Roomburg; *Alba(ni)anis* = Alphen aan den Rijn; *Nigropullo* = Zwammerdam; *Lauri* = Woerden; *Fletione* = Vechten; *PATAVI(A)* = Batavia; *Foro Adriani* = Arentsburg (Voorburg); *Flenio* = Helimium (Mouth of the Meuse?); *Tablis* = ?; *Caspingio* = ?; *Grinnibus* = probably Rossum (Gld.); *Fl. Patabus* = the Meuse (Below it the *B* of Belgium). Distances in Leagues (which is approximately 2200 m; Stuart, 1991).

Ems Estuary (cf. Van Es, 1981)[2]. Similar conditions existed in the NE of The Netherlands (cf. Van Es, 1981; Kooi, 1988). In 98 A.D. Tacitus mentioned that the people of Friesland lived around vast lagoons (*Tacitus, Germania, 34*). Unfortunately no maps are known which depict the northern Netherlands in some detail[I,3]. After a few attempts to expand their Empire beyond the Rhine, the river was finally considered the northern border of the Roman Empire (Fig. 2).

To summarize: the picture, emerging from these written sources, in combination with geological reconstructions (Zagwijn, 1985), is as follows:

The NW Wadden Sea consisted of an irregular micro-topography of Pleistocene deposits, cropping out at many places. These high areas were partly surrounded by eutrophic and oligotrophic peats (Zagwijn, 1985). To the SE lakes (Lake Flevo area) were present. These

[2] Pliny, Hist. Nat., XVI, I (Ed. Loeb), most likely describing the area directly E of the Ems (Van Es, 1981): "*....quaenam qualisque esset vita sine arbore ulla, sine frutice viventium. Diximus esse in oriente quidem iuxta oceanum complures ea in necessitate gentes; sunt vero et in septentrione visae nobis Chaucorum qui maiores minoresque appellantur. vasto ibi meatu bis dierum noctiumque singularum intervallis effusus in inmensum agitur oceanus, operiens aeternam rerum naturae controversiam dubiamque terrae an partem maris. illic, misera gens, tumulos optinent altos aut tribunalia extructa manibus ad experimenta altissimi aestus, casis ita inpositis, naviganti-bus similes cum integant aquae circumdata, naufragis vero cum recesserint, fugientesque cum mari pisces circa tuguria venantur. non pecudem his habere, non lacte ali ut finitimis, ne cum feris quidem dimicare contingit omni procul abacto frutice. ulva et palustri iunco funis nectunt ad praetexenda piscibus retia, captumque manibus lutum ventis magis quam sole siccantes terra cibos et rigentia septentrione viscera sua urunt. potus non nisi ex imbre servato scrobibus in vestibulo domus. et hae gentes, si vincantur hodie a populo Romano, servire se dicunt! ita est profecto: multis fortuna parcit in poenam.*"
(....what is the nature and what are the characteristics of the life of people living without any trees or any shrubs. We have indeed stated that in the east, on the shores of the ocean, a number of races are in this necessitous condition; but so also are the races of people called the Greater and the lesser Chauci, whom we have seen in the north. There twice in each period of a day and a night the ocean with its vast tide sweeps in a flood over a measureless expanse, covering up Nature's age-long controversy and the region disputed as belonging whether to the land or to the sea. There this miserable race occupies elevated patches of ground or platforms built up by hand above the level of the highest tide experienced, living in huts erected on the sites so chosen, and resembling sailors in ships when the water covers the surrounding land, but shipwrecked people when the tide has retired, and round their huts they catch the fish escaping with the receding tide. It does not fall to them to keep herds and live on milk like the neighbouring tribes, nor even to have to fight with wild animals, as all woodland growth is banished far away. They twine ropes of sedge and rushes from the marshes for the purpose of setting nets to catch the fish, and they scoop up mud in their hands and dry it by the wind more than by sunshine, and with earth as fuel warm their food and so their own bodies, frozen by the north wind. Their only drink is supplied by storing rain-water in tanks in the forecourts of their homes. And these are the races that if they are nowadays vanquished by the Roman nation say that they are reduced to slavery! That is indeed the case: Fortune oft spares men as a punishment.)

[3] For the Roman numerals, see appendix 1: overview of maps.

were connected to the Wadden Sea (via Vlie). If the island Flevum, situated in a river delta was indeed the barrier island Vlieland this might explain the aberrant sea-ward position of the island in later periods (cf. Ligtendag, 1990).

Along the rest of the Dutch Wadden Sea (from present-day Vlieland/Terschelling towards the Ems) barrier islands protected the extensive sub- and intertidal flats and estuaries (especially Boorne, Hunze, and Ems) behind them. The supratidal zone on the mainland was a low lying area, open to storm surges and dissected by creeks. Barrier islands were present towards the North Sea. If the islands mentioned by Pliny include all the barrier islands along the coast from the Rhine to Esbjerg, the low number suggests that the semi-diurnal tides might have been of a somewhat smaller amplitude than at present (cf. Hayes, 1975; cf. Wolff, 1986). In that case, the, presently micro- to meso-tidal, Dutch Wadden Sea, had a more micro-tidal character at that time. This is in accordance with the increase in tidal amplitude during the last 2000 years, indicated by modelling studies (Franken, 1987; De Ronde, 1983).

5 - 8th century A.D.

Around 500 A.D. the supratidal area of the mainland coast had prograded and was only some 2-4 km S of the present-day shoreline of the mainland. Many estuaries had been largely filled up. In front of the mainland an extensive intertidal area was present. In the area of the Lauwerszee the mainland coast was circa 4 km from the present-day shoreline; only two large rivers had an estuarine connection with the sea (Bosch & Vos, 1992).

The '*Frisian islands*' are mentioned in the Finnsburgh fragment of the Beowulf saga (origin in the 5th-6th century, composed in the 8th century (Alexander, 1972), which probably relates to the Dutch part of Friesland (Fenger, 1935). The '*Frisian Islands*' may refer to parts of the Frisian mainland, which were separated from each other by swamps and rivers, as well as the barrier islands. In the latter case this would be the only source, known to the author, mentioning these islands in the period.

Already before 720 A.D. the Wadden Sea was an important trading route as is mentioned in the vita describing the live of Wulfram (Lang, 1976). In the Post-Roman to Early Medieval period the tides gradually started to invade the Boorne area again (Van der Spek, 1994). However, between 734 and 738 A.D. Karel Martel had a camp along the '*river*' Boorne (Kusternig, 1982), suggesting that the area was not strongly marine.

In 779 A.D. the '*river Lauwers*' was crossed by the missionary Willehad on his way from the town of "*Dockum*" to "*Humarcha*" (Wattenbach et al., 1888, 22, p. 96). This indicates that the southern Lauwerszee embayment was small and had not yet reached its full dimensions (Andreae, 1881). The mention of the '*river Lauwers*' slightly after 782 A.D. (together with the backbarrier island Bant[4]; Wattenbach et al., 1888, 22, p. 75) and in the

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[4] Either NE of the present province of Friesland, or S of Borkum.

Lex Frisonium (compiled circa 802; Blok et al., 1896, no 2; Eckhardt & Eckhardt, 1982), supports the view.

However, marine influences were already noticed in the Lauwerszee area. After the murder of St. Bonifatius in 754 A.D, a church was built in Dockum. Between 754-768 A.D. it was written that the church was placed on an artificial mound to '*keep out the invading ebb and flood*' (possibly neap and spring tide; Gottschalk, 1971), which '*caused in continuous alternation the rise of the sea level and the retreat of the ocean, the decrease and increase of the waters*'[5] (Tangl, 1920, 1st vita, 9). In the period 800-850 A.D. it was mentioned that in that time the ground water of the area around Dockum was brackish (Tangl, 1920, 2nd vita, 16). This suggests that already in the second half of the 8th century marine influences reached the town of Dockum, and that the sea was gradually invading the area. This is supported by ^{14}C data of in-situ (Griede, pers. comm.) marine bivalves. These indicate that the southern Lauwerszee area was inundated between, or after, 727 and 959 A.D. (Griede, 1978; Knol, 1992), mainly in the second half of the 8th and in the 9th century (Roeleveld, 1974; Griede, 1978; Bosch & Vos, 1992; Knol, 1992). In first instance the inundation probably happened along the rivers, in an estuarine way. The depressions in the landscape were easily flooded as can be concluded from the extensive marine deposits along the Reitdiep, the former Hunze and other rivers (cf. Stiboka, 1973).

9th - 13th century A.D.

9th century A.D.

In the beginning of the 9th century the Frisian islands were mentioned several times. At least some of the written sources refer to the barrier islands bordering the Wadden Sea. In 810 A.D. large parts of the Frisian islands and the mainland coast were plundered by Vikings (Rau, 1955, p. 95). Einhard probably referred to the same raid when he said that during the reign of Charlemagne (who died 814 A.D.) several East(?) Frisian barrier islands were plundered by Vikings (Abel, 1893, Kap. 17, p. 26; Rau, 1955, p. 189). After the death of Charlemagne large parts of the Dutch and German coast were ruled by baptized Vikings, although this did not entirely stop the plundering of the Frisian coast in the 9th century by other Vikings. In the Codex Eberhardi[6] (8th-12th century) the barrier island Ameland is mentioned several times (Friedlaender, 1881, p. 786, 788?, 792). Radio-carbon and pollen data indicate habitation of the island in or before 800 A.D. (De Jong, 1984). The names of the villages on the island indicate that they were founded in the 10th or 11th century (Van

[5] "*....propter immensas ledonis ac malinae inruptiones, quae diverso inter se ordine maris aestum oceanumque cursum, sed et aquarum infusionesque commovent*" (in: Gottschalk, 1971).

[6] This is a list of gifts provided during the period 750-<1150 A.D. (Blok et al., 1896, no. 1, note 1). According to Gijsseling (1960) the first mention ("...*in insula que dicitur Ambla...*") dates from the second half of the 8th century.

Oosten, 1986). Remnants of several old churches found on Ameland (Edes & Glashouwer, 1990) indicate a continuous habitation since at least the 11th century.

At the end of 9th century the lower reach of the Boorne was flooded and changed into the 30 km long Middelzee embayment in the 10th century (Van der Spek, 1994). The inundation of the Lauwerszee continued. An important storm surge occurred in almost the whole Frisian territory in 838 (Rau, 1958, p. 41[7], p. 343,; Gottschalk, 1971). Another inundation possibly occurred in 868 (Rau, 1958, p. 361; 1960, p. 73[8]; Gottschalk, 1971). Artifacts of Carolingian Age (725-900 A.D.) show that by that time the permanently inhabited coastline was much farther landinward than nowadays (Knol, 1991). The Lauwerszee is thought to have reached its maximal (southern) dimensions around 1000 A.D. (Bosch & Vos, 1992). Part of the inundation of the area was probably caused by erosion of peats, comparable to the situation in the Jade Bay. Curiously enough Andreas Cornelius[9] (1597) mentioned that in 806: '*a part of the place Ezonstad drifted away with 35 houses on it*' (Gottschalk, 1971). This indeed suggests the erosion of peat. Southeast of the Lauwerszee, subsidence (compaction and oxidation), and almost complete erosion of the original peat layer is thought to have been partly due to agriculture. The area was gradually flooded between 850-1000 A.D. (Stiboka, 1973). Peat excavation for salt production and for heating enhanced the flooding of the area as well (Griede, 1978; Bosch & Vos, 1992). Peat excavation may have become quite important in Carolingian times, when the population grew considerably (Knol in: Bosch & Vos, 1992).

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[7] Ann. Bertiniani, 26th dec. 838: "*Praeterea die septimo Kalendas Ianuarii, die videlicet passionis beati Stephani protomartyris, tanta inundatio contra morem maritimorum aestuum per totam paene Frisiam occupavit, ut aggeribus arenarum iliic copiosis, quos dunos vocitant, fere coaequartur, et omnia quaecumque involverat, tam homines quam animalia caetera et domos, absumpserit.*" (On the day of suffering of the blessed martyr Stephan, the 26th december, the sea inundated, in contrast to a normal flood, almost all of Frisia, so that the water almost reached the height of the mighty sand hills, called dunes, and where the flood came it dragged along humans, as well as animals and houses.)

[8] The flood is either a marine flooding, perhaps associated with a comet (Ann. Xantenses; also observed by other sources; Buisman, pers. comm.), or a river flood (Ann. Fuldensis; Rau, 1960). Ann. Xantenses, 869: "*....et XV. Kal. Martii, id est nocte sancta septuagesimae, stella cometes visa est ab aquilone et occidente, cui statim nimia tempestas ventorum et inmensa inundatio aquarum est subsecuta, in qua multi inprovidi interierunt.*" (....and on the 15th of February, this is in the holy night after Septuagesimä, a comet appeared in the NW (of Nijmwegen), and was immediately followed by a terrible storm and an immeasurable flooding, in which many imprudents died.)

[9] Gottschalk (1971) citing other authors states that Andreas Cornelius, who claims to cite older sources (Okko van Scharl & Johan van Vlietarp) is an unreliable source.

The following sites and indications of peat excavation have been reported:

1) in the Anjumer Kolken, W of the Lauwerszee; done in the tidal marshes (Bosch & Vos, 1992). These may have started around 0 A.D. and were ended when the area was flooded (833-891 A.D.; Griede, 1978),

2) near Obbemasate. The excavations were ended when the area was flooded (727-814 A.D.; Griede, 1978),

3) 250 to 625 m east of the coast of the Engwierderpolder and north of Dockumerdiep. Pottery from the site has been dated partly 10th century, mainly 11-13th century (Elzinga, 1969; Griede 1978; Knol, pers. comm.). These excavations were probably made during low tide to win salt from the peat (Elzinga, 1969),

4) near Dokkum (Friesch Dagblad, 5/12/1991),

5) in the Kollumer Waard, southern Lauwerszee, in the tidal marshes, and later covered by Lauwerszee deposits (Bosch & Vos, 1992),

6) near Burum, southern Lauwerszee salt peat excavations dating from 1300-1400 (Friesch Dagblad, 5/12/1991),

7) perhaps also the settlement traces near Vierhuizen can be related to peat excavation, because many pits and peat were found (Andreae, 1881),

8) in 1575 and several years later peat has been excavated to win salt in the Ruigewaard (SE Lauwerszee; Elzinga, 1969),

9) furthermore the name 'Zoutkamp' refers to salt, and is an indication for the winning or trading of salt, and

10) remnants of peat indicating large-scale peat excavation were also found below Thesinge, NE of Groningen and near Garmerwolde (Edelman, 1952; Stiboka, 1973).

Griede (1978) calculated that in the first three areas $9.5*10^6$ m^3 of peat has been excavated, lowering the area locally (over 19 km^2) with half a metre. At least the same amount of overlying clays had to be removed. The clay was dumped in the old peat pits. Apart from the removal of the peat, the removal of the overlying protecting clay layer will have considerably enhanced erosion and marine incursions (Bosch & Vos, 1992).

10-13th century A.D.

In approximately 1075 A.D. the barrier islands of Friesland were mentioned by Adam von Bremen (Steinberg & Schmeidler, 1926, Scriptores, B. 4,3, p. 209). As discussed for Ameland, the eastern Dutch barrier islands were inhabited and of economical interest. This is illustrated by a conflict about the island Cornasant (Blok et al., 1896, no. 379 (1344) and a war about an island, probably Rottumeroog, (Emo, 1231-1250; Blok et al., 1896, no. 429 (1354); Formsma, 1966). Saxo Grammaticus described, between 1145-1185, the meandering tidal channels and the intertidal flats of Friesland (Elton, 1894, ch. 46), for the area between Vlie and Weser, or more probably for the area between the Vlie and the Lauwers (Gosses, 1923).

According to Mansholt (1929) several villages S of present-day Ameland were flooded and lost during the 12th and 13th century (amongst others Oosthuizen S or E of Oerd). However, Van Oosten (1986) states that this happened as late as 1570 or even later, which is more likely (see 1550-1600).

Inundation and strong erosion of the islands and the mainland by the storm surges of 17/18-2-1164 (N storm, Friesland? & Groningen; Schmeidler & Lappenberg, 1910, 97, p. 227; Stoob, 1963, p. 338[10]; Emo, 1219), 2-11-1170 (Staveren in Friesland; Platner, 1867, p. 82; De Groot, 1992[11]), 6-12-1196 (Fig. 3; Middelzee and Dollard; Menko, 1237; Coronike van Vrieslant, 1450, in: Gottschalk, 1971), 16-1-1219? (almost springtide, NW storm, Friesland and Groningen; Emo, 1219), 28-12-1248 (Fig. 4; NW storm, Friesland and Groningen), 28-1-1262 and 14-12-1287 (Fig. 5; Northern Holland, Friesland and Groningen) is suggested by several, sometimes disputed, historical sources (Gottschalk, 1971).

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[10] Helmold, Chronica Slavorum, 1164: "*....in mense Februario, hoc est XIIII Martii, orta est tempestas maxima ventorum, procellae, fulgorum choruscatio et tonitrui fragor, quae passim multas edes aut incendit aut subruit, insuper tanta maris exundacio oborta est, quanta non est audita a / diebus antiquis, quae involvit omnem terram maritimam Fresiae, Hathelen et omnem terram palustrem Albiae et Wirrae et omnium fluminum, qui descendunt in occeanum mare, et submersa sunt multa milia hominum et iumentorum, quorum non est numerus.*" (....in the month February, that is the 17th, occurred a large thunderstorm with strong storm wind, flashes of lightning and roaring thunderclaps, which far and wide set fire to many houses or destroyed them, besides developed such flooding by the sea, which had not been heard of since ancient times. It inundated the whole coast of Friesland, Hadeln and the whole marsh area along the Elbe, Weser and all those rivers, which debouch into the Ocean, and many thousands of people and innumerable amounts of cattle drowned.)

[11] A large area of land near Staveren was lost in the storm of 1170 (Chronica Regia Colonienses, in: Gottschalk, 1971), thus enlarging the Zuider Zee area. During this storm, and perhaps another in 1173? (Buisman, in prep.), the city of Utrecht and its surroundings were flooded probably through the Zuider Zee and Vecht (Gottschalk, 1971). Gottschalk (1971) doubts whether sea water flooded the area. However, a contemporaneous source (Annales Egmundenses, in: Gottschalk, 1971) mentions the catching of a sea fish and the observation of ebb and flood near Utrecht. The flooding of Utrecht must have been severe. Afterwards it the city was hastily heightened on many places with all kinds of material available (up to 3 m.; De Groot, 1992).

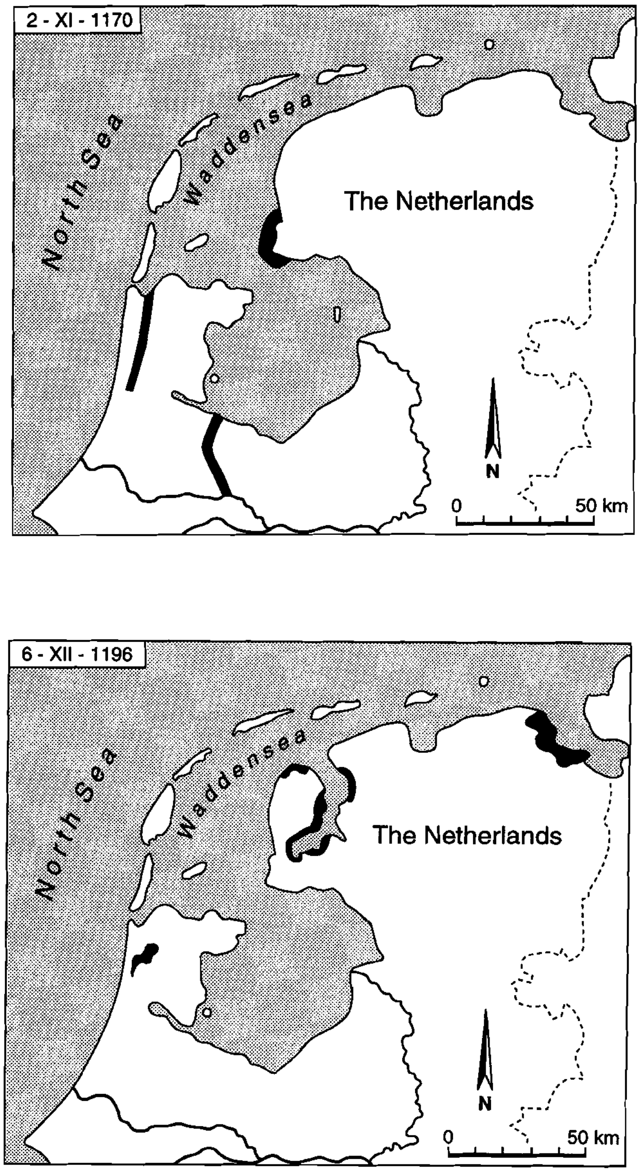

Figure 3: Areas along the Wadden Sea probably hit by the storm surges of 1170 and 1196, given in black (mainly after Gottschalk, 1971).

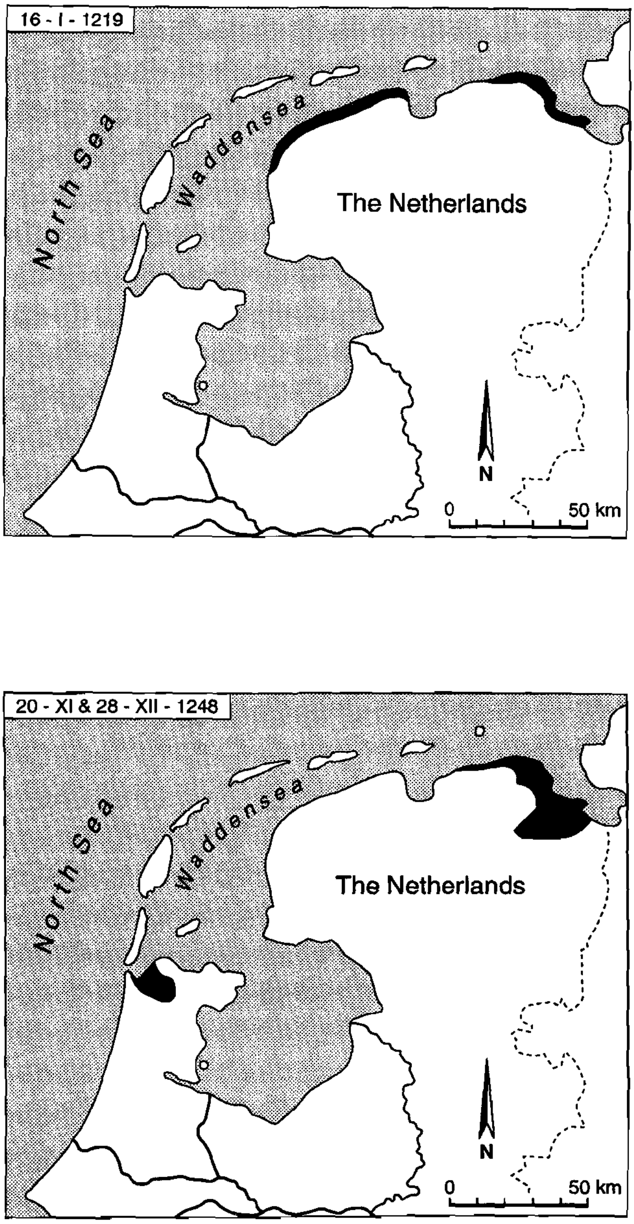

Figure 4: Areas along the Wadden Sea known to be hit by the storm surges of 1219 and 1248, given in black (redrawn after Gottschalk, 1971).

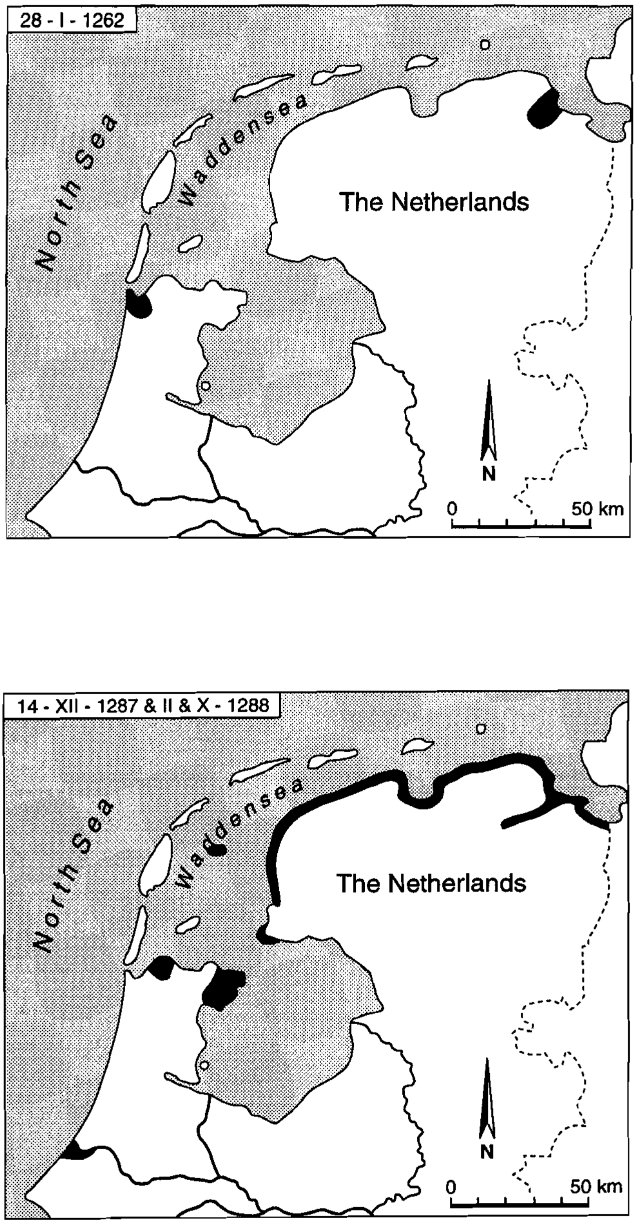

Figure 5: Areas along the Wadden Sea known to be hit by the storm surges of 1262, 1287 and 1288, given in black (redrawn after Gottschalk, 1971).

Pedological research in North Holland and the Zuider Zee area indeed indicated that the peat fen in the area was destructed at a very fast rate after circa 1150 (Pons & Wiggers, 1960; Gottschalk, 1971). ^{14}C-dates of shells indicate that the entrance of the Zuider Zee became marine around circa 1200 A.D. (Koopstra et al., 1993). These observations are in good accordance with the above reported loss of land in the Zuider Zee area.

The exact position of the mainland coastline around the year 1000 is not known. Documents indicate several inhabited places (mainly terps). Comparison of various Frisian Codes shows that in the 8th and 9th century no laws existed with regards to dykings (Acker Stratingh, 1866; Van Giffen, 1964). Dyke building probably started in or after the 10th century and became important later (Acker Stratingh, 1866; Rienks & Walther, 1955; Edelman, 1974). The earliest dykes in the Lauwerszee area may have been built in the 11th century, but they have been largely destroyed by storms and man (Andreae, 1881). Laws concerning dykings in the Frisan Codes must have been added in the 12th and 13th century (Acker Stratingh 1866)[12], and show that, around that time, a system of dykes surrounding

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[12] Rüstinger Law (probably compiled in the first half of the 12th century; Acker Stratingh, 1866; Barentsen, 1961): "*Thet is ac londriucht: Thet wi Frisa hagon ene seburch to stiftande, and to sterande ene goldene hop, ther umbe al Frislond lith; the skil on wesa alleraierdik iuin har oron, ther thi salta se, betha thes dis an tes nachtes, to swilith. Ther skil thi utrosta an thi inrosta thos wigis plichtich wesa, tha strete thes wintres and thes sumures, mith weyke and mith weine to farande, thet thi wein tha oron metha mugi.*

Alsa thi inrosta to tha dike cumth, sa hage re alsa gratene fretho opa tha dike, alsa re oua tha wilasa werpe, and alsa re ona tha weida stherekhoui; heth there thenne buta dike alsa felo heles londes and grenes turues, thet ter ne dik stathul mithi halda mugi; ac neth there nauwet sa felo buta dike heles londes and grenes turues, thet ter ne dik mithi halda mugi; sa hage re binna dike thritich fota turues and thritich fetma to gerse; thet skil wesa alla fennon and fili, er Sante Vitus di.

Vta skilu wi Frisa vse lond halda mith thrium tauwon, mith tha spada and mith bera and mith there forke; ac skilu wi use lond wera mith egge and mith orde, and mith tha bruna skelde, with thene stapa helm and with thene rada skeld, and with thet unriuchte herskipi. Aldus skilu wi Frisa halda use lond fon oua to uta, ief us god helpa wili and sante Pederr."

(This is also law of the land: That we, Frisians, have to found a stronghold against the sea, a golden hoop, which lies around all of Friesland, of which every dijksroede (measure used for the dykes) shall be equal to the other, (and) where the salt sea, both by day and by night, swells against. Therefore shall the outermost and innermost (those who live farthest from and nearest to the dyke) be obliged to go over the streets by winter and summer with a sled and chart, (so) that one may avoid/meet the other. (The article gives the impression that a continuous system of dykes was present protecting Friesland against the high waters).

If the innermost, comes to the dyke, he will have the same peace on the dyke, as he would have on the unhallowed yard, and as he would have on the consecrated churchyard. He has (can use) then outside the dyke as much land and green turfs (so) that he may hold the dyke with it. If he has not so much land and green turfs that he can hold the dyke, so he has within the dyke thirty feet for turves and thirty feet for greenland. That shall be ready before Saint Vitus day (15 June; the possible presence of supratidal marshes outside the dyke clearly indicates that the dykes were built at or above the MHW line.)

Also will we Frisians, hold our land with three tools: with the spade and the barrow and with the fork; also we shall defend our land with sword and with spear and with the brown shield, Friesland became established (cf. Van de Ven, 1993). Originally dykes were probably only meant to protect land against erosion, whereas in the 12th or 13th century active polderisation started (Edelman, 1974). From that time on, the supratidal areas of the Lauwerszee and along the mainland coast were dyked, and thus the drainage area and the tidal prism of the related inlets gradually decreased. The Middelzee was largely reclaimed between the 10th and 13th century (Van der Spek, 1994). According to Andreae (1881) the Lauwerszee may have expanded its E-W dimensions until the 13th century due to storm surges. Indeed, storm surges occurred in the area in 1219, 1220, 1221, 1246, 1248, 1249, 1257?, 1262, 1268, 1285, 1287, 1288, and 1290? (Figs 4 and 5; Gottschalk, 1971), but the above discussed influence of human-induced subsidence of the area (peat-excavation, drainage) must have also have been substantial. After the 13th century the Lauwerszee was gradually dyked and reclaimed (Eibergen, 1941; Bosch & Vos, 1992). During the enlargement of the Lauwerszee several places were lost, as is suggested by:

1) the drowning of the village Wartena/Wardenae/Werdina/Varden, S of Ezumazijl in 1230(?) (Andreas Cornelius[9] (1597) in: Halbertsma, 1969; Gottschalk, 1971). The place is mentioned in 839-849 A.D. as the place where the parents of Saint Liudger lived, and was also visited by him (Wattenbach et al., 1888, 21, p. 75, 80). The Codex Eberhardi possibly mentions it for the period 750-1150 A.D. as: "*in pago Ostrache, in villa que dictur Werba*" (Friedlaender, 1881, p. 790). Hamconius (1609) mentions it as the birthplace of Anskar, the first bishop of Bremen (Halbertsma, 1969).

2) the drowning of the hamlet/farm/castle Ezonstad/Esonstede (Anonymous (probably Andreas Cornelius[9] (1585), going back on Martinus Ylstanus, circa 1507, in: Halbertsma, 1969), in 1230 (year probably incorrect (Gottschalk, 1971, p. 161). The village was said to[13] have been '*totally destroyed and floated away, as well as drowned entirely*'. It was perhaps situated E of Ezumazijl, where ancient bricks were found 300 m E off the coast

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against the high helmet and against the red shield (attributed to the Vikings) and against unjust ruling. So shall we, Frisians, hold our land from high to outside, when God will help us and Saint Peter.)

[13] Andreas Cornelius, probably (1585), claiming he was citing Martinus Ylstanus, c. 1507: "*Ick hebbe tot Dockum int clooster van de Premonstratenser op haer libarie aengeteeckent gevonden zeer oude monumenten, waaroonder nog verhaalt stondt dat, hoe als men duizent twee hondert ende dertich screef, is Ezonstadt aen de Louwerts zee gelegen deur een ongehoorden zeer hoogen waetervloedt ende stormachtigen tempeest geheelyck vergongen ende wech gedreven, oeck alsoe geheelyck verdroncken, zoe datter geen thien huijsen staande bleven.*" (I have found noted in Dockum in the cloister of the Premonstratenser in their library, very old documents, in which it was said, that when it was written thousand two hundred and thirty, Ezonstadt near the Lauwers Sea was, by a very high storm surge and a storm, totally destroyed and floated away, as well as drowned entirely, so that less then ten houses remained).

(Andreae, 1881; Dijkstra, 1959; Van den Berg, 1983). The existence of Ezonstad is disputed; it may be a mystification, although names as Ezumazijl, Ezumburen and Ezumakeeg suggest differently (Elzinga, 1969; Gottschalk, 1971; Van den Berg, 1983); no traces of such settlement were discovered after the dyking of the Lauwerszee.

3) the terps Oosthalm and Midhalm which are close to the dyke at the eastern side of the Lauwerszee, suggesting a 'Westhalm' west of it (Andreae, 1881).

4) settlement traces recovered in 1720 west of Vierhuizen (Groningen) on the tidal flats (Andreae, 1881; Elzinga, 1969).

Reconstructions 14th-18th century A.D.

1300-1350; Fig. A1

The earliest, still known, maps giving details of the Frisian Wadden area are the Mediterranean sea charts (Portulanos)[14]. From at least 1321 onwards the names of several barrier islands and inlets are mentioned and the Zuider Zee area[15] and the sailable backbarrier area were depicted (Table I; Figs 6 and 7; cf. Lang, 1955, 1958; Koeman, 1985). The geographical sites were probably mentioned, because they all lie along an important waterway (partly inland, via the Zuider Zee) from Flanders to Ribe (Lang, 1955). The lack of adequate detail does not allow to use these charts directly for the reconstructions. Therefore reconstructions were mainly based on previous studies of the area (Roeleveld, 1974; Griede, 1978; Knol, 1992) based on geological and archaeological data in combination with historical sources. For the reconstruction of the Lauwerszee area of 1300, the maps of Bosch & Vos (1992) of 1000 A.D. and 1500 A.D. were used, assuming open waterways to Dockum and Groningen, and continued reclamation and dyking. Also several other sources have been used for the reconstruction (Isbary, 1936; Kooper, 1939; Rienks & Rijkswaterstaat, 1948, 1961; Walther, 1955; RWS, 1948, 1959; Stiboka, 1973, 1976, 1981; Griede, 1978; Knol, 1991; Sha, 1992; Van der Spek, 1994).

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[14] Several sailing directions for North European (especially nordic) waters were written down in the 10th-14th century (Lang, 1955). Sailing directions and sea charts were probably common for North European sailors as early as the 14th century (Lang, 1955). During that and the following century the first charts of the Wadden coast must have been made.

[15] In 1340 the Zuider Zee area is for the first time mentioned as the sea Sudersee (Koopstra et al., 1993); before that it was called Aelmere (Ael-lake).

Table I. Places mentioned on some Portulanos (Figs 6, 7, and 12)

Vesconte, 1321[II]	Atlas Catalan, 1375[III]	Mecia de Viladestes, 1413[IV]	Petrus Roselli, 1462[VI]	C = City B = Barrier I = Inlet*
Veret**				C Utrecht
Sca forda	Scalingue	Scalinga		B Terschelling/-Coevorden[16]
	ardrohic	ardoich		C Harderwijk
masdiepay	masdiepa	masdiepa	masdiepa	I Marsdiep
			caldiepa	I Keeldiep
			StobaiaSti	I Scholbalg
vangaroza	vuangroga	vuangaroga		B Wangeroog
	ollanda	olanda	olanda	Holland
lelie	le ulie	leulie		B Vlieland/I Elbe?
frislanda	ffrixa	fixa	frixa	Frisia
flisilanda				Frisia
	aqua vllie	aqaullie		I Vlie/Zuider Zee/-German Bight?
			vestermeza	I W. Ems
			ostermossa	I E. Ems
			Vessa	I Weser
	ripiss	fipis		C Ribe
			eleua	I Elbe
	Insula Sce			Helgioland
Dancsmarch	dacja		costa denermech	Denmark

*): Interpretations of modern names partly after Keuning (1914), Lang (1955, 1968) and Koeman (1985).

**): Underscored names, in red on the charts, mostly indicate lands, provinces or capitals.

[16] Edrisi (1139) mentioned a city Schkela or Skela and a city Sikela or Sikla, which is some 80 miles east (or north?) of Ghand (Lelewel, 1966). Sca forda may have been Coevorden. The barrier island Terschelling is known as die Sc(h)elinghe/a in charters of 1296, 1320, 1324, 1337 (Wumkes, 1900).

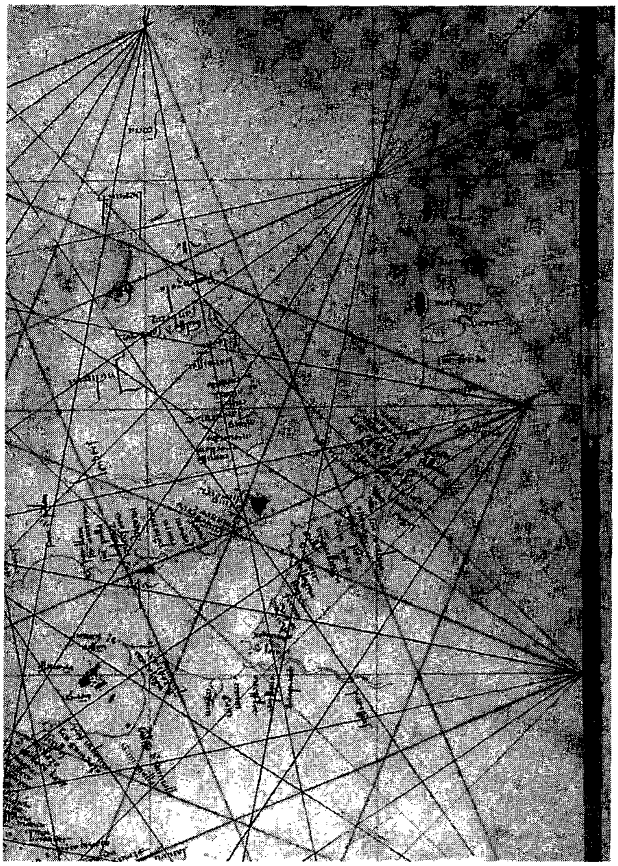

Figure 6: Petrus Vesconte (1321): Portulano atlas, page 8: The coast of France, England and Ireland. Straight coast line from below the Zuider Zee (the embayment left of Veret) to Denmark. For names, see Table I.

In 1307 the barrier island "*Ammeland*" (Ameland) was mentioned (Hohlbaum, 1879, n. 111). The island was separated from the barrier island east of it (Schiermonnikoog, anonymously mentioned in 1323; Hamb. Urkb., II, 3, 587[17]) by the Zoutkamperlaag.

[17] The letter is a defence of the friars or lay brothers of the convent of Klaarkamp: that they did not steal a mast of a ship which was wrecked on their island (Schiermonnikoog), but legally bought it after it was recovered on Tornasand (Re-reading of the original showed that it was most likely "*Cornasand*" (Pers. comm. K.J. Lorenzen-Schmidt).

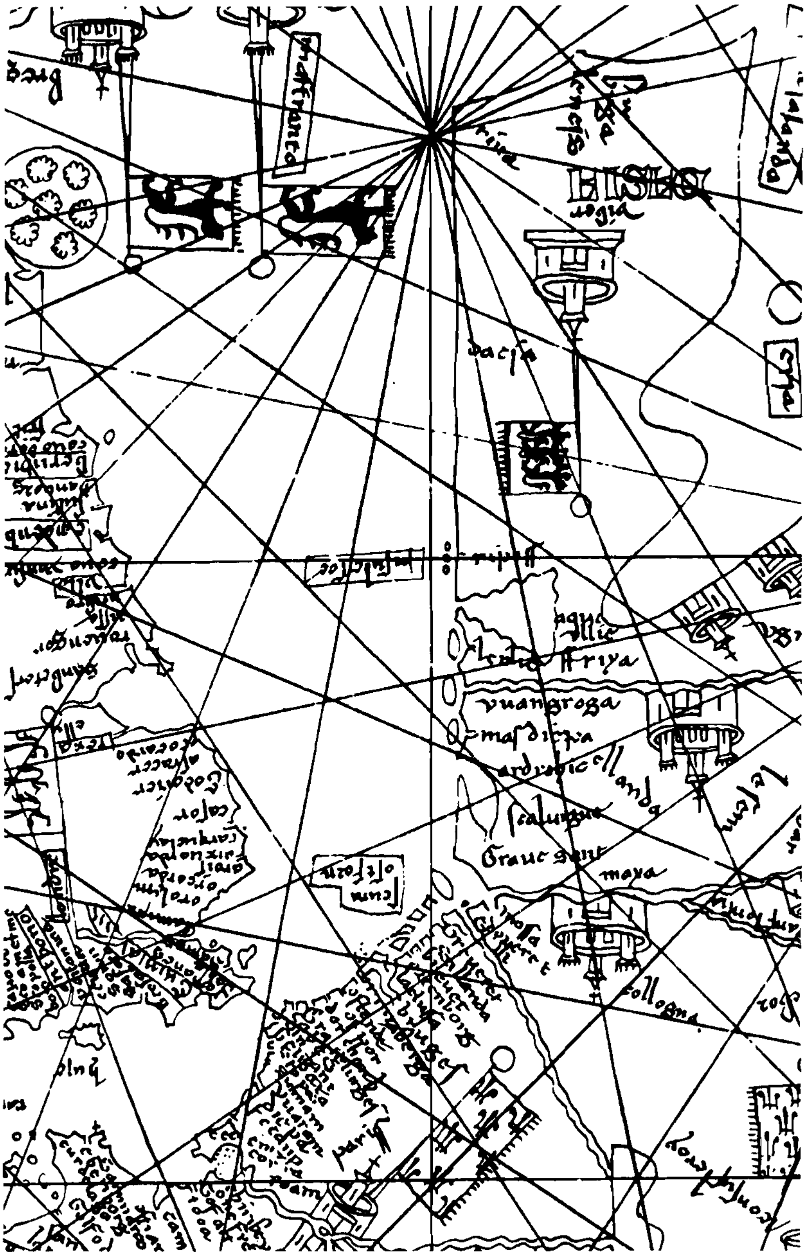

Figure 7: Abraham Cresques (1375): part of the Atlas Catalan, N to the top. Note the N-S-oriented coastline from Gravesant (The Hague) to Denmark, the overall strong contortion of the area, and the depiction of the barrier islands (centre of the chart: the ovals in front of the coast), the backbarrier area and the inlets (indented in the coast). Also clearly visible is the Zuider Zee area (N of Gravesant). For names, see Table I.

It was then called Scudbalke (1307, 1311, Close Rolls, (Hohlbaum, 1879, no. 202; Blok et al., 1899a, no. 239), Sculbalgh (in 1483; Sipma, 1927, no. 326), Schulbalg or Schoelbalg (in 1551; Huusen, 1979), or Scholballich (Wagenaer, 1584/85[XXIX]). The word "schol" or "sceald" (anglo-saxon) means shallow (Schönfeld, 1951), whereas "balg" stands for an inlet, blindly ending in the Wadden Sea; in contrast with river names, such as Boorne, Lauwers, Ems, or "Gat", which indicate inlet systems with a connection to the mainland (cf. Verwijs & Verdam, 1889). The names Scudbalke and Scholbalg thus indicate that the original channel was a shallow inlet ending in the Wadden Sea, and not or poorly connected with the Lauwerszee. Below the Engelsmanplaat, W of the inlet, at a depth of 5-10 m below DOL, a large marine, early Holocene clay plug was encountered in cores (Sha, 1992) and seismics (Oost & De Haas, 1992), indicating that this must have been a high area also in those days. To that side the drainage area of the Zoutkamperlaag must have been restricted. Thus, the Zoutkamperlaag originally must have been a small inlet system; it may have been an important harbour for the people of Paesens. The name Lauwers for the inlet E of Schiermonnikoog indicates that this was originally the main drainage channel for the Lauwerszee (cf. Emo, 1217). The fact that parts of a ship which was wrecked in ± 1320 on Schiermonnikoog, W of the Lauwers Inlet, were recovered on the island Cornasant, (south)east of the inlet, suggests that by that time it was still the passage way for ships (Hamb. Urkb., II, 3, 587, 1323).

Around 1300 the barrier island Schiermonnikoog was situated more to the west than at present, i.e., north of the coast of Friesland[18]. This is also indicated by the fact that Schiermonnikoog has been and still is part of province of Friesland and under the jurisdiction of Oostergoo (the most eastern part of Friesland). It is interpreted that, in contrast to the present situation, the watershed of the barrier island was connected with Friesland around 1300[19]. Indeed, it has been said that animals were driven over the watershed from Friesland to Schiermonnikoog, to graze there during the summer (De Weerd, 1963; after 1165, probably 1306; cf. Westendorp, 1832). Because the watershed in these areas is situated normally at 2/3 of the length of the island, measured from the westpoint, and taking into account the reconstructions of later years, the western end of Schiermonnikoog must have been located somewhat E of the present Engelsmanplaat in 1300. The eastside was located east of the present village; radiocarbon dates (Cleveringa, pers. comm.) show that the area must have been higher subtidal to supratidal in that time. In the period 1350-1550 the drainage of the Lauwerszee was taken over by the inlet system W of Schiermonnikoog (the Scudbalke). The inlet was located directly E of the Engelsmanplaat (Fig. A1).

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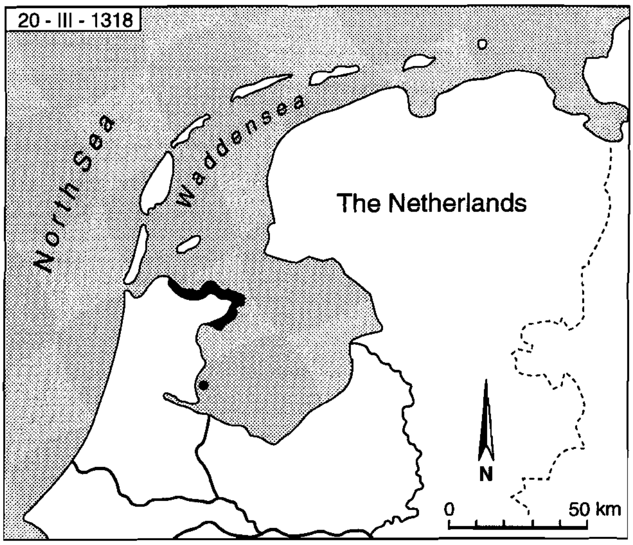
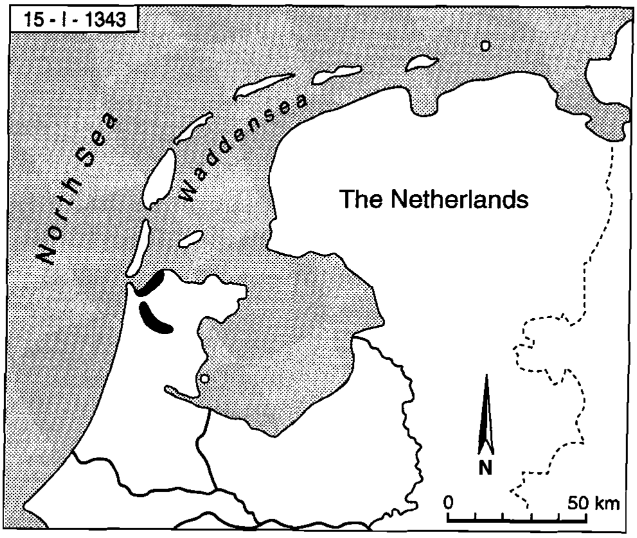
[18] Tradition tells that Schiermonnikoog extended so far to the W and SW that one could originally go to Ameland using a plank, and that one could hear the clocks of Anjum (Winkler Prins, 1867).

[19] De Haan et al. (1983) and Reitsma (1991a) reached the same conclusion. Unfortunately the sources mentioned by them could not be recovered.


20 - III - 1318
North Sea
Waddensea
The Netherlands
N
0
50 km


15 - I - 1343
North Sea
Waddensea
The Netherlands
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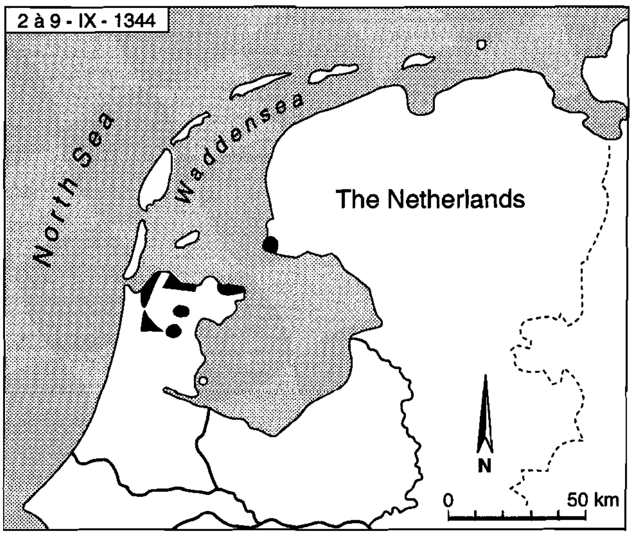
0
50 km


2 à 9 - IX - 1344
North Sea
Waddensea
The Netherlands
N
0
50 km

In the period 1300-1550 the Lauwers Inlet must have lost its drainage function for the Lauwerszee (cf. Overdiep, 1958). Little direct information is available about storm surges in the period (Fig. 8), but from the available data Gottschalk (1971) concluded that storm activity was stronger during the 14th century than in the previous centuries[20].

1350-1400

In october 1375 a storm surge caused severe damage to the barrier islands Huisduinen, Texel, Terschelling and probably also to the other barrier islands; large parts of land near Enkhuizen and Hoorn (Zuider Zee area) were probably lost (Fig. 9; Gottschalk, 1971; Niemeijer, 1975; Buisman, pers. comm.).

In a charter of 1400 (Blok et al., 1899b, no. 1060; Sipma, 1933, no. 4) inhabitants of Ameland got permission to sail or drive through the area west of the Lauwerszee (Oostergoo) to and from Groningen for trade. This gives the impression that it was still difficult to sail via the Frisian Inlet to the Lauwerszee. Nevertheless the "*Lauwerze*" was mentioned in 1389 with reference to shipping and toll (Blok et al., 1899b, no. 1004, 1005). An explanation is that the Lauwers Inlet was still the most important sea way, towards Lauwerszee, and that the Zoutkamperlaag (Scholbalg) was of minor importance. This is also suggested by Das Seebuch[V], a sailing direction describing the inlet configuration of the area somewhere between 1400 and 1500 (Koppmann, 1876). The text does mention the "*Lawerse*" and "*Lauwers*" (most likely Lauwers Inlet, perhaps the Lauwerszee) and the Ameland Inlet, but not the Scholbalg.

Radiocarbon data show that formation of peat in dune valleys, W of the present village of Schiermonnikoog, started 525 ± 30 B.P. (Core 2G/286, Westerplas 1406-1428 A.D.) to 515 ± 55 B.P. (Core 2G/287, Westerduinen 1344-1440 A.D.; Cleveringa, pers. comm.). In humid dune valleys a peaty layer of up to 10 cm can form in a few decennia, when the dunes have become stable after their initial development (Van Oosten, 1986). Thus, around circa 1400 A.D. a supratidal dune area must have been firmly established at that position.

Figure 8 (opposite page): Areas along the Zuider Zee known to be hit by the storm surges in 1318, 1343 and 1344, given in black (redrawn after Gottschalk, 1971).

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[20] As she points out, the detailed, contemporaneous accounts of Emo and Menko in combination with uncritical writings have resulted in a high number of so-called storm surges in the 13th century for the northern Netherlands. Critical study of the historical information shows that the amount of 13th century surges did not deviate from that in the 12th century (Gottschalk, 1971). Even though the sources of information about the northern Netherlands are largely lacking, the amount of registered 14th century surges is comparable to that of the 13th century. For the S and SW area of Holland the amount of registered storm surges increased markedly in the 14th century.

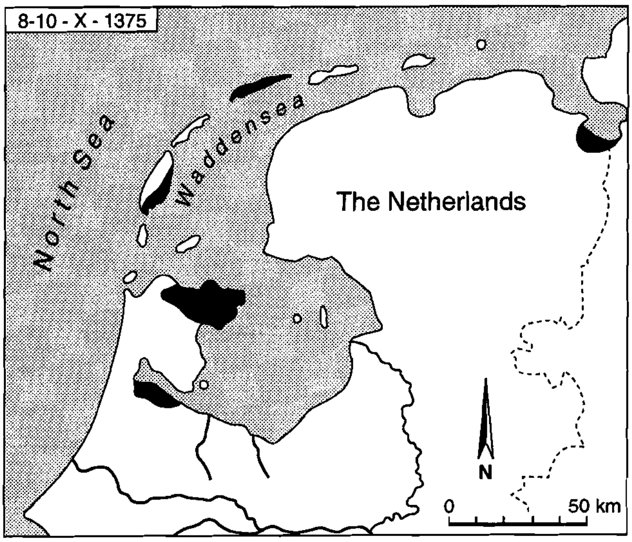

Figure 9: Areas along the Wadden Sea known to be hit by the storm surge of 1375, given in black (redrawn after Gottschalk, 1971).

East of the Lauwers Inlet, the island Bosch was probably situated. East of it was the barrier island Rottumeroog (see above). In the backbarrier area E of the Lauwers Inlet, the islands Cornasant[21] (1323; 1343; 1344; Hamb. Urkb., II, 3, 587; De Vries, 1936; Blok et al., 1899a, no. 379), and S of it, the island Heffesant were situated near the Groningen coast. Heffesant may have formed from sediments eroded from the Lauwerszee and laid down in the inner bend of the channel connecting the Lauwerzee with the Lauwers Inlet (Vos, pers. comm.). Another part of the sediment may have been derived from the river Hunze. The river originally debouched into the Wadden Sea slightly E of the island. Because of

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[21] The islands Cornasant and Bosch were close to each other. Probably they have partly merged in the early 16th century, since Van Deventer (1559; field observations 1536/45) indicates on his chart: "*Bosch als/atq(ue)? Coornsant*" (Bosch, but also Cornasant), Waghenaer (1592): "*....dat Eylandt d'welcke de Schippers ende Stuerlieden den Bosch noemen/ ende van andere Corensant ghenoemt wort....*" (....the island which is called Bosch by the Skippers and the Steersman, and by others Cornasant....). In contrast with this, a chart of 1586 (field observations 1565/75) shows the very small island "*Korsant*", S of Bosch with dunes and a beacon. It is likely that the remnant turned into a higher intertidal to supratidal sandy shoal, mentioned in 1659 ("*Koornheff*"), 1695 ("*Cornsant*") and 1707 ("*Corn en Sijmens Sant*"; cf. De Vries, 1936) and still depicted on a map of Beckeringh in 1745 ("*t Korensant*"). Ownership of such shoals gave right to stranded goods.

compaction and oxidation of inland polders (Edelman, 1974) and ongoing sedimentation, the river Hunze was finally no longer capable of running from S to N and took the reverse route in 1371 (Eibergen, 1941), after which it debouched into the Lauwerszee. The islands Heffesant and Cornasant were later reworked by currents and wave action and gradually disappeared (see below).

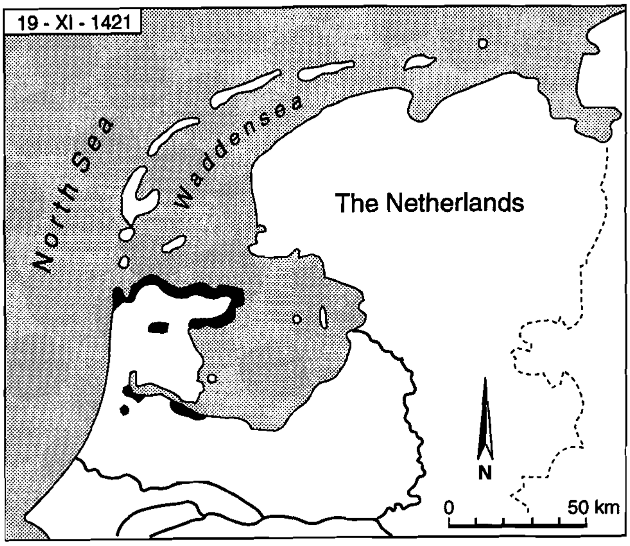
1400-1450

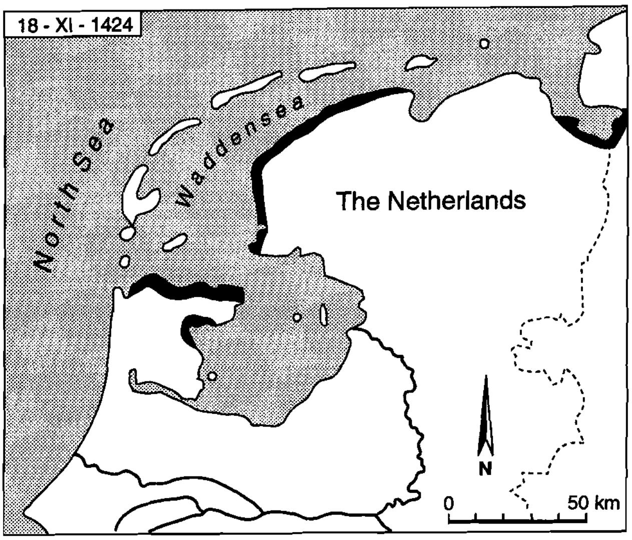
Major storm surges hit Friesland and Groningen on 18-11-1424, and Groningen (NW storm) on 10-4-1446, and caused loss of land in the Dollard area (Fig. 10; Gottschalk, 1975). Also in the area of the Zoutkamperlaag changes occurred, which point to the shift of Schiermonnikoog. A charter of 1440 states that Schiermonnikoog was situated east of Ameland and that the nearest mainland was Paesens[22], SSE of Engelsmanplaat (Westerhoff, 1866; Formsma, 1956). In 1465 the distance between the nearest church and the island is said to have been approximately 2 miles[23] (= 3 km). From Paesens it must have been some 4 km to the island, from Bandt, which was also inhabited, it was circa 3 km. Radiocarbon data (GrN 7267) show that formation of peat in a dune valley started around 430 ± 45 (1444-1592 A.D.) at the location of the present village of Schiermonnikoog (De Jong, 1975). The younger age of peat under the present village indicates that dunes were formed later in that area than in the west, and that the dunes expanded eastward. The downdrift expansion of the dunes was most likely associated with erosion at the updrift side of Schiermonnikoog, as is often the case on the barrier islands. The fact that the nearest mainland was Paesens, suggests that the eastern end of Schiermonnikoog was probably positioned slightly E of the Kooiduinen. These dunes nowadays form the eastern side of the polder area of Schiermonnikoog. If Schiermonnikoog migrated eastward during the period 1350-1450, this may have resulted in an increasing easier drainage of the Lauwerszee embayment by the Zoutkamperlaag.

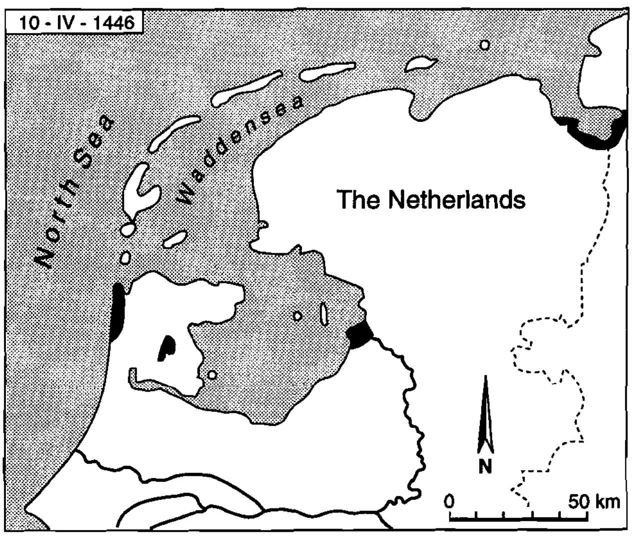
Figure 10 (next page): Areas along the Wadden Sea and Zuider Zee known to be hit by the storm surges of 1421, 1424, and 1446, given in black (redrawn after Gottschalk, 1975).

[22] "*Paidza in Oostergoo landen*"

[23] Johannes van Bethanien & Jacobus, 1465 (in: Winkler Prins, 1867; Mellema, 1981): "*....dicta insula per duo fere milliaria ab omni vicinio hominum et ecclesiarum distare....*" (....the aforesaid island is approximately at two miles (mile of, most likely, 1000 double passus = 1.5 km; cf. Koeman, 1985) distance from all neighbouring places of man and churches....)


19 - XI - 1421
North Sea
Waddensea
The Netherlands
N
0
50 km


18 - XI - 1424
North Sea
Waddensea
The Netherlands
N
0
50 km


10 - IV - 1446
North Sea
Waddensea
The Netherlands
N
0
50 km

1450-1500; Fig. A2

In Holland people complained[24] in 1494 about extensive loss of land along the Zuider Zee and North Sea coast due to increasing erosion and flooding by the sea, especially during storms[25] (Fig. 11). Dyke maintenance is said to have become more expensive since 1477 , '*because the inlets between the barrier islands have become deeper and wider*'[26] (Enkhuizen; Gottschalk, 1975). The inlets must have become deeper and wider, due to the increase in tidal prism (cf. Jarrett, 1976; Sha, 1990; Flemming & Davis, 1991). This may be due to an increase of tidal amplitude or of drainage area, or both. In 1511 it was indeed stated that the height of the tides and the dimensions of the inlets were increasing day by day (Gottschalk, 1975). Tidal amplitude may have increased due to the closure of more southern inlets in Northern Holland (cf. Schoorl, 1973; Van de Plassche, 1982). Increase of tidal amplitude due to enclosures were observed also in the 20th century (Sha, 1989a; De Ronde, 1993). Alternatively, the tidal amplitude may have increased due to a slight change of the tidal system in the North Sea (shift of the amphidromic point; cf. De Ronde, 1993).

Another possibility is that the size of the drainage area changed (cf. Sha, 1989a). The complaints about loss of land indeed indicate that the tidal basin became larger. Erosion of peaty areas has been shown (Schoorl, 1973; Gottschalk, 1975). Especially at the eastern side of the Zuider Zee this may have played an important role; at the western side the land was better protected by a system of dykes and dams (Van de Ven, 1993). An important factor in the loss of land was certainly the gradual lowering of the bottom by Man-induced erosion, oxidation, compaction (especially of peats and clays) and the mining of peat (Schoorl, 1973; Van de Ven, 1993). The oxidation and erosion was further enhanced by the seawater, the effects of tidal differences in water level and the tidal currents. The drainage area may have increased further due to an increase in mean sea-level height resulting from an asymmetric increase in tidal amplitude (cf. De Ronde, 1983).

A strong import of sand must have occurred into the new tidal basins (due to the flood dominance in the area, and the underfilled character of the basins (cf. Sha, 1990). The sand must have been derived from the coasts of the barrier islands of Northern Holland, which were strongly eroded (cf. Schoorl, 1973).

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[24] Enqueste up 't stuck der verpondinghe... van Hollant ende Vrieslandt, gedaen in den jaere 1494 (Gottschalk, 1975).

[25] In chronicles of about 1400 and 1466 the widening of the inlets is already mentioned; an increasing number of surges in the Zuider Zee is reported in 1447, 1466 and 1514 (Gottschalk, 1975; Buisman, in prep.). Around 1600 the Zuider Zee had more or less the same dimensions as in 1900 (Koopstra et al., 1993).

[26] In a charter of 5-7-1466 about the Westfrisian dyke in the Zuider Zee it is stated that the inlets Marsdiep, Heersdiep and Vlie had become deeper and wider and that the floods became higher (in: Gottschalk, 1975). Schoorl (1975), however, shows that the Heersdiep reached its maximum dimensions in the 14th century and had almost disappeared in 1526.

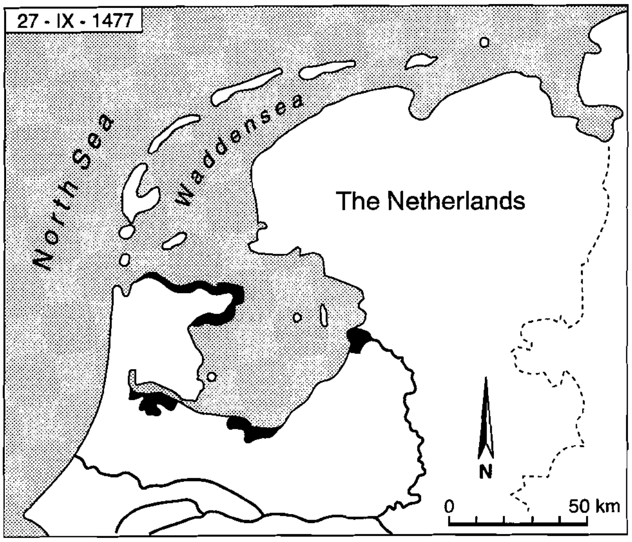


Figure 11: Areas along the Zuider Zee known to be hit by the storm surge of 1477, given in black (redrawn after Gottschalk, 1975).

The Zoutkamperlaag ("*Stobaiasti*" (Roos & Bruggeman, pers. comm.) = Scholbalg) was mentioned by Petrus Roselli in 1462^{VI} (Fig. 12; Table I). It indicates that by that time the inlet was already nautically important. Several sources mention the installation of buoys by the city of Groningen in the period, carried out by the people of Schiermonnikoog (e.g., 1453 (Van Buijtenen, 1954), 1493, 1496 (Westerhoff, 1866). Since at least ±1490/1500 buoys were installed in the Zoutkamperlaag (Huussen, 1979). The mention of the Zoutkamperlaag ("*Sculbalgh*") as a border of Ameland[27] in a charter of 1483 (Sipma, 1927, no. 326) gives the impression that the Pinkegat Inlet was insignificant; its presence is not mentioned in sailing directions of the 15th and early 16th century. Nevertheless, an inlet must have been

[27] 21-2-1483 (In: Sipma, 1927): "....*up Aemland moegha gheyja wrt land fan Nes twischen dae Rijd en(de) den Sculbalgh om hieare(n) toe voermeythien Mer jelkers scil dy abt int land fors. als twischen dae Rijd en(de) den Sculbalgh dat riucht jeyjen selff haulda offte toe des claesters profijt verheera ffoerdmaer alla renthen deer ko(m)med fan dae pijnken als foer dae fiskenye....*" (....on Ameland (he) may go over the land of Nes between the Rijd and the Sculbalgh to disport himself. But also will the abbot keep the right of the aforesaid land between the Rijd and the Sculbalgh for himself or hire it out to the profit of the cloister, furthermore all interests which come from that "pijnken" (probably fishing with a boat; the inlet E of Ameland is still called 'Pinkegat') as well as for that fishing....). The 'Rijd' in the charter is most likely the present 'Nieuwlandsrijd', which is a low-lying washover complex (Rijt means split or crack) between Buren and Oerd on Ameland (cf. Isbary, 1936).

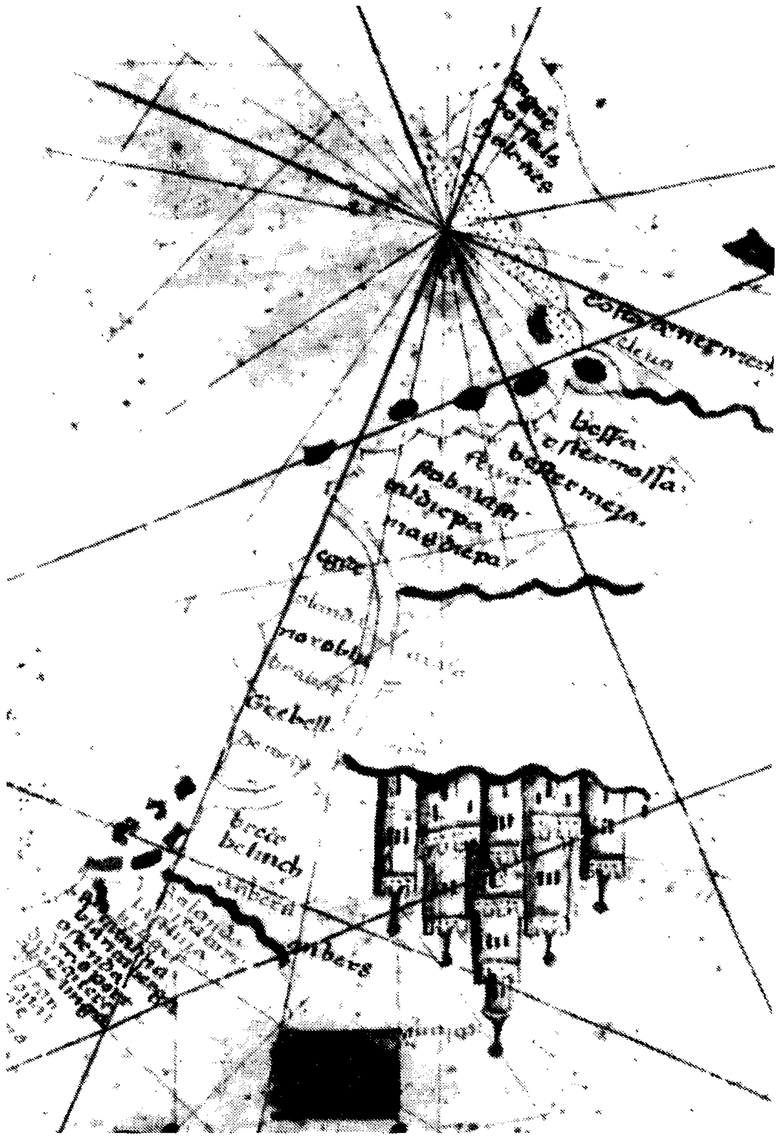


Figure 12: Petrus Roselli (1462): part of the Portulano: Atlantic, Mediterranean and Black Sea, N to the top. The first known, more realistic depiction of the German Bight (Koeman, 1985). The chart gives a series of names of the inlets and depicts many barrier islands. For names, see Table I.

there, because behind it a large tidal drainage basin was present (cf. Van der Spek, 1994)[28]. Moreover, a map of sGrooten (1592)[XXVIII] shows its existence. The typical pointed form of the eastern end of Ameland (due to lateral accretion) on the map of Van Deventer (1559, field observations 1536/45)[XX] indicates that an inlet must have existed.

The Lauwers Inlet was shown on charts slightly younger than the one of Petrus Rosseli, such as the Portulano attributed to Ch. Columbus[VIII], 1492 (Lang, 1955; De La Ronciere & Mollat du Jourdin, 1984). Maps of the late 15th and early 16th century show that the Lauwerszee embayment was the waterway between the North Sea and the cities of Dokkum[VII] and Groningen[IX]. Small tidally influenced meandering rivers formed the connection between

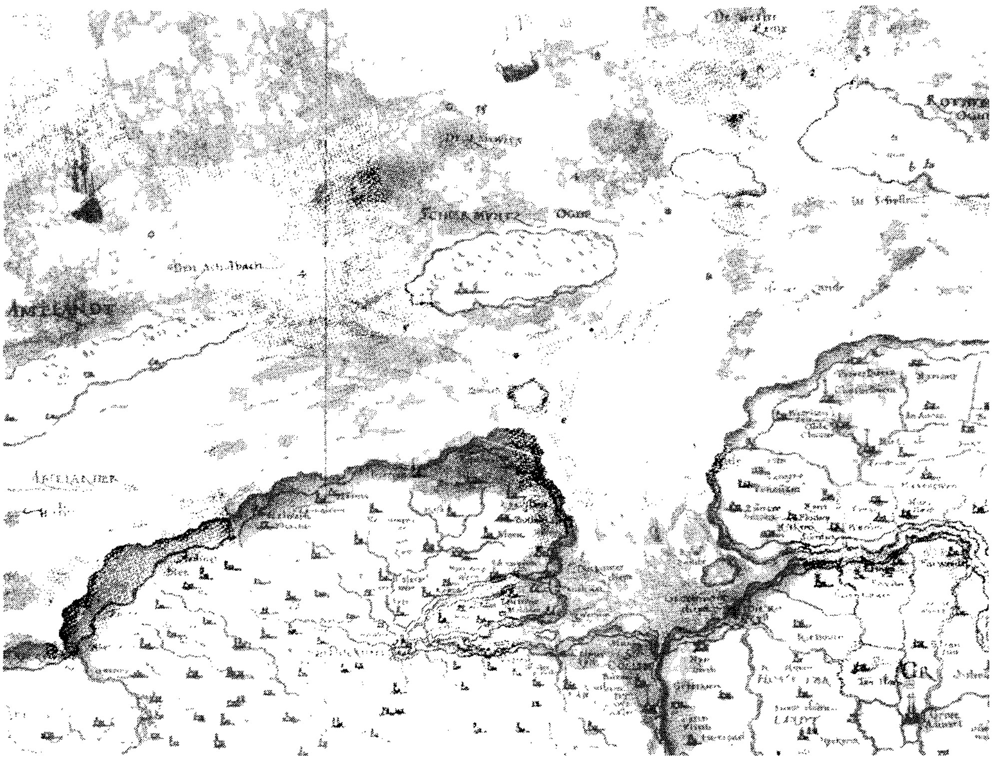


Figure 13: Christiaan sGrooten (1579-92): "Madrilenean Atlas", showing the two waterways from the Lauwerszee embayment to the North Sea, via the Lauwers Inlet and the Scholbalg (Zoutkamperlaag Inlet).

[28] Drainage via the Zoutkamperlaag Inlet would be difficult, because the inlet system was separated from the Pinkegat area by the Engelsmanplaat and the Early Holocene clay plug.

these cities and the Lauwerszee, as is clearly shown on a fragment of the first Dutch chart of 1526[XI]. Another map fragment[X] from about the same period mentions both the Lauwers Inlet and the Zoutkamperlaag and gives the impression that both were important. This is supported by copies of a chart of Anthonisz, printed around 1543[XIV, XIX, XXIII] (Lang, 1986). These charts show them as separate outlets from Groningen to the sea and from Dokkum to the sea. How the situation must have been in reality is shown on maps of 1573 and 1579-92[XXVII, XXVIII] of sGrooten (Fig. 13based on older sources). In all his depictions of the waterways to the Lauwerszee, sGrooten showed two inlets: one W of Schiermonnikoog and one E of it. Both inlets were connected with the Lauwerszee. The same is indicated by the sailing direction of 1541[XIII]. It gives sailing directions for both inlets and warns for their ebb-tidal deltas. Also the third print of Anthonisz' chart of 1560[XXI], shows both inlets connected with the Lauwerszee. The last major dyke over the former Middelzee was closed in 1505; the area must have been supratidal around 1500.

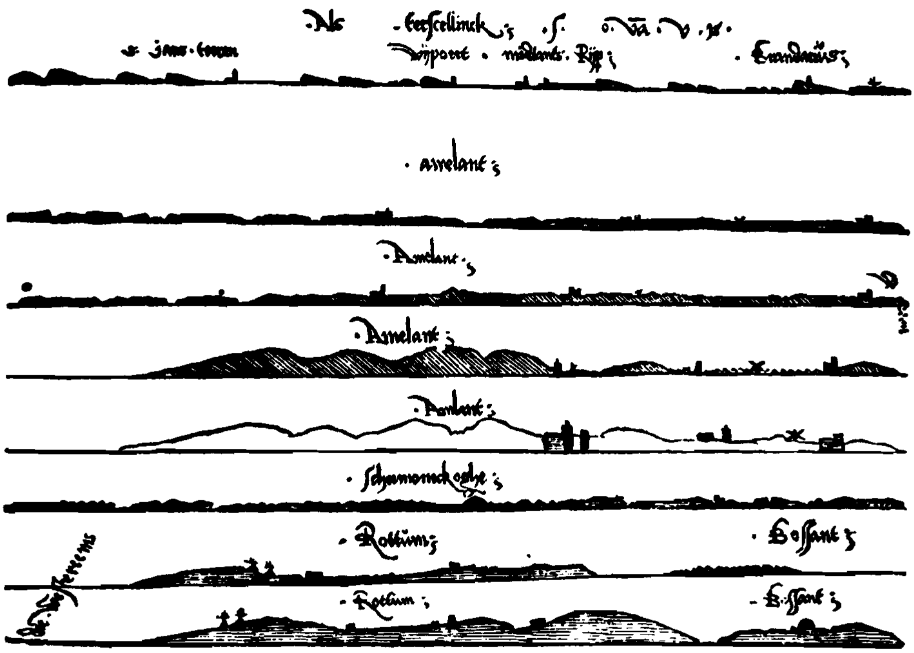


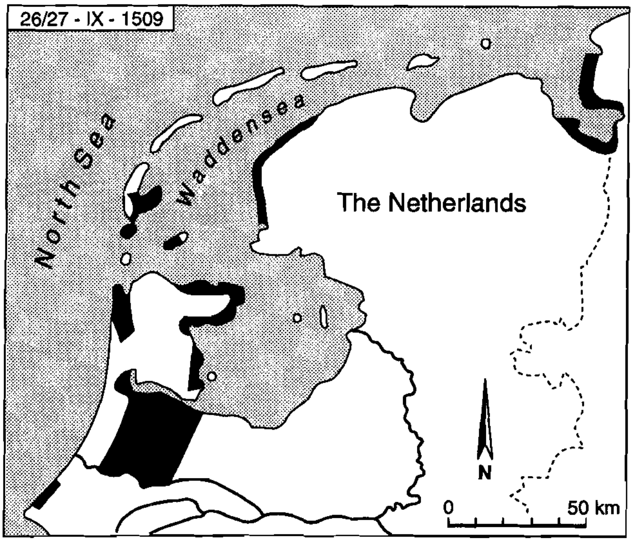
Figure 14: Cornelis Anthonisz (1558): "Caerte van die Oosterse See", side-views of the islands from the North Sea, showing the islands Terschelling to Rottumeroog.

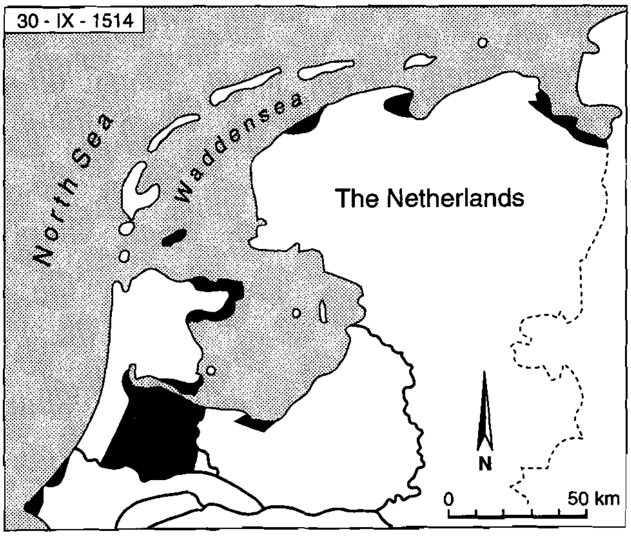
The reconstruction of 1500 is as follows: The eastern end of Terschelling was still almost 11 km west of its position of 1976. The Borndiep Inlet still consisted of two separate channels draining, in an estuarine way, the remnant of the once larger Middelzee (Van der Spek, 1994). The most western channel was perhaps still the most important one (Seebuch, circa 1450[V]). At the western end, Ameland was positioned approximately 0.5 km W of its present position. At its eastern end it was more than 1.5 km west of its position of 1976. Perhaps it was even positioned further to the E, if Oosthuisen indeed was located E of Oerd. Its HW line at the northern coast was probably almost at the same position as nowadays. At the southern side the HW-line was positioned up to 2 km southward of the HW-line of 1976 (in the Ballumerbocht and Oerd). The Pinkegat was small or insignificant. East of it a large shoal was present (Malesant, the precursor of the Engelsmanplaat), which probably could be reached from Ameland.

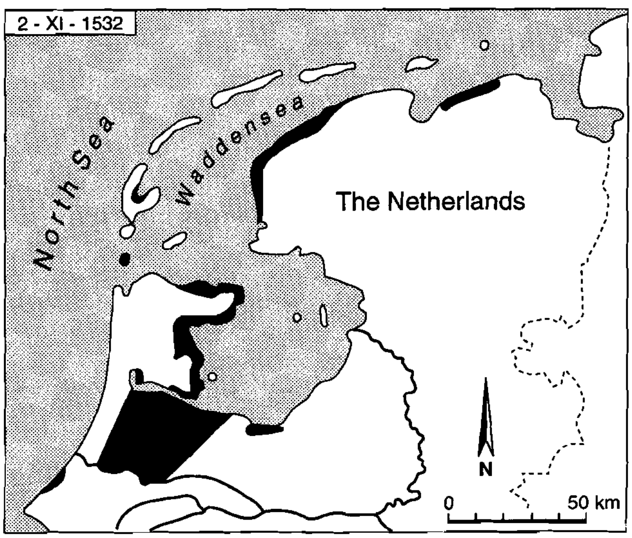
The Zoutkamperlaag was oriented to the NW and was the main drainage channel of the Lauwerszee. The Lauwers Inlet was less important for the drainage. In the ebb-tidal delta a shoal (Dircsant) was present along the NW-oriented channel. The western HW-line of Schiermonnikoog was positioned 2.5 km farther to the W, whereas its eastern end was positioned more than 6 km westward, of its position in 1976. According to sGrooten (1579/92)[XXVIII], a bar was attached to the island at the NW side. The eastern small island/shoal of Simens Sandt was separated from Schiermonnikoog by a small inlet channel ("*de Knock, or Knockebalg*"). East of it, the Lauwers Inlet (circa 10.5 km W of its position in 1976) entered. East of the inlet, the island Bosch was situated, probably with steep high dunes at its western side (sailing direction, 1541[XIII]; Fig. 14). East of Bosch the Schild Inlet entered, with E of it the large island Rottumeroog. At that moment Rottumeroog probably still had a village, and had a considerable size (according to the sailing direction of 1541[XIII] and the map of sGrooten (1579/92)[XXVIII] it was ±11 km long). On the reconstruction, the island has been copied from sGrooten (1579/92), but re-oriented with its length axis ENE-WSW, in order to fit the other maps of the period, the later reconstructions, and those of Lang (1958). The western HW-line of the island was reconstructed to have been more than 11.5 km farther to the west and its eastern HW-line 4 km more to the W with reference to the position in 1975.

Within the Lauwerszee embayment the Scholbalg bifurcated into two channels. A southern channel which further upstream turned west, towards the city of Dokkum (Dokkumer Diep), and a SE-NW oriented one towards the city of Groningen (Groninger Diep). The two channels were separated by a triangular shaped, intertidal flat.

Figure 15 (next page): Areas along the Wadden Sea known to be hit by the storm surges of 1509, 1514, and 1532, given in black (redrawn after Gottschalk, 1975).


26/27 - IX - 1509
North Sea
Waddensea
The Netherlands
N
0
50 km


30 - IX - 1514
North Sea
Waddensea
The Netherlands
N
0
50 km


2 - XI - 1532
North Sea
Waddensea
The Netherlands
N
0
50 km

1500-1550; Fig. A3

The configuration of the inlets and ebb-tidal deltas which drained the Lauwerszee changed in the period 1500-1550 (Figs A2 and A3). The change may well have been encouraged by the storm surges in the first half of the 16th century[29] (Fig. 15; cf. Gottschalk, 1975), stimulating downdrift sediment transport (cf. Steijn et al., 1992). In 1532[XII] a warning was given that a shoal of the Zoutkamperlaag ebb-tidal delta expanded seaward. This was likely the result of a net growth of the ebb-tidal delta. In his atlas in Brussels[XXVII] sGrooten already names the connection between Lauwerszee and Lauwers Inlet: "*Dat olde Groeninger Diepe*" (the old/former Groningen channel), most likely referring to the decline of that channel. The Lauwers Inlet was mentioned in sailing directions of 1541[XIII] (Rogge, 1885) and 1558[XVIII] (Keunig, 1950), but not any more in one of 1566[XXV] (Knudsen, 1920). This suggests that by

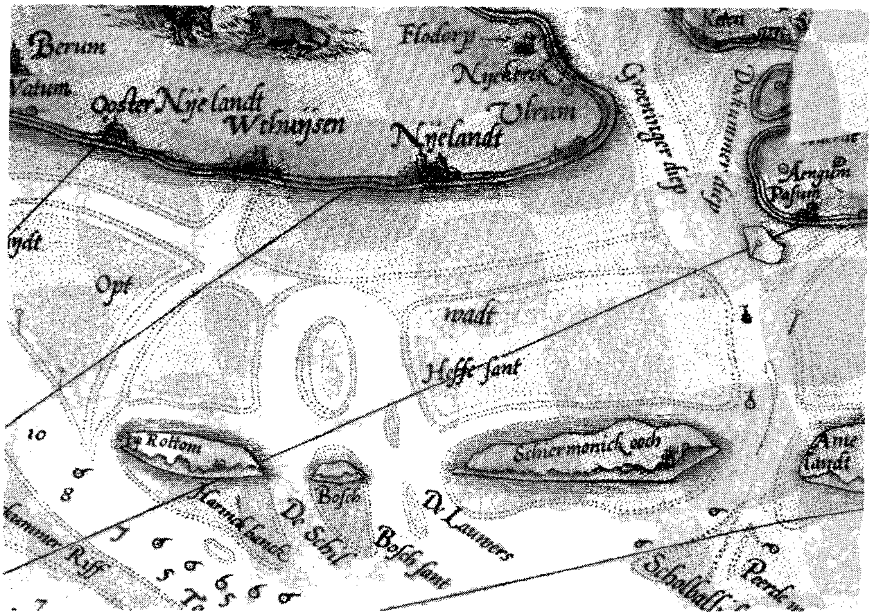


Figure 16: Part of the chart of Waghenaer (1584/85) showing an intertidal channel, between Lauwers Inlet and the Lauwerszee embayment (N of the Nijelandt).

[29] On 26/27-9-1509 a major storm surge occurred. The Zuider Zee area was extremely badly hit, as well as the coast of Friesland; from the Groningen coast there are only reports of the Dollard area (Gottschalk, 1975). On 2-11-1532 a major storm surge (NW storm during neap tide) inundated the northwestern part of Friesland; Groningen had likely also been damaged by a storm (Gottschalk, 1975). Minor storm surges in the area occurred in 1502, 1505?, 1507, 1508, 1509, 1512, 1514, 1516, 1517, 1524, 1538? and 1543.

that time the Lauwers Inlet had become insignificant, although a small ebb-tidal delta remained (Waghenaer, 1584/85). Indeed, after 1530 a dispute started between the city of Groningen and emperor Charles V about the ownership and rights to put beacons in the Zoutkamperlaag Inlet. In 1551 it was stated that the Lauwers Inlet could only be sailed by small vessels during high water, whereas during low water one could go on foot from Schiermonnikoog to the mainland of Groningen (although this was disputed by people of Friesland, and that only the Zoutkamperlaag Inlet was wide and deep enough at that time for sea-going ships (Van Buijtenen, 1954; Formsma, 1954; Overdiep, 1958; Huussen, 1979). Thus, around 1550 the Lauwers Inlet was only connected with the Lauwerszee by a small inter- or just subtidal channel. An intertidal channel was also observed by Waghenaer around 1570/79 (Lang, 1957), and indicated in his famous sailing direction of 1584/85[XXIX] (Fig. 16). Slightly later sailing directions[XXX, XXXI] even indicate the absence of a connection between the Lauwerszee and the Lauwers Inlet. On one of the charts of the sailing direction of 1586[XXXI] it is written in the Lauwers Inlet: *end of the Lauwers Inlet*' (*"Die Ende vanden Lauwers"*), indicating the abandonment of the Inlet. Slightly afterwards, the Lauwers Inlet was only mentioned as a save, wind-sheltered harbouring place[XXXIII, XXXV, XXXVI], vital to the wooden ships of those days.

The reconstruction of 1550 is as follows: The island Terschelling was probably still more or less at the same position as in 1500 (cf. Ligtendag, 1990). From the map of Van Deventer (1559)[XX] it appears that east of the dune area a large sandy shoal may have been present (given here as intertidal). For the position of the islands Ameland and Schiermonnikoog in 1550 the fairly accurate maps of Van Deventer (1559)[XX] were mainly used[30]. Around 1550, the length of the barrier island Ameland was about 22 km (nowadays, 23.5 km). The HW-line of the western and eastern end were 0.5 respectively 1.5 km more W then in 1975.

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[30] On the mainland the accuracy of Van Deventers maps was unsurpassed until 1616, when Wicheringe produced his map (Van der Top, 1992). For the depiction of the Wadden Sea islands, however, the accuracy Van Deventers work was only surpassed by local island maps and the province maps in the late 18th to early 19th century (Athallin, 1811 & Krayenhoff, 1809/1823). The mapping of the barrier islands up to Schiermonnikoog was quite accurate. The map showed a natural form of the barrier islands (Koenders, 1986) and topographical details on the islands which are confirmed by descriptions. The western side of Ameland must have extended more to the W than drawn on Van Deventers map, because W of Hollum the hamlet Sier was located; also the dunes were probably not depicted fully correct. The low, eastern end of Schiermonnikoog extended probably somewhat more to the east. The island of Bosch did show a survey mark, but was most likely displaced to the E to create space for the title-cartouche (cf. Koenders, 1986). The position of the western end of Bosch is given by Haeyen (1585) on the line over the churches of Leens and Hornhuizen; the only churches in the area with a high tower (Van Deventer, 1556). The E-W dimensions have been obtained from the dimensions on the maps of Van Deventer (1536-1545, c. 5.4 km) and Waghenaer (1570-1579, c. 3.6 km). The position of Rottumeroog has not been copied, because it showed no survey mark. The maps of many later authors, such as Camotius (1566), Abaraham Ortelius (1568), Gerard de Jode (1578), Gerard Mercator (1585) and Johannis Quad (1600) are, directly or indirectly, copies of Van Deventers maps as far as the barrier islands are concerned (cf. Vredenberg-Alink, 1974; cf. Koenders, 1986).

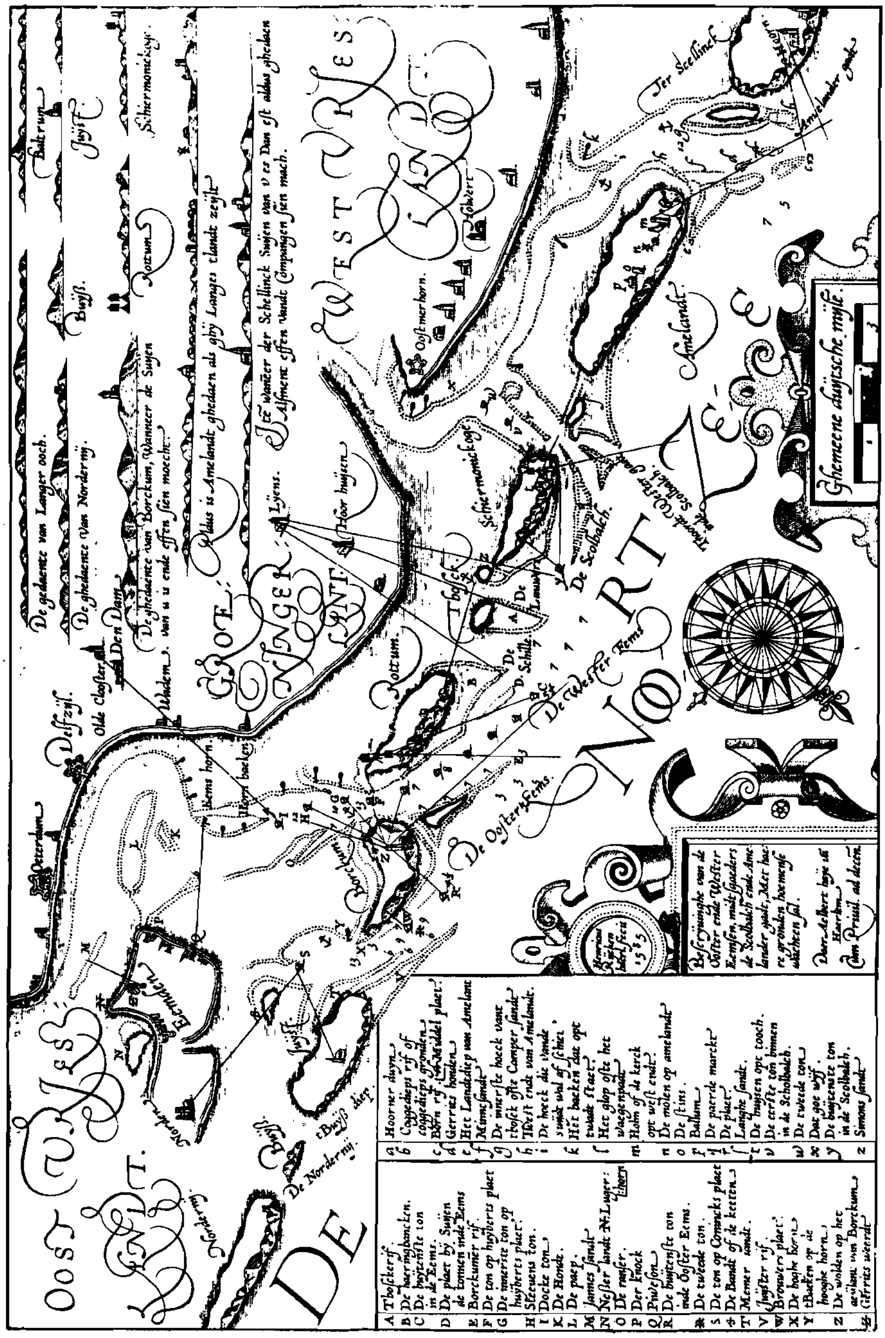

Figure 17: Chart of Haeyen (1585) of the eastern Dutch Wadden Sea.

The width of the island was about the same as at present, except for the northwestern part (nowadays situated 1 km further seaward), the SE part (near Oerd), which in 1500 was extending more than 0.5 km further to the S, and the south-central part of the island, which extended at maximum, almost 2 km more to the S. Nowadays a channel bend the 'Ballumer Bocht' is present at the latter position, with its erosive outer bend towards the island. Along large parts on the northern side of the island the minimal HW-line (close to the dunes as possible) coincides with the present HW-line.

The Pinkegat must still have been insignificant around 1550. Van Deventer (1545) did not mention it. Waghenaer (1584/85)[XXIX] did not indicate an inlet, but Haeyen (1585)[XXX] suggested an inlet by drawing a supratidal shoal "*Langhe Sant*", the precursor of the Engelsmanplaat. East of the shoal the Scholbalg was situated; by then the major drainage channel of the Lauwerszee. In 1551 it was stated that the main outer channel was NW-oriented (with a depth of 5-6 m around 1541[XIII]; Rogge, 1885). A new shallow, NE-oriented outer channel (most likely a marginal flood-channel) had formed (Huussen, 1979). North of these channels a shallow, subtidal to intertidal shoal ("*Paardemarkt*") occurred. The Paardemarkt shoal consisted of two parts, separated by a shallow subtidal channel.

Comparison of the maps of Van Deventer (1559)[XX] and Waghenaer (1584/85)[XXIX] shows that erosion had occurred at the NW side of Schiermonnikoog (see below). An important cause was probably the formation of the new NE-oriented outer channel along the NW side of the island. The position of the western HW-line of Schiermonnikoog was less than 2 km more to the W, whereas its eastern end was more than 6 km more to the W, than in 1975. The length of the island was said to be 14.8 km in 1541[XIII], and 11.1 km in 1585[XXX] (the latter by Haeyen, who knew that the Simens Sandt was a separate island). The reconstruction indicates a length of 13.1 km in 1550 for Schiermonnikoog including Simens Sandt and 10.7 km without Simens Sandt). The eastern small island/shoal of Simens Sandt was separated from Schiermonnikoog by a small inlet/washover channel around 1550 ("*de Knock, or Knockebalch*"; pers. comm. Roos & Bruggeman). East of it the Lauwers Inlet entered (Formsma, 1954), which by 1551 had become relatively small and had only an intertidal connection with the Lauwerszee (Huussen, 1979). The position of the small, probably uninhabited island Bosch was reconstructed from the charts of Haeyen (1585)[XXX]. The western side of the island was positioned on the line over the churches of Leens and Hornhuizen (Fig. 17). On the reconstruction (Fig. A3) the island has been drawn with steep high dunes at the western side. This still must have been the case in 1550[XV, XIX, XXV]. East of Bosch the Schild Inlet entered, with E of it the still relatively large island Rottumeroog (between the HW lines almost 7.5 km long; Haeyen (1585) gives a length of 7.4 km). The position of the island was reconstructed mainly after the reconstruction of 1580 by Lang (1958), but 1 km more to the E. This is more realistic considering the dimensions of Bosch, which had been eroded at both sides after 1500. The decrease of its length will partly have resulted from the frequent storms in the 16th century (Gottschalk, 1975) and the probable absence of an island keeper on Rottumeroog before the 17th century (cf. Ufkes, 1989). East of Rottumeroog the estuarine Western Ems Inlet entered.

In the backbarrier the islands Heffesant and Cornasant were still present. In 1503 the island Heffesant was used for sheep herding: a sign of deterioration (Eibergen, 1941). In 1535 the monastery of Aduard partly gave away the islands Heffesant, Cornasant and Bosch (De Vries, 1936); at that time Heffesant may still have been inhabited. Only a small supratidal shoal Cornasant was drawn on a chart of a MS sailing direction of 1586[XXXI] (field observations 1565/75). Cornasant probably disappeared largely by merging with the small island Bosch before 1575, (see above; Van Deventer, 1559[XX]; Edelman, 1964).

Near the hamlet Ferwerd, between Holwerd and Ternaard, and near the hamlet Anjum, the Frisian coastline was situated further landward than at present. Around 1550 the hamlet Wierum was still larger than nowadays (Poortinga, 1956). Just off the northeastern coast of Friesland the small island Bandt still occurred, separated from the mainland coast (perhaps connected by a dam; RWS, 1948). Around 1550 the configuration of the Scholbalg within the Lauwerszee embayment had remained largely the same as in 1500. In 1550 the coastline of Groningen was situated 1.5-2 km more to the S than at present.

1550-1600; Fig. A4

During the period 1550-1600 many storm surges occurred (Fig. 18). Major storm surges occurred in 1552 and 1570 (Gottschalk, 1975). For the area studied it is only known that during the surges of 1552 dykes were breached in Friesland and Groningen (Gottschalk, 1975). More damage may have occurred on 15-9-1559, when Texel was flooded and dykes in Friesland were breached (Gottschalk, 1975). The 'All Saints day'-surge of 1-11-1570 has been quite influential on the Dutch coast. The spring flood, in combination with a storm from the SW turning to the NW, resulted in one of the worst storm surges (lasting two tides) ever to hit the Dutch coast (Gottschalk, 1975; Buisman, in prep.). The water reached a height of around +4 m DOL from Scheveningen to Friesland (Gottschalk, 1975). This is higher than the maximum levels reached in the whole period 1900-1994 in that area and occurs only several times per thousand years (cf. RWS, 1994). Strong coastal erosion occurred in Holland, amongst others in Scheveningen, Noordwijk, Berkheide, the barrier islands Callantsoog[31] and Huisduinen[32] (Schoorl, 1973; Gottschalk, 1975; Bremer,

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[31] Dunes were breached from the NW and the village and church were destroyed (Schoorl, 1973).

[32] Waghenaer (1592, data probably collected before 1586), in his chapter entitled: "*Hoe sommige gaten, gronden ende santberghen veranderen, ende op sommige plaetsen also verstuyven ende met storm ende onweer overloopen, dat sy ten laetsten in een bloote strant verandert worden*" (How some inlets, shallows and dunes change, and are blown away at some places, and flooded during storms, to such extent, that they change finally into a bare strand plain):

"....*dat sommige custen, die niet al te wel met duynen beset en zijn, somtijts also veranderen ende verstuyven, datmen daer aff gheen kennisse en can ghenemen. Als by exempel: De Keteldüyn*

1983). Damage also occurred on the barrier islands Texel and Vlieland, and the backbarrier islands Wieringen, Marken, Urk, and Schokland (Gottschalk, 1975).

The coasts of the Zuider Zee, of Friesland and of Groningen were flooded over a large area (Fig. 18)[33]. Around Wierum the coastline retreated due to strong erosion. This resulted in the destruction of the northern side of the village[34], probably mainly during the 'All Saints day'-surge of 1570 (Poortinga, 1956). A connection of the island Bandt with the mainland was totally destructed (RWS, 1948). Of the barrier islands in the area studied, Ameland may have been damaged[32]; the hamlets Swartewolden[35] and Oerd/Oosterhuizen/-

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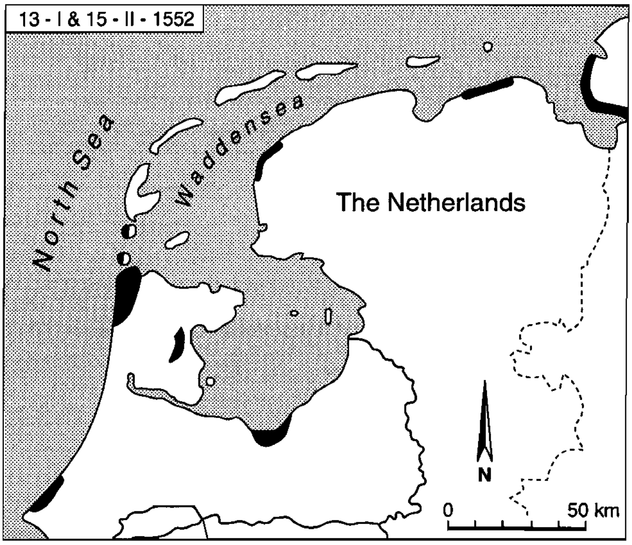
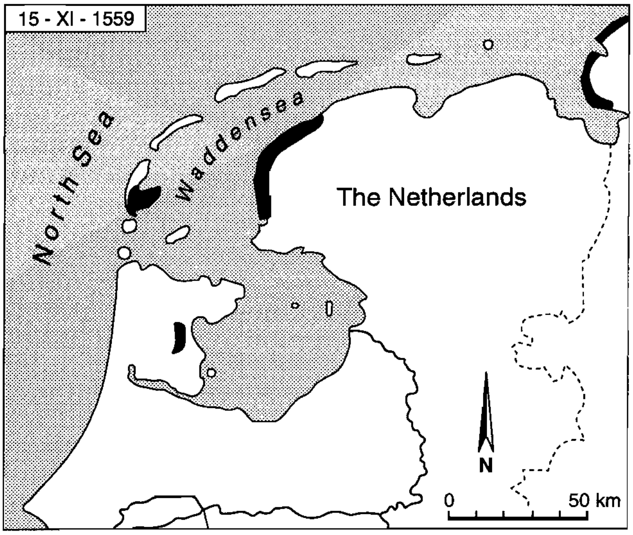
plach over 24 iaren de geheele kennisse van Westvrieslandt (....of oock Noordthollandt) te wesen: ende is nu ter tijt alsoo vergaen ende verstoven, ende met de zee overloopen, dat dese voorseyde uitnemende hooghe duyn in een slichte ende effen strandt verandert is. Van ghelijcke oock aende custe van Oostvrieslandt, tusschen die Lauwers ende die Scil, dat Eylandt d'welcke de Schippers ende Stuerlieden den Bosch noemen, ende van andere Corensant ghenoemt wort, binnen tien oft twaelf iaren herwaert alsoo verstoven is, dat in een slichten strandt is verandert. Dergheliјcx tot meer ander plaetsen, als zijn Amelant, Schimmelickoogh, Huysduyne, ende dat Hontsbosch, alwaer die stranden ende duynen seere verandert zijn...." (....that some coasts, which have not too much dunes, sometimes change and are blown away to such extent, that it is not possible to recognize them (any more). As an example: The dune 'Keteldunyn' (the dune complex was already eroding, as mentioned in the sailing directions of 1541 and 1566, but was destroyed in 1570 and the years thereafter; Schoorl, 1973) used to be for 24 years the identification mark for all 'Westvrieslandt' (....also called 'North-Holland'), and nowadays it has decayed and blown away, and is flooded by the sea to such an extent, that this aforesaid outstanding high dune, has changed into a low and even beach. The same also (happened) on the coast of 'Oostvrieslandt', between the Lauwers (Inlet) and the Schil, the island which is called Bosch by the Skippers and the Steersman, and by others Corensant, which ten or twelve years ago has been blown away to such and extent, that it changed into a low beach. Similarly (happened) on other places, such as Ameland, Schiermonnikoog, Huisduinen, and the Hontsbosch, where the beaches and dunes have changed strongly....).

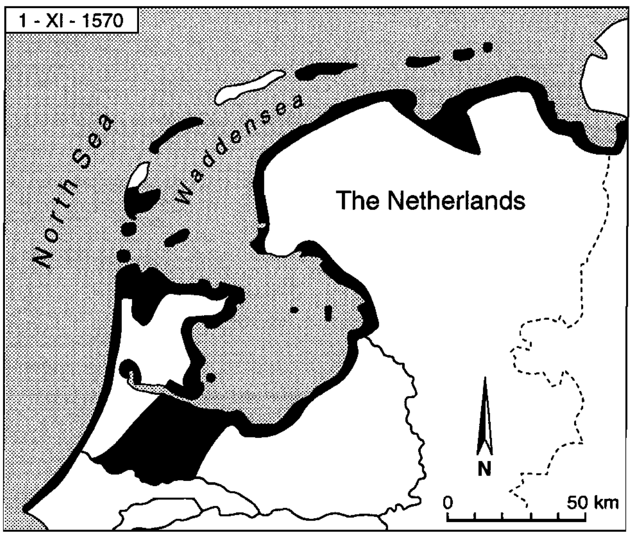
33 From the family register of the family Elema, Uithuizen, Groningen in: Buisman (in prep.): "*Den 1 November is 1570, zynde Allerheiligen des avonts omtrent 9 uiren verhieff sich de zee alhier over te loopen, alsoo starck dat het solte waater an myn groepen stondt ende dat noch myn achterdeur wilde inslaan ende gansch Groningerlandt, dyken ende dammen in stucken....*" (The 1st november 1570, being All Saints-day, in the evening around 9 o'clock the sea rose here and flooded over (the dyke), so strongly that the salty water reached my stable and wanted to destroy my backdoor and everywhere in Groningen, dykes and dams (were) broken....).

34 The church of Wierum, originally in the middle of the town, stands near the dyke; the Wadden area N of it was in 1956 still known as: '*the graveyard*' (Poortinga, 1956).

35 Originally located S of Kooiplaats.


13 - I & 15 - II - 1552
North Sea
Waddensea
The Netherlands
N
0
50 km


15 - XI - 1559
North Sea
Waddensea
The Netherlands
N
0
50 km


1 - XI - 1570
North Sea
Waddensea
The Netherlands
N
0
50 km

Oldhuys[36] were not mentioned any more after 1558 A.D. (ordonance, ARA, 25 March 1558, Brouwer, 1936; Van Oosten, 1986).

Probably Schiermonnikoog also suffered from erosion. In 1585 Kempius stated that the western end of the island had been eroded by storms (Feenstra, 1990). Albert Haeyen was on the island between 1580-1584 A.D., and he was told that the western end had been eroded strongly in the past[37]. Furthermore, comparison of his descriptions with others indicates strong erosion around that time (cf. Feenstra, 1990). In 1559 Jacob van Deventer (field

Figure 18 (opposite page): Areas along the Wadden Sea and Zuider Zee known to be hit by the storm surges of 1552, 1559, and 1570, given in black (partly after Gottschalk, 1975).

[36] Maps of the late 16th and early 17th century show no village on the eastern side of the island. A map of 1665 indicates that the Oldhuysder duinen, and thus likely also Oldhuys/Oosterhuizen, were located E of the dunes of Oerd. According to Allan (1857, in: Bakker, 1970) Oerd and Oosterhuizen were synonyms, but the map suggests that this is incorrect. The map of 1650 shows that by that time only two small buildings remained in the Oerder dunes. Oerd did not vanish completely in 1570. After the storm surge of 1825 an anti-catholic commemorative medal of 1584 was found at the site of Oerd (Van Leeuwen, 1826, in: Van der Molen, 1968). The medal suggests that Oerd may have been a hide-out for the "Watergeuzen" ('Sea-Beggars': a Dutch seagoing resistance movement/pirates against the Spanish troops), who used the islands as a basis in those days. The Pinkegat was still called Geuzegat around 1700 (Guitet, 1708/1710). The writings of Van Leeuwen show that finally the remnants of the village have been partly covered by dunes. By 1956 the fishermen of Wierum still had a memory of a walled place "*Hoarn*" on eastern Ameland (Poortinga, 1956).

[37] Albert Haeyen (1585, field observations 1580/84): "*....datter wel sijn sommighe Eylanden die in etlicke iaren meer dan die helft afgespoelt sijn, als het Eylant van Schirmonickoogh. My is aldaer van oude lieden onderrecht datmen eertijts van d'West-eynde van het Oogh op het Oosteynde van Aemlant met eenen bal gheslaghen heeft, d'welcke nu meer dan twee mijlen wijt is. Jae dat meer is, my is al daer voor seker vertelt, datter noch persoonen binnen vier Jaer herwaerts ghestorven sijn, die ghedincken mochten dat die huysen die nu op d'West-eynde staen, in henlieder ieught midden in t'lant stonden.*" (....that there are some islands of which half has been washed away in several years, like the island of Schiermonnikoog. I was told by old people that originally one could hit a ball (according to Schoorl, 1973, p. 86) some 340-480 m) from the western end of Schiermonnikoog onto the east end of Ameland (perhaps a memory of the original situation around 1300, and even then only possible if Engelsmanplaat was considered part of Ameland), which is now more than two miles wide (approximately 15 km, for German miles, which is a too large distance, or 12.7 km for Spanish miles). Yes, moreover I was told for sure that people, who died less than four years ago, could remember that the houses which now stand on the western end, were standing in the middle of the (is)land when they were young.)

observations 1536/45)[XX] showed a tower[38] standing a quarter from the western end of the island. Waghenaer (1584/85, field observations 1570/79)[XXIX] showed that the tower already stood behind a row of dunes in his time. The straight line formed by the NW coast of Schiermonnikoog on his map indicates that the coast was eroded. In 1585 Haeyen (field observations 1580/84)[XXX] stated: '*on the western end a tower stands near to the beach*'[39], suggesting strong erosion between 1536/45-1570/79 and 1570/79-1580/84. Coastal erosion probably continued afterwards. In 1592 Waghenaer (compiled after 1586)[XXXIII], mentioned the tower, but in 1613 Haeyen[XXXVI] did not mention it any more; it had probably been eroded or dismantled. The strong erosion at the NW side had likely been strongly enhanced by the NE-oriented outer channel, which had newly formed.

Almost certainly the barrier island Bosch also suffered from the storm surge of 1570 (Lang, 1958). The steep (indicative of coastal erosion) high dunes on the western end of Bosch were mentioned in sailing directions of 1541[XIII] (Rogge, 1885), 1558[XVIII], and 1566[XXV] (Knudsen, 1920; (Lang, 1958), an inspection in 1556 (Formsma, 1954) and the manuscript sailing direction of 1586 (field observations 1565/75)[XXXI]. In a manuscript sailing direction for which the data were collected slightly later than for the sailing direction of 1586[40] (field observations 1565/75)[XXVI] the island Bosch with its dunes is shown as a beacon (Lang, 1958). On another, somewhat later copy of the sailing direction[XXIV], Bosch is not shown, but on the same position it is stated: "*hier heft den bos ghestaen*" (here *was* situated Bosch; Lang, 1958). Haeyen (1585, field observations 1580/84, Fig. 17)[XXX] stated: "*that between Rottum and Schiermonnikoog a small island used to be situated, called Bosck, but it has been blown away, or washed away by a storm; one can now only see a low beach*"[41]. Waghenaer probably even changed the copper engraving of his map (1584/85, field observations 1570/79)[XXIX] to remove the high dunes of the island Bosch, and showed it as a low island in his sailing direction (Lang, 1958). The destruction of the island thus must have occurred between 1565 and 1575. Since the only major storm surge during that period was the 'All Saints day'-surge of 1570, and only some minor storm surges occurred in 1566, 1568?, 1573 (Gottschalk, 1975), it is likely that the larger part of the dune erosion

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[38] The tower was probably built by the Convent of Klaarkamp as part of a granary (comparable to the Schierstins in Veenwolde, which has an identical massive refuge tower and was also built by the same cloister). It already existed in 1465 (Winkler Prins, 1867; Mellema, 1981; Feenstra, 1990).

[39] Albert Haeyen (1585, field observations 1580-1584): "*....op d'West eynde staet een Tins dicht by de strandt....*".

[40] On the map seven buoys were present instead of six.

[41] Haeyen, 1585: "*Tusschen Rottum ende het Oogh pleeght een Eylandeken te ligghen, ghenaemt de Bosck, maer het is verstoven, oft met eenen storm wegh ghespoelt, men siet nu daer op niet anders dan een slichte strant.*"

is due to the surge (cf. Lang, 1958). By comparison: a smaller surge of 3/4-1-1976 (Hoek van Holland +2.98 m DOL; Den Helder +2.97 m DOL; Harlingen, Wadden Sea +3.69 m DOL; Delfzijl +4,35 m DOL) caused local dune erosion of at maximum 35 m on Texel, 15 m on Vlieland, 20 m on Terschelling, 92 m on Ameland and an average dune erosion of 4 to 5 m on Schiermonnikoog (Van der Malde, 1976). During the second half of the 16th century storm surges occurred frequently (Gottschalk, 1975). This will have hampered the re-establishment of the dune area of Bosch, because erosion exceeded the supply of sand, and vegetation had insufficient time to re-establish. The decreasing dimensions of the ebb-tidal delta of the Lauwers Inlet in that period must also have been influential (see below in discussion). Sailing directions of the year 1588 indicate that Bosch had become somewhat higher again (Lang, 1958). Maps of Emmius (1595 in: Lang, 1958) and Blaeu (1608)[XXXV] even show dunes. According to Blaeu (1608) these small dunes had resulted from the planting of bent-grass, but the grass had already been destructed, leaving a shoal just below spring-flood level. The island Rottumeroog may also have been hit in 1570, since the most western manor house is depicted more to the W on the chart of Haeyen (1580-84)[XXX] than on the chart of Waghenaer (1570-79)[XXIX], which implies dune erosion (Lang, 1958). The dyke guards of Usquert complained in 1578 that the dimensions of the inlets increased (Acker Stratingh, 1866). This may indirectly have been caused by the storms, which helped to erode former peat areas, especially in the Dollard, thus increasing tidal volume.

In 1600 the backbarrier island Heffesant had probably disappeared as is indicated by the charts of Waghenaer (1584/85, field observations 1570/79)[XXIX]; Lang, 1958; Edelman, 1964), showing "*Heffe sant*" as a sandy intertidal shoal (on the wrong place), and those of Haeyen (1585, field observations 1580/84)[XXX] and of Waghenaer (1592)[XXXIII], which do not show it at all. After 1570 it has not been mentioned any more as an island, even not in the list of ex-cloister properties of 1594 (Reitsma, 1979). Heffesant was probably also largely destructed during the 'All Saints day'-surge of 1570.

Above, a large part of the description of the situation around 1600 has already been given (Fig. A4). The eastern end of Terschelling was still almost 11 km W of the position of 1975 (Ligtendag, 1990). The large, mainly intertidal (Winsemius, 1622) shoal "*Kampersant*" still divided the inlet into two separate branches. The exact dimensions of Ameland around that year are not very well known. Haeyen's description (1585)[XXX] suggests that Ameland was 22.2-22.6 km long and Emmius (1592) stated that the island was ± 22 km. The dimensions of the island (>22 km HW-line) were deduced by interpolating between the positions in 1550 and 1650, for both the HW- and the LW-water line.

A single-channel Pinkegat is clearly shown on the chart of Bleau (1608)[XXXV]. It ended blindly in the Wadden Sea and was positioned 1 km W of its position in 1976. On the same chart the supratidal part of the sandy shoal E of the Pinkegat had become smaller, and the shape had become round and probably high ("*T Hooge Zand*" = high-sand), instead of elongate (Lange Sandt = long sand) as in 1550 (De Haan et al., 1983). The channels of the ebb-tidal delta of the Zoutkamperlaag had most likely shifted downdrift[42]. The dunes on the western end of Schiermonnikoog were high; the eastern end consisted of a beach plain with a few isolated low dunes (Haeyen, 1585[XXX]; Blaeu, 1608[XXXV]). Not much care was taken of the dunes as follows from reports in 1598[43] (Mellema, 1981). The Knokkebalg, the channel separating Schiermonnikoog from Simens Sandt, ended blindly in the Wadden Sea in 1600. It had been reduced to a washover channel. The Lauwers Inlet formed a small entrance to the Wadden Sea with a depth of 7.1 m in the entrance and some 13 m S of Schiermonnikoog (due to eddies?) (Waghenaer, 1592[XXXIII]; Haeyen, 1613[XXXVI]). Simens Sandt more or less kept its position. The supratidal shoal which formed the remainder of Bosch had shifted to the E. The Schild Inlet may have migrated slightly to the W (cf. Lang, 1958).

Along the Frisian coast, between the hamlets of Blija and Oostmahorn, large parts of the Wadden Sea had been reclaimed in the period between 1550 and 1600. In 1592 the Anjumer and Lioessener polder were reclaimed. The remainder of the former island Bant strongly eroded at its N-side and migrated to the S. In 1600 the coastline of Friesland nearly resembled today's coastline. In the Lauwerszee also some parts were reclaimed. In 1583 a sluice was built in the centre of Dockum, thereby closing off the more western part of the sea arm (Van der Wal, 1988). In the SE part of the Lauwerszee an area was closed off. A small island in the Dokkumer Diep had been poldered in 1600. The seaway to Groningen did not change strongly in the period 1550-1600. In contrast to the Frisian coastline, the position of the Groninger coastline remained much the same with reference to 1550.

1600-1650; Fig. A5

On 23-1-1610 a major storm surge occurred, after persistent gales, during NW winds, between neap and springtide (Fig. 19). Texel and the Zuider Zee were badly hit, and dykes breached on many places in Friesland and probably also in Groningen (Gottschalk,

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[42] It should be noted that the sailing directions only mentioned the main channels. Most sailing directions explicitly warned for the quick changes in the Zoutkamperlaag Inlet. From contemporaneous observations it is clear that the outer channels migrated in the same way as they do now, i.e. to the north at the westside and to the east at the northside of the ebb-tidal delta.

[43] Groot Plakkaat en Charterboek van Friesland, 29-3-1598, in: Mellema (1981): "*Van gelycken sal myn Heeren Staeten believen, mede alzodanige goede ordre te stellen op te Waranden van de Eylanden van Schiermonickooch, ten eynde die helm op te duinen aldaer wassende van de beesten niet vertreeden en affgebeten ofte andersins uitgepluckt worden, waer deur die duinen metter tyd zolde geraecken te verderven, dewijle men bevindet dat deur die verpachtinge der Waranden van het voorschreven Eyland, quade toesicht en weynich achtinge daer op genomen word.*" (Also my Lords of State demand that order should be established on the grounds of the island of Schiermonnikoog, so that the bent-grass, which grows on the dunes there will not be trodden upon by animals and bitten off, or plucked in another way, so that the dunes in time can be ruined, because it has been found that this is badly supervised and little attention is given to it, because the grounds of the island are leased.)

1977). In the winter of 1612-1613 the island Rottumeroog and perhaps the coast of Friesland was damaged by storms (Gottschalk, 1975; Ufkes, 1989). On 8-3-1625 another major storm surge occurred, during the onset of springtide in combination with NW storm (Fig. 19). The barrier islands Texel, Vlieland, and Terschelling were damaged, as well as the coast of Friesland (Den Bilt) and Groningen (Marne area, Reitdiep; Gottschalk, 1977).

The reconstruction is given in Figure A5. Comparison of the reconstructions of the eastern end of Terschelling around 1600 (Ligtendag, 1990, basing on Sybrandt Hansen, 1602[XXXIV]) and between 1623-1650 (Beckering Vinckers, 1943; Van Oosten, 1986, probably mainly based on Blaeu, 1623[XXXVIII]) shows that the island expanded eastward over almost 3.5 km (cf. Blaeu, 1608[XXXV] and 1623[XXXVIII]). Between Terschelling and Ameland two inlets were present, separated by a broad supratidal shoal.

An excellent map of Ameland of 1665[XLV] allows a detailed reconstruction of the island at that time (Fig. 20). It does not show a HW-line at the North Sea side of the island. The inferred minimal HW-line (a rather southern position as close to the dunes as possible!) is positioned more or less at the same position as in 1536/45 (Van Deventer, 1559[XX]); some seaward progradation may have occurred in the east. The washover channel on Ameland, between Ballum and Nes, was mentioned in 1627 for the first time (Van Oosten, 1986). In the south-central part of the island the Ballumer Bocht had shifted as much as 500 m to the N since 1550 (HW-line).

No data are available on the Pinkegat and the Engelsmanplaat in that period. Based on the erosion of the eastpoint of Ameland, the Pinkegat probably extended further into the Wadden Sea as compared with the 1600 situation. Several charts indicate the existence of a just supratidal shoal ("*T Hooge Sandt*") E of the Pinkegat Inlet (that part had to be interpolated). The shape of the ebb-tidal delta of the Scholbalg Inlet had gradually changed[XXXVIII, XLII]. The eastern part of the Paardemarkt shoal had become more elongated and had migrated towards Schiermonnikoog, i.e., to the SE, and was attached to a shoal in the inlet (Roode Hooft shoal). The rest of the Paardemarkt shoal became larger at both its eastern and western side. Thus, there were two separate NE-oriented channels. The easternmost channel, which had caused so much erosion on Schiermonnikoog, was largely abandoned. Already in 1608 Blaeu warned that the NW-SE-oriented channel was too shallow to sail[44]. In 1613[XXXVI] Haeyen does not mention the channel any more and warns that the channel west of it was only some 2.3 m deep at low water. In 1643 Blaeu[XLII] did not mention the latter any more, but still depicts both channels on his chart.

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[44] Bleau, 1608 (In: Jansen, 1941): "*Tusschen Amelandt en Schiermonicooghe gaet de Schol in, men plach die van beoosten in the seylen by Schiermonicooghe langhs. Dan dat gat is binnen toegheslaghen en gantsch tot niet.*" (Between Ameland and Schiermonnikoog the Zoutkamperlaag (Scholbalg) enters. One could originally (still mentioned in 1592 by Waghenaer) sail into it along Schiermonnikoog from the East. But now the channel has been filled up inwards (in the backbarrier area) and has become insignificant).

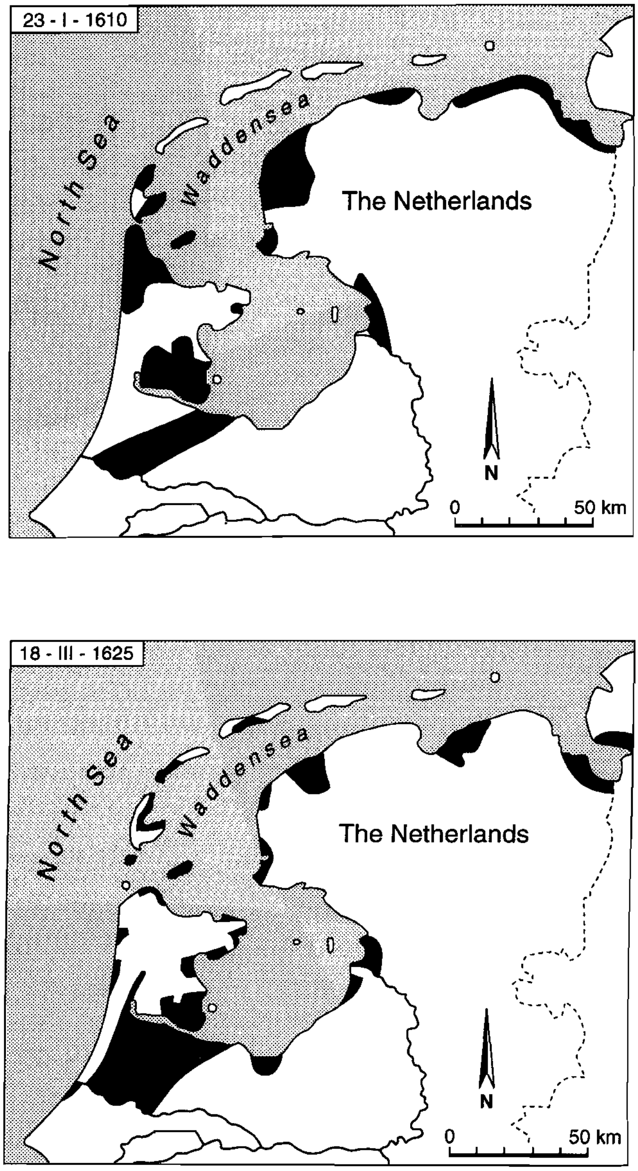

Figure 19: Areas along the Wadden Sea and Zuider Zee known to be hit by the storm surges of 1610 and 1625, given in black (redrawn after Gottschalk, 1977).

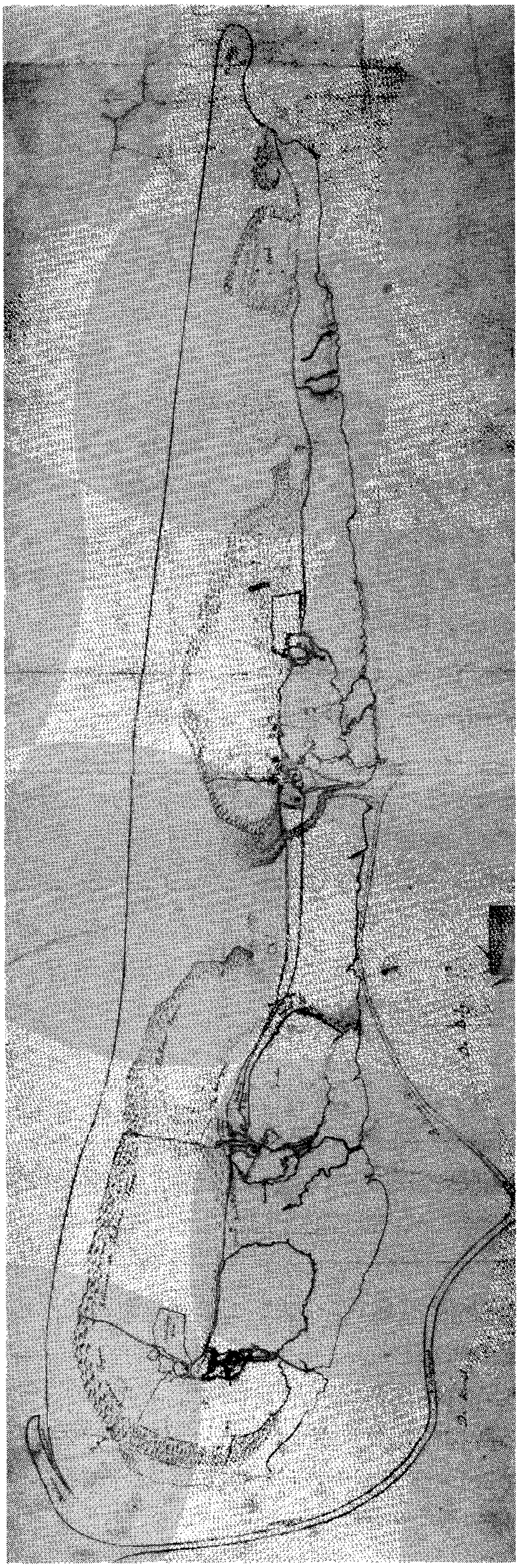

Figure 20: Ameland, circa 1665 (Anonymous)

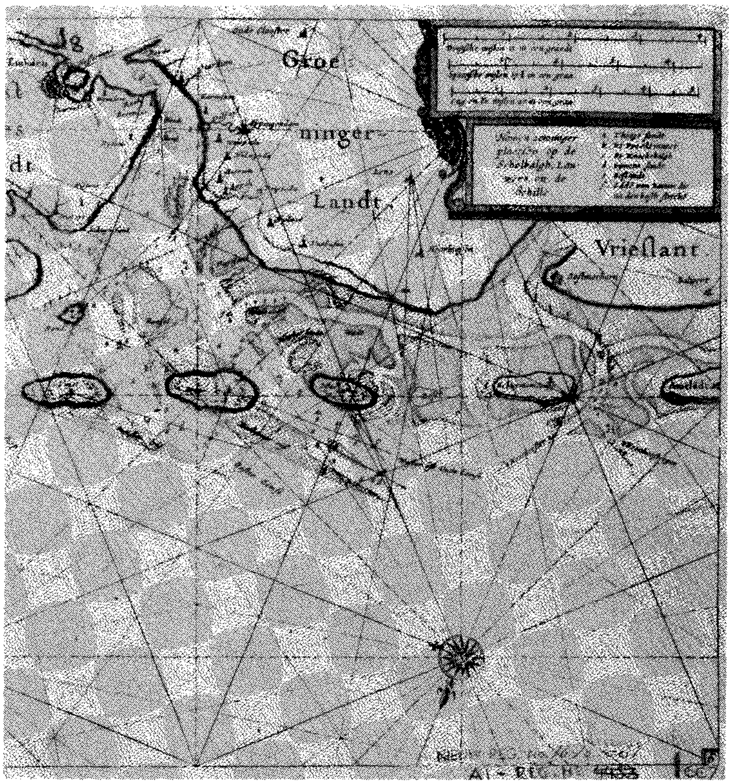

Figure 21: Willem Jansz. Blaeu (1643): "The Sea-Beacon", with changes E of the shoal Bosch after the excellent chart of Faber (1642).

At the NW side of Schiermonnikoog, the LW line continued to retreat. A map of circa 1625[XXXIX], also indicates some dune erosion at the SW side. Around 1650 the island Schiermonnikoog was some 13 km long (including Simens Sandt, which had completely attached to the island[45]. Bleau (1643)[XLII] mentioned a length of two leagues (of 20 in 1 degree = 11.1 km), which may refer to the old situation without Simens Sandt (Fig. 21). The Lauwers Inlet had shifted to the east as appeared from the precise mapping by Faber (1642[XLI]; Lang, 1968). Bleau (1643)[XLII] stated that the inlet was largely filled up with

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[45] Ufkes (1988): In a complaint in 1644, Lambert van As, on behalf of the owner of Schiermonnikoog, J. Stachouwer, stated: "....*dat voor enigen tijt een straet vaerder (was) gestrandet aen Simons sant vast zijnde aen het....eilant Schiermonnicke oge....*" (....that some time ago a merchandise ship (had) stranded on Simens Sandt, which was attached to the....island Schiermonnikoog....).

sediment, due to westward growth of the beach of Bosch, leaving only a narrow, shallow subtidal channel between Bosch and Schiermonnikoog. An almost entirely abandoned Lauwers Inlet is also visible on the chart of Faber (1642)[XLI]. Letters of 1636, 1639, 1643, and 1648, refer to Bosch as an island (Ufkes, 1988). This indicates that it was supratidal and perhaps had some dunes, although it is drawn and mentioned by Blaeu (1643)[XLII] as being a supratidal shoal[46]. The intertidal part of Bosch experienced a net grow on the eastern side. The Schild consisted of two inlets which migrated downdrift (to the E). The easternmost inlet eroded part of the intertidal shoal of Rottumeroog. The island itself may have been somewhat eroded at its western end. In 1628 it was still inhabited by a probably small population (Isbary, 1936). Much work was done to restore storm damage on the island. Repairs at both the east and west side are known from 1613, 1626, 1630-1634 (15 times!), 1635, 1636, 1637, 1640, 1643 (western dunes), and 1649 (Ufkes, 1989).

The Frisian coast remained at the same position since 1600. Bant became attached to the coast (Haacma et al., 1664)[XLIV]. In the Lauwerszee the Scholbalg shifted its course more to the east. No important changes took place in the Dokkumer Diep. In the Groninger Diep a river bend was cut off by the construction of a canal (Kooper, 1939). The position of the Groninger coastline in 1650 was largely the same as that in 1600.

1650-1700; Fig. A6

The storm surge of 5-12-1665 (not due to springtide, but to a NW storm, +3.6 m DOL near Scheveningen; Gottschalk, 1977) caused major damage on the backbarrier islands in the Zuider Zee: Marken was flooded, Wieringen was said to be ripped into two pieces (Fig. 22; Buisman, 1984). Texel and Terschelling (2.4 m above MHW of that time (Schoorl, pers. comm.) were flooded, and so was the coast of Friesland (Harlingen, Het Bildt, Holwerd, the Lauwerszee area, and Dokkum) and Groningen (Gottschalk, 1977; Buisman, 1984). On 5-11-1675 another major storm surge occurred, and caused much damage to Northern Holland and the Zuider Zee. Western Ameland seems to have lost part of its undyked land at that occasion (Fig. 22; Gottschalk, 1977). In 1686 a local storm surge (almost at neaptide and with NW wind) caused the breaching of dykes and major damage all along the Wadden Sea and the Lauwerszee coast of the province of Groningen; large areas were covered by sand and clay (Niemeijer, 1975; Gottschalk, 1977). During the same storm, Terschelling was flooded (2.1 m above MHW of that time; Schoorl, pers. comm.). On Ameland a ship even floated through the washover channel between Ballum and Nes from the North Sea into the Wadden Sea. The surge caused major damage to the island (Van Leeuwen, 1826; Isbary, 1936; Edelman, 1964).

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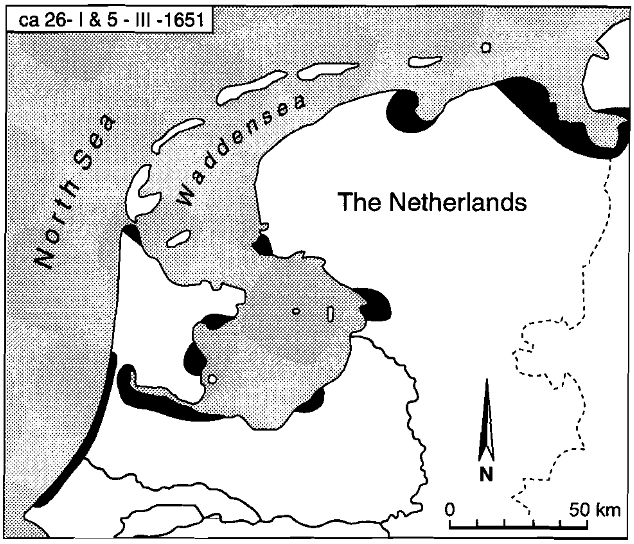
[46] That island was too small and too low to be inhabited permanently is also indicated by the fact that the newly installed island keeper of Bosch lived on the nearby island Rottumeroog in 1636 (Ufkes, 1988).


ca 26- I & 5 - III -1651
North Sea
Waddensea
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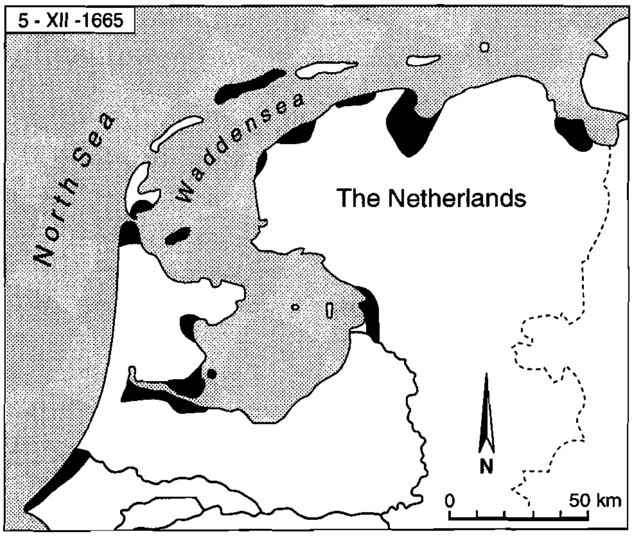
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Waddensea
The Netherlands
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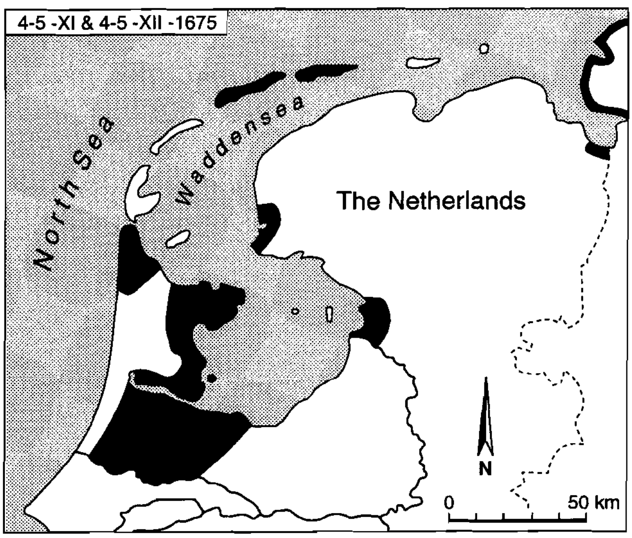
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North Sea
Waddensea
The Netherlands
N
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The reconstruction of 1700 (Fig. A6) is as follows: The eastern end of Terschelling had shifted another 4.3 km more to the E (reconstruction of Schoorl, in prep.). The inlet between Terschelling and Ameland by now more or less consisted of one main channel; the eastern inlet had become quite shallow. On the ebb-tidal delta, the main channel had oriented to the NE. By 1700 a large part of the shoal, which formerly separated the two channels, had become part of the eastern end of Terschelling, or, alternatively, the western end of Terschelling had grown in a spit-like way. Maps of 1695[XLVIII], 1708/1710[XLIX], 1745[LVI], 1762[LXI], and 1794[LXVII] show that regularly parts of shoals must have merged with the eastern end of Terschelling, whereas new shoals formed east of them.

The coastline of Ameland was interpolated between the positions of the coastlines on maps of 1665[XLV] and 1749[LVIII] (original 1731). The data show a strong westward shift of the LW-line at the western end of the island, a slight seaward shift of the dunes (and the minimal HW-line) at the North Sea side, and continued erosion at the Ballumer Bocht. The available data[XLIX] indicate that the Pinkegat Inlet had become a multiple inlet system, and that channels were present some 2-2.5 km west of the inlet of 1650. The high shoal (Engelsmanplaat, formerly "*T Hooge Sandt*") between the Pinkegat and the Zoutkamperlaag had changed in form considerably.

In the ebb-tidal delta of the Zoutkamperlaag three outer channels[XLIX] were present. The NW-oriented channel, directly N of Engelsmanplaat must have been newly formed. The channel in the shoal N of this outer channel may have been a former NW-oriented outer channel, which had shifted to the N (similar to present-day channels). In 1700 the NE-oriented outer channel had a more eastern position than in 1650. A large bar had been attached to Schiermonnikoog. This was most likely the northern part of the elongate shoal of 1650, which had separated from the Roode Hooft shoal in the main channel. Perhaps even several shoals became attached, considering the shallowness of the NE-oriented outer channel W of the elongate shoal around 1613[XXXVI]. During the attachment of the shoal(s), currents through the channel may have become strong, and the erosion of the western end of Schiermonnikoog probably continued. It was followed/accompanied by eastward dune migration (see 1700-1750), and perhaps a straightening of the dune rows.

The Lauwers had migrated some 2 km to the E, in combination with the formation of a large subtidal shoal NE of Schiermonnikoog and a strong shift of Bosch to the SE (Guitet, 1708/10[XLIX]; Lang, 1958). Its drainage area had expanded to the E and its dimensions had increased. In 1654, 1655, 1659 (Ufkes, 1988), 1695?, and 1707 (De Vries, 1936) Bosch was mentioned as an island, indicating that it was still supratidal. At the original position of Bosch, a new supratidal shoal (Koeplaat) had formed by 1707 (Lang, 1958).

Figure 22 (opposite page): Areas along the Wadden Sea and Zuider Zee known to be hit by the storm surges of 1651, 1665, and 1675, given in black (partly after Gottschalk, 1977).


Wester Eems
Ooster Eems
t Friese Gatt
Voor de Friese Gronden is Steeck Grond
Schilt
Banc
Kop Plaat
Borkomer Riff
Borkommer Banck
Juster Riff
Homme Gatt
d'Iming
Nordney
Bus
Wadde
de Meem
de Band
Norder wadden
Meeren Steen
De Ranjels
de Wester Eems
de Ooster Eems
Uythuyser Wadden
Eems Horn
de Eems Ru
Groeninger Diep
Korehause
Warffe
Uythuysen
Umpieda
Spik
Nye landt
Griet Zyll
Pilsem
Marienho
Tyschend
Pesens
Wierom
Oostmer horn
Kolwert
Fernaert
Anniam

The reconstruction, based on interpolation of the maps of 1650 and 1750, the chart of Guitet[XLIX] (Fig. 23), and the reconstructions by Lang (1958) suggest that the Schild Inlet, as well as its main backbarrier channels, and the island Rottumeroog had shifted considerably to the E in the period 1650-1700. Indeed, already in 1652 it was stated that the dunes of Rottumeroog were high and steep at the westside (Reenders, 1986), an indication of (dune) erosion.

Southwest of Schiermonnikoog a large subtidal channel (De Noorman) was present. It extended further than the easternmost dunes of Schiermonnikoog. In the 18th century it would influence the development of Schiermonnikoog considerably. The Frisian coast had remained largely at the same position as in 1650; the remainder of the island Bant was reclaimed slightly later[LI]. Along the coast of the Lauwerszee small parts of land were reclaimed[L]. The coast of Groningen remained nearly the same, but in front of it a large summer dyke was constructed. A part of the area between Vierhuizen and Zoutkamp was reclaimed.

1700-1750; Fig. A7

On Christmas 1717, a catastrophic storm surge ("Kerstvloed") took place. Again, like in 1570, the wind turned from SW to NW, this time during the latest quarter of the moon (Buisman, 1984). Dunes and dykes along large parts of the coast of Holland and the Zuider Zee were damaged, and several areas were flooded (Burger, 1967; Schoorl, 1973; Buisman, 1984). Especially in the province Groningen, but also in Friesland (among others 't Nieuwe Bildt, Ferwerderadeel, Dongeradelen, Kollumerland) dykes were breached or they were flooded (the water level being up to 1.7 m above the dykes) on many places. The sea flooded the mainland over large distances (Fig. 24), leaving a sediment cover after it retreated (Anonymous, 1717, in: De Leeuwarder Courant, 1967). Several places (Kommerzijl, Morra, Lioessens, Engwierum, and Anjum) were partly damaged, but Holwert and Ferwerd were totally destructed (Van Leeuwen, 1826, in: Van der Molen, 1967). After the disaster, the seaward summerdyke of Groningen was changed into a new sea dyke (Van de Ven, 1993). It was also decided that the Dokkumerdiep would be closed off. This happened between 1725 and 1729 (RWS, 1948; Buisman, 1984).

Figure 23 (opposite page): Mathurin Guitet, 1708/10[XLIX], Amsterdam: "Wad en Buytenkaart...", showing the area from Ameland to Germany.

Texel, Vlieland[47], Terschelling[48], Ameland[49], Schiermonnikoog[50], and Rottumeroog[51] were flooded, and strong dune erosion occurred (Buisman, 1984).

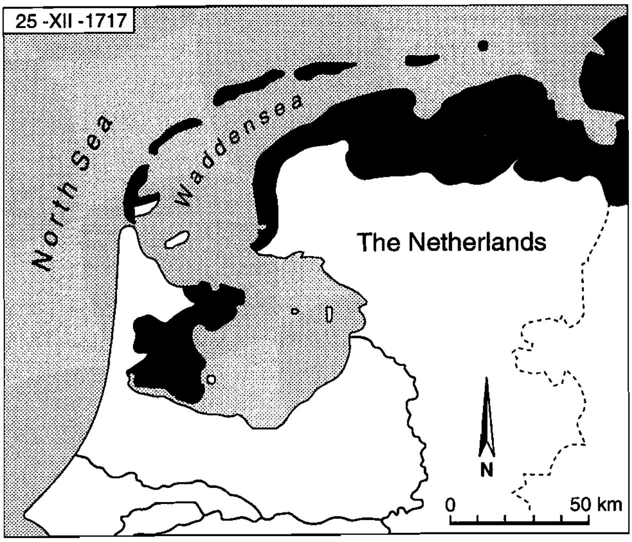

Figure 24: Areas along the Wadden Sea and Zuider Zee known to have been hit by the storm surge of 1717, given in black (Vredenberg-Alink, 1974).

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[47] Anonymous (C.J.), 1717 (in: De Leeuwarder Courant, 1967): "*De Noordzee is op Vlie-land over de Duynen geloopen zoo dat er ook veel schade is geschied....*" (On Vlieland the North Sea has flooded over the dunes, so that much damage has been caused....). Also, the dykes on the island were breached (Anonymous, 1946).

[48] Gerhardus Outhof, 1718 (in: Niemeijer, 1975): "*En op 't eiland der Schelling quamen de wateren aan twee kanten, van 't Noorden en 't Zuiden, ter dijken ingebroken....*" (And on the island Terschelling the waters came at two sides from the north and the south breaking through the dykes....).

[49] Niemeyer, 1975 (source unknown): "*Van 't Eiland Amelandt wierde ook berigt dat het water aldaar ook verscheide dyken verbrak! zeer hoog quam, de menschen op de zolders deede vlugten....*" (From the island Ameland it was reported that also there the water broke several dykes! became very high, forcing the people to fly to the attics....).

[50] Winkler Prins, 1867: "*De kersvloed van 1717 en de vloed van nieuwjaar 1720 hadden een aanmerkelijk gedeelte van de westelijkste duinen verzwolgen.*" (The Christmas flood of 1717 and the flood of new year 1720 had swallowed a considerable part of the westernmost dunes).

[51] The storm surge of 1717 caused such devastation that the inhabitants were forced to move to the mainland (Isbary, 1936; Edelman, 1964).

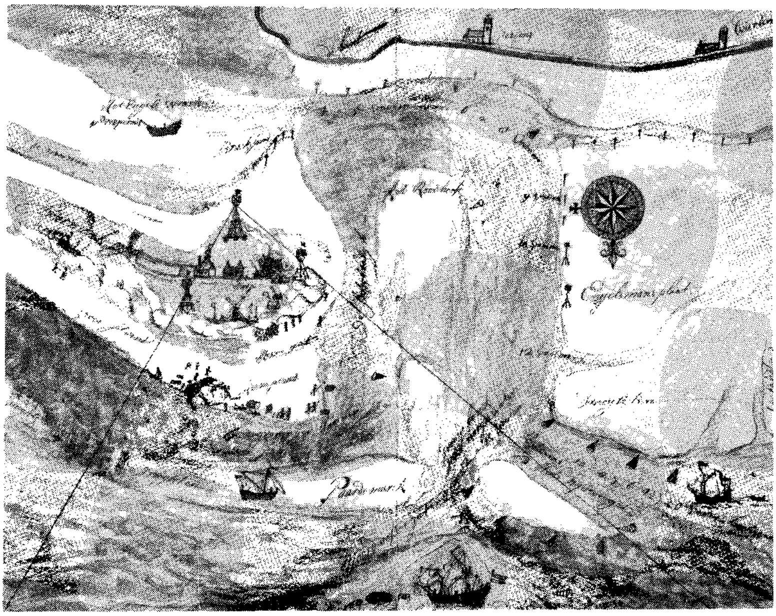

Figure 25: Anonymous, probably P.W. Donama (1730-1735?): "The pirate map", showing the old (right side of the island) and the new village (left side) on Schiermonnikoog. Clearly visible is also a large swash bar attached onto the beach (where a ship is wrecked N of the old village).

Schiermonnikoog was strongly eroded in this period. First at the WNW side, due to eastward migrating dunes in combination with tidal and storm surge erosion (storm surges of 1717 and 1720; Winkler Prins, 1867). The eastward erosion of the dune foot was estimated "*200 roede*" (630, 720 or 780 m; Verhoeff, 1983), probably in the period 1690-1760[52]. People

[52] Hayes, 1762, in: Winkler Prins (1867): "*Nu zoude ik u dan bekend maaken die tyd, wanneer ik op het eiland gekomen ben, dat was woensdag den 27sten January 1762, en den 3^{den} February hebben Pieter Jans, oude Dros (drost of drossaart, vertegenwoordiger van den Heer) en ik het eyland bezien en hy heeft my alles bekend gemaakt, hoe dat 't eyland voor dien tyd geweest is. Men schryft hem oude Dros, dat geeft zoo veel te kennen, dat hy lange jaren het eyland regeeren en bestieren moest voor die Heer, omdat die doe nog klein was. Hy is een oud man byna 77 jaar, hy hadde het eyland wel in zyn fleur gekend, daarom konde ik nooit beter onderrigtinge krygen, dan ik nu heb gekregen. Doe gingen wy eerst naar de oude kerk, die was afgebroken. Die konde daer niet langer staen om het stuiven van de duinen. Het fondament van de westend van de kerk heb ik doe nog gezien, die was omtrent 5 voet binnen de wal, zoo na op de strand, en het fondament van de pastorye heb ik ook gezien, die was digt by de kerk op de kant van de strant. Daar was doe een schip in de zee, doe zeide de oude man tegen my, dat het vaste lant met hooge duinen nog al verder hadde geweest als dat schip was, doe zeide ik tegen hem, dat zal wel 200 roede weezen, en hy antwoordde my van ja dat zal wel weezen. Gy moet my wel verstaan, het is nu deze oude man zyn oordeel, hy heeft het niet gemeten. Dit land is hier aan de westkant weggespoeld.*

Daar de kerk gestaan heeft, daar is nu een groote duin, en daerbij om zijn nog meer duinen. Als men op het strand staat, dan is 't, of men bij een muur opziet. Als het water hoog floeit, dan spoelt het onder weg, daerom staen de duinen zoo regt op en kan gy het wel denken, als de Heere het niet verhoet, dat het eylant dan nog in een groot gevaer is, want als het zand droog wort, dan loopt het na beneden, en als het water dan weer hoog vloeit, dan spoelt het onder al weer weg. De dooden hebben zy lange jaaren op dat kerkof begraven, die staen zoo hoog op malkander, omdat het zand daer zoo met der tyd al hooger by op stuifde, maer nu begraven zy de dooden daer beneden op de vlakte. Doe zyn wy daervan af gegaen naer de plaats, waer de tweede kerk gestaan heeft en onderweegs onderrigtte hy my van datgeene dat ik schryven zoude. Deze tweede kerk is gezet in het jaar 1717, die is een end van de oude af gezet in het zuidwesten na myn oordeel en op het vlakke veld, daer waer doe het beste land. Daer zagen zy doe geen gevaer om het stuven van de duinen, en om zoo weggespoeld te worden, daer waren de minste gedachten niet om. Doe hebben zy daer zoolang veilig gewoond aen het jaer 1737 toe. Doe hebben zy het voor de eerste keer vernoomen. Dit is nog verhaalt van Fokke Thomas, veerman op Dokkum, dat daer had een schipper geweest, die was beschonken en hy was niet in staat om zyn schip te regeeren en evenwel trok hy het zeil op en doe zeilde hy op zyn eigen anker; doe kreeg hy een gat in zyn schip, doe dreef hy op de strand en doe zonk hy naer beneden. Hy was met balvlinten gelaaden, doe begon het rondom het schip weg te scheuren, dat is het eerste begin geweest daer zy vernomen hebben en van die tijd af aan is 't eyland al trekkende verloren de eene tijd wat meer en de andere tijd wat minder. Het is wel geweest, dat zy het niet konden zien dat 't verloor, maar den 7 September 1756, doe die harde storm was, doe heeft het ongemeen veel verloren en doe zyn daer huizen omgewaaid en schepen met volk weggescheurd, die zy nooit weer gezien hebben, en van dien tijd af aan het het ongemeen veel verloren, maer den eenen tyd wat meer als den ander, maer byzonder in het jaer 1760 de tweede kersdag, doe is de kerk omgespoeld, zoo hoog vloeide het water, en verscheiden huizen. De banken en de predikstoel hadden zij een dag of drie vooraf uit de kerk gehaald en de doden uit de graven. Elk borg doe zyn eigen doden. Het is daer een gebruik als daer een sterft, dat haer naam op de kiste gezet wordt, en de andere dooden die spoelden van het kerkhof af in de zee; de kisten, die zy weêr krygen konden, die hebben zy aangehaald en op het land weer begraven, en van die ryke lieden, die in de kerk begraaven waren, die hebben zy by die plaats gebragt, daer de nieuwe kerk staan zal. Als die gemaakt is, dan zullen zij daer weer begraven worden; zij staan nu zoo lang onder een afdak die daertoe gemaakt is. Wat was daar doe een naare tijd. De menschen die moesten vlugten uit hare huizen en zy wysten niet, waar da zy vlugten zouden, en dat in den winter in koude en harde wind en zy moesten schielijk hare huizen afbreken of zy spoelden weg, en vele arme menschen, die niet betalen konden om haar huis af te breken en dan weer te zetten; daer zyn wel menschen die haer huis verkocht hebben aan een ander. Die het dan afgebroken hebben, die daar een huis wilden zetten, die moest eerst 15 gulden geven en dan alle jaren 2 gulden grondpacht en dan hebben zy niet meer grond als daar een huis op staan kan. Als zy zoo veel grond hebben willen, dat zij daar schol op droogen kunnen dan moet zij eerst 3 gulden geven. Ik heb by verscheiden menschen geweest die het my zelve gezegd hebben, dat zy zoo lang in haar huis gebleven waren, dat het eene end van haar huis al omviel en dat het water het geheele huis al langs spoelde, en dat het zand tegen de onderdeur aanspoelde en dat zy die niet konden open krygen, dat zy moesten over de deur heen. Het water dat vloeide zoo hoog op dat, 't is my verhaald, dat het op het dak van een huis geweest is, maer het was een lage muur, maer met het spatten ging het over de huizen heen. De schelvissen die spoelden op 't land. Tjeerd Jans heeft het my zelve gezegd, dat hy met zyn beiden 20 gevonden heeft. Men kan het zoo niet beschryven of de een of den ander zal hier nog wel wat byvoegen kunnen van diegene, die daer op 't land woonen, maar ik beschrief het u anders al redelyk klaar. Die huizen zyn daer allemaal afgebroken en weggespoeld, daer staan nog wat buiten af. Het fondament van deze kerk heb ik ook gezien, dat was op de strand, en de regenwatersbak van de pastorie heb ik ook gezien, die was digte by de kerk op de strand. De oude *man zeide tegen my, daar zoude wel een klein uur gaans vast land weggespoeld wezen, daer nu de schepen al zeilen, en aan de buitenkant hebben twee regels groote duinen geweest, daer hadde hy meenig mael heen geweest om konynen te vangen. Het meeste land is daer weggescheurd in het zuidwesten van de kerk, maer nu hebben zy die huizen weer gebouwd, maer allegaar niet. Die zijn gezet een klein half uur daervan af in het noordoosten. Daer stonden ook een groote menigte van huizen, die staan principaal in twee regels oost en west naer myn oordeel, en daer is nu nog een lange regel bygezet van 70 huizen op een regel. In het jaer 1756 is daer één opgezet, dat was de eerste, en 1757 zijn daer 6 opgezet en 1758 eene groote menigte, want zy braken hare huizen af, zoo trekkende al af, omdat zij daer vooraf al begonnen te vreezen, dat het zoo gaan zoude. Vier weken voor Kersdag, daer het alles zoo weggespoeld is, doe was daer nog 77 treden vast land van de kerk af; het is my van den schoolmeester verhaald, die hadde het land zelve getreeden, en nu staat het fondament van de kerk een end op de strand. En in het jaer 1759 en 1760 en 1761 doe zyn daer ook vele huizen gezet, doch van het jaer 1756 af zijn daer 94 gezet al ik wel geteld heb, en die tyd wanneer dat die gezet zyn dat heeft Marten Lammerts, baas timmerman my gezegd, en die huizen zyn byna allegaer met twee opstaande gevels en twee kamers, dat heb ik met myn eigen oogen gezien, want ik ging het eene huis uit en andere weer in met boeken te verkoopen. Zij beginnen met haar huizen al wat in orde te komen, maer die kerk is nog niet gebouwd; die grond hebben zy al begonnen te bereiden, die zal midden in de buuren staan, maer ik weet nog niet de tyd, wanneer zy beginnen zullen te bouwen met de kerk, dat is de 3de kerk al die deze oude man op het eyland gekend heeft. Zy hebben een mooi groot huis daer zy haer godsdienst in verrigten. Het eyland dat zal nog omtrent 3 uur gaans lang wezen met de strand en omtrent een klein uur breed. Ik heb verscheiden menschen gevraagd, en die hebben het my zoo berigt en op die getuigenisse maak ik staat op."*

(Now I will make you acquainted with that time, when I came on the island, that was wednesday, January the 27th (of) 1762, and on the 3rd february Pieter Jans, old Dros (drost or drossaart, the island-keeper for the Owner) and I saw the island, and he told me everything, how the island was originally. One calls him 'the old Dros', which means as much that he during many years has ruled and governed the island for the Owner, because he was still (too) young (the Owner was Johan Willem Stachouwer IV, 1722-1784. Was Pieter Jans the same person as P.W. Donama, who drew several maps of the island and who was 'Drossaart' in 1736 (Winkler Prins, 1867), as suggested by Reitsma (pers. comm.)?). He is an old man, almost 77 years, and knew the island when he was young, so that I would never have got more accurate information, than I have now.

First, we then went to the old (first) church, which has been dismantled. It could stand there no longer, because of the shifting sand from the dunes. I then also saw the foundation of the church, which was about 5 feet within the wall, so near to the beach, and the foundation of the vicarage I saw also, which was near to the church at the side of the beach. Then there was a ship at sea, and the old man said to me, that the land with high dunes had been further than the ship was, and I said to him, that will be some 200 roede (630, 720 or 780 meter; Verhoeff, 1983), and he answered me: "yes that might well be". You have to understand, it was the judgement of the old man, he had not measured it. The land has here been washed away on the west side. Where the church once stood, there is now a large dune, and around it are more dunes. If one stands on the beach, it is as if one is looking up a wall. The dunes stand so steep, because the lower part is eroded, if the water flows high and one can well imagine, if the Lord doesn't prevent it, that the island is still in much danger, because if the sand becomes dry, it flows down, and as the water becomes high again, it is all washed away at the lower part. They have buried the dead for many years in that churchyard, and they (the dead) stand so high on top of each other, because the sand was blown higher and higher in time, but now they bury their dead down below on the plain.

Then we went from there to the place where the second church had stood, and on the way he taught me what to write. The second church was built in the year 1717, and was placed at some distance from the old one to the southwest, as far as I could judge, and on the plain, where the best land was then. There they saw no risk from shifting dunes, and nobody thought about being washed away. There they have lived safely until 1737. Then they experienced it for the first time.

This has been told by Fokke Thomas, ferryman on Dokkum, that there had been a skipper, who had been drunk and unable to sail his ship. Nevertheless he raised (the) sail and sailed on his own anchor. Then he got a hole in his ship, floated to the beach and (then) sunk. He was loaded with boulders (used for dykings), and then erosion started around the ship. That was the first time they had noticed, and from that time on the island has lost by and by, sometimes more, sometimes less. Sometimes they could see no loss, but on 7 september 1756, when that hard storm was, it has lost much and at that time houses were blown down and ships with people were ripped away, which they have never seen again. From then on very much has been lost, but at one moment more, than at the other moment, but especially in the year 1760 at the second Christmas day, then the water flooded so high, that the church and several houses were destroyed (by the water). They had removed the benches and the pulpit some three days earlier and had taken the dead from the graves. Everybody rescued his own dead. It is a practice there that, if somebody dies, their name is put upon the coffin. And the other dead were washed from the churchyard into the sea; the coffins they could catch were gathered in and burried once again on the land. And those (bodies) of rich people, which were burried within the church, they have transported them to that place where the new church will be. When it will have been built, then they will be buried there again; in the mean time they stand below a shed which was made for that purpose.

What a terrible time that was. People had to flee from their houses and they did not know where to flee, in the winter in a cold and hard wind. And they had to take apart their houses quickly or they were washed away. And many poor, who could not afford to pull down their houses and rebuild them; there have been people who sold their house to another. Those who had pulled it down, and who wanted to build a house there, should first pay 15 guilders and each year 2 guilders for the lease of the ground, and they had no more ground than needed for a house, If they wanted enough ground to dry plaice they firstly had to give 3 guilders. I have visited several people who said to me that they stayed so long in their houses, that one end of their house was already falling apart and that the water was washing along the whole house, and that sand was washed against the lower hatch and that they could not open it so that they had to go over the door. The water flowed so high, I have been told, that the water came onto a roof of a house, but it was (had) a low wall, but the splashes went over the houses. The haddocks were washed onto the land. Tjeerd Jans himself has said so to me, that he altogether found 20 of them. One cannot describe it in such a way that one or the other can add something to this, from those who live on the land, but I have described it to you fairly clear. Those houses have all been broken down and washed away, (but) there are still several standing at the outskirts. The foundation of the (second) church I have also seen, that was (lying) on the beach, and the rain-water tank of the vicarage I have also seen, that was near to the church on the beach. The old man said to me, there should be washed away nearly a small hours walk (less than 5 km (Verhoeff, 1983); a strong over-estimate, since this would place the western end of the island on Engelsmanplaat. Based on the available maps of that time a minimal estimate was made of the position of the western end) of solid ground, where now the ships were sailing. And at the outer end there had been two rows of large dunes, where he went many times to catch rabbits. Most of the land SW of the church has been washed away, but now they have rebuilt the houses, but not all of them. They have been placed at a small half hour walk (less than 2.5 km; Verhoeff, 1983) from there to the northeast. There stood also a large number of houses in two rows east and west, as far as I could judge, and now they have placed another 70 houses on a row (Langestreek). In the year 1756 one was built, that was the first one, and 1757 6 were built and in 1758 a large row, because they took their houses apart, and thus left the place, because they feared beforehand that it would go like that. Four weeks had to give up their houses west of the old church[53]. The large part presumably moved to the present-day village (Fig. 25). Feenstra (1990) reports a considerable growth of the population in the years 1721 and 1722. The old church had to be given up in 1715 (Winkler Prins, 1867). The remnants of it were seen by Hayes in 1762[52], partly covered by dunes[53] and near to the beach, whereas the foundation of the vicarage was on the beach. Strong aeolian sedimentation at the nearby graveyard, east of the church, enabled the natives to bury their death atop of each other (Hayes, 1762[52]). On the map of 1769[LXIV] it is called "*Het Oude Kerkhof*" (the old graveyard). It is situated in the dunes, west of the original hamlet Oosterburen (eastern neighbours).

A new church was built in 1717, somewhat SW of the old one, (Hayes, 1762)[54]. This is in close agreement with the map of 1769[LXIV]. The church was standing in, or N of the hamlet 'Westerburen' (W of the manor house of the island owner). Westerburen and the hamlet 'de Dompen' were gradually destroyed by erosion by the channel De Noorman in the period 1737-1761 (Hayes, 1762)[52]. The channel gradually migrated towards the island. It thus formed the shortest, most direct connection between the backbarrier area and an inlet channel, which migrated downdrift. Eastward erosion at the SW side was at least 2 km in the period 1650-1800, mainly in the period 1690-1762. The inhabitants moved (after the fourties of that century, especially after 1758; Hayes, 1762; Mellema, 1981; cf. Feenstra, 1990) to the more eastern village, at present called Schiermonnikoog (called Oosterburen in

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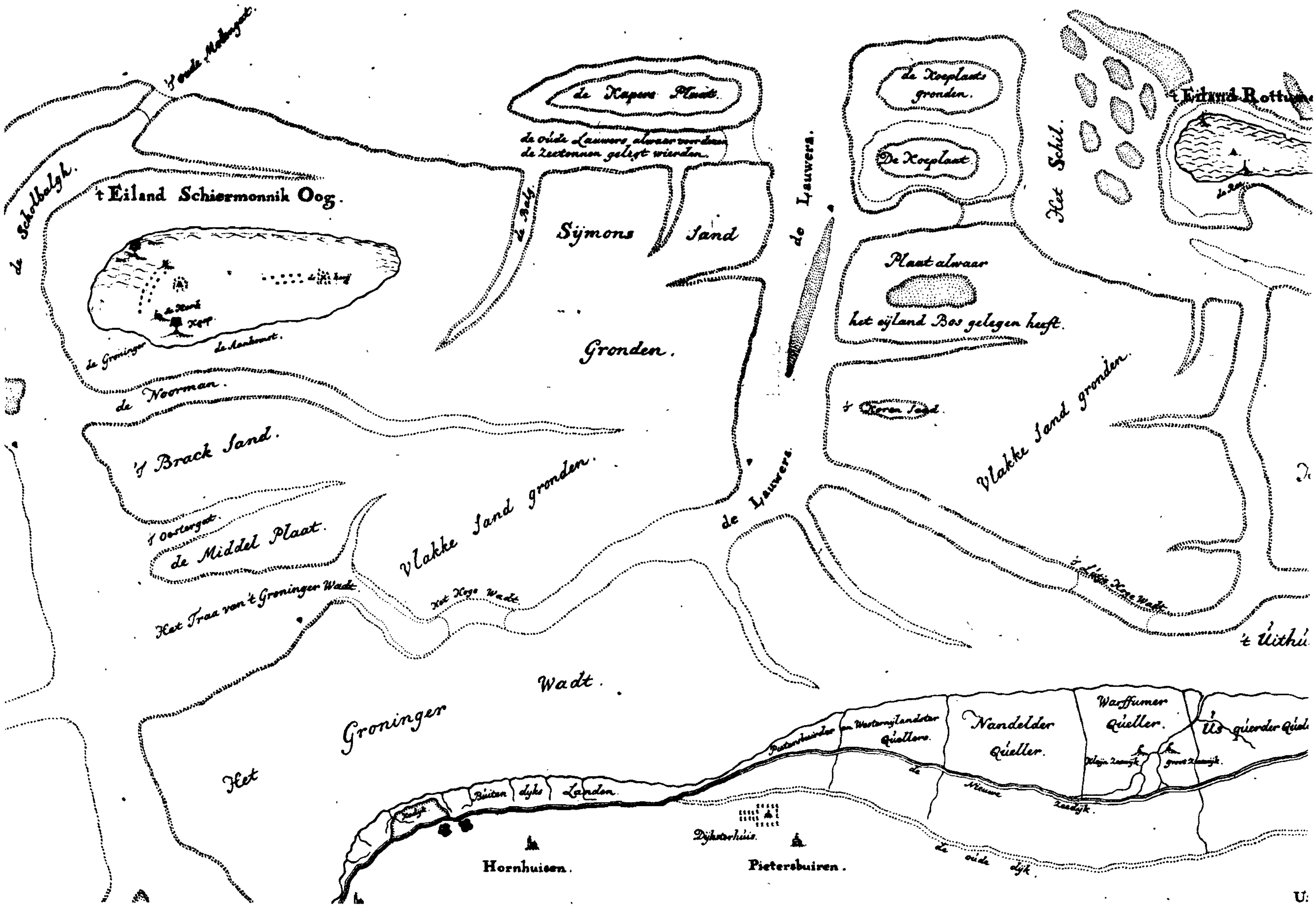
before Christmas day, when all was washed away there, there was still 77 treads (most likely 77 treads of 2.5 voet of 29.2 cm = c. 56 m) solid ground from the church on; this has been told to me by the schoolmaster, who himself had measured off the land, and nowadays the foundation of the church stands at some distance away on the beach.

And in the year(s) 1759 and 1760 and 1761 many houses were placed there, but from the year 1756 onwards 94 houses were placed there, if I have counted well. And the time when those were placed was told to me by Marten Lammerts, the master carpenter. And the houses are almost all with two raised fronts and two rooms, that I have seen with my own eyes, because I went from one house to another to sell books. They have their houses already somewhat in good order, but the church has not yet been built. They have already started to prepare the ground. It will stand in the middle of the neighbourhoods, but I do not know the time that they will start to built the church. That is the 3rd church that the old man has known on the island. They have a beautiful large house where they practice their religion. The island will be some 3 hours walk (c. 15 km) long including the beach and about a small hour broad (less than 5 km). I have asked several people, and they have told me so, and on their accounts I rely.)

[53] P. Gerlofs & W. Fokkes, 1713 (R.A.F., Heerlijkheid Stachouwer, no. 158/59, in: Mellema (1981): "*....de tegenwoordige ouwde kerk af te breken, ..., omdat zij op een ongleegene plaats en onder Duins staat, waar wij 't vreesen als dat zij met een korten gelijk veel ander Huysen is geschiedt onder de duynen sal bestolpt worden....*" (....dismantle the present-day old church, ..., because she stands on an inconvenient place and near to the dunes, so that we fear that in a short time, like happened to many other houses, she will be covered by dunes....).

[54] Not to the SE as many will have it!


't Eiland Schiermonnik Oog.
de Schol balgh.
'T oude Molengat.
de Kapers Plaat.
de oude Lauwers alwaar voordeen de Zeetonnen gelegt wierden.
de Balg.
Sijmons Sand
Gronden.
de Lauwers.
de Koeplaats gronden.
De Koeplaat.
Het Schil.
't Eiland Rottum
Plaat alwaar het eijland Bos gelegen heeft.
't Koren Sand.
Vlakke Sand gronden.
de Groninger
de Noorman.
't Brack Sand.
't Oostergat.
de Middel Plaat.
Vlakke Sand gronden.
Het Hoge Wadt
Het Traa van 't Groninger Wadt.
de Lauwers
Wadt.
Het Groninger
Buiten dyks Landen.
Westernijlandster Quellers.
Nandelder Queller.
Warffumer Queller.
Us querder Queller
Kleijn Zeesijl
groot Zeesijl
de Nieuwe Zeedijk.
Dijkstershuis.
de oude dijk.
Hornhuisen.
Pietersbuiren.
't Uithu

the 19th-early 20th century, and in the 18th century Hoogdorp). Erosion at the western side was not directly compensated by sedimentation at the eastern end of the island.

At the position of the large shoals at the eastern end of Schiermonnikoog, depicted by Guitet in 1708/1710[XLIX], a new shoal Simonszand, located E of the original Simens Sandt, developed. The map of 1745[LVII] depicts the new Simonszand, as an intertidal area (Fig. 26). However, in 1744 Catharina Stachouwer of Schiermonnikoog stated that Simonszand (or Simens Sandt) was fixed to the island Schiermonnikoog, and was therefore a part of her island[55]. Judging from the reconstructions before and after 1750, from the length of Schiermonnikoog mentioned in 1762 by Hayes (with beaches '*about a three hours walk long*' = circa 15-16.7 km, including the new Simonszand), and the partial sales of Simonszand in 1707 (De Vries, 1936), it is concluded that the area was supratidal. It was separated from Schiermonnikoog by the 'Balg', a small inlet/large washover channel. North of Simonszand the large supratidal shoal "*Kapersplaat*" was situated.

East of the Simonszand the Lauwers Inlet was situated. According to the map of 1745[LVII] the Lauwers Inlet had shifted almost 3 km eastward, thereby increasing its drainage area considerably. Directly east of the inlet, the shoal Koeplaat was located. The island Bosch, now located in the backbarrier area, probably got its final blow during the storm surge of 1717 (cf. Reitsma, 1991a&b). In 1707 it was still mentioned, but around 1719 it was stated that the island Bosch had been washed away totally[56] (De Vries, 1936).

Figure 26 (opposite page): A.T.B. (probably Ass. Theodorus Beckeringh, possibly A.T. Bontekoe (Vredenberg-Alink, 1974)), 1745: Map of the coast of the province Groningen from Zoutkamp to Enkhuizen and the adjacent Wadden Sea. N for the Wadden is to the top left. Note the text: "*Plaat alwaar het eijland Bos gelegen heeft*" ('*Shoal where the island Bosch was (originally) situated*').

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[55] Maria Catharina Stachouwer, c. 1744 (in: Mellema, 1981): "*Corenheff is helemaal niet meer te vinden; en wat het Symonsant aenbelangt, dit is vast aen der Supplianten strand alwaer de ingesetenen konnen met paard en wagen henerijden, en alsoo een gedeelte van het eyland Schier; het eyland Busch is niet meer bekent waer gelegen of geweest is en de Koeplaat is een vrije zeeplaat....*" (Corenheff is not to be found any more; and as far as Simonszand (or Simens Sandt) is concerned, this is fixed to the beach of the petitioner (Catharina herself) and to be reached by natives by horse and cart, and is therefore part of the isle of Schiermonnikoog; of the island Bosch it is not known any more where is has been located and the Koeplaat is a free sea shoal (supratidal).

[56] Spanheim Staat Boek, c. 1719, in: De Vries (1936): "*T Eylandt Bosch genaamt.... is te eenemaal weg gespoelt*". Minor storm surges occurred in 1714 and 1715 (Buisman, 1984).

In 1744 the States of Groningen had an argument with the owner of Schiermonnikoog, Catharina Stachouwer, because people of her island had illegally obtained goods from the 'Anna'. It had stranded on the Koeplaat, belonging to Groningen. Letters concerning the argument (Mellema, 1981)[55] show that, by that time, Corenheff (probably the remnants of a part of Cornasant) and the island Bosch had disappeared. This is in agreement with the original MS map of Beckeringh (1745[LVII]). It shows that the island Bosch had turned into an intertidal or just supratidal shoal (Fig. 26).

Erosion also occurred on Rottumeroog (e.g., in 1717), where a large shoal, migrating eastward together with the more western Schild Inlet approached the island. The channel E of the shoal eroded the island, since at least 1707 (Isbary, 1936). By 1730 it was found that a large part of the island had been eroded, and the dunes were said to be so badly damaged that it was feared that the island might be lost soon (Schortinghuis, 1975; Reenders, 1986). A new storm surge in 1738 caused such strong erosion at the western end that the island was bought back by Groningen for 40% of its 1659 value (Isbary, 1936; Schortinghuis, 1975; Reenders, 1986). Although bent grass was planted from 1741 onwards by the island keeper, the island continued to migrate to the E (Isbary, 1936). The strong erosion is clearly visible on the maps of Beckeringh (1745[LVII], 1781[LXV]), showing a large area at the western side of the island, consisting of isolated patches of sand; the remnants of the western side.

Above, a large part of the description of the situation around 1750 (Fig. A7) has already been given. The eastern end of Terschelling had probably not shifted strongly in the period 1700-1750 (mainly reconstruction Ligtendag, 1990). Around 1723 the NE-oriented main connection to sea of the Ameland Inlet was displaced strongly to the east and had become quite shallow. A new NW-oriented connection had formed. The new channel was deeper than the NE-oriented one, and the skippers of Harlingen advised to put buoys in it[57] (Beckeringh Vinckers, 1943). A map of 1745[LVI] shows that this was indeed done. The LW-line of the western end of Ameland had migrated to the E. The HW-line at the western end may have migrated slightly to the W: The map of la Rive (1731; copy 1749)[LVIII] indicates some dunes W of the old ones. They were likely newly formed, because they lack the characteristic straight seaward front of eroded dunes (cf. Isbary, 1936). The situation of 1731, depicted on the reconstruction of 1750, changed quickly in the period until 1800. Soon after 1731, the last houses of Sier at the W side of Ameland were abandoned, because the hamlet had gradually been covered by eastward migrating dunes (Bakker, 1970). The length of the island was about 21.5 km. The position of the LW-line at the North Sea side was about the same as in 1665[XLV]. Erosion of the LW line occurred at the eastern end of Ameland, due to the erosion of the westernmost channel of the Pinkegat Inlet System, which had several inlets[LXI]. The Engelsmanplaat was broad and cut by several abandoned(?) channels.

[57] Skippers guild of Harlingen, 1723, R.A. Leeuwarden, register van resoluties en placaten van H.E. Mog. H. Staten van Friesland (In: Beckering Vinckers, 1943).

The Frisian coast did not change over the last period. In the Lauwerszee extensive tidal flats were present. The channels in the Dokkumer Diep and the Groninger Diep became in increasingly narrower as large parts of the Lauwerszee were reclaimed (Fig. 27). After the disastrous storm surge of 1717 the summer dyke along the coast of Groningen was changed into the new dyke.

1750-1800; Fig. A8

During the storm surge of 1775 a large part of Friesland near the Zuider Zee was flooded. The inundation of Friesland in 1776 was even more extensive (Buisman, 1984). During that latter storm, Texel was flooded, and also other Wadden Sea barrier islands were badly damaged (Buisman, 1984).

Around 1800 several accurate maps have been made[LXVIII, LXIX, LXXI, LXXII, LXXIII, LXXIV, LXXV]. They allow a detailed reconstruction (Fig. A8). In 1800 Terschelling was only 1 km W of the position of 1975. The easternmost part consisted of a many km long supratidal beach plain (Athallin, 1811; Fig. 28). Hydrographical maps[LXVIII, LXXIV] of the Ameland Inlet show that, by that time, there was only one inlet present, 1.5 km W of its position of 1975. A map of circa 1762[LXI] shows that the NE-oriented channel in the ebb-tidal delta had become quite shallow. At that time there were two NW-oriented channels. The southernmost was the deepest. The shallower one may either be the original NW-oriented channel of 1745 or a new channel. In 1781 the NE-oriented channel is not mentioned any more and had probably (almost) disappeared (C. du Moulin, 1781[58], in: Jansen, 1941; Beckering Vinckers, 1943). In 1781 the main connection was again oriented to the NW. This is also shown on the map of Buyskes (1798)[LXVIII].

To the W, the HW-line of Ameland had migrated westward at the SW side of the island, with reference to 1731 (copy 1749[LVIII]). The dunes had shifted eastward. The HW-line at the northern side of the island is also depicted on the map of 1809[LXXIII]. Its position was further north than the position of the inferred minimal HW-lines (as close to the dune as possible) of 1665[XLV] and 1731[LVIII]. Comparison of the latter two maps with earlier maps (Van Deventer, 1559)[XX] and later maps suggests an ongoing slight seaward progradation of the inferred minimal HW-line until 1809. The LW-line has more or less the same position on the maps of 1665, 1731, and 1809 (except between Ballum and Nes, where the LW-line was positioned several hundreds of metres more seaward in 1809 than before). This leads to the conclusion that the inferred minimal HW-lines in the reconstructions for the period prior to 1800 are perhaps too close to the dunes. If the HW-line is taken 100 m S of the LW-line it may have shifted seaward in the periods 1545-1665 and 1731-1809 (central part) and remained fairly stable between 1665-1731 and 1809-circa 1850, after which it retreated, especially in the centre (see below).

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[58] C. De Moulin, 21-03-1781: Rapport wegens de verdediging van de eilanden Ameland en Schiermonnikoog, ARA, Raad van State, no. 2205 (in: Jansen, 1941; Beckering Vinckers, 1943).

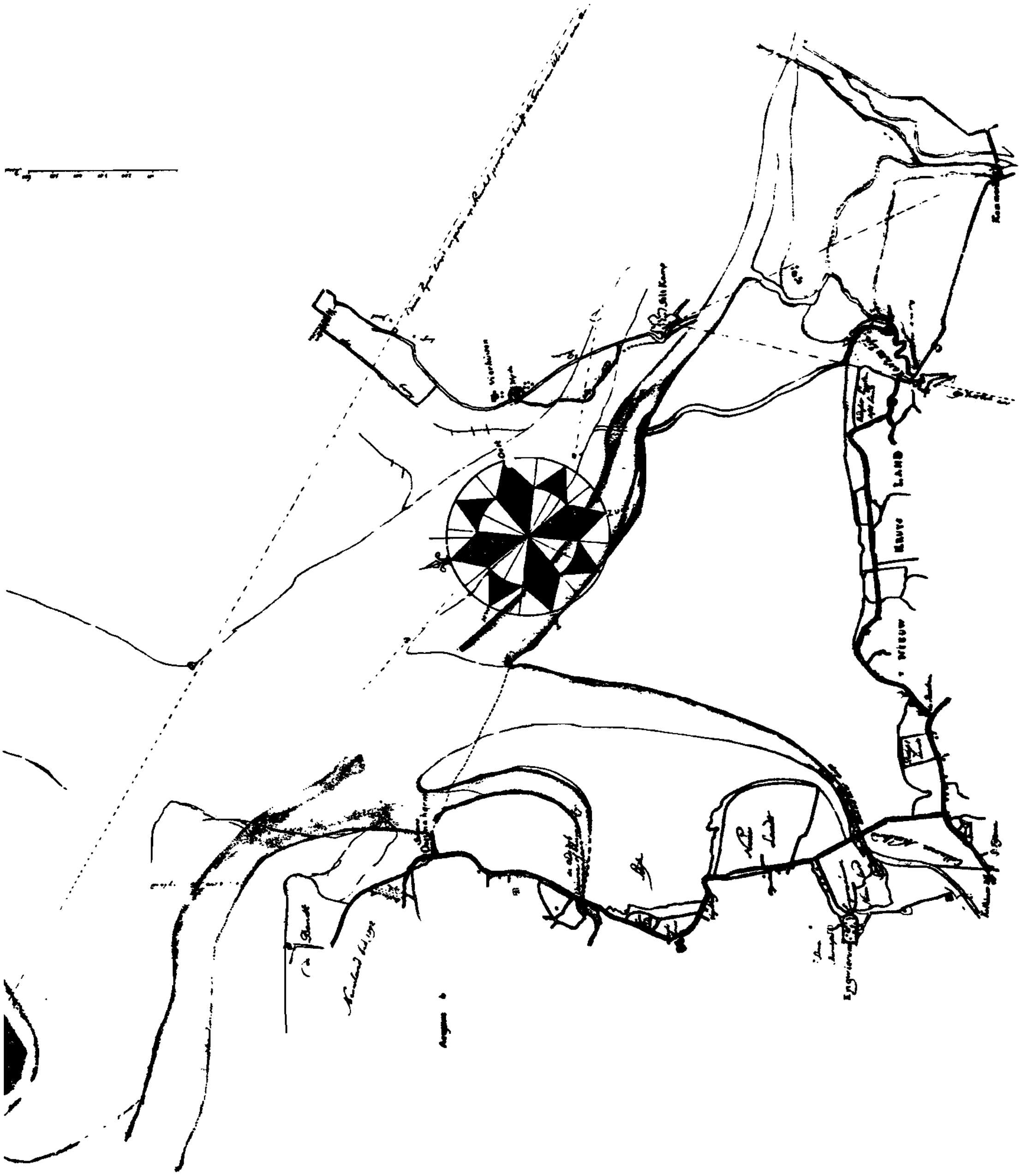

Figure 27. Anonymous, 1757: Chart of the Lauwerszee (copy by Kooper, 1938).

The erosion at the south-central part (Ballumerbocht) still continued; in 1787 it was reported that extensive, clay-rich land areas were eroded (Isbary, 1936). In 1790 an attempt was made to close the washover channel between Nes and Ballum, but this failed (Athallin, 1811). At its eastern end Ameland had migrated eastward over more than half a kilometre. Strong eastward shift had been observed in the mid 18th century (Overdiep, 1964).

By 1806/11 the Pinkegat had become an almost single inlet system, with an abandoned channel separating the two supratidal shoals of the Engelsmanplaat[LXIX, LXXIII, LXXIV]. Limited information is available about the ebb-tidal delta of the Zoutkamperlaag. The NW-oriented outer channels had shifted clockwise and downdrift. A new outer channel was probably formed at the SW side of the ebb-tidal delta. The most northeastern outer channel, and the channel De Noorman which was connected to it, were almost entirely separated from the main inlet channels in 1811 (Athallin, 1811)[LXIX, LXXV]. Erosion of west Schiermonnikoog continued in this period and the island lost several km^2 at its SW side[LXXI, LXXV]. The manor house was probably dismantled in 1761 (Mellema, 1981), the barns in 1787 (Reitsma, 1988), whereas the remaining dyked land was destroyed gradually by migration of dunes and backbarrier channel erosion in the period 1786-1843[59, LXXXII] (Kros, 1848). In the period 1750-1800 the southwestern HW-line became positioned more than 0.5 km E of its location in 1975. The dunes were badly maintained as shown by documents of 1757[60] (Mellema, 1981). The island had expanded some 2 km to the E and was 13.4 km long. The shoal east of it, Simonszand, was now clearly supratidal. The channel between Schiermonnikoog and Simonszand had become a small inlet system.

Comparison of the maps of Beckeringh (1745)[LVII], De Gross (1792)[LXVI], and the one of Athallin (1811)[LXXV] shows that the Lauwers Inlet system had formed a more westward oriented inlet (5.5 km W of its position in 1976) by breaching part of the shoal Simonszand. The more westward orientation may well have resulted from the increase in drainage area (Sha, 1989b; Sha & Van den Berg, 1992). In the process of westward migration of the inlet, the Kapersplaat, located N of Simonszand in 1750, 'crossed' the inlet and amalgamated with the downdrift shoals of Koeplaat and the remains of Bosch. They thus amalgamated into the supratidal shoal, which was again called Bosch. The drainage area of the Schild Inlet was partly taken over by the Lauwers system. The Schild Inlet and the barrier island Rottumeroog migrated to the east. From at least 1770 a long shoal (mostly called: Noorderrif, sometimes: Rottumerplaat; Lang, 1958) approached the N to NW side of Rottumeroog. The channel in between the shoal and the island caused strong erosion of the island (Reenders, 1986). The old house of the island keeper, whose job it was to maintain the bent-grass, was destroyed in 1799 after 56 years, because it had become positioned at the western side of the

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[59] The map of L. den Berger (1809) shows a remnant of the dykes surrounding the eastern land of the manor house.

[60] RAF, Heerlijkheid Stachouwer, no. 180.

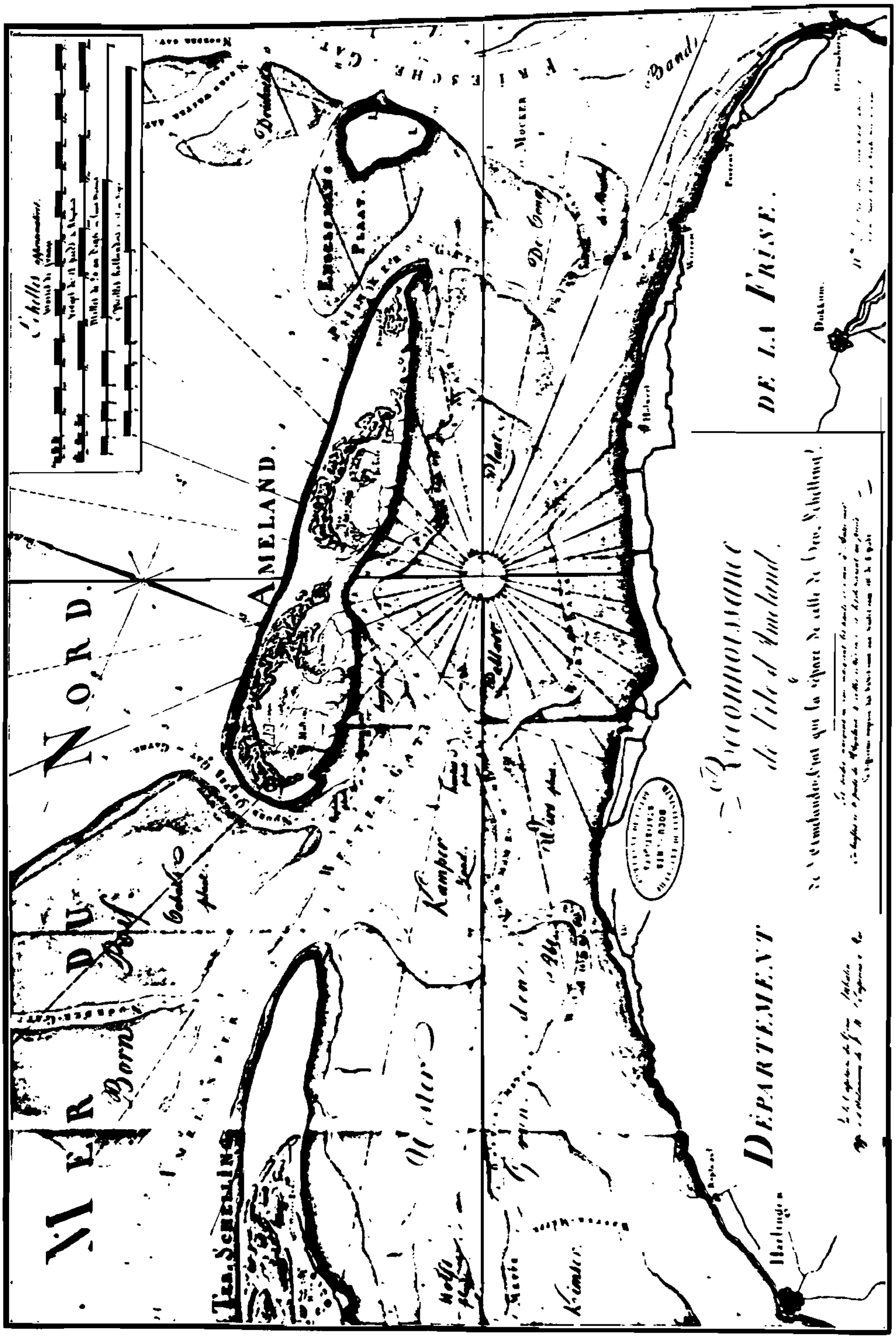

Figure 28: Atthalin, 1811[LXXIV]: "Reconnoissance de l'ile d'Ameland", and surrounding seas.

island (Abrahamse, 1990; Data RWS). The erosion at the NW side was compensated by sedimentation at the E and S side (Reenders, 1986). The erosion towards the ESE/E direction amounted 265 m between 1771-1788, 375 m between 1788-1805, 95 m between 1805-1809[LXXII] (on average 19 $m.yr^{-1}$; Isbary, 1936; Jansen, 1941; Ligtendag, 1990).

Parts of the Frisian coast, NW of Hallum, and along the Lauwerszee were reclaimed. The larger part of the coastline did not change between 1750 and 1800.

1800-1850; Figs A9 and A10

On 3-2-1825 a catastrophic storm surge occurred. The wind, turning from SW to NW, in combination with springtide (Niemeijer, 1975) pushed the waters to great heights (Harlingen +2.93 m DOL; Terschelling +2.95 DOL; Zoutkamp +4.45 m DOL; Delfzijl +4.60 m DOL; Nieuwe Statenzijl +5.09 m DOL; RWS, 1948, 1959). Five out of six successive tides were above the storm-surge level, on average to be only exceeded twice a year (Buisman, 1984). All along the Dutch coast large areas were inundated. Two thirds of the province of Friesland (RWS, 1948), and large parts of land along the Lauwerszee and the Groningen coast were flooded (Niemeijer, 1975). Most barrier islands were flooded as well (Buisman, 1984).

The reconstruction of 1850 (Figs A9 and A10) is as follows: The eastern end of the barrier island Terschelling remained more or less in the same position (that is about 1 km W of the position of 1975), and consisted of a large supratidal beach plain. By 1831[LXXVII] a large supratidal hook had formed to the SE. From 1800 onwards the Ameland Inlet had a western orientation and always consisted of two major outer channels, from which the most western channel was most important most of the time (except in 1831, in 1892, and in 1914; Beckering Vinckers, 1943). In the period 1798[LXVIII] to 1831[LXXVII] the main channel shifted to the E, and from 1831 to 1854[LXXXV] the main channel was oriented again more to the W, thereby probably eroding the eastern end of Terschelling.

The western end of Ameland eroded up to 0.5 km on many places in the period 1809[LXXIII]-1854, the westernmost side being only 0.5 km west of the position in 1975. Of the northern side little is known, but erosion must started locally before 1854, because the map of Eekhoff[LXXXVII] of that year already shows erosional features. A map of 1866 clearly shows a coastal retreat over circa 200 m W of pole 14. The erosion at the south-central part (Ballumerbocht) continued: sand was washed out by the channel and the overlying clays were underscoured, after which they collapsed (Kros, 1848; Isbary, 1936). The development was stopped in 1843 (Isbary, 1936), when a large dam was built. In 1809[LXXIII] the barrier island Ameland still consisted of three separate dune complexes, with in between beach plains and washover channels. According to the descriptions[61], these washover channels were still active in the winter season. The washover channel on Ameland, between Ballum

[61] Athallin on special request of Napoleon.

and Nes, was artificially closed with a weak aeolian dyke in 1807. It was heightened in 1851 (Brouwer, 1936; Van Oosten, 1986). During the storm surge of 1825, dykes on Ameland were breached on several places, and strong erosion occurred at the North Sea side. The fields were covered with dune sand and sea sand (Van Leeuwen, 1826, in: Van der Molen, 1963). The Oerder dune parabola was strongly eroded, and the blown-out sand formed a new dune up to some +24 m DOL (Van Oosten, 1986). In the period 1800-1866 the eastern end of Ameland became positioned 1 km more to the E.

For details about the development of the Pinkegat in the period after 1800 the reader is referred to chapter 3. Since at least 1800 the Pinkegat shows a distinct cyclic development pattern from a single inlet into a multiple inlet and back. A cycle covers several decennia. During the development, the eastern end of Ameland shifts to the east if the channels move into that direction, and retreats to the west if new channels erode the eastern end. In the maps before 1800 sometimes a multiple inlet and sometimes a single inlet is observed. The eastern island end retreats also westward and subsequently shifts eastward (cf. the maps of 1665, 1745 and 1809). These observations strongly suggest that before 1800 inlet development may also have been cyclical, probably since 1600. As a result of the dynamical behaviour of the Pinkegat, the Engelsmanplaat was eroded at its western side in the period 1800-1850[LXIX]. The Zoutkamperlaag gradually split up into two main inlets (for details see chapter 4).

In this period erosion of Schiermonnikoog continued, especially after 1814 (Van Leeuwen, 1826, in: Van der Molen, 1963). The channel De Noorman (with a depth up to 13 m below MHW in 1832!) caused a strong erosion (0.5 km to the NE in 1809-1832/34[LXXI, LXXVIII] and 0.15 km in 1832/34-1843[LXXXII]) of the SW side of Schiermonnikoog, mainly until 1835, after which erosion decreased (Ackersdijck, 1826; Kros, 1848)[62]. After that year erosion became more important in the W and NW (Kros, 1848; the NW coast eroded in the SE direction over 0.2 km in 1809-1832/34, 0.45 km in 1832/34-1843, and 0.2 km in 1843-1854[LXXXII, LXXXVIII]; Berger, 1809; Keuchenius, 1832/34; Boltz, 1842; Min. of War, 1854). Erosion at the NW end occurred due to the downdrift shift of the easternmost outer channel of the Zoutkamperlaag ebb-tidal delta (i.e., to the E). Due to the shift the channel De Noorman[63], was 'pulled' towards Schiermonnikoog (i.e., it migrated towards its inner bend!) and eroded the SW part of Schiermonnikoog. At the S side of the island outer bend erosion occurred (cf. Ackersdijck, 1826). Eye-witness reports (Ackersdijck, 1826; Boltz, 1842; Winkler Prins, 1867) state that erosion along the channel proceeded in two ways:

[62] Loss of farmland was said to have amounted 6.43 km^2 over the period 1786-1837 due to channel erosion and dune migration (Winkler Prins, 1867), of which 0.22 km^2 by channel erosion in the years 1830-1837 (Kros, 1848). Sloppy maintenance will have stimulated the dune migration (Ackersdijck, 1826; Kros, 1948; Winkler Prins, 1867; Mellema, 1981).

[63] In 1832 the channel at the SW point of the island was connected to an outer channel to the NE, E of the main inlet, and by a westward oriented channel to the main inlet (Boltz, 1842)

1) Gradual erosion occurred during every flood: sand was eroded from beneath the soil so that the clay-rich topsoil was undercut and collapsed, and

2) Fast erosion happened due to strong NW and W winds. These strongly increased the 'tidal' volume and hence the current velocities. Additionally erosion by waves occurred[64] (Boltz, 1842; Kros, 1848).

During high floods the grass land was flooded (Athallin, 1811; Kros, 1848) and dunes were sometimes eroded. During the storm surge of 1825, the sea water flooded most of the island through the dunes at the NW side and through the dunes at the southern side, thereby eroding several of them. Several old dykes were breached (Winkler Prins, 1867), others were simply flooded (the water rose more than 1 m above the dykes). Like in Ameland, the fields were found to have been covered with sand afterwards (Winkler Prins, 1867; Van Leeuwen, 1826, in: Van der Molen, 1963; Ackersdijck, 1826; Stachouwer, 1825, in: Reitsma, 1988).

Between 1835 and 1843 the part of the channel De Noorman at the SW end of the island became abandoned and filled up (Boltz, 1842; Kros, 1848). The closure of the channel was caused by a new connection with the main channel, forming a more direct transport path for the tide, in combination with the merger of the shoal SW of the channel De Noorman with Schiermonnikoog (cf. Abrahamse, 1994). After abandonment, erosion came to an end, and a large beach formed at the SW side from circa 1842/1843[LXXXII] onwards (Boltz, 1842; Kros, 1848). Nowadays the coastline lies some 3 km SW of that of circa 1840. As far as can be judged from the maps of 1811[LXXV], 1843[LXXXII], and 1854[LXXXVIII] and from the above descriptions, the NE-oriented outer channel managed to form a direct connection with the main channel after the abandonment of the channel De Noorman. The new connection may well have contributed to the abandonment of De Noorman.

At the eastern end of Schiermonnikoog sedimentation occurred around 1826 (Ackersdijck, 1826), but the new area had already been eroded around 1832/34[LXXVIII], after which again a slight eastward expansion occurred. The shoal Simonszand was in 1832/34 positioned more into the Wadden Sea than in 1811[LXXV], and had become intertidal. The, originally quite deep (before 1832, data RWS), channel between the shoal and the island Schiermonnikoog (Eilanderbalg or Balg) had become intertidal by 1845 and supratidal by 1848 (Kros, 1848). In the period 1848-1873/74[XCIII], the channel deepened and the shoal Simonszand increased in height, and became gradually supratidal.

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[64] J. Ackersdijck, 8 september 1826, during the onset of stormy weather: "*Dezen morgen hebben wij nog eene wandeling gemaakt naar de zuidwestkust van Schiermonnikoog. De vloed kwam nu met geweldigen stroom uit het westen, geheel in de rigting van Groningen. Daar zagen wij nu hoe het land met groote stukken door de zee wordt weggeslagen, juist van den besten kleigrond.*" (This morning we made a walk to the SW coast of Schiermonnikoog. The flood came with a powerful flow from the west, in the direction of Groningen. There we saw how the land, the best clay ground, was ripped away in large chunks by the sea.)

The gorge of the Lauwers Inlet system shifted circa 3.5 km to the E in the period 1811-1873/74[LXXV, XCIV], mainly in the period 1811-1833/34[LXXV, LXXIX]. In the backbarrier area the Lauwers Inlet system had contact with the Schild Inlet system and expanded eastward in 1811-1833/34. Between 1833/34 and 1873/74 the Lauwers Inlet system strongly retreated westward, and became disconnected from the Schild Inlet system. Comparison of the surveys of 1833/34 and 1873/74 suggests that already in 1833/34 an additional channel may have been present W of the Lauwers (Binnen Spruit of 1873/74; Jansen, 1941). The western channel became quite pronounced in the period 1833/34-1873/74, suggesting an increase in drainage volume at the western side. In 1826 'new' Bosch was a sandy shoal which was regularly flooded (Ackersdijck, 1826). In 1833/34 Bosch was a 5 km broad supratidal shoal. It was positioned considerably coastward (an effect of the storm surge of 1825 and/or partly due to inaccurate mapping?). Until 1873/74 Bosch decreased in size and shifted in the direction of the North Sea.

From 1825 onwards, a new wave- and flood-built shoal was formed, also called Rottumerplaat (Isbary, 1936). On the soundings of 1833/34[LXXIX] no shoal Rottumerplaat is shown yet, but afterwards it gradually increased in height until it became supratidal around circa 1860 (cf. Brilhuis et al., 1990). Between the shoal and Rottumeroog, the relatively large Schild Inlet was present. It became smaller during the period 1811-1873/74[XCIV].

In 1805 strong erosion was observed at the north side of Rottumeroog (Reenders, 1986).

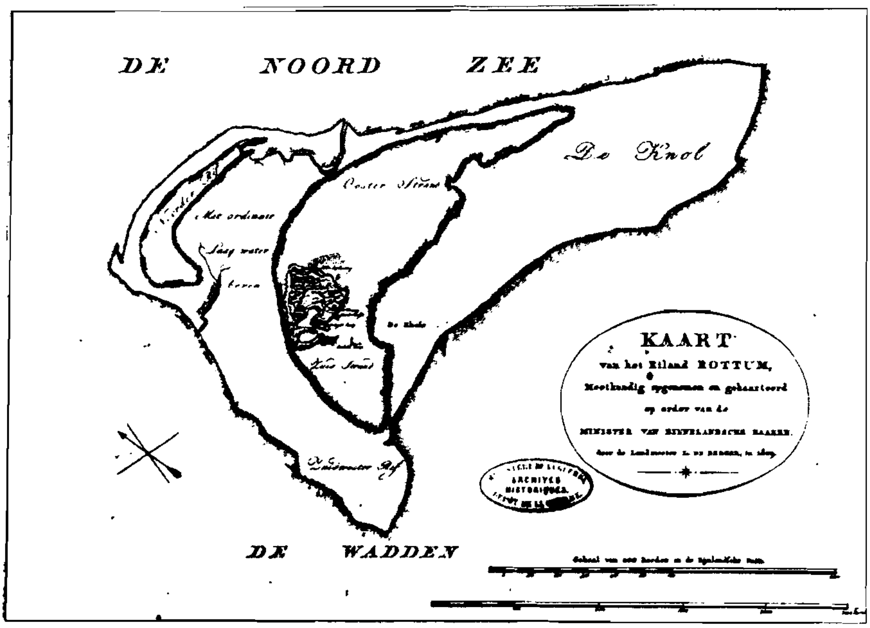

Figure 29: L. den Berger, 1809: "Kaart van het Eiland Rottum", map of the barrier island Rottumeroog.

Over the period 1799-1826 (mainly until 1809, since the map of 1833[LXXIX] does not differ much of the one of 1809[LXXII]; Fig. 29) eastward erosion is said to have amounted 340 m. The now supratidal shoal Noorderrif, NW of Rottumeroog, became partly connected by an intertidal zone with the island in 1805 (Reenders, 1986). On the 3.5 km long supratidal Noorderrif even birds were breeding in 1808 (Isbary, 1936). Due to the steady erosion of the large shoal in the period 1809-1839, enough sand was supplied to Rottumeroog to temporarily stop the net erosion of the W side of the island (cf. Isbary, 1936; Jansen, 1941). During the storm surge of 1825 the dunes[65], from the SW to the NE, were eroded over 18 m (Niemeijer, 1975) and locally the shore was eroded over 40 m (4 roeden of 10 m; Verhoeff, 1983) over a length of 1 km (Ackersdijck, 1826). The bent-grass vegetation of Rottumeroog was, however, well kept[66], and new dunes formed at the S and E side (Ackersdijck, 1826; Abrahamse, 1990). However, little or no new land was gained (Reenders, 1986).

1850-1900; Figs A10 and A11

The eastern end of Terschelling remained a sandy, just supratidal shoal (beach plain) with some isolated dunes. The gorge of the Ameland Inlet shifted towards the east in the period 1854-1903/4[CI]. In addition, marginal channels at the NW side of Ameland were repeatedly pushed towards the island by downdrift migrating shoals (e.g., 1865-1866[XCII], around 1892[XCVII]). These inlet and channel movements caused a considerable erosion and retreat of the dune line at the NW coast of Ameland, and of the HW-line at the SW end. At the North Sea side, especially the western part, erosion was strong; up to 0.5 km. Considerable growth occurred at the eastern end of the island before 1873. On 20-12-1862 Ameland was largely flooded during a strong storm. It caused breaching of the southern dykes (Brouwer, 1936). The washover complex east of Buren was, after several attempts in 1855 and 1880, closed by an aeolian dyke between 1882-1893 (Brouwer, 1936; Van Oosten, 1986).

For details on the development of the Pinkegat, Engelsmanplaat and Zoutkamperlaag the reader is referred to chapters 3 and 4. The Pinkegat most likely developed a single inlet phase slightly before 1873[XCIII] (Fig. 30). It was replaced by a multiple inlet in 1891[XCV]. In the process, the Engelsmanplaat was strongly eroded at its western side. The easternmost inlet of the Zoutkamperlaag became gradually more isolated.

In this period the westernmost coast of Schiermonnikoog continued to retreat irregularly, especially due to a channel which was present between circa 1873 and 1903 (data RWS). Around 1903 the westernmost HWL was some 1 km E of its position in 1975. Drawings of RWS suggest a considerable retreat (300 m) of the North Sea coast in the centre of the island in the period 1854-1873/74[XCIII]. At the end of the 19th century the new dune row of

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[65] In 1811 Athallin reported that the dunes at the NW side of Rottumeroog were some 9 m high (Abrahamse, 1990).

[66] The island keeper even managed to exterminate all rabbits on the island in 1839.

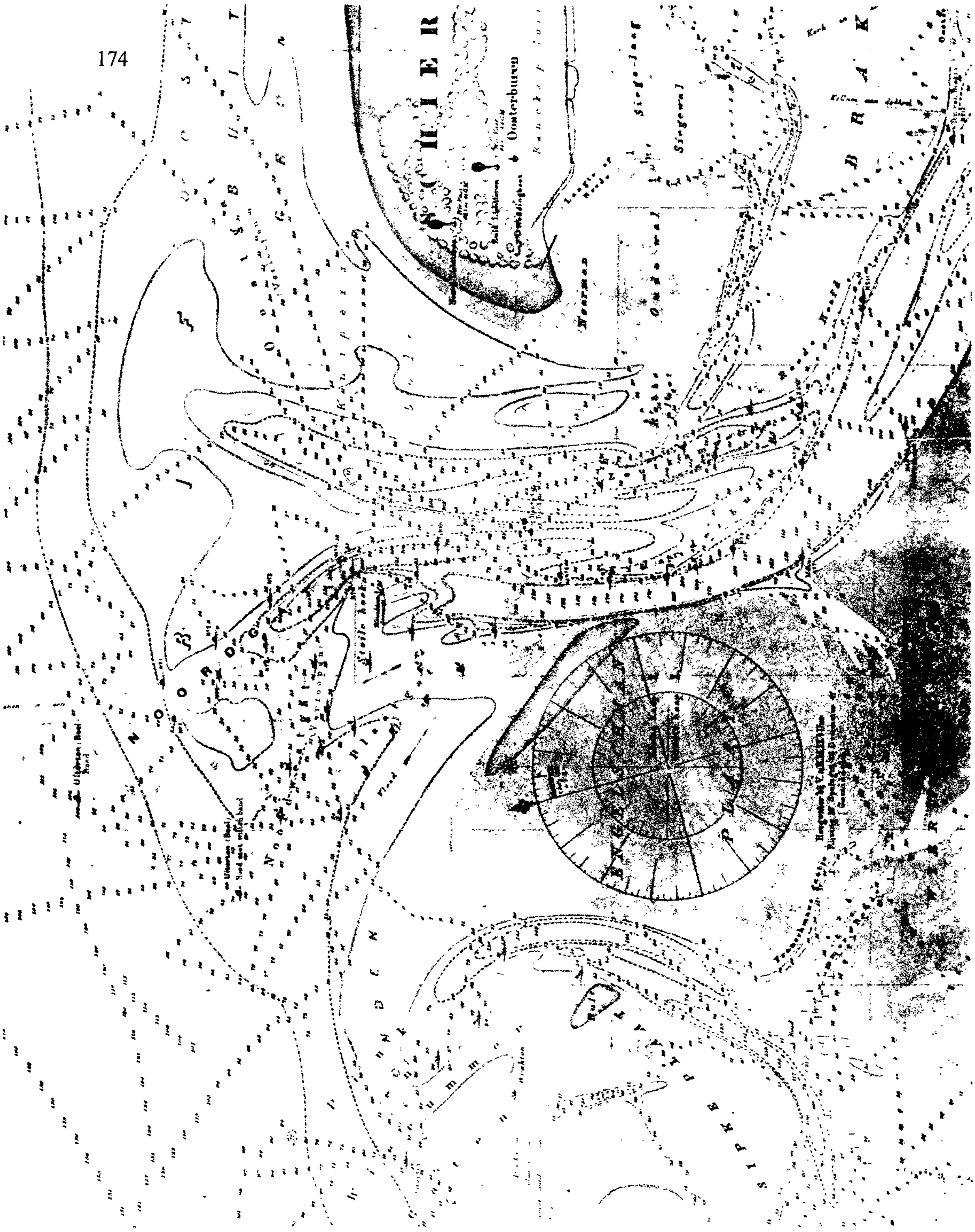

Figure 30: P.J. Buijskes & T.E. de Brauw, 1873/74: Illustrating the strong improvement in depth-sounding techniques after the 18th century.

Kobbeduinen formed E of the row Kooiduinen. These were the most easterly dunes until than (Van Oosten, 1986). A large part of the polders at the southern side of Schiermonnikoog were protected by a large dyke in 1860, built by the new island owner Banck. The supratidal area S of the dyke was gradually eroded, especially in the period 1891-1960. This was probably caused by the increase in wave height due to the approach of channels of the Zoutkamperlaag system, due to wave reflection against the dykes and/or to the effects of the increase of sea-level rise. In addition, the removal of sediment for the dyke along Southern Schiermonnikoog may have played a role. During the period 1850-1900, the eastern end became broader, and a dune complex formed on it.

In 1867 the shoal Simonszand was separated from the barrier island Schiermonnikoog by a shallow, just subtidal channel (Eilanderbalg; Van Buijtenen, 1954). This indicates that the channel had become deeper since 1848 (cf. Kros, 1848). After 1867 the channel became still deeper and shifted some 1.2 km to the E between 1873/74-1891[XCIV, XCVI]. A new main channel opened W of the original one (Jansen, 1941). At the same time the backbarrier channels of the Zoutkamperlaag retreated westward. Simonszand became again a supratidal shoal.

The Lauwers Inlet shifted some 0.5 km to the E between 1873/74-1891. It was only 1.5 km E of its position in 1978. It expanded its drainage area to the E. The westernmost backbarrier channel (Binnen Spruit) merged with the main channel of the Lauwers Inlet by 1890 (Jansen, 1941). The Boschplaat decreased in size until 1887, and afterwards it increased again (Brilhuis et al., 1990). The supratidal shoal Rottumerplaat increased gradually in size from at least 1874 onwards, and expanded westward and eastward over some 0.6 km in the period 1873/74-1890 (Fig. 31; Jansen, 1941; Brilhuis et al., 1991). After 1890 the westward growth continued[XCVIII]. In 1872 natural dunes had formed on the island, but they were destroyed already in 1887/88 (Isbary, 1936; Abrahamse, 1983). In 1888/91[XCVI] Rottumerplaat had become a long wave-dominated, supratidal shoal. On the shoal RWS tried to establish dunes, but these were washed away each winter (Abrahamse, 1983).

The Schild Inlet system formed an almost straight channel into the backbarrier area in 1873/74[XCIV]. It became more meandering towards 1888/91[XCVI]; at the same time the inlet shifted some 0.8 km to the E (48 $m.yr^{-1}$; Jansen, 1941; Abrahamse, 1990). In 1854 a shoal was present N of Rottumeroog (Reenders, 1986). Between 1854-1861 the shoal probably merged (cf. Jansen, 1941), so that the western side of the island grew. The shift of the Schild Inlet caused strong erosion at the western side of the island (Abrahamse, 1990) in the period 1873/74-1889 (the HW-line retreated ± 820 m in 1861-1888/91[67] and ± 600 m in 1888-1899; on average 37 $m.yr^{-1}$; data Jansen, 1941).

[67] Occasional storm surges, such as on 19/20-12-1862 (20 m dune erosion), 3-12-1863 (>2.3 m above MHW), 1887 (destruction of the second home of the island keeper) contributed considerably to the erosion (cf. Isbary, 1936; Reenders, 1986).

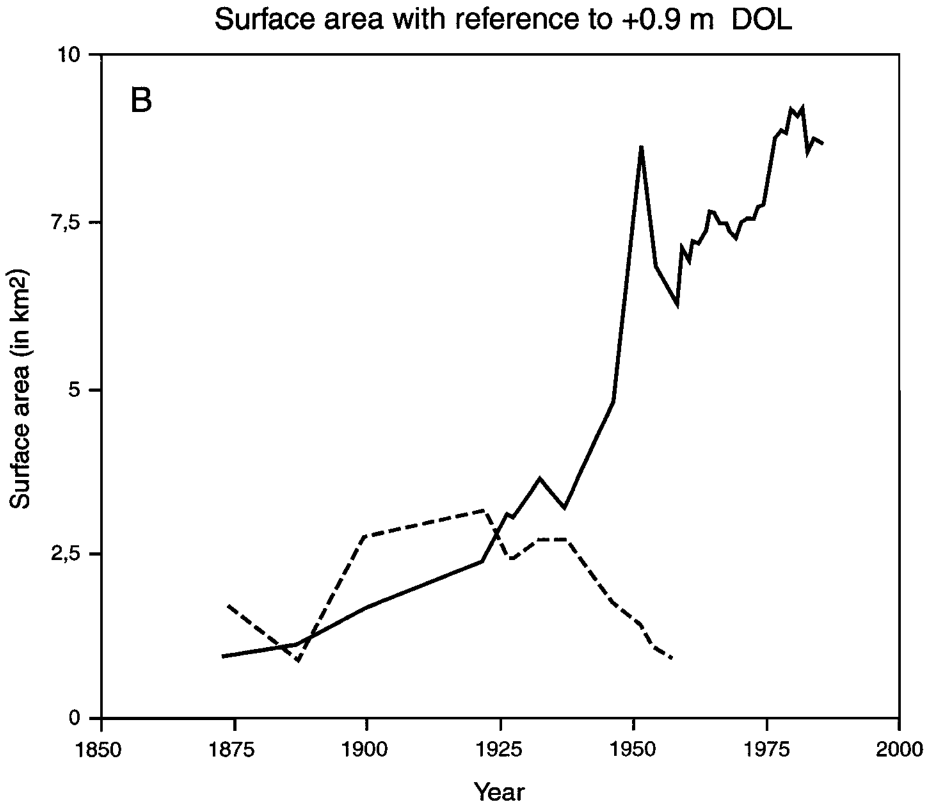

Figure 31: Increase of surface area (above +0.9 m DOL) of Rottumerplaat since 1875 (solid line) and growth and decline of Bosch (broken line) (after: Brilhuis et al., 1990).

Over the period 1850-1900 the size of the island Rottumeroog remained fairly stable (Fig. 32; Jansen, 1941; Brilhuis et al., 1990), because erosion at the western side was compensated by sedimentation at the eastern side. The shift was primarily made possible by a gradual shift of the northern part of the Western Ems to the E. The maintenance of a thick cover of bent-grass (Abrahamse, 1990) will have stabilized new dunes. Clay-rich sands (up to 30 cm thick) were deposited S of the dunes (Isbary, 1936). In 1861 the dunes were 6 m to 13.5 m high, and in 1887 they were 11 m to 13.5 m high at the NW side (Isbary, 1936; Reenders, 1986), while only 9 m were reported in 1811 (Abrahamse, 1990). Obviously, the massive supply of sand due to the erosion at the western side allowed a strong vertical growth.

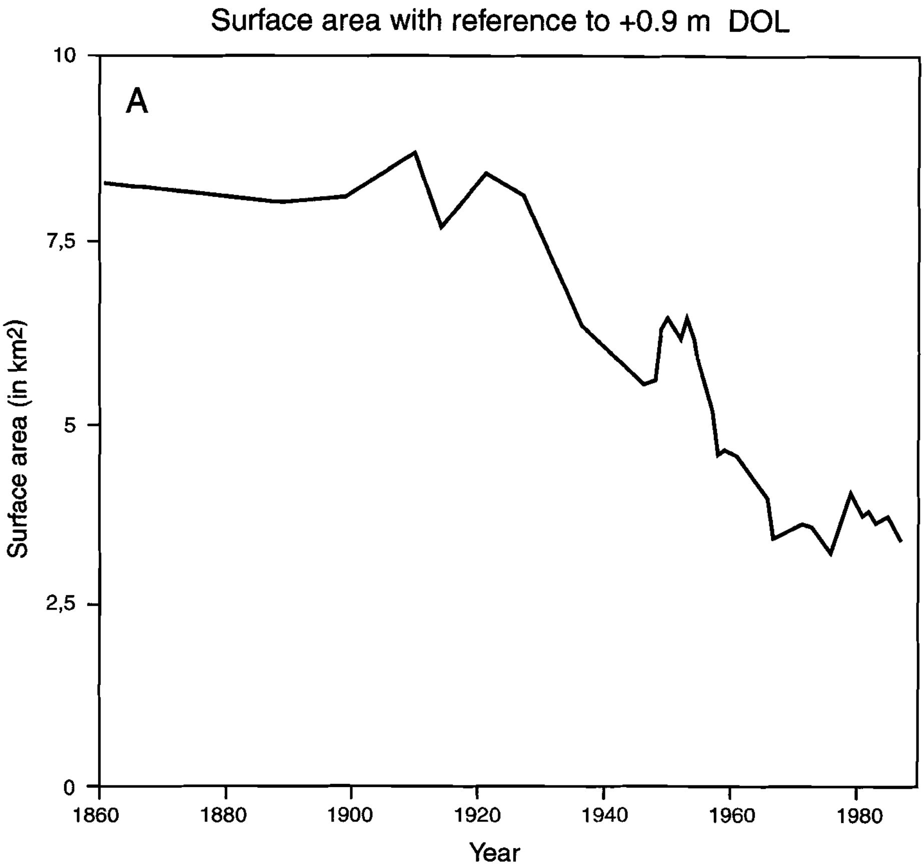

Figure 32: Decrease of surface area (above +0.9 m DOL) of Rottumeroog since 1860 (after: Brilhuis et al., 1990).

1900-1950; Fig. A12

The eastern end of Terschelling (Boschplaat) did not change much in size during this period[XCVII, CIX]. Around the first half of the 20th century, when the dune-covered area increased in size, wash-over channels became more pronounced. To prevent the formation of washover channels, aeolian dykes were constructed between 1923-1929, which were not oriented well to the wind and washed away. Between 1931-1938 an E-W trending, new aeolian dyke was formed. Behind the dyke a salt marsh established (Van Oosten, 1986). After eastward expansion in the period 1880-1922, the eastern part of the island was slightly eroded, when a small backbarrier channel migrated to the N and gradually evolved into a separate inlet channel (around 1934; Beckering Vinckers, 1943).

Ameland Inlet shifted strongly to the E, and the orientation changed from NW-SE to N-S (Joustra, 1971). As a result strong erosion occurred, especially at the SW side of Ameland (Fig. A12). On the island erosion of the dunes N of Hollum and Ballum continued (Brouwer, 1936). Also the HW-line at the North Sea coast retreated slightly between pole 10 and 20, whereas west of pole 7 sedimentation occurred. New rows of dunes formed at the eastern side of the island in the 20th century (Van Oosten, 1986). After an initial westward retreat between 1900-1917, the eastern end of Ameland shifted strongly to the E in the period 1917-1949[CIX] (the LW-line over circa 3.6 km in the period 1910-1927 (probably after 1917), which is at least 210 $m.yr^{-1}$ on average, and the HW-line over 4.5 km in the period 1917-1949, which is almost 140 $m.yr^{-1}$!). The Pinkegat developed (perhaps even twice) from a multiple inlet into a single, and back again, in the period 1894-1949[XCV, CIX] (chapter 3). Due to the dramatic changes in the Pinkegat, strong erosion occurred at the western side of Engelsmanplaat.

The easternmost inlet channel of the Zoutkamperlaag became gradually abandoned. From detailed surveys it appears that the outer channels of the inlet experienced a strong downdrift shift (see chapter 4). Between 1918 and 1922, strong channel-induced coastal erosion occurred at the NW side of Schiermonnikoog (cf. Mellema, 1981). This part of the island retreated eastward over at average 0.7 km in the period 1854-1929 (Isbary, 1936). New dune parabolas formed at the eastern side of the island. The island expanded over 2 km to the E in the period 1891-1927[C, CVI]. Also the inlet of the Eilander Balg shifted strongly downdrift, but its drainage area did not shift much. In the same period Simonszand migrated downdrift to the E over 3.1 km. The shift continued until at least 1937/39 (Jansen, 1941).

The Lauwers Inlet shifted slightly to the E (circa 0.9 km between 1888/91 and 1927), but became less wide due to the migration of Simonszand. It also took over some of the drainage area of the Schild Inlet system. With the increase in drainage volume the backbarrier channel became increasingly oriented to the NW (cf. Sha & de Boer, 1991; Sha & Van den Berg, 1992). In the period 1888/91-1927/28 the points of bifurcation of the backbarrier channels shifted seaward (Jansen, 1941) together with a more seaward alignment of the shoals between Schiermonnikoog and Rottumeroog.

In this period Boschplaat, between Schiermonnikoog and Rottumeroog, firstly increased in size, but after 1925/1933 it started to become smaller (Fig. 31; Isbary, 1936; Brilhuis et al., 1990). The shoal gradually migrated to the ESE (the western HW-line between 1888/91-1927/28 over 0.86 km with an average speed of 21 $m.yr^{-1}$, and between 1927/28-1950 over 0.91 km with an average speed of 40 $m.yr^{-1}$; Jansen, 1941). The shoal Rottumerplaat increased in surface area since 1875, due to the amalgamation of sandy shoals at its western side (Fig. 31; Brilhuis et al., 1991). Therefore, its HW-line shifted westward over 3.6 km in the period 1888/91-1950 (of which 2.9 km in the period 1927/28-1950; an average migration of 127 $m.yr^{-1}$), despite of the fact that erosion at the NW side occurred repeatedly (Isbary, 1936). Annual observations over the period 1930-1939 show that the eastern end of Rottumerplaat (LW-line) shifted some 1.4 km towards the east (155 $m.yr^{-1}$; Jansen, 1941). Between 1904-1911 the artificially encouraged aeolian sedimentation resulted in the formation of more permanent dunes (Abrahamse & Luitwieler, 1983)[68]. Little maintenance was done between 1914-1950 (Abrahamse, 1983), but the dunes survived. By 1933 3-4 m high dunes were present, which were eroded at the NW-side (Isbary, 1936).

The tendency of the Schild Inlet system to meander continued between 1888/91-1927/28. In that period the inlet gorge migrated downdrift (1.6 km). Its drainage area decreased and the inlet and ebb-tidal delta reoriented increasingly more to the downdrift side. In the period 1927-1939 the channel hardly migrated eastward (150 m). It became less wide (-1.4 km), because the two inlet channels merged.

Over the period 1900-1950, especially after 1930, Rottumeroog decreased in size[69], because it still migrated eastward and was disappearing gradually into the Western Ems (Fig. 32; Brilhuis et al., 1990). Erosion at the western side could not be compensated any more by sedimentation at the eastern side. The northern part of the Western Ems could not shift much more to the east, because the island E of it, Borkum, had been totally stabilized by massive protection works. Erosion of Rottumeroog proceeded at a relatively regular pace, although sometimes the occasional attachment of sandy shoals at the westside temporarily reversed the erosional trend (Isbary, 1936; Brilhuis et al., 1990). Reversely, storm surges accelerated the erosion[70]. Aeolian erosion became especially strong during 1940-1947.

1950-2000; Fig. A13

The erosive channel S of Terschelling had become abandoned around 1949, and the shoal to the S merged with the eastern end of the island. Then, as part of a cyclic development, a new channel evolved in the backbarrier area and also shifted towards the N, thereby eroding the shoal (De Boer et al., 1991a). In 1981 the channel had reached the southern HW-line of the island, and in the period 1981-1984 the eastern end was eroded over circa 700 m (De Boer et al., 1991a). The gorge of Ameland Inlet continued its eastward shift, thereby eroding

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[68] In 1911 Jac. P. Thijsse visited the shoal and stated that the dunes were up to 4 m high, and were covered with a dense vegetation of *Elymus farctus*, *Leymus arenaria* and, at the northern sides, *Honkenya peploides*, while on the lower places *Cakile maritima* grew. In 1917 he revisited the island: the dunes had become higher; two dune complexes were present. New plants had settled, such as: *Plantago maritima*, *Armeria maritima*, *Pucinella maritima*, *Sonchus arvensis*, *Senecio sp.*, *Spergularia media*, *Aster tripolinium* and *Sedum acre* (Abrahamse & Luitwieler, 1983).

[69] Moreover, after 1930 a small shoal (Zuiderduintjes) detached from the island, and migrated S over 1.5 km in the period 1930-1990 (Brilhuis et al., 1990). This also decreased the size of the island.

[70] Storm surges with major damage occurred on 30-9-1911, 13-1-1916 (+3.3 m above MHW, 25 m dune erosion) and on 17-1-1931 (Reenders, 1986).

the SW side of Ameland. In the sixties the process was stopped by massive underwater protection (De Boer, 1991b). Thereafter the inlet gradually deepened, because the drainage area became larger towards the east (Fig. A13). From 1988 the recurved bar NW of Ameland shifted and strong erosion occurs to the southwest and the east of it.

The aeolian dykes of Ameland were re-enforced in the second half of the 20th century (Reitsma, 1984). At the North Sea coast of Ameland, erosion occurred between pole 10 and 15. The eroded sand was most likely deposited in the backbarrier area. New sand was artificially nourished to the eroded beach regularly since 1979 (Roelse, 1988). For the area, RWS has chosen this flexible (natural) way to defend the coast. The HW-line at the eastern end of Ameland has retreated strongly between 1949 and 1967 (1.8 km). Afterwards it did not show large fluctuations until 1982, after which it migrated to the east, together with the shift of the Pinkegat Inlet. The inlet developed from a multiple inlet (1958) to a single (1967), to a multiple (1972/75), and back to a single inlet phase (1987). From 1987 onwards a new multiple inlet phase is developing (for details see chapter 3).

A new shoal was formed by waves and flood currents at the northern side of Engelsmanplaat, thereby adding to the abandonment of an outer channel, between the new shoal and Engelsmanplaat, which was filled up. Both shoals likely will merge within the next decennia. The Engelsmanplaat has become so small from W to E, that it may well disappear if it will again be eroded by channels at its western side[71]. Similar changes were observed after the disappearance of the island Buise in the East Frisian barrier chain (Luck, 1975). In the period 1650-1860 the island disappeared, leaving too broad an inlet. This was compensated by the merger of the inlet systems Noorderneyer Tief and Buisetief and by strong eastward shift of the eastern end (circa 5.3 km in the period 1650-1960) of the updrift barrier island Juist (Luck, 1975). Also the watershed S of Juist shifted eastward, due to the close interrelation between the position of the watersheds and the eastern ends of the barrier islands. Similarly, upon disappearance of the Engelsmanplaat, the Pinkegat Inlet and the Zoutkamperlaag Inlet will likely merge, and form one inlet system. The too large width of that inlet may well be compensated by a downdrift expansion of the updrift island Ameland. Because Ameland cannot shift at its western side, it may become too long, if the observed relation between the island length and tidal amplitude is considered (Wolff, 1986). This may lead to a tendency to a split up of the island if artificial beach nourishment is not continued[72],

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[71] Artificial coastal protection prevents further erosion of the updrift end of Ameland. This likely decelerates the net expansion of the eastern end of Ameland, because the length of a barrier island is related to tidal amplitude (chapter 1). However, the relation is not very strict (Fig. 6, chapter 1), and the still continuing cyclic development of the Pinkegat Inlet may lead to the elimination of last part of the Engelsmanplaat shoal. Indeed, strong erosion at the western side was observed in the period 1987-1991 (chapter 3).

[72] It should, however, be kept in mind that also Terschelling is 'too long' since at least several centuries (chapter 1, fig. 6), but has never been breached. Apparently the length of a barrier island is allowed to deviate somewhat from the ideal.

most likely over the then newly formed easternmost end. The, with the eastward extension of Ameland related eastward shift of the watershed will increase the surface area of the tidal basin of Ameland Inlet, and hence lead to an increase of the dimensions of that system. The substitution of the clay core beneath Engelsmanplaat by sand and additional sand needed to make the large-scale changes possible, will most likely require amounts of the order of 10^7 m^3 (Oost & De Haas, 1993). Considering the dynamics of the system, this will largely be extracted from the updrift coasts[73], i.e., especially from Ameland.

The Zoutkamperlaag has changed considerably, because of the closure of the Lauwerszee in 1969. The reduction of the tidal volume led to a change of the volume of the ebb-tidal delta, to the formation of a very large recurved swashbar NW of Schiermonnikoog (Fig. 33), to the partial infill and the downdrift shift of the inlet, to a partial fill of the main channel, and to the downdrift shift of the watershed S of Schiermonnikoog (for details see chapters 4 and 5). The shift of the watershed was temporary. At present the watershed is moving back to the W (pers. comm. De Vries) and the main channel is filling up. The large recurved bar NW of Schiermonnikoog is being eroded and will disappear in the near future. The high supratidal shoal SW of Schiermonnikoog decreased in height in the period 1958-1970. It increased in height again from 1970-1987, the sediment being derived from the erosion of the ebb-tidal delta. The measurements over the period 1987-1991 show a new decrease in height. This will probably continue, whereas the ebb-tidal delta is delivering less sand than shortly after the closure of the Lauwerszee.

In the period 1949-1970 a large bulge-formed shoal gradually spread along the WNW side of Schiermonnikoog, mainly by wave action. In the period 1958-1982 strong sedimentation occurred at the NNW side of the island, mainly because the large recurved bar merged with the island. After 1982, the wave of sedimentation moved gradually to the east. In the period 1950-1978 an aeolian sand dyke was constructed along the northern side of the eastern part of Schiermonnikoog. In the shelter, S of the dyke a salt-marsh vegetation developed on top of the sandy beach deposits. The salt-marsh deposits wedge out towards the North Sea. Originally the artificial dune extended further to the east, but part of it was destroyed by storm surges. Repairs were stopped after 1984, so that washovers became active again. Several of the washover areas increase each winter half year (November-March) in their areal extent, thereby eroding the dune row and part of the tidal marshes. In the summer half year (May-September) aeolian sediments partially fill the eroded area. The eastern end of Schiermonnikoog remained largely in place although there has been some expansion (circa 0.5 km) around 1967 and 1991, with in between periods of erosion and the formation of small channels, which eroded the eastern end. At present an increasingly natural development is allowed to occur on the eastern part of Schiermonnikoog; the whole island is nowadays a National Park.

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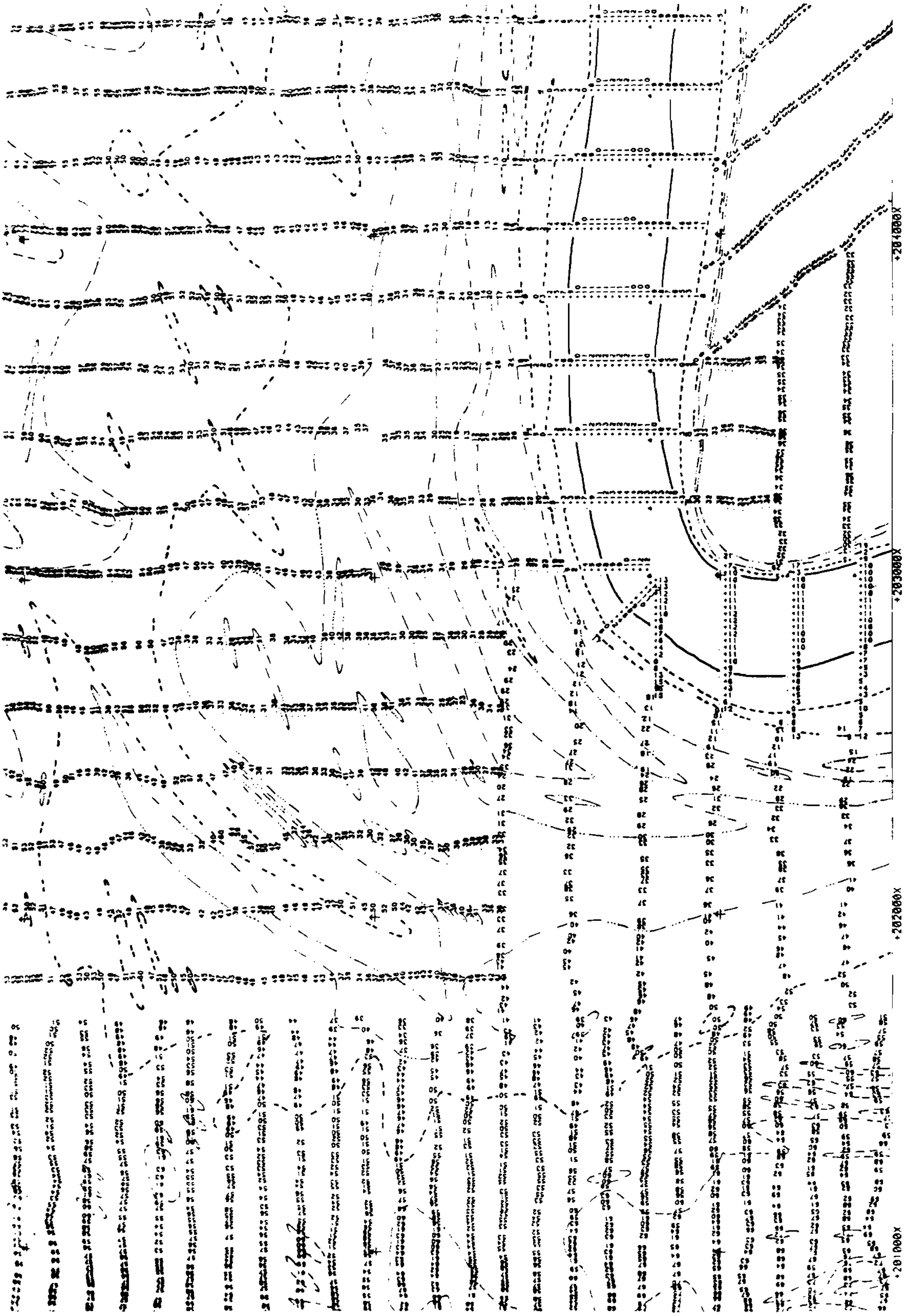
[73] Another minor part of the sediment will most likely be derived from the offshore area.


+204000X
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+201000X

The Eilanderbalg Inlet, E of Schiermonnikoog, shows a cyclic behaviour. The inlet channels shift downdrift and new outer channels are formed at the updrift side. After 1950 the position of the inlet as a whole did not change considerably. In the period 1969-1987 its cross-sectional area, however, decreased, due to a decrease of the tidal drainage area when channels of the Zoutkamperlaag Inlet migrated to the east. At present the watershed S of Schiermonnikoog is moving westward, and it is to be expected that also the dimensions of the Eilanderbalg will increase again.

The shoal Simonszand has been relatively stable in this period. The tidal volume of the Lauwers Inlet has increased during the last 100 years, because its drainage area has expanded to the E (Brilhuis et al., 1990; data RWS). The increase from $160*10^6$ m^3 to $187*10^6$ m^3 in the period 1954-1980 (Brilhuis et al., 1990) is likely an important reason for the increasingly updrift orientation of the ebb-tidal delta. Considering the threshold values given by Sha (1990), it passed a critical value defining the transition from downdrift to updrift oriented tidal inlets. Considering the fact that a slight updrift orientation was already present earlier it is concluded that, apart from tidal volume, also the large Western Ems ebb-tidal delta is an important factor determining the orientation of the ebb-tidal delta of the Lauwers Inlet.

Rottumerplaat expanded to the S (especially after 1927/28) and, as predicted by Isbary (1936), merged with the shoal Boschplaat in 1959 (Brilhuis et al., 1990; Unpublished data RWS). Although erosion occurred at the NW side between 1950-1978, the island gradually increased in size, stimulated by the construction of aeolian dykes (Fig. 31; Abrahamse, 1983), and was some 8.6 km^2 large in 1985 (Brilhuis et al., 1990). It is to be expected that the island will continue to increase in size (Brilhuis et al., 1991). The Schild Inlet system is gradually decreasing in size, because the Lauwers Inlet is taking over its drainage area. It is illustrated by the decrease in cross-sectional surface area of the inlet in the period 1887-1987 from 5,300 to 1,880 m^2 (Brilhuis et al., 1990).

Detailed measurements show that Rottumeroog has shifted some 3 km to the E during the last 100 years (Brilhuis et al., 1990). In 1952-1953 eastward erosion of the HW-line even was 100-200 m (Reenders, 1986). Such rates illustrate the impressively high migration velocity of such small barrier islands. The decrease in size of the island accelerated from 1930 onwards. In 1987, Rottumeroog was about 3.4 km^2 (Fig. 32; Brilhuis et al., 1990)[74] large.

Figure 33 (opposite page): Part of modern depthsounding chart, Frisian Inlet, 1987, Noordblad 131,11, (Mapno. 79 10 129, register 88.0086). Detailed depthsounding of the NW point of Schiermonnikoog, N to the top. Visible is the large swash-bar, which formed after 1967.

[74] Frequent storm surges in combination with aeolian erosion of the dune cliffs resulted in a dune erosion of some 500 m in the period 1955-1990 (data Reenders, 1986; Brilhuis et al., 1990).

If erosion continues at the same rate, the island will disappear in the next 30 to 40 years (Brilhuiset al., 1990). The total intertidal area of Boschplaat, Rottumerplaat, Rottumeroog, and Zuiderduintjes has been fairly constant, between 12-13 km^2, over the period 1900-1985 (Brilhuis et al., 1990). An ongoing decrease of Rottumeroog may well be compensated by a growth of (the) other islands. Another possibility is that Rottumerplaat will merge with or feed sand to Rottumeroog in the future, after abandonment of the Schild Inlet (Brilhuis et al., 1990). The courageous decision has been made to let Rottumerplaat and Rottumeroog develop in a natural way, with only a minimal interference of man (Brilhuis et al., 1991; Brilhuis pers. comm.). This leads to a natural development of a morphodynamically active landscape, rare in The Netherlands (Revier, 1991).

The changes during history per area: a summary

A short summary is given of the changes, which appear from the above pre-historical and historical data. For details and references the reader is referred to the above chronological discussion.

Geological information reveals that barrier islands have been present since at least 5,300 B.P., and intertidal flats, tidal channels, estuaries, and tidal marshes since at least 6,500 B.P.. Many changes have occurred in the Dutch Wadden Sea over the past 2 millennia. The small amount of historical information available for the first 1500 years, confirms that (from at least 0 B.C. onwards, possibly since the 4th century B.C.) the Wadden Sea had barrier islands, channel systems, and subtidal and intertidal flats. Along the coast, estuaries were present. Around 0 A.D., the western part of the Dutch Wadden Sea was still a high and supratidal area, and the Zuider Zee lakes drained through the Vlie Inlet. During Roman times the tidal range may have been smaller than at present.

The southern Lauwerszee embayment was formed during the 8-9th century, and its maximal E-W dimensions probably increased until the 13th century. The Middelzee was formed when the Boorne valley was flooded in the late 9th-10th century. It was filled in from circa 1000 A.D. onwards.

During the last five centuries the changes in the eastern Dutch Wadden Sea can be followed in fairly great detail. From the eastern end of Terschelling to the Western Ems most islands and inlet systems experienced a net downdrift, eastward shift, interrupted by phases of subordinate updrift motion. In addition, erosion and sedimentation occurred, so that the coast lines of the islands prograded and retreated. From W to E, along the coast, the following changes occurred after 1300 (Figs A1-A13):

1) Eastern end of Terschelling

The eastern end of Terschelling became 10.5 km longer after 1550 (Figs A3-A13). This happened for the larger part before 1800. In the process the westernmost inlet channel of Ameland Inlet (Koggediep) was closed (see 2).

2) Ameland Inlet

Ameland Inlet changed from a double inlet system into a single inlet system. After an initial stage of abandonment (c. 1550-1700), the remainder of the westernmost inlet had been filled up, and the shoal east of it merged with Terschelling (Figs A3-A8). Subsequently, downdrift migration of the remaining eastern inlet occurred (period 1850-1950; Figs A9-A13).

3) Western end of Ameland

The western end of Ameland remained fairly stable until the first half of the 19th century (Figs A1-A13). The reconstructions suggest minor westward growth and subsequent eastward retreat of the HW-line, but a stronger growth and retreat of the LW-line. After circa 1850 (Figs A10-A13) an eastward retreat was observed of 0.9 km at the NW side, due to erosion by marginal channels. After ca. 1988 strong erosion occurs at the NW side. The SW side retreated over 2 km due to the eastward shifting inlet.

4) Seaward side of Ameland

At the seaward side of Ameland the LW-line likely prograded seaward in the periods 1545-1665 and 1731-1809 (central part of the island). The LW-line was fairly stable during the period 1665-1731 and between 1809-1850 (Figs A3-A9). Around 1850 the coastline was positioned on average 0.5 km seaward from the present day LW-line (Figs A9-A13). Unfortunately, the HW-line at the North Sea coast of the islands is not indicated on maps prior to 1809. If the position of the HW-line is supposed to have remained at a fixed distance from the LW-line, the HW-line would have followed the LW-line. If the position of the HW-line is supposed to have been as close to the dunes as possible, the maps even suggest a continuous seaward progradation of the HW-line in the period 1536/45 to circa 1809. From circa 1850 onwards, the HW-line and LW-line shifted landwards. The process has now been stopped by beach nourishment.

5) Landward side of Ameland

At the landward side Ameland has lost a considerable area of land (Figs A1-A13). Erosion occurred in the south-central part of the island until 1843. Originally around 1545 it extended, at maximum, almost 2 km more to the S than in 1975. A large outer bend of a backbarrier channel gradually eroded the south-central part of the island, until the channel was artificially blocked. Erosion also occurred at the SE part (near Oerd), which, originally in 1545 extended almost 1 km further to the S than in 1975. South of Oerd, repeated erosion occurred by E-W trending laterally migrating backbarrier channels of the Pinkegat Inlet system. S of Ameland no substantial erosion has been observed between 1731 and 1865, except close to channels.

6) Eastern end of Ameland

The eastern end of Ameland retreated westward and shifted eastward several times in the last five centuries (Figs A3-A13). Net eastward migration occurred especially after 1800.

7) Pinkegat

Before 1600 A.D., the Pinkegat was smaller than at present (Figs A1-A4). The scarce, available data suggest that the inlet developed cyclically from a single inlet into a multiple inlet, and back again since at least 1600 (Figs A4-A13). The cyclic development of the inlet can be observed in great detail for the period after 1800 (chapter 3). During eastward shift of the westernmost inlet, Ameland has grown for several decennia with velocities of 100 $m.yr^{-1}$ for the HW-line, and even the double amount if the LW-line is taken into account. After 1800, a net downdrift, eastward shift of the Pinkegat Inlet system, and an eastward shift of the watershed S of Ameland took place.

8) Engelsmanplaat

Both the early Holocene clay core and the historical observations indicate that Engelsmanplaat has been a high since long. Throughout the last five centuries, large supratidal shoals have been built up at the N-side of the shoal, by waves and flood currents (Figs A1-A13; for details see chapter 3). This led to a cyclic pattern: The Engelsmanplaat increases in surface area and height as soon as a shoal in the north amalgamates with it, thereby closing the crossing channel. Subsequently, the enlarged Engelsmanplaat is eroded at its northern side by the formation of a channel and becomes smaller. North of the channel a new shoal is formed, and the Engelsmanplaat decreases in size, and becomes lower (see also De Haan et al., 1983). Especially after 1800, a strong erosion is observed at the western side of the shoal, due to the net downdrift shift of the Pinkegat Inlet system (Figs A8-A13).

9) Zoutkamperlaag

From the historical data it is inferred that around 1300 the Zoutkamperlaag Inlet system was rather small, and had no or only a minor connection to the Lauwerszee (Fig. A1). After 1300 the Zoutkamperlaag started to take over the drainage of the Lauwerszee from the Lauwers Inlet. Around 1500 both inlets together took care of the drainage (Fig. A2). By 1550 the Lauwerszee was drained mainly by the Zoutkamperlaag (Fig. A3). Inlet channels and separate outer channels shifted downdrift and eroded the coast of Schiermonnikoog (Figs A3-A13). Strong changes also occurred after closure of the Lauwerszee in 1969 (chapter 5).

10) Western end of Schiermonnikoog

The western end of Schiermonnikoog has been eroded considerably. Around 1550 the position of the HW-line was at least some 2.5 km more to the W than at present (Fig. A3). Erosion occurred at the NW-side in the period between circa 1530 (or earlier) to circa 1600 (Figs A2-A4). The erosion, enhanced by the storm surges of that time, must have been caused by a newly formed, easternmost outer channel of the Zoutkamperlaag system. Channel erosion most likely continued at the NW side, during the abandonment of this new channel (Fig. A5). After the abandonment a large shoal merged with the island (Fig. A6). The erosion was followed (from about 1700 onwards) and/or accompanied by eastward dune migration. As a result, the original church had to be abandoned in 1715, and a new one was built somewhat to the SW. In the period of 1700-1835 the SW side of the island was eroded due to backbarrier channel erosion over some 4 km (LW-line Figs A6-A9). Eastward migration of the outer channel between 1835-1929 caused erosion of the NW side of the island over a length of about 2 km. After the abandonment of the channel De Noorman and of the eastern outer channel connected to De Noorman, a substantial southwestward growth of the island occurred (Figs A10-A13). The shoal thus formed is being eroded at present. A huge bar became attached to the island at the NW side around 1970. In the lee of the recurved bar, the North Sea beach of NW Schiermonnikoog gradually expanded seaward. A part of the sediment (mainly sand) has been derived from the ebb-tidal delta, which became smaller after, and due to, the closure of the Lauwerszee.

11) Landward side of Schiermonnikoog

The southern side of Schiermonnikoog was eroded since at least 1550. This occurred in particular after 1750. The erosion is mainly due to erosion by laterally migrating channels. After 1865 the supratidal area S of the dyke, built in 1860, was subject to erosion.

12) Eastern end of Schiermonnikoog

The HW-line at the eastern end of Schiermonnikoog migrated more than 6 km eastward in the period 1550-1975 (Figs A3-A13). Migration was partly due to repeated amalgamation of (parts of) the shoal Simens Sant and its successors, and partly due to lateral accretion.

13) Simens Sant

The shoal Simens Sant and its successor(s) were formed by wave action and flood currents at the border of the North Sea and the Wadden Sea. These shoals were supratidal most of the time and shifted downdrift in front of the eastern end of Schiermonnikoog. They became attached to Schiermonnikoog when the channel in between became was filled and abandoned; for instance around 1600 (Fig. A4) and temporarily around 1848 (Fig. A10).

14) Lauwers Inlet

It is inferred that around 1300 the Lauwers Inlet drained the Lauwerszee (almost) solely, and had a more westerly position than at present (Fig. A1). After the take-over of the drainage of the Lauwerszee by the Zoutkamperlaag Inlet system (1300-1550; Figs A1-A3), the Lauwers Inlet became gradually detached from the Lauwerszee embayment (1500-1550; Figs A2-A4). It decreased in size to such an extent, that it was reported that the inlet gorge had almost totally disappeared in the first half of 17th century (Figs A4-A5). Then it shifted downdrift, and its size again increased, due to expansion of its drainage area to the E. The inlet gorge shifted considerably to the E (10.5 km in the period 1600-1975; Figs A4-A13).

15) Bosch

Originally, the island Bosch, between Schiermonnikoog and Rottumeroog, must have been a fairly large island of at least several km length (Figs A1-A3). In 1541 sailing directions mentioned that steep high dunes occurred at its western end. This indicates that the dunes and the western end of the island were already being eroded at that time. Historical evidence suggests that the dunes of the island were severely or totally eroded during the storm surge of 1570. The stormy climate of the 16th century prevented the re-establishment of a new dune area. Attempts to establish a new bent-grass vegetation failed, and the island became a sandy supratidal shoal between circa 1600 and 1642 (Figs A3-A5). The shoal migrated to the SE into the backbarrier area between 1642 and 1710 (Figs A5-A6). In this period the Lauwers Inlet increased in size again. North of the migrating shoal Bosch, the supratidal shoal Koeplaat built up between 1642-1710 (Figs A5-A6), comparable to the formation of the present-day shoals in front of Engelsmanplaat by wave and flood processes. Between 1792 and 1811 the Kapersplaat (from the W), the Koeplaat and the remains of the shoal Bosch amalgamated into a new shoal Boschplaat (Fig. A8). The available maps suggest a large supratidal shoal, which was positioned far into the Wadden Sea by 1832/34 (Fig. A9). Afterwards the shoal decreased in size, and migrated probably seaward (Fig. A10). After 1872/74 it migrated to the ESE (with speeds of up to 40 $m.yr^{-1}$) and merged with Rottumerplaat by 1959 (Figs A10-A13).

16) Rottumerplaat

Rottumerplaat has existed as an intertidal shoal since about 1825 and became supratidal probably around 1860. Dunes were present in 1872; permanent dunes became established between 1904-1911. The island has grown since at least 1873/74 (Figs A10-A13), and expanded with rates of up to 127 $m.yr^{-1}$ to the west until about 1950, when erosion started at the western side. The island continuously increased in surface area since 1875, or earlier.

17) Schild Inlet

The Schild Inlet, E of Rottumerplaat, has shifted some 11 km to the E between 1550-1976 (at an average velocity of 25 $m.yr^{-1}$; Figs A3-A13). Over shorter spans of time, migration velocities of up to 40 $m.yr^{-1}$ were observed at the eastern side. At the western side of the inlet it was up to 120 $m.yr^{-1}$. At present, the Schild Inlet system is losing drainage area to the Lauwers Inlet, and the dimensions of its ebb-tidal delta and cross-sectional area decrease.

18) Rottumeroog

The shift of the island Rottumeroog is one of the most dramatic. Between 1500 and 1550 the island has most likely been eroded at both its eastern and western end (Figs A2-A3).

Around 1550 the western side was almost 10.5 km further to the west than in 1976. Especially after 1650 the island shifted downdrift (to the east) and became smaller (Figs A5-A13).

19) Heffesant and Cornasant

The backbarrier islands Heffesant and Cornasant were eroded after 1550. Heffesant disappeared probably around 1600 (Figs A2-A3). A part of Cornasant merged with Bosch and its other part formed a supra- to intertidal shoal.

20) Reclamations

Large parts of the backbarrier area of the Frisian Inlet System have been reclaimed and dyked after 1300 (Figs A1-A13). At first, reclamation went rapidly, but it decelerated towards the 19th century. The reclamations led to a considerable decrease in tidal prism of the backbarrier basin.

DEVELOPMENT IN HISTORICAL TIMES: DISCUSSION

The main trends in the development of the eastern Dutch Wadden Sea in historical times and variables involved are:

1) A net downdrift shift of the inlets and the barrier islands. These shifts are controlled by tidal currents, waves, and changes in basin configuration,

2) Coastward shift of the barrier islands and the formation and closure of tidal embayments. These shifts are controlled by sea-level movements, tidal wave distortion, sediment supply, wave-attack and man.

Downdrift shift of the inlets and the barrier islands

From west to east, the eastern Dutch Wadden Sea is here divided in an area from the barrier island Terschelling to the shoal Engelsmanplaat, and another area from the Zoutkamperlaag Inlet to the Western Ems Estuary.

Terschelling to Engelsmanplaat

In the 10th century, the Middelzee Estuary was a 30 km long funnel-shaped embayment. The Middelzee drained through the Ameland Inlet, which at that time consisted of two channels (Van der Spek, 1994). Between the 10th and 13th century the Middelzee embayment had been largely filled up (Van der Spek, 1994), so that only a small bay remained around 1500. The infill of the Middelzee led to a strong change in the configuration of the drainage area of Ameland Inlet, from a N-S elongated basin to a rectangle-shaped basin, which was oriented in a W-E direction (Van der Spek, 1994). The change of basin configuration must have resulted in hydrographical changes. Elongate, funnel-shaped basins often have separate flood- and ebb-dominated channels. In some cases, such as the Ems Estuary, ebb and flood channels may even flow along either side of the island (Borkum). When the Middelzee estuary thus altered into a more lagoonal configuration, where ebb and flood currents used the same channels, the updrift inlet channel of Ameland Inlet was abandoned. This occurred at a slow pace (from 1500 to 1800), because abandonment was retarded by the temporary expansion of the drainage area to the southwest (Van der Spek, 1994). During closure of the westernmost channel, the eastern end of Terschelling shifted eastward over 10.5 km.

The drainage basin of the remaining inlet of the Ameland Inlet system was mainly positioned S of the island downdrift of it (Ameland). The downdrift shift of the eastern inlet resulted in the erosion of the western end of Ameland after 1800, especially during periods in which the main inlet channel was positioned to the NE. Due to the erosion, the drainage of the eastern backbarrier area became swifter and the drainage area of Ameland Inlet expanded eastward. The NW side of Ameland was eroded by marginal channels. The erosion at the western end of Ameland was compensated by sedimentation at the eastern end. The Pinkegat system, east of Ameland, migrated eastward after 1825. This eastward shift resulted in erosion of the Engelsmanplaat.

Zoutkamperlaag to Western Ems Estuary

Drainage of the Lauwerszee embayment

Around 1300 the Lauwerszee embayment is inferred to have been (mainly) drained by the Lauwers Inlet. When the drainage of the Lauwers Inlet was taken over by the Zoutkamperlaag, a chain-reaction set off: the Zoutkamperlaag system increased in dimensions. The western end of Schiermonnikoog was eroded by laterally migrating outer channels and backbarrier channels. At the downdrift, eastern end Schiermonnikoog grew by lateral accretion (downdrift sediment supply) and amalgamation of shoals, such as Simenssant. In the same period huge parts of the Lauwerszee embayment were reclaimed. In analogy to the recent closure of the Lauwerszee in 1969, the tidal wave could travel faster to the watershed, and the watershed shifted to the east.

Lauwers Inlet and Eilanderbalg

The transition of the drainage of the Lauwerszee, followed by the eastward expansion of the Zoutkamperlaag system, resulted in a decrease of the drainage basin of the Lauwers Inlet. At first, the Lauwers Inlet system decreased dramatically in size, and the inlet filled up strongly, so that around 1640 it had been almost abandoned. Due to the decrease of tidal forces through the inlet, downdrift sediment supply could result in a more effective sedimentary infill of the inlet. The smaller drainage volume must, moreover, have resulted in a decreasing size of the ebb-tidal delta (cf. Dean & Walton, 1975; Steijn, 1991). Erosion started to exceed sedimentation on the downdrift island Bosch, as evidenced by the decay of the island after at least 1500. The net erosion in combination with the frequent storm surges of the 16th century, of which the surge of 1570 was one of the largest in known history, destroyed the dunes[75] of Bosch. The shoal shifted landward, due to the wave- and flood-induced sediment transport, comparable to the landward shift of the shoals N of Engelsmanplaat (see summary historical changes, point 8). Within the backbarrier area the prevailing western winds forced the shoal to shift to the E. The remaining drainage area of the Lauwers Inlet was mainly east of the inlet. Moreover, the influence of waves and longshore sediment supply increased relative to the decreasing tidal prism of the Lauwers Inlet (cf. Sha, 1990). As a result the inlet shifted downdrift and thus provided a faster connection to the eastern backbarrier area than through the Schild Inlet system, downdrift of it, and thus the Lauwers Inlet partly took over the drainage area from that inlet[76]. This, in combination with the wind-driven movement, resulted in a SE-directed shift of the shoal Bosch into the backbarrier area. From circa 1640 onwards the tidal volume of the Lauwers increased. In 1710 large new shoals (Koeplaat) were present downdrift of the ebb-tidal delta of the Lauwers Inlet.

After 1700, and more especially after circa 1750, the Eilanderbalg Inlet formed through the shoals E of Schiermonnikoog, and drained the backbarrier area SE of Schiermonnikoog more efficient than did the Lauwers. At the same time a supratidal shoal (Simonszand) was formed downdrift of the channel. During a period of abandonment of the Eilanderbalg from circa 1811 to 1848, the shoal Simonszand became intertidal, and migrated into the backbarrier. Afterwards the drainage volume and the ebb-tidal delta of the Eilanderbalg increased again, and the shoal Simonszand became larger and migrated seaward. These developments

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[75] During storm surges the water height along a dune row increases and the coastal profile tends to adapt to the prevailing water level and wave dynamics. The dunes are eroded to an almost vertical profile. This can be further accentuated by aeolian erosion. Normally, the lowest, seaward parts of the dune have a low angle and can thus be covered by *Elymus farctus*, which has a shallow root system and can withstand flooding by salt water (Van Oosten, 1986). The higher parts of the dunes are covered by, amongst others, *Ammophila arenaria*, which avoids salt-rich environments and *Elymus arenarius* (Van Oosten, 1986). These plants with deep vertically penetrating roots form a dense mat and enable the formation of a relatively steep dune profile, because the plants can keep up vertically with the sedimentation. At first an erosion profile formed during a storm will become more outspoken, because the lower part is too steep for *Elymus farctus* and too salty for *Ammophila arenaria*, and thus remains uncovered, whereas vertical accretion proceeds in the higher parts (Bleuten, 1971). This leads to the typical '*steep high dunes*', which are so often mentioned in the sailing directions of the 16th century. The lower part of the steep erosion profile can only be restored if sand is provided, to allow the formation of a low angle profile and the growth of *Elymus farctus* (Bleuten, 1971). If sand supply is insufficient, the losses during storm surges cannot be restored, and the dunes will finally be completely eroded.

[76] It should be noted that an increase in downdrift drainage area may result in a more updrift orientation of the inlet and ebb-tidal delta with increasing tidal volume. This happened with the Lauwers Inlet between 1750 and 1800 and between 1888/91 and 1980 (especially after 1954).

are quite comparable to the changes during the abandonment and the re-opening of the Lauwers Inlet around 1640 (see above).

The formation of the Eilanderbalg (1708/10-1750) resulted in a decrease in dimensions of the western part of the drainage area of the Lauwers Inlet. In combination with an eastward shift of its gorge, the Lauwers Inlet could more then compensate the loss by increasing its drainage area to the east. Due to its large tidal prism, the Lauwers Inlet became increasingly more updrift-oriented (cf. Sha, 1989b). The temporal westward expansion of the Lauwers Inlet, in combination with the eastward shift of the watershed S of Schiermonnikoog, will have enhanced the temporal abandonment of the Eilanderbalg (1811-1848). After circa 1848 the Eilanderbalg increased again in size, most likely due to the renewed downdrift shift of the Lauwers Inlet. That the Lauwers system continues to expand to the east is most likely because it has approached the Schild Inlet so closely that one inlet suffices to drain the area (cf. Fig. 6, chapter 1; Wolff, 1986).

Schild Inlet

The eastward shift of the Lauwers Inlet system after 1650 resulted in an irregular decrease of the tidal volume of the Schild Inlet and of its ebb-tidal delta volume. The high sediment demand of the large Ems Estuary[77] which is filled in (chapter 1) and the downdrift shift and infill of the Western Ems, especially after circa 1860[78] has probably played and still plays an important role in the sedimentary development of the coast. At present the outer channels and ebb-tidal delta of the Western Ems Estuary still reach up to East-Schiermonnikoog, that is 26 km W from the inlet gorge. The ebb-tidal delta of the Lauwers Inlet and especially that of the Schild Inlet directly border these outer channels, and thus loose sand to them (cf. Van de Kreeke & Robaczewska, 1993). The sediment is not available any more to the downdrift islands. Especially after 1650 the erosion of Rottumeroog proceeded at a high velocity. This coincided with the observed eastward shift of the Lauwers Inlet system during which the dimensions of the Schild Inlet were repeatedly reduced.

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[77] In addition to the natural net infill of the Dollard a considerable volume ($40*10^6$ m^3, Gussinklo pers. comm., NAM) of sand, and, in the landward part of the Dollard, clay is needed to compensate the subsidence by gas extraction from the giant Slochteren Field in the coming centuries (cf. Dijkema & Oost, 1993). Moreover, the removal of dredged sediment from the estuary generates an additional sediment demand.

[78] For the period 1580-1650 the downdrift shift of the Western Ems is mainly due to the formation of a shoal at its western side (Lang, 1958). In the period after 1800 the downdrift shift of old outer channels and the updrift opening of new ones, in combination with inertia effects, provided space for the island Rottumeroog to shift downdrift. The net downdrift shift of the Western Ems can probably be attributed to the decreasing tidal volume of the Ems-Dollard Estuary after circa 1600. Furthermore, the increase in drainage by the Eastern Ems (18th and early 19th century) will have further reduced the tidal flow through the Western Ems. At the moment the Western Ems cannot shift any more to the east, because the downdrift island Borkum has been artificially protected against erosion.

From 1873/74 onwards the annexing of the drainage area of the Schild Inlet by the Lauwers Inlet can be traced quite well. From the observations it follows that such a take-over is a gradual process, which takes several centuries. The Lauwers Inlet system increased in size during the process, and shoals built up and grew at its downdrift side (especially Rottumerplaat; Brilhuis et al., 1991). The Schild Inlet became smaller, and its gorge shifted downdrift, due to a relative increase in wave force (cf. Sha, 1989b; see below). An additional effect might be the development of shoals at the updrift side, enhanced by the growing ebb-tidal delta of the updrift Lauwers Inlet. At present the eastward shift of Schild Inlet cannot compensate the reduction in drainage volume by gaining drainage area to the east. This is due to the presence of the fixed Western Ems, and the Schild Inlet thus continues to be reduced in size. Also the island Rottumeroog is still decreasing in size. The islands Rottumerplaat and Rottumeroog are both too small for the prevailing tidal range, in which islands of some 13-14 km length are to be expected (cf. Wolff, 1986; chapter 1). The logical future development thus will be that finally the Schild Inlet will close, or merge with the Lauwers Inlet or the Western Ems. Thus, Rottumerplaat and Rottumeroog may merge or Rottumerplaat may form a new, large island (Brilhuis et al., 1990).

Coastward shift of the islands and the formation and closure of tidal embayments

As described above, the Eastern Dutch barrier islands have shifted eastward quite strongly. Only the barrier island Ameland remained largely in its original position. At its seaward side, the LW-line of Ameland likely prograded seaward in the periods 1545-1665 and 1731-1809 (central part of the island), and it was fairly stable between 1665-1731 and between 1809-1850. After 1850 it retreated. As stated, the islands Schiermonnikoog and Rottumeroog have shifted downdrift strongly, and therefore their evolution perpendicular to the coast cannot be followed in detail. Comparison of the reconstructions, however, suggests that their seaward side (especially of Schiermonnikoog) remained more or less on the same position.

The most impressive change on the mainland was the formation and closure of the embayments. The Lauwerszee embayment expanded strongly southward during the 8-9th century, and probably increased its maximal E-W dimensions until the 13th century. Afterwards the Lauwerszee embayment was gradually reclaimed and the remaining embayment was closed off by a dyke in 1969. The Middelzee was formed due to the flooding of the Boorne valley in the 9th-10th century. It was filled in from circa 1000 A.D. onwards.

Mechanisms related to the downdrift shift of the inlets and the barrier islands

Small-scale and large-scale developments are observed during historical times. The small-scale developments concern individual inlet systems. They are superimposed on, and form part of the large-scale developments. The latter are the general trends of development of several adjacent basins.

Small-scale developments

From the observations of the historical development and from the literature it appears that the following variables are of influence on the small-scale development of the Dutch barrier coast: 1) residual sediment transport, 2) channel shifts, 3) amalgamation of swashbars, 4) the influence of the ebb-tidal delta, and the response to these variables by 5) the backbarrier channels, 6) the updrift and downdrift ends of the islands and 7) the watershed.

1) Residual sediment transport

In the North Sea, ebb and flood currents are of roughly the same strength, if the influence of winds and limited water depths are excluded. Along the coast the tidal-current-velocity vectors at one point form an almost symmetrical ellipse. Thus, these currents generate only a limited residual water transport (of the order of $cm.s^{-1}$), directed mainly to the E (Otto et al., 1990; Steijn & Louters, 1992; Steijn et al., 1992). Along the higher part of the shore the tidal current becomes asymmetrically directed to the E (Van Straaten, pers. comm.). A residual flood dominance occurs generally along the North Sea coast at either side of the inlets along the island tips (Smith, 1984; Steijn, 1991; cf. Steijn et al., 1992). In addition, sediment is transported by wave action and wind. Average wind speed along the coast was 6.5-7 $m.s^{-1}$ over the period 1931-1960, with winds blowing dominantly from the SW (Vroom et al., 1989) and during storms especially from the NW. Commonly, waves come in from WNW to NE sector, and high waves from the NW (Steijn et al., 1992). In such cases the waves generate a strong littoral, eastward drift (cf. Steijn & Louters, 1992; Steijn et al., 1992). In addition waves generate a wave-induced suspension of sediment, due to which sand transport by tides can increase by an order of magnitude (Steijn et al., 1992). Wind shear is especially important on the beaches, and causes a substantial aeolian sand transport. Apart from duneward transport, also longshore aeolian transport (net towards the E) takes place (Eysink, 1979; Steijn, 1991). All these mechanisms result in a residual sediment transport to the E. As a result of the residual sediment transport rapid expansion of the downdrift (eastern) ends of the islands can be observed from time to time, for instance Terschelling (1550-1800), Ameland (1850-1976), Schiermonnikoog (1550-1976), Rottumerplaat (after 1950).

2) Channel shifts

Tidal forces tend to keep the outer channels and the main inlet of the Wadden Sea Inlets in an updrift orientation (Sha, 1989b, 1990; Sha & Van den Berg, 1992). However, downdrift sediment supply by waves and tides, lateral accretion of channels and, in some cases, the inertia of the outflowing ebb current, tend to force the outer channels and the inlet gorge into a downdrift orientation (see Chapter 3 and 4). For all tidal inlets the position and orientation of the main channels and the related ebb-tidal delta is determined strongly by the balance between tidal energy and wave energy: dominance of the tide results in an orientation into the updrift direction, whereas wave dominance results in a orientation into the downdrift direction (FitzGerald et al., 1984; FitzGerald, 1988; Sha, 1989b)[79].

The continuous rivalry between these processes often leads to an irregular downdrift shift of the outer channels and sometimes of inlets (Joustra, 1971; Sha, 1990; chapters 3 to 5). If channels shift downdrift, the updrift island can expand downdrift as well (i.e., to the E). At the downdrift side of the ebb-tidal delta marginal flood channels can be generated from time to time. Finally, channels at the downdrift side become hydraulically inefficient for the drainage of the related backbarrier area and are abandoned. During the final stage of abandonment the outer channel becomes quite narrow. Observations on Texel, Ameland (after 1800), Schiermonnikoog (1650, 1843-54), and Rottumeroog (1800-1976) show that this may result in powerful currents and related strong erosion of the northwestern[80] sides of the islands. Prior to and during abandonment one or several new channels are generated at a more updrift position (chapters 3 and 4). This causes erosion of the downdrift end of the island (updrift of the inlet; chapter 3).

3) Amalgamation of shoals and bars

If a most downdrift channel on an ebb-tidal delta becomes abandoned, the shoal updrift of it merges with the downdrift island, as is, for instance illustrated in the developments at Schiermonnikoog around 1700 (Figs A5 and A6) and around 1970. As a result of the merger, the island grows at its updrift side. The amount of growth strongly depends on the size of the attaching shoal. The size of the shoal in its turn, depends on the size of the ebb-tidal delta, and thus on the tidal prism. For example, the Noorderhaaks shoal (above -5 m) in Texel Inlet has a greater volume than the complete ebb-tidal delta of the Pinkegat.

Besides the larger bars, also smaller swashbars merge with the island, if this is not hindered by large channels (FitzGerald et al., 1984; Steijn, 1991). The zone where the bars merge (here called bar-amalgamation zone) with the island is determined by the position of the contact zone of the ebb-tidal delta with the shoreline of the downdrift island. For the East Frisian Islands FitzGerald et al. (1984) found that the bar-amalgamation zone is positioned more eastward if the main inlet channel has an easterly position, or if the ebb-tidal delta is large. The bar-amalgamation zone is the zone with net sedimentation (FitzGerald et al., 1984). As discussed above (point 2), the main inlets and the related ebb-tidal deltas of the Frisian barrier islands are oriented in an more increasingly updrift direction with increasing tidal prism. The tendency is an increasingly updrift position of the bar-amalgamation zone. This is counteracted by the increase in size of the ebb-tidal delta with increasing tidal prism, which promotes a more downdrift position of the bar-amalgamation zone. Thus,

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[79] This implies that, upon decreasing tidal volume, ebb-tidal deltas and their main channels will become more dominated by wave forces, and oriented in a more downdrift direction.

[80] For Texel the SW side.

within relatively broad constraints, the position of the bar-amalgamation zone will be rather stable and of little influence on the migration of the island. Only for relatively small islands with a small, downdrift asymmetric tidal delta at their updrift side the bars might merge at the most downdrift part of the island and thus result in erosion of the updrift part of the island.

The attachment of smaller shoals to an island may not be enough to compensate the erosion caused by the channels in front of the shoals. In case of the attachment of a larger shoal, the volume of sand exceeds the amount of sand eroded by the flood-dominated channel separating such a shoal from the island (cf. Texel; Sha, 1990).

4) Influence of the ebb-tidal delta

Ebb-tidal deltas increase in size with increasing tidal prism (Dean & Walton, 1975; Dean, 1988). They form seaward protrusions, and thus protect the downdrift island to some extent from waves coming in from the W-NW sector. The seaward protrusion deflects incoming waves at the downdrift side of the ebb-tidal delta. The resulting, updrift residual current transports material to the updrift side of the downdrift island (Hayes & Kana, 1976; FitzGerald et al., 1984; Sha, 1990; Steijn et al., 1992). The amount of protection and the strength of the residual updrift current increase with increasing size of the ebb delta. In several cases a clear decrease of the size of the downdrift island can be observed in the development in historical times, after a reduction of the size of the ebb-tidal delta (e.g., Bosch, 1500-1650; Rottumeroog after 1650, and in more detail after 1800 (Brilhuis et al., 1991); Simonszand 1811-1850). Reversely, a growth of islands occurs after an increase of the updrift ebb-tidal delta (e.g., formation of Simonszand after 1700, and again after 1850; formation of the new shoal of Bosch and the Rottumerplaat). All implies that net erosion of the downdrift island increases if the ebb-tidal delta decreases in size. This is at least partly to be attributed to the decreasing shelter provided by the ebb-tidal delta, and due to a decreasing deflection of waves at the downdrift side. Moreover, smaller ebb-tidal deltas will, because of their limited seaward protrusion, not be able to 'catch' as much downdrift migrating sand (for instance during northwestern storms; cf. Steijn et al., 1992) as do larger ebb-tidal deltas. Hence, the larger the ebb-tidal delta, the more it will contribute to the maintenance of the downdrift island.

5) Response of backbarrier channels

Where downdrift migrating outer channels or inlets drain backbarrier channels directly S of the island, the downdrift shift of the outer channel or inlet results in a migration of the backbarrier channels. The latter tend to be located and orientated in such a way that they form the shortest connection between the inlet and the drainage area. In the Wadden Sea, erosion at the SW part of a barrier island may result from such an orientation. This is dramatically illustrated by the strong erosion of SW Schiermonnikoog during the period 1736-1843. Erosion, caused by the migration of the backbarrier channel due to the gradual shift of the complete inlet is illustrated by the erosion at the SW end of Ameland in the period 1854-1949.

6) Response of the updrift and downdrift ends of the islands

The dynamics of erosion and sedimentation differ for the updrift and downdrift ends of the islands. The West Frisian barrier islands typically have a drumstick form, with the updrift end being much broader than the downdrift end. It is rare that one channel erodes the full updrift end of an island simultaneously. Mostly only the NW side or the SW side is eroded, which results in a slow downdrift erosion of the western part of the island (e.g., Ameland and Schiermonnikoog). Moreover, most larger West Frisian tidal inlets are oriented to the updrift side of the ebb delta, or at least the larger part of their outer channels is. This implies that channels at the seaward side are mostly more influential at the downdrift side of the barrier (= the updrift side of the inlet) then at the updrift side. At the updrift side of the barrier island commonly only relatively small marginal flood-channels occur.

At the downdrift side of the barrier islands small inlets instead of large inlets are sometimes present. The small inlets often curve tightly around the island (e.g., Pinkegat during single inlet configurations and Eijlander Balg). The downdrift end of a barrier island is mostly relatively sharp-angled, so that erosion or sedimentation may occur over the full length and can be generated by the activity of one channel simultaneously. The above discussed strong longshore sediment transport is directed towards the downdrift end of the island, which thus can grow rapidly by spit accretion (FitzGerald et al., 1984; Steijn et al., 1992; chapter 4).

Due to the differences between the updrift and downdrift ends, the downdrift end of a barrier island is much more dynamical than the updrift end. This is, for instance, illustrated by the strong dynamics of the eastern end of Terschelling, Schiermonnikoog and especially Ameland between 1800-1976, which all show a repeated rapid growth and erosion. The net longterm shift of both ends of barrier islands, however, is normally more or less similar. This fits to the observation that the length of barrier islands is correlated with the tidal range (chapter 1).

7) Response of the watershed

In the Wadden Sea, sediment transport in the backbarrier area is also largely determined by the combined forces of tides and winds. The tidal wave enters the inlet at the updrift side of the barrier island earlier than the inlet at the downdrift side. Also, the wind generates a current which is dominantly directed to the E (De Boer, 1979; De Boer et al., 1991a, cf. Hayes, 1979). As a result the morphological watershed occurs S of the island at circa 2/3 from the updrift end. Whenever a phase shift occurs between the tidal waves from either side (due to a change of the distance to the inlet, or to the propagation velocity of the tidal wave, etc.) the watershed reacts and shifts. This implies that an eastward shift of an inlet results in an eastward shift of the watershed and a reduction of the drainage area of the inlet east of it (e.g., Ameland and Pinkegat Inlet after 1800; Lauwers and Schild Inlet after 1650). Such a reduction may be compensated by an expansion of the drainage area into the downdrift direction.

Large-scale developments

From the observed developments it appears that the developments of adjacent inlet systems (each consisting of an ebb-tidal delta, an inlet, the adjacent island ends and the drainage basin), and the barrier islands between them are closely linked. This can be clearly demonstrated for the Ameland and Pinkegat Inlet, and for Zoutkamperlaag to Schild Inlet.

The closure of the Middelzee led to a change from an estuarine to a lagoonal configuration and to the abandonment of the westernmost of both inlets. The remaining drainage basin of Ameland Inlet was located mainly S of the downdrift island Ameland. During the repeated downdrift shift of the remaining inlet a shorter way to the drainage basin was formed by erosion of Ameland. The short-cut resulted in the eastward shift of the watershed, thus reducing the westward extent of the drainage area of the Pinkegat system. The cyclic shift of the inlet channels of the Pinkegat, in combination with the expansion of the drainage basin of Ameland Inlet resulted in a net downdrift shift of the whole system in the 19th and 20th century. During and by these shifts the shoal Engelsmanplaat, E of the Pinkegat, is being eroded. If the process continues it is to be expected that the Engelsmanplaat will be fully eroded, and that the Pinkegat Inlet and the Zoutkamperlaag will merge in the near future. This is also suggested by model calculations, which show that the present double inlet situation (Pinkegat and Zoutkamperlaag) in the Frisian Inlet is not naturally stable (Wang, 1991). Then, the drainage area of the combined Pinkegat-Zoutkamperlaag Inlet will be large, and the inlet will become more updrift-oriented again.

The changes in the configuration of the drainage area of the Zoutkamperlaag and the Lauwers Inlet resulted in a chain reaction (a knock-on effect) which continues to the present day. Upon reduction of the backbarrier area of the Lauwers Inlet by the take-over of the drainage of the Lauwerszee embayment by the Zoutkamperlaag Inlet, the Lauwers Inlet decreased in dimensions and became almost abandoned around 1650. As a result also the ebb-tidal delta must have decreased in dimensions and the island Bosch, downdrift of it, could not be maintained any more. During the downdrift shift of the Lauwers Inlet its drainage area expanded in the same direction. Due to the increase the inlet and ebb-tidal delta became larger again (1700) and new, large shoals were formed E of it. After a short interruption (1745-1811) the Lauwers drainage system continued its eastward expansion, and also the Lauwers Inlet shifted eastward. With the eastward expansion of the drainage area of the Lauwers Inlet a new large shoal Rottumerplaat was built up (after 1800) at the eastern side of the inlet. The Schild Inlet lost a considerable part of its drainage basin to the Lauwers system. As a result the ebb-tidal delta of Schild Inlet decreased in size and Rottumeroog could no longer be maintained and was eroded. The Schild Inlet became smaller and will likely disappear in the near future (cf. Brilhuis et al., 1991).

The mutual dependence of the development of adjacent inlet systems and the islands between them is also clearly demonstrated by the developments after the closure of the Zuider Zee and of the Lauwerszee embayment (chapters 1 and 5). After the closure of the Zuider Zee, Texel Inlet expanded to the east and changed the configuration of Vlieland Inlet and Terschelling Inlet (Klok & Schalkers, 1980). After the closure of the Lauwerszee embayment the Zoutkamperlaag drainage basin expanded eastward and (temporarily) took over part of the drainage basin of the Eilanderbalg.

Mechanisms related to the coastward shift of the islands and the formation and closure of tidal embayments: variables

The morphological development of the barrier islands and the backbarrier areas depends on the balance between variables, such as 1) climate, 2) relative sea level, 3) storm-surge frequency, tidal range and the related tidal prism. To a large extent these parameters are mutually dependent. Climate change may lead to temperature-induced and wind-induced sea-level rise. Sea-level rise may increase the tidal drainage area if the mainland coast is low or consists of easily erodible peats (as in e.g., the Jade Bay in the German Wadden Sea), and thus also tidal prism. Climate change can be considered to be a central 'motor' driving other changes. Another factor which has become increasingly important is man.

Climate

Slight changes in average global temperature (of the order of a degree) may significantly influence the climate conditions (Lamb, 1977; Buisman, 1984; Warrick & Oerlemans, 1990). Since 1706, detailed instrumental meteorological observations have been made on various places in The Netherlands (1706-1734, Delft; 1735-1849, Zwanenburg; 1848-1899, Utrecht; 1899 onwards, De Bilt; Labrijn, 1945). These data were all reduced to station De Bilt to produce a homogeneous sequence (Labrijn, 1945, revised 1991). Based on the sequence IJnsen (1974, 1988, 1991a,b) characterized the winter period (November-March) with a winter index C_v and the summer period (May-September) with a summer index C_s (Table II). Increasingly higher numbers (1 to 9) signify an increasingly less maritime character of the climate, whereas low numbers indicate a more maritime character. A less maritime climate is characterized by hot summers and cold winters, whereas a maritime climate is characterized by cool summers and soft winters. In general storm surges occur more often during soft, more maritime winters than during cold, less maritime winters (see below).

The climate over the year can be characterized by the so-called maritimity of the weather. The maritimity can be expressed by the so-called thermic maritimity index (I_m; IJnsen, 1994):

$$I_m = 20 - (C_v + C_s)$$

Table II: Winter indices, based on the winter temperatures, reduced to De Bilt, of 1707-1990 (IJnsen, 1991a), and the summer indices based on the summer temperature sequence reduced to De Bilt 1706-1990 (IJnsen, 1991b).

Winter index C_v or summer index C_s	Description winter index	Description summer index	Theoretical frequency of occurrence
1	extremely soft	extremely cool	1%
2	very soft	very cool	3.8%
3	soft	cool	11.1%
4	fairly soft	fairly cool	21%
5	normal	normal	26.2%
6	cold	fairly warm	21%
7	severe	warm	11.1%
8	very severe	very warm	3.8%
9	extremely severe	extremely warm	1%

The more maritime the climate, the higher the thermic maritimity index is. Based on an extensive study of the Dutch winter and summer climate from 760 A.D. onwards, estimates were made of the winter and summer indices C_v and C_s for the pre-instrumental period for the western European climate province (Buisman, 1984; Buisman, 1995). For the period 1591-1706 these indices were published by IJnsen & Buisman (1993), and IJnsen (1994), and for the period before 1591 by Buisman (1995; in prep). An overview for the period after 1300 is given in Fig. 34, and an overview of the maritimity index between 761 and 1990 is given in Figure 35.

Global temperatures increased in the period 900-1200 (Fig. 36; Barth & Titus, 1984). Also in the Alps, summer temperatures (based on dendro-climatological data) were above average in the period 1070-1220 (Schweingruber et al., 1988). For the Western Europe climate province, Buisman (1995) and Buisman & IJnsen (pers. comm.) found increasing 30 yr average temperatures and increasing maritimity indices for the period circa 900-1200, with an interruption around 1150 (Fig. 35). The temperature increase resulted in the medieval Climatic Optimum (1000-1350), during which the average global temperatures were high (Fig. 36; Barth & Titus, 1984).

After the medieval Climatic Optimum the climate deteriorated. According to Hofstede (1991) global temperature dropped between 1250-1500 by about 0.9 ^{0}C. For the Western Europe climate province the summers became more maritime (less warm) between 1360 and circa 1580 (Fig. 34; Buisman, 1995). Winter climate became less maritime from circa 1400 onwards (Fig. 34; Buisman, 1995). The Little Ice Age started around 1430 A.D. in Western

Europe (IJnsen, 1995b). As a general trend the annual climate in Western Europe became less maritime until circa 1700 (Fig. 34; Buisman, 1995). A slight amelioration in the world climate occurred between 1500-1570 (Hofstede, 1991).

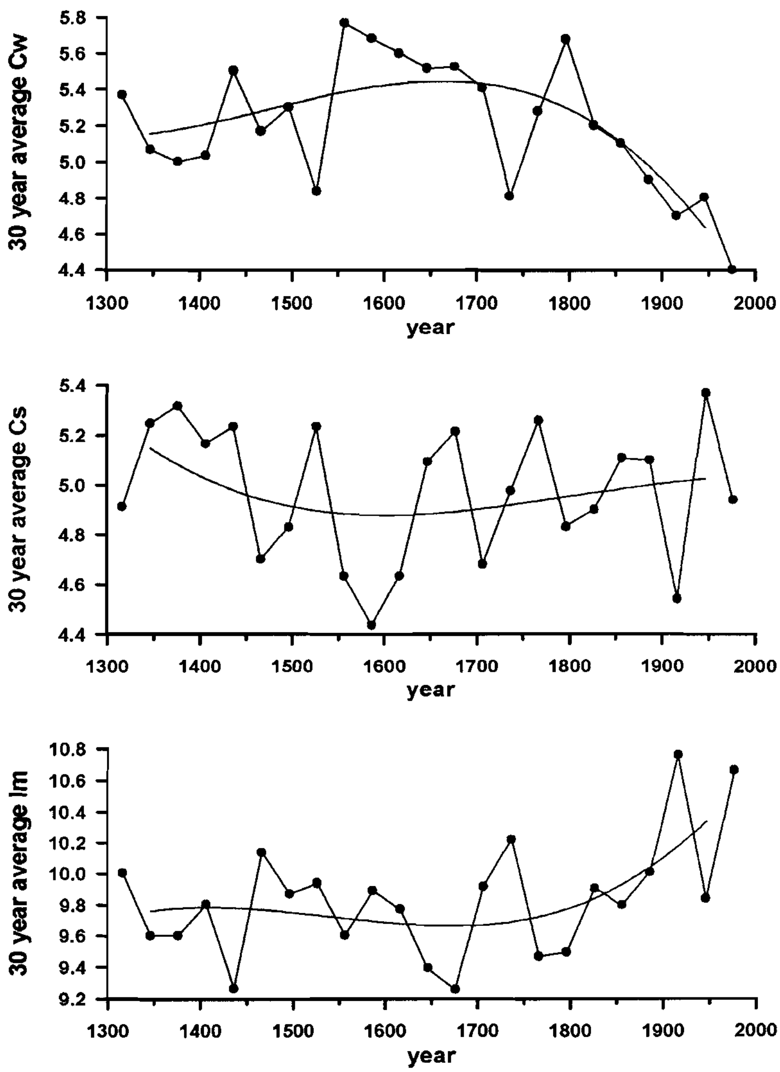


Figure 34: Winter (Cv), summer (Cs), and maritimity (Im) indices, for NW Europe from 1300 A.D. onwards (data courtesy Buisman & IJnsen; IJnsen, 1994; Buisman, 1995, in prep.). Solid line with marks gives 30 yr averages. Other line is 3th order polynomal fit.

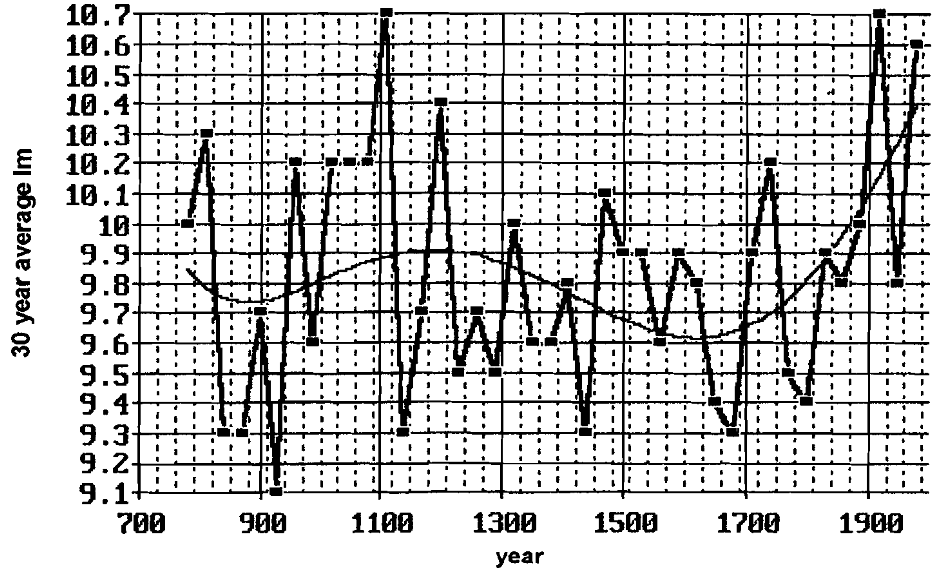


Figure 35: 30 yr averages of the maritimity index for the period 761-1990, with 5th order polynomal fit (IJnsen, 1994; Buisman, in prep.; Buisman & IJnsen, pers. comm.)[81].

Figures 34 and 35 clearly show the cold period of the Little Ice Age until circa 1820 (IJnsen, 1995b), especially in the 17th century. This was especially brought about by the cold winter climate. Although the average winter temperatures during the coldest decennia of the period were less then 1 ^{0}C below the present ones, the colder climate of the Little Ice Age resulted in a more frequent and longer lasting ice cover of shallow shelf seas and lagoons such as the Wadden Sea (Gottschalk, 1971-1977; Buisman, 1984, 1995).

After 1790, temperatures in Western Europe increased, and the climate got a more maritime character again (Figs 34 and 37; IJnsen, 1994; Buisman, in prep.). Although these changes in temperature and wind pattern were relatively small, they influenced other parameters, as will be shown below.

[81] The record until 1350 shows a decreasing amount of hiatuses. IJnsen (1994) emphasizes that in incomplete records extreme summers and winters will get more attention than the average weather types. The hiatuses have been partly corrected for by substituting the expected values and correcting for the standard deviation, with help of the standard deviation for the series after 1300 (IJnsen, pers. comm.). The record before 1300 is to be considered as an indication of the trend. Also, the record between 1400 and 1591 is still subject to minor alterations (Buisman, pers. comm.).

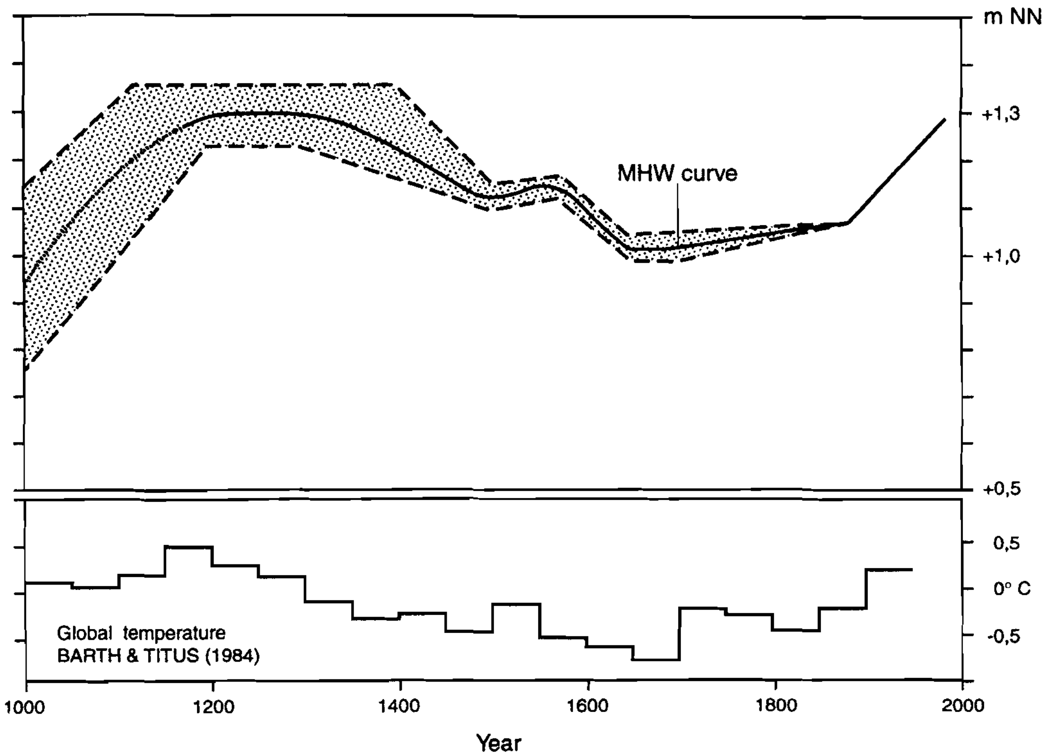
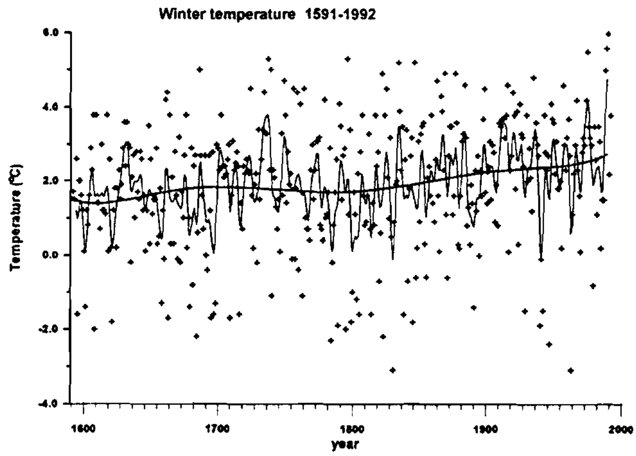


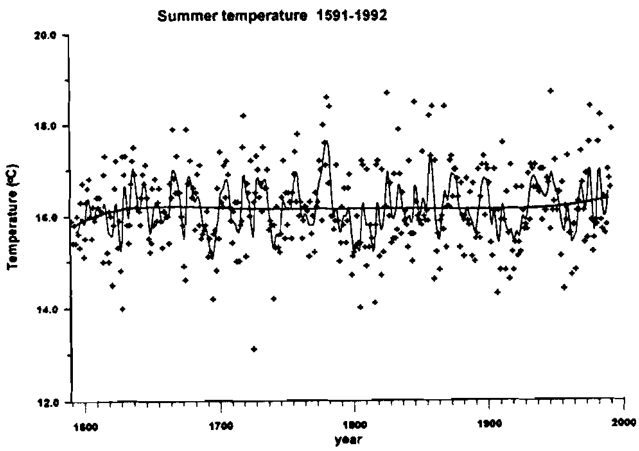
Figure 36: Average world temperature, with above it a reconstruction of the German MHW-level since 1000 A.D., with 1 sigma error band (Redrawn after Jensen et al., 1993).

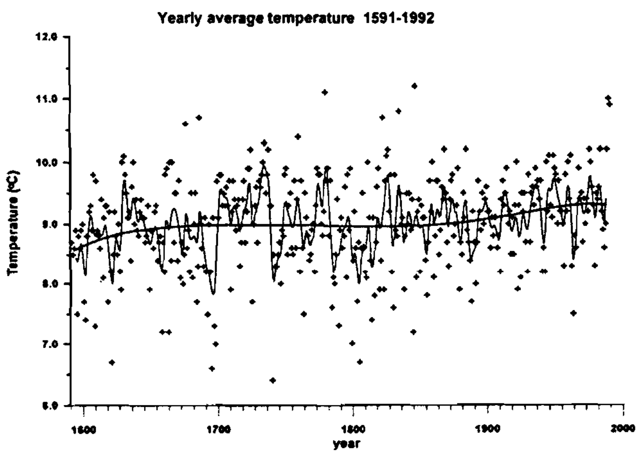
Relative sea level[82]

As a general trend, eustatic sea level fluctuated with global mean average temperature (Hofstede, 1991; Jensen et al., 1993). A correlation is observed over periods longer than several decennia (cf. Mörner, 1995). Over shorter spans of time (several decennia or less) there is little straightforward correlation between global temperature and eustatic sea level (Mörner, 1995); most likely resulting from non-linear dynamics. Moreover, basin configuration, air pressure, wind effects, tidal cyclicity, shifts of amphidromic points, and dredging and dyking influence relative sea level (Van de Plassche, 1982; Jensen et al., 1993; Franken, 1987; De Ronde, 1993). Many opinions have been expressed about sea-level history during the last millennia. Here, discussion is restricted to data relevant to the northern Netherlands.

[82] Unless stated otherwise the notion 'sea level' implies relative sea level.


Winter temperature 1591-1992
Temperature (°C)
6.0
4.0
2.0
0.0
-2.0
-4.0
1600
1700
1800
1900
2000
year


Summer temperature 1591-1992
Temperature (°C)
20.0
18.0
16.0
14.0
12.0
1600
1700
1800
1900
2000
year


Yearly average temperature 1591-1992
Temperature (°C)
12.0
11.0
10.0
9.0
8.0
7.0
6.0
1600
1700
1800
1900
2000
year

In the northern part of The Netherlands, MSL rose from 2000 B.C. to 1200 B.C., after which sea-level rise decreased slightly until 900 B.C. (Van de Plassche, 1982). After 900 B.C. it rose again until 750 B.C. The main trend was sea-level rise from 2000 to 750 B.C. (Van de Plassche, 1982)[83]. MHW was about -0.4 m DOL, and MSL was circa -1.3 m DOL around 750 B.C. (Van de Plassche, 1982). Van de Plassche (1982) remarks that the fluctuations closely follow the eustatic sea-level curve of Mörner (1980). Alternatively, the curves of Jelgersma (1979) do not show the small variations shown by Van de Plassche (1982).

For the period after 3000 B.P. only limited data are available. The curves of Roeleveld (1974) and Louwe Kooijmans (1974, 1976) indicate that MSL rose more or less at the same rate (slightly decelerating) from 2500 B.P. onwards, and that MHW may have experienced some fluctuations of the order of decimeters (Van de Plassche & Roep, 1989). A slight lowering of sea level may have occurred around 2000 B.P. (Streif, 1989).

For the last two millennia, a limited amount of reliable MHW indicators can be radiocarbon dated on the barrier islands. Fortunately, these few indicators are not influenced strongly by the effects of dykings, as is the case on the mainland. The resulting sea-level curves during the last two millennia for the East Frisian barrier islands (Streif, 1989) and those of the West Frisian barrier islands (De Groot et al., in press) show a comparable development. They show fluctuations of the MHW-level, but it is stressed that within the error band of the MHW curve of De Groot et al. (in press) also a MHW curve can be drawn which rises without interruption from 1550 B.P. until present.

Around 1965 ± 130 B.P. (105 B.C.-210 A.D) MHW was on a level below +0.4 m DOL on the East Frisian islands (Streif, 1989). MHW rose or lowered slightly to around DOL until 1500 B.P. (500 A.D.; Streif, 1989; De Groot et al., in press). Salt-marsh rootlets indicate that MHW in the East Frisian Islands has risen to at maximum +0.4 m DOL around 1185 ± 125 B.P. (675-990 A.D.; Streif, 1989). On the West Frisian islands it was found that MHW rose slowly since 1550 B.P. (circa 455-579 A.D.[84]) until circa 820 B.P. (circa 1173-1262 A.D.; De Groot et al., in press), although a high-stand around 675-990 A.D. cannot be excluded (De Groot, pers. comm.). Subsequently MHW along the East Frisian

Figure 37 (opposite page): Winter temperature, summer temperature and annual temperature recalibrated to De Bilt (data: Labrijn 1945, revised 1991; IJnsen, 1994). Solid line gives 6^{th} order polynomal best fit; zig-zag line is 3*3 point moving average.

[83] With the exception of the first part (around 2000 B.C.), the MHW curve for the southern part of the former Old Rhine Estuary shows the same trend (Van de Plassche, 1982).

[84] All values ± 1 sigma.

Islands rose quickly to a level of 1.2-1.4 m DOL before 545 ± 80 B.P. (1305-1435 A.D.), perhaps even before 690 ± 110 B.P. (1125-1395 A.D.; Streif, 1989). For the West Frisian barrier islands, De Groot et al. (in press) found an increase to present-day levels around 400 B.P. (circa 1460-1605 A.D.). Between the maxima of 545/690 B.P. and 400 B.P. a lower water level may have occurred (Fig. 36; cf. Hofstede, 1991; Jensen et al., 1993).

For the period after 400 B.P. no sedimentary sea-level marks are available, only tide-gauge observations since 1683. The most likely high MHW[85] before about 1400 A.D. and the rise of sea level since at least 1890 indicate that in between the MHW level in the Wadden Sea was most likely about 0.2 to 0.3 m lower (De Groot et al., in press; cf. Jensen et al., 1993). In the same period, circa 1400-1700, land ice in many parts of the world reached its greatest extent since the last Ice Age (Lamb, 1977). The two phenomena are obviously related (cf. Warrick & Oerlemans, 1990).

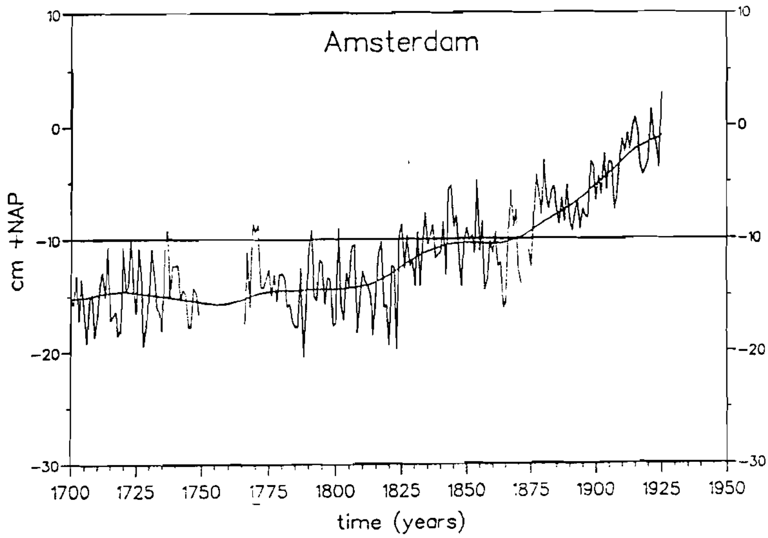


Figure 38: Record of the MSL at the tide gauge of Amsterdam, 1700-1925, corrected for subsidence (after De Ronde, 1993).

The AP (Amsterdams Peil) mark had been officially measured in 1683/84, and represented MHW in that period (Van Veen, 1945). Comparisons with the period 1700-1719 learned that the MHW level had not changed significantly (Van Veen, 1945). The MSL curve of Amsterdam covers the period 1700-1930. Corrected for subsidence, the curve shows a small rise

[85] On both the West and West Frisian islands many observations of a highstand around that time are available (Streif, 1989; De Groot et al., in press).

over the period 1700-1810 (Fig. 38; cf. De Ronde, 1993). After an initial rise average temperatures in Western Europe decreased over this period (Fig. 37; cf. IJnsen, 1994, 1995b; Libby, 1987). After 1810, a faster rise in sea level can be observed, largely coinciding with a rise in average temperatures in Western Europe (IJnsen, 1994, 1995b; Libby, 1987).

Along the Dutch coast relative MSL has risen some 18 cm in the period 1890-1990 (Fig. 39; De Ronde, 1993). Tide-gauge measurements in the Wadden Sea, at the stations Den Helder, Harlingen and Delfzijl suggest that there MSL has been rising since 1830 (Fig. 40). Average global and Western European temperature also rose during this period, especially after circa 1860-1870 (cf. IJnsen, 1994, 1995b; Hansen & Lebedeff, 1987). The MHW rose faster than MSL (circa 0.05 $cm.yr^{-1}$ faster over the period 1940-1990; De Ronde, 1993).

Research on present mean sea-level rise shows that, apart from thermic expansion of the ocean water, also the volume of the land ice is important (Warrick & Oerlemans, 1990). Indeed, the melting of land ice and the retreat of glaciers started after 1850 (Jensen et al., 1993). At present it is anticipated that average global temperature will continue to increase due to the man-induced green-house effect (Warrick & Oerlemans, 1990).

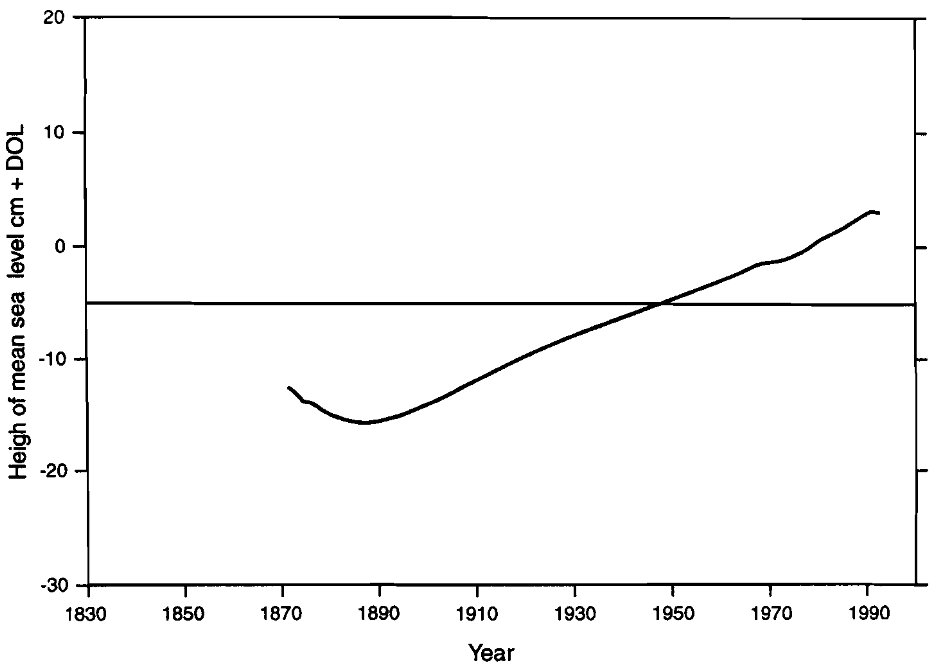


Figure 39: Record of the Dutch MSL since 1880, based on tide gauges along the coast. The lowering in MSL until 1880 is mainly caused by stations Delfzijl, Vlissingen and IJmuiden (after De Ronde, 1993).

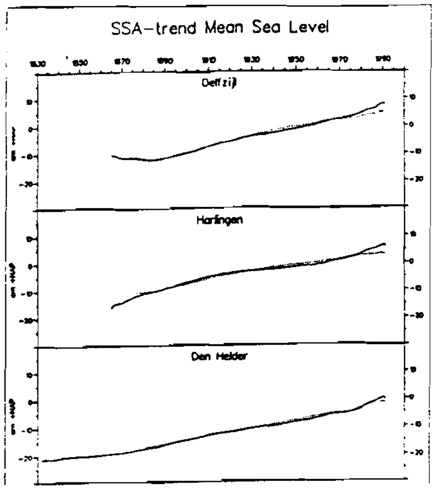


Figure 40: Record of the MSL in three tide gauges along the Wadden Sea mainland coast.

Figure 41: Copper engraving illustrating the effects of the storm surge of 1686 in Groningen, by Jan Luyken (in: Gottfried, 1698). During storm surges, the drainage area may temporarily increase considerably, due to the flooding of supra- and intertidal areas. The strong currents and the lift force on buoyant materials, such as peat, may cause strong erosion, and thus increase the drainage volume of the related tidal inlet system.

Storm surge frequency

The above historical observations show that storm surges have a devastating effect on the dunes and on the North Sea side of the barrier islands. An increase in storm frequency thus may shift the balance between coastal erosion and sedimentation (Fig. 41). As discussed, storm surges considerably increase erosion by forcing migration of channels and inlets, and they thus accelerate erosion and sedimentation at the ends of the barrier islands. Moreover, sudden blockage of backbarrier channels (pers. obs.) and outer channels (Anonymous, 1877) during storms occurs, forcing the channels to find a new course. Important is also the temporary overdeepening of channels by a strong flow. After a storm channels need sand to re-establish their equilibrium dimensions.

Fluctuations in storm-surge activity occurred in relation to climate changes (Fig. 42; Gottschalk, 1971-1977). The changes in the amount of recorded storm surges, resulting in flooding events are for an important part artificial. Gottschalk underestimated the effects of subsidence due to human interference (Van de Ven, 1993). In case of the drainage of peat, for instance, subsidence may have been up to several metres. Also, the improvement of dykes will have diminished the effects of storm surges. For instance, the increase in the number of storm surges during the period 1500-1600, especially in 1550-1600, is partly due to the bad conditions of the dykes as a result of the 80-year independence war (1568-1648) of the Dutch against the Spanish (cf. Gottschalk, 1975, 1977). However, the bad quality of the dykes does not totally explain the observed strong increase of storm-surge floodings in 1500-1600, because this happened simultaneously in Groningen, Friesland, Zuider Zee and Holland (cf. Gottschalk, 1972). Moreover, dykes were also low and weak before 1500, when not so many storm surges were experienced.

Climatic conditions play a significant role in the storm-surge frequency. Storm surges in the Wadden Sea are mainly brought about by western to northern winds, pushing the water into the backbarrier area (IJnsen, 1977). During the last 110 yrs 87% of all storm surges in The Netherlands occurred during 4 out of 30 different types of atmospheric circulation patterns (IJnsen, 1992). Of the four, the two most important (67% of all storm surges), generate more maritime climatic conditions[86] (IJnsen, 1992). In the Wadden Sea (station Harlingen) 87% of all storm surges in the last 110 yrs occurred during the winter halfyear (IJnsen, 1992, pers. comm.). Especially in the winter halfyear, oceanic depressions (low atmospheric pressures) occur frequently under the present, more maritime climatic conditions. They increase the chance of storm surges, and cause a rise of the sea level (1 mm less air pressure results in circa 1 cm extra sea-level height; cf. Lamb, 1980). Figure 43 suggests this relation between sea-level height, climate and storm surge activity for the period 1880-1990. A more maritime (winter) climate coincides with higher MHW and more frequent storm surges.

[86] During other atmospheric circulation patterns, which bring about a more continental climate, eastern to southern winds often cause a lowering of the water levels in the backbarrier, or result in quiet wind conditions during which storm surges are rare (IJnsen, 1987, 1992).

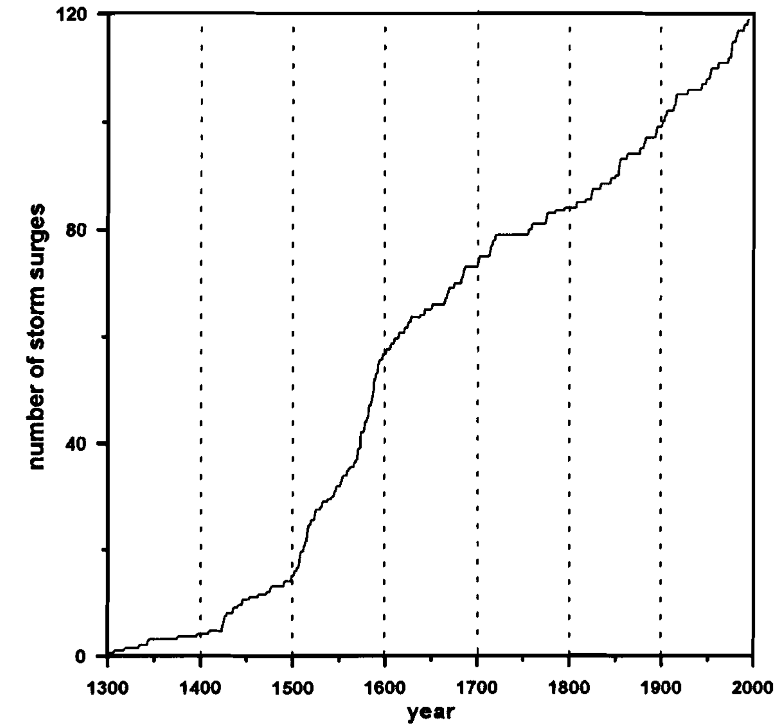


Figure 42: Cumulative storm surges in Friesland and/or Groningen, since 1300. Until 1825 storm surges were counted if damage to the islands or the mainland had occurred. All storm surges were counted as 1, irrespective of their strength; only for uncertain or very weak surges a 0.5 was counted. Data from 1300-1700 were mainly derived from the critical review of storm-surge records by Gottschalk (1971, 1975, 1977), and Buisman (1995, in prep.). The storm surges of the period 1700-1825 were derived from Buisman (1984; in prep.) and own data. Data after 1825 were derived from tide-gauge measurements of Min. of Waterworks. These were corrected for relative mean sea-level rise. They were counted as storm surges, if they reached a level higher than about 0.5 m below the height of the dykes before 1717, and, on average 1.2 m below the dykes of and after 1717. Short-lasting gales and waves may push the high water over such dykes. Note the high amount of storm surges during the 16th century.

After 1600, during the Little Ice Age, which started 1430 to 1550 (Buisman, 1984, pers. comm.), and ended around 1790 (Lamb, 1977) to 1820/1860 (IJnsen, pers. comm.; Strahler & Strahler, 1987), atmospheric circulation patterns led to a climate with a less maritime character, especially in winter. As discussed above such circulation patterns lead to less storm surges, which was indeed observed. After 1800 the climate became more maritime again, and the number of storm surges increased slightly.

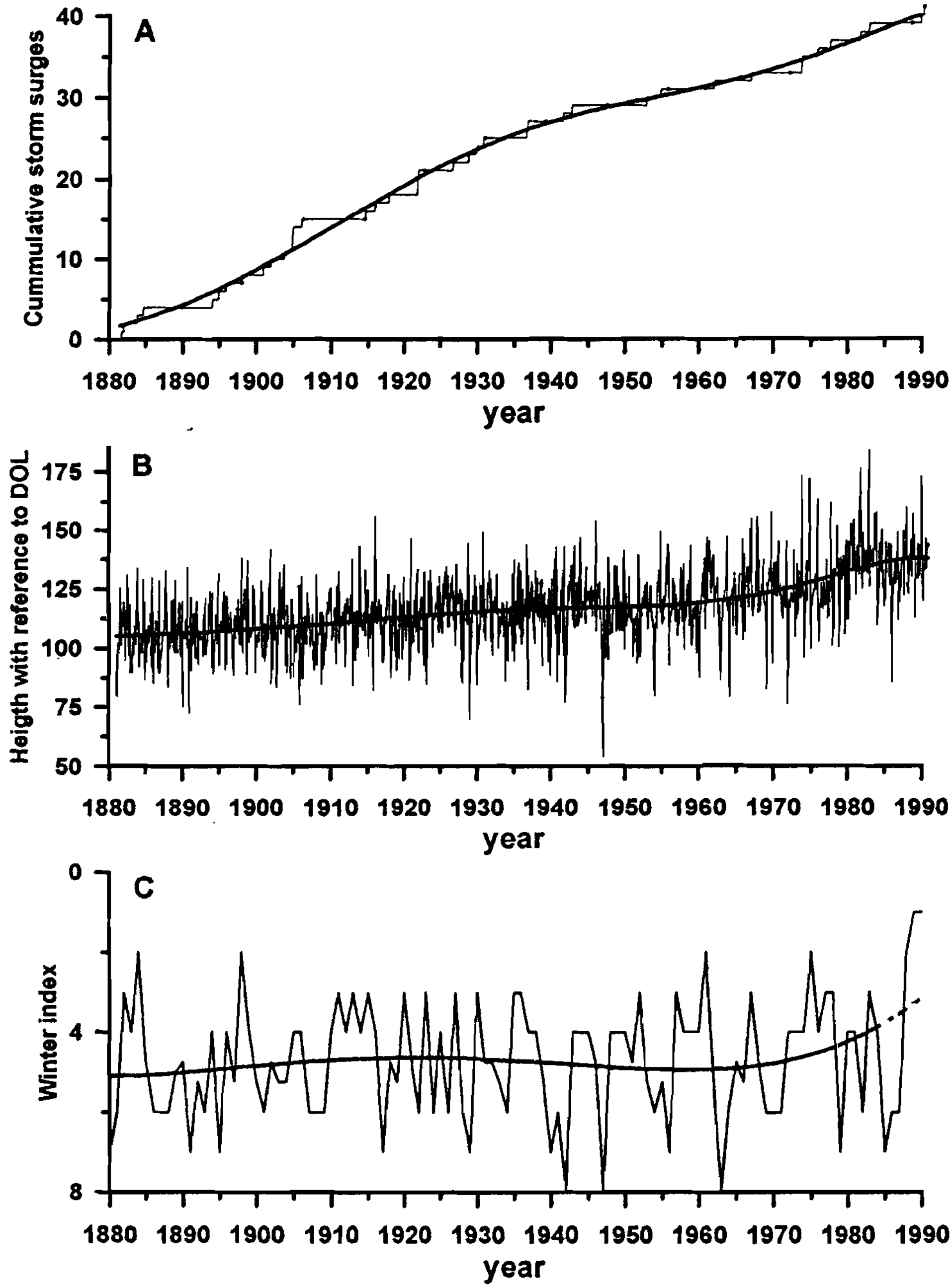


Figure 43: A) Cumulative storm-surge heights (m above storm-surge level 1991, corrected for MHW of the year in which it occurred, data IJnsen, 1977), B) MHW, and C) winter index for the period 1880-1990, all with 6th order best fit. Tide gauge Delfzijl.

The observed increase in the number of storm surges in the period 1550-1600 cannot be fully explained by the poor quality of the dykes. A climatic factor seems likely. The increase cannot be attributed to the winter climate, which became less maritime (less storms). Perhaps storm surges were enhanced by the strong contrast between the winter and summer climate. This indeed is suggested by the observation that flooding due to storm surges in the period 1550-1600 occurred also in the summer period, which, as far as information is available, never happened in other periods. Lamb (1980) suggested that the extreme amount of storm surges was generated by an intensified thermal gradient in our latitudes, due to the advance of the Arctic sea ice.

Dykes and other human influences

From circa the fifth century B.C. onwards people lived on artificial dwelling hills (terps). Often the vertical growth of these terps was sufficient to keep up to or outpace relative sea-level rise (mainly by subsidence of the land), but sometimes the terps were flooded and abandoned or re-established (Venema, 1993). Comparison of various Frisian Codes shows that in the 8th and 9th century no laws existed with regards to dykings (Acker Stratingh, 1866; Van Giffen, 1964); dykes were probably not built, due to the vast amounts of uncultivated peat lands still available for agriculture (Edelman, 1974). Dyke building probably started in or after the 10th century, and became important in the 12th and 13th century (Acker Stratingh, 1866; Andreae, 1881; Rienks & Walther, 1955; Edelman, 1974). Around that time, a system of dykes surrounding Friesland became established (cf. Venema, 1993). Originally dykes were probably only meant to protect land against erosion, whereas in the 13th century active polderisation started (Edelman, 1974; Van de Ven, 1993). From that time on, the supratidal areas of the Lauwerszee and along the mainland coast were dyked, and thus the drainage area and the tidal prism of the related inlets decreased.

The islands were partly protected by dykes at the Wadden Sea side and by dunes along the North Sea side. Terschelling was protected locally by small dykes and dams, also to keep cattle out of the agricultural areas. In 1506 the continuous dyke on Terschelling is mentioned for the first time. It probably existed already for quite some time, because it should be maintained according to the '*old habit*' (Van Oosten, 1986). According to Reitsma (1984) the '*old habit*' of maintaining dykes on Ameland, up to a height of 4 feet, is mentioned in judicial archives of the 14th century of Ferwerderadeel. It can thus be concluded that the relatively small agricultural parts of the island (Van Oosten, 1986) were protected by low and weak (Huber, 1879; Brouwer, 1936) dykes since at least the 14th century, probably since the onset of the settlement in the 10th century. Near the Wadden Sea, outside the agricultural parts, the communal meadows were located, mainly supratidal marshes, which were not or hardly protected by dykes (Van Oosten, 1986; Reitsma, 1984). After 1806 poorly successful attempts were made to protect these lands against further erosion. In 1930 a dyke was established (Reitsma, 1984). The early dykes on Schiermonnikoog were comparable to those of Ameland, but they were destroyed during the island migration (1700-1850). After the strong migration in the 18th century, low dykes were built in 1758 and 1767. Only since 1860 a continuous high dyke protects the polder on Schiermonnikoog (Abrahamse & Koning, 1969).

Erosion of dunes is largely prevented by plants, especially by bent-grass (Ammophila arenaria). Regulations concerning the harvesting of bent-grass in the dunes of the barrier islands, forbidding the destruction of the root-system, date from at least 1354 (e.g., Blok et al, 1896, no. 429).

Mechanisms related to the coastward shift of the islands and the formation and closure of tidal embayments: discussion

Coastward shift of the islands

From the maps of Ameland it follows, that the period during which the coast did not retreat or even prograded (MHW-line, 1536/45-1850), mainly coincided with a low or even negative sea-level rise. A lower sea level, in combination with the eastward shift of the western watershed of Ameland Inlet (Van der Spek, 1994), most likely generated a surplus of sediment, becoming available to the North Sea coast. From probably 1830 onwards, relative sea level rose considerably, and a strong retreat (circa 0.5 km) of central Ameland can be observed since 1850. This is in line with the expectation that sea-level rise results in the erosion and coastal retreat of barrier islands.

From the less maritime climate during the 17th and 18th century it follows that the development of the barrier coast was relatively less wave-influenced in that period than before and after. The weaker wave attack is probably an additional reason why Ameland did not retreat or even prograded seaward in that period. For the other islands the lower sea level and the weaker wave attack in the period 1536/45-1850 must have had similar effects on the coastlines, which may explain their relatively stable position during these periods. Indeed, along other islands of the Wadden Sea strong morphological changes (mostly sand supply or island growth; e.g., Schiermonnikoog & Rottumerplaat) have been observed after 1850, which have been attributed to the ending of the Little Ice Age (Ehlers, 1988).

Formation and closure of the tidal embayments

The main flooding of the Boorne valley (Middelzee) occurred at the end of the 9th century into the 10th century. In the period 750-900 A.D. the Lauwerszee was flooded and expanded to the south. Its E-W dimensions increased until the 13th century (Andreae, 1881). The initial flooding of both embayments thus coincided with a sea-level rise, that reached a possible maximum between 780/911 A.D. After a possible slight decrease until about 1173/1262 A.D., sea level rose to, at maximum, higher than present-day heights around 1125/1395 A.D[87]. The ongoing flooding after 780/911 A.D. indicates that the Middelzee and the Lauwerszee embayment were quite vulnerable to flooding. Human-induced lowering of the

[87] Alternatively, sea level may have been rising from 1500 B.P. until 500 B.P..

areas as, for instance, evidenced by the many peat excavations in the Lauwerszee embayment, played an important role. Moreover, the flooding (ingression; Van der Spek, 1994) of the mainland was enhanced by storm surges, as many direct and indirect evidence shows (such as the fast erosion of land in the Zuider Zee in the 12th century; Fig. 42). Gottschalk (1980) even attributed the majority of the Subatlantic ingressions to storm surges.

Important is also that the flooding of embayments can be a self-propelling process. Tidal amplitude increases towards the inner parts of funnel-shaped areas (Van de Plassche, 1982) such as the Lauwerszee[88]. At present this can still be observed in the Ems-Dollard Estuary, where the tidal amplitude increases from 2.2 m at the inlet to 3 m in the Dollard embayment (cf. Vroom et al., 1989). The larger tidal range results in a stronger erosion of peat areas. Due to erosion, the size of the tidal area increases, and the tidal amplitude in the back of the embayment can increase further, so that the process can continue[89]. The erosion will stop if peat is not available any more, or if the flood-basin effect becomes dominant (this is the reduction of the tidal amplitude if a tidal wave, after passage through a narrow entrance, enters a wide basin; cf. Leenders, 1986; Van de Plassche, 1982, in press). The flood-basin effect caused by the eastward and westward expansion of the Lauwerszee, and the lack of erodible peat must have put an end to the flooding of the Lauwerszee. Some expansion at the end of the tidal basin was also observed in the Middelzee (Boorne) and the flood-basin effect may have counteracted the flooding there.

The reclamation of the Lauwerszee more or less coincided with the lowering of the sea level after circa 1125-1395 A.D. (cf. Jensen et al., 1993, Fig. 36). In the 14th century the reclamation advanced quickly. Afterwards it slowed down. This cannot be explained by sea-level height, which most likely decreased in that period. The reclamation led to a decrease of the width of the area, and to a decrease of the flood-basin effect. The decrease in area thus most likely led to an increase of the HW level. The same must have occurred in response to the dyking of the broad Dollard embayment. Indeed, in addition to compaction, the higher HW-level is thought to be one of the main reasons for the greater height of the younger polders (Van de Ven, 1993).

The above indicates that even small sea-level changes can exert a great influence on the extent of the drainage basin if the mainland is vulnerable to flooding. Once flooding starts, the process becomes self-propelling and large areas can be ingressed, even if sea level is falling.

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[88] In general, the tidal wave in the Wadden Sea is distorted in such way that tidal amplitudes increase towards the landward part of the back-barrier area (Vroom et al., 1989).

[89] The complaints about increasing tidal levels and the increase of inlet dimensions for the Zuider Zee area around the 15th century may well illustrate the process in action. Also, the complaint of the dyke guards of Usquert in 1578, about the increasing dimensions of the inlets and the resulting insufficiency of the dykes (Acker Stratingh, 1866) points to such a process, probably especially for the Eems-Dollard and adjacent regions.

The historical reconstructions show that man is able to influence the development of the barrier islands with simple measures such as planting and maintaining bent-grass, extermination of rabbits, and artificial stimulation of dune growth. However, once such activities are neglected, strong erosion and migration may undo the previous results. Large-scale measures do have significant, but frequently unwished effects. For example, the protection of Borkum has halted the eastward migration of the Western Ems, by which the small island Rottumeroog can not migrate further to the east, and will most likely disappear. It appears that both doing nothing and taking strong unnatural measures is no good. This rises the question what to do now sea-level rise will probably accelerate? The growing tendency to 'work along with nature' and even to allow, where possible, a controlled natural development, may provide the answer to the problem.

From the above discussion it is clear that the observed changes in the eastern Wadden Sea are rarely to be understood in terms of a simple cause-effect relation. Mostly there are several causes which together generate the changes. In their turn, the changes in the morphology also influence the hydrodynamics of the system. The result is that (series of) changes may last many centuries. Moreover, the development in historical time shows that, in spite of the changes of the external and internal variables (sea level, climate, basin form), the morphology of the barrier system was maintained.

CONCLUSIONS

Pre-historical and historical developments

Despite the changes of external conditions, the main morphological characteristics of the Wadden Sea were maintained through time. Historical evidence shows the continuous presence of supratidal marshes, creeks, estuaries, sub- and intertidal flats, tidal channels and barrier islands in the eastern part of the Dutch Wadden Sea since at least 0 A.D., perhaps even since the 4th century B.C. This is in accordance with geological data which reveal the presence of barrier islands since at least 5,300 B.P., and the presence of intertidal flats, tidal channels, estuaries and tidal marshes since at least 6,500 B.P.. The western part of the Dutch Wadden Sea was still a high and supratidal area around 0 A.D. The Zuider Zee lakes were, by then, drained through the Vlie Inlet and gradually became larger. During Roman times, tides were, just as today, semi-diurnal and the tidal range seems to have been smaller than at present, as was also suggested by theoretical calculations.

The initial southward expansion of the Lauwerszee embayment, during the 8-9th century, and of the Middelzee in the late 9th-10th century coincided well with a possible highstand of the MHW. The Boorne was filled in from circa 1000 A.D. onwards. The Lauwerszee embayment probably increased in maximal in E-W dimensions until the 13th century. The initial formation of the embayments and the later expansion were enhanced by the substantial man-induced lowering of the area, especially by the excavation of peat. Moreover, the increase in length of the embayments must have led to an increase of the tidal amplitude towards the S of the basins. During E-W expansion of the basin, the amplification was reduced by the flood-basin effect.

The historical data show that the system of barrier islands, tidal inlets and backbarrier channels and tidal flats is highly dynamical, and shows an overall tendency to shift toward the east. The tendency is, for the larger part, caused by wind- and tide-driven eastward sediment transport, the eastward propagation of the tidal wave, and by the tidal-prism-related size and orientation of the ebb-tidal deltas and the orientation of the inlet systems. These effects make it easy for a tidal system to shift eastward and relatively difficult to shift to the west. Superimposed on the eastward drift of the systems, high-frequency, cyclic, sometimes periodical changes of the position and orientation of the outer channels occur. Also the island ends expand and contract over periods of several decennia, with velocities of up to several hundred metres per year.

It is inferred that the Lauwerszee embayment was drained mainly by the Lauwers Inlet east of Schiermonnikoog around 1300. Historical evidence suggests that the Lauwerszee drainage was gradually taken over by the Zoutkamperlaag, W of Schiermonnikoog, between 1300 and 1550. The watershed S of Schiermonnikoog, i.e., towards the Frisian mainland, was breached in the process. The western side of the island was eroded considerably by the large inlet system of the Zoutkamperlaag. The change caused a chain-reaction: The Lauwers Inlet was largely abandoned in the period 1300-1640. Due to the decrease of the related ebb-tidal delta, the downdrift island Bosch changed into a sandy shoal (also due to the storms in the 16th century). After 1640 the Lauwers system migrated downdrift, and took over parts of the drainage of the Schild system. The downdrift expansion was further enhanced by the formation of the Eilanderbalg, W of the Lauwers Inlet. As a result the drainage of the Lauwers increased, and new shoals started to form at its downdrift side. The drainage of Schild Inlet was gradually reduced, and the downdrift island Rottumeroog consequently became smaller.

The change in the basin configuration due to the infill of the Middelzee Estuary led to the closure of the westernmost inlet of the Ameland Inlet system. The closure was accompanied by the eastward shift of the eastern end of the barrier island Terschelling over 10.5 km, due to eastward sediment supply and amalgamation of shoals. After the closure the remaining easternmost inlet shifted to the E, and eroded the western end of the barrier island Ameland from circa 1800 onwards. This shift is attributed to the decrease in the drainage volume and the downdrift position of the drainage basin.

The morphological changes are rarely to be understood in terms of a simple cause-effect relation. The interactions of processes with each other and with the morphology result in series of changes that may last many centuries.

Climate and sea level

Part of the developments is controlled by sea-level height and storm regime. The data and interpretations presented in the chapter may help/serve to predict the natural future development of the system, and also to estimate risks in case of extreme weather conditions, such as occurred during the All Saints Flood of 1570. As already stated above (formation of embayments), a part of the observed changes can be attributed to sea-level fluctuations. For example, the seaward progradation and subsequent retreat of the barrier island Ameland coincided with the probable lowering and rise of sea level during and after the Little Ice Age, respectively. Storms have a strong influence on the development of the barrier islands, as is clear from the historical reconstructions. Especially important is the enhancement of erosion of the islands by migration of channels during storms. Embayments in the mainland are enlarged by storms. Subtle changes of climate (circulation patterns) influence the number of storms and the height of MHW and MSL. It is inferred that part of the increased stability of Ameland during the Little Ice Age resulted from the smaller wave-attack.

Future developments

From the trends in the historical development, the future development can be extrapolated to some extent. Several significant changes are likely to occur. The Engelsmanplaat may disappear, so that the adjacent tidal inlets, Zoutkamperlaag and Pinkegat will merge. The sand needed for the change may lead to increased erosion of especially Ameland. After a merger of the inlets, the eastern end of Ameland may expand over another 5 km to the east. In that case, the logical natural development would be that a new inlet forms over this new easternmost end of Ameland. More to the east the Schild Inlet will most likely be abandoned in the first half of the next century, and the formation of a larger island ('New Rottumeroog') is to be expected.

Influence of man increased over the last 2000 years, and has become quite strong to dominant in many places. The problems related to the sea-level rise in the near future will most likely be met best, if the natural development of the system is respected and made use of.

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REFERENCES

Abel, O., 1893: Kaiser Karls Leben von Einhard (p. 26: vita Caroli Magni by Einhard, written between 814-820 A.D.). Die Geschichtschreiber der deutschen Vorzeit, 13, 9th Cent., 1st vol., 3rd Ed., Verlag der Dutschen Buchhandlung, Leipzig, 62 pp.

Abrahamse, J., 1979: Inhammen en oude kaarten. Waddenbulletin, 14, 2, 110-111.

Abrahamse, J., 1990: Rottumeroog. Waddenbulletin, 25, 2, 82-87.

Abrahamse, J., 1992: Rottumerplaat. Waddenbulletin, 27, 4, 222-228.
Abrahamse, J., 1994: Waterstaat maakte omgeving Westerplas tot fantastisch natuurgebied. Schiermonnikoog Journaal, jrg. 2, nr. 1, 4-10.
Abrahamse, J. & Buwalda, J.D., 1964: Wadlopen, 'land'-loperij bij laagwater. In: Anderson, W.F., Abrahamse, J., Buwalda, J.D. & Van Straaten, L.M.J.U. (Eds): Het Waddenboek. Thieme, Zutphen, 9-36.
Abrahamse, J. & Koning, H., 1969: Schiermonnikoog, van Westerstrand tot Willemsduin. Knoop & Niemeijer, Haren, pp. 112.
Abrahamse, J. & Luitwieler, F., 1983: Rottumerplaat. Waddenbulletin, 18, 2, 35-37.
Acker Stratingh, G., 1866: Twee hoofdstukken uit de geschiedenis van ons dijkwezen. J.B. Wolters, Groningen, 98 pp.
Alexander, M., 1972: Beowulf, Penguin Classics 176 pp.
Andreae, A.J., 1881: De Lauwerszee, nagespoord in hare wording, haren omvang en hare verschillende bedijkingen, met een aanhangsel betreffende de bedijking van het "Cruijslandt" in Kollumerland. H. Kuipers, Leeuwarden, 179 pp.
Anonymous, 1532, Amsterdam, printed by Jan Zeuerszoon: De kaert vander Zee. Reprint, Knudsen, J., 1914 (Ed.), Nijhoff,'s-Gravenhage, 99 pp.
Anonymous, 1541: Dit is die Caerte van der zee: om Oost en West te zeijlen. Reprint, Rogge, H.C., 1885 (Ed.), 86 pp.
Anonymous, 1566: Amsterdam, by Jan Roelants: Die Caerte van der Zee. Reprint Knudsen, J., 1920 (Ed.), Nijhoff, 's-Gravenhage, 221 pp.
Anonymous (C.J.), 1717: Korte beschrijvinge van de schrikkelijke water-vloed, veroorzaakt door een sterke storm-wint, voorgevallen tusschen den 24 en 25 december 1717, ok betrekking hebben op "Holland Groninger-, Embder- en Oost-Friesland, 't Stigt Bremen, Oldenburgerland, Holsteyn en meer andere Landen, Thijssen, Leeuwarden, 31 pp. PBF.
Anonymous, 1978: Getijtafel voor Nederland, 1979. Staatsuitgeverij, 's-Gravenhage, 141 pp.
Athallin, A.M., 1810-1811: Military reports on the Wadden Sea, manuscript, Archives du Ministère de la Guerre, Chateau de Vincennes no. 1471:
- Notes sur les batteries de Zoltkamp et Oostmahorn ansi que sur les îles de Rottum et Schiermonnikoog,
- Sur l'île d'Ameland et sur la passe dite Amelander gat,
- Iles des Watten
- Sur l'île de Ter-schelling et sur la passe du Vlie
- Sur l'île d'Ameland et sur la passes et les Wades qui l'avoissint
- Sur les batteres d'Oostmahorn et de Zoltkamp
- Sur l'île de Schiermonnikoog et la passe de Frieschegat
- Sur l'île de Rottum et description des Wades, du Lauwers et du Schildt.
Avis, J.G., 1934: Het auteurschap van de 16e eeuwse kaarten van het Friesche Bilt. Tijdschrift voor Geschiedenis, 11, 403-415.
Bakker, H., 1970: Ameland van Hollum tot de Hon. Triangelr., Knoop & Niemeijer, Haren, 99 pp.
Barentsen, W., 1961: De zeedijk van zijn ontstaan tot het jaar 1730. OTAR, 45, 9, 195-205.
Beckering Vinckers, J.A., 1943: Nota Amelander Gat, Rijkswaterstaat, studiedienst Hoorn, 181 pp.
Beets, D.J., Van der Valk, L. & Stive, M.J., 1992: Holocene evolution of the coast of Holland Mar. Geol., 103, 423-443.
Beets, D.J., Van der Spek, A.J.F. & Van der Valk, L., 1994: Holocene ontwikkeling van de Nederlandse kust. RGD rapport 40.016 - Projekt Kustgenese. Haarlem, Rijks Geologische Dienst, Hoofdafdeling Ondiepe Ondergrond, 53 pp.
Bleuten, W., 1971: Longshore en ripcurrents op het strand van Schiermonnikoog. Int. Rep. Fac. of Geogr. Sci., Univ. of Utrecht: 16 pp; Utrecht.
Blok, P.J., Feith, J.A., Gratema, S., Reitsma, J. & Rutgers, C.P.L., 1899a: Oorkondenboek van Groningen en Drenthe, eerste deel, J.B. Wolters, Groningen.

Blok, P.J., Feith, J.A., Gratema, S., Reitsma, J. & Rutgers, C.P.L., 1899b: Oorkondenboek van Groningen en Drenthe, tweede deel, J.B. Wolters, Groningen.

Bodel Nyenhuis, J.T., Eekhoff, W., 1846: De algemeene kaarten van de provincie Friesland verzameld, beoordeeld en geschiedkundig beschreven. 1-106.

Boltz, L., 1842: Rapport wegens de gesteldheid van het eiland Schiermonnikoog. Report Rijkswaterstaat, no. 241/1, Leeuwarden 10 November, 11 pp.

Borger, G.J., 1967: De betekenis van de Kerstvloed van 1717 voor de gebieden rond het IJ. KNAG, Geog. Tijds., 2, 97-103.

Bosch, A. & Vos, P.C., 1992: Paleogeografische reconstructie van het Lauwersmeergebied, concept, RGD-Project 40009, 22 pp.

Brandon, P.F., 1971: Late-medieval weather in Sussex and its agricultural significance. Inst. British Geographers, Transactions, 54.

Brilhuis, R., Bossinade, J.H., Van den Bergs, J., De Vlas, J., 1990: Een meer natuurlijke ontwikkeling van Rottumerplaat en Rottumeroog: aanvaardbaar of niet? RWS, Nota GRAN 1990-2003, 55 pp.

Brilhuis, R., Bossinade, J., Van den Bergs, J. & De Vlas, J., 1991: De toekomst van Rottumerplaat en Rottumeroog. Waddenbulletin, 26, 2, 69-71.

Brouwer, D.A., 1936: Ameland, een sociaal-geografische studie van een waddeneiland. Amsterdam: H.J. Paris, 51 pp.

Buisman, J., 1984: Bar en boos. Zeven eeuwen winterweer in de lage Landen. Bosch & Keuning NV, Baarn, pp. 317.

Buisman, J., 1995: Duizend jaar weer, wind en water in de Lage Landen, deel I: 764-1300. Van Wijnen, Franeker.

Buisman, J., in prep.: Duizend jaar weer, wind en water in de Lage Landen, deel II: 1301-1550.

Buisman, J., in prep.: Duizend jaar weer, wind en water in de Lage Landen, deel III: 1551-1800.

Buisman, J., in prep.: Duizend jaar weer, wind en water in de Lage Landen, deel IV, 1801-1995.

Burton, J. & Marrow, P., 1989: Seismic hazard and earthquake source parameters in the North Sea. Nato Advanced Study Institutes Series. Series C, 266, 663-664.

Coronike van Vrieslant, c. 1450: probably written in the Cloister Thabor at Tirns (near Sneek) and composed of parts of Beka and the major chronicles of the Middle Ages (Gottschalk, 1971).

De Boer, M., 1979: Morfologisch onderzoek Ameland. Verslag van het onderzoek op het Amelander Wantij in 1973. Rijkswaterstaat, Directie waterhuishouding en waterbeweging, district kust en zee, studiedienst Hoorn, Nota WWKZ-79.H005.

De Boer, M., Kool, G. & Lieshout, M.F., 1991a: Erosie en sedimentatie in de binnendelta van het Zeegat van Ameland, 1926-1984, Deelonderzoek no.4. Directie Noord Holland, Rapport ANVX-91.H202.

De Boer, M., Kool, G., Lieshout, M.F. & Ulm, D.L., 1991b: Erosie en sedimentatie in de ebgetijde delta van het Zeegat van Ameland, 1926-1982, Deelonderzoek no.8. Directie Noord Holland, Rapport ANVX-91.H205.

De Groot, H.L., 1992: Terugblik op Traiectum, een archeologische schets. Ter gelegenheid van het zeventig jarig jubileum van de vereniging Oud-Utrecht. Stichting publ. Oud-Utrecht, 31 pp.

De Groot, Th.A.M., Westerhoff, W.E. & Bosch, J.H.A. in press: Sea level rise during the last 2000 years as recorded on the Frisian Islands, The Netherlands. Med. RGD, 3 pp..

De Haan, H., IJbema, R. & Reitsma, D. Th., 1983: Engelsmanplaat. Geschiedenis van...en gebeurtenissen rond... een zandbank. Uitgave Stichting 't Fiskershuske, 168 pp.

De Jong, J., 1975: Pollen-Analysis and radiocarbon dating of a peat deposit from the island of Schiermonnikoog, The Netherlands. Berichten van de Rijksdienst voor het oudheidkundig Bodemonderzoek, 25, 183-186.

De Jong, J., 1984: Age and vegetational history of the coastal dunes in the Frisian Islands, the Netherlands. Geol. Mijnb., 63, 269-275.

De la Roncière, M. & Mollat du Jourdin, M., 1984: Portulane, seekarten vom 13. bis 17. Jahrhundert, Office de Livre, Fribourg.

De Ronde, J.G., 1983: Changes of relative mean sea-level and of mean tidal amplitude along the Dutch coast. In: Ritsema, A.R. & Gürpinar, A. (Eds), Seismicity and seismic risk in the offshore North Sea area. Reidel Publ. Cie., 131-141.

De Ronde, J.G., 1993: De toestand van de zee van morgen. Verwachtingen rondom zeespiegelstijging, getij, wind, stormvloeden en golven. RWS, Int. Rep., GWAO-93.148x, 10pp.

De Smet, L.A.H., 1965: De bodem van Groningen. Toelichting bij blad 1 van de bodemkaart van Nederland schaal 1 : 200.000. Stichting voor Bodemkartering, Wageningen, Meppel, Noord Nederlandse Drukkerij, 117 pp.

De Vries, D., 1983: Nieuwe caert van Frieslant, 1739, heruitgave van de wandkaart van Bernardus Schotanus à Sterringa, toegelicht door. Canaletto, Alphen aan den Rijn, 24 pp.

De Waard, D., 1948: Urk omstreeks het jaar 1000. K.N.A.G., 2nd ser., 65, 626-630.

De Weerd, H., 1963: Waddenlegenden, met illustraties van P. Pouwels, 3e druk. Uitg. de Boer Jr. Hilversum, 134-146.

Dean, R.G. & Walton, T.L., 1975: Sediment transport processes in the vicinity of inlets with special reference to sand trapping. In: Cronin, L.E. (Ed.): Estuarine Research: Vol. II, Geology and Engineering. Acad. Press, NY, 129-150.

Denucé, J. & Gernez, D., 1936: Le livre de mer, manuscript de la bibliothèque communale d'Anvers, No. B.29166, 2 Vol., De Sikkel, Antwerpen, 91 pp.

Dijkema, K.S., 1987: Changes in salt-marsh area in the Netherlands Wadden Sea after 1600. In: Huiskes, A.H.L., Blom, C.W.P.M. & Rozema, J. (Eds): Vegetation between land and sea. Junk, Dordrecht, 42-49.

Dijkema, K.S., Reineck, H.E. & Wolff, W.J. (Eds), 1980: Geomorphology of the Wadden Sea Area. Rep. 1 of the Wadden Sea Working Group. 135 pp.

Donkersloot-de Vrij, Y.M., 1981: Topografische kaarten van Nederland vóór 1750, handgetekende en gedrukte kaarten - aanwezig in de Nederlandse rijksarchieven - toegelicht en beschreven. Thesis Utrecht, 223 pp.

Dijkstra, F., 1959: Esonsted, leginde of wierheit? Friesche Plaknammen, XI, 19-30.

Eckhardt, K.A. & Eckhardt, A., 1982: Lex Frisonium, Monumenta Germainia Historica, Fontes Iuris Germanici Antiqui XII, Hanover.

Edelman, T., 1952: Exploitatie van turf ver voor de Middeleeuwen. Tijdschrift K.N.A.G., 2nd ser., 4, 69, 476-477.

Edelman, T., 1964: De historische veranderingen in de natuurlijke gesteldheid van het Nederlandse Waddengebied. In: Anderson, W.F., Abrahamse, J., Buwalda, J.D. & Van Straaten, L.M.J.U. (Eds): Het waddenboek. Nederlandse Geologische Vereniging, Thieme & Cie, Zutphen, 37-65.

Edelman, T., 1974: Bijdrage tot de historische gegrafie van de Nederlandse kuststrook. - Rijkswaterstaat, Dir. Waterhuishouding en waterbeweging, The Hague, 84 pp.

Edes, B. & Glashouwer, C., 1990: Uit de geschiedenis van de Amelander kerken, 152 pp.

Ehlers, J., 1986: The morphodynamics of the Wadden Sea. Balkema, Rotterdam, 397 pp.

Eibergen, P., 1941a: De grenzen van de provincie Groningen. Tijdschrift van het Ned. Aardr. Gen., 2e reeks, 58, 173-180.

Eibergen, P., 1941b: De stroomrichting der wateren in de provincie Groningen. Tijdschrift van het Ned. Aardr. Genootschap, 2e reeks, 58, 466-475.

Eijndhoven, M.J., 1992: De Hollandse kust. Overzicht van het geologisch onderzoek naar de kustontwikkeling van de laatste 10.000 jaar, concept. RWS, DGW, afdeling AOFM, 35 pp.

Eisma, D. & De Wolff, W.J., 1980. The development of the westernmost part of the Wadden Sea in historical time. in: Dijkema, K.S., Reineck, H.E. & Wolff, W.J. (Eds) Geomorphology of the Wadden Sea Area. Rep. 1 of the Wadden Sea Working Group, 95-103.

Elton, O.B.A., 1894: The first nine books of the Danish History of Saxo Grammaticus (translated in english), London David Nutt 270-271, Strand.

Elzinga, G., 1969: Het verdronken Esonstad: werkelijk "Dichtung"? Waddenbulletin, 4, 1, 8-12.

Eysink, W.D. 1979: Morfologie van de Waddenzee, gevolgen van zand- en schelpenwinning. Verslag literatuuronderzoek R 1336. Waterloopkundig Laboratorium, Delft, 92 pp

Eysink, W.D. & Biegel, E.J., 1992: Impact of sea level rise on the morphology of the Wadden Sea in the scope of its ecological function. Investigations on empirical morphological relations, Rijkswaterstaat Dienst Getijdewateren, ISOS2 Project, Phase 2, 73 pp.

Feenstra, H., 1990: Eerherstel voor de Schiere monniken. Bijdragen ta Pleatslike skiednis III, 1990, Fryske Akademy, 117-130.

Fenger, H., 1935: Friesland und England in ihren Kulturellen und wirtschaftlichen Beziehungen. Bonner Studien zur englischen Philologie, Heft XXV. Bonn, Peter Hanstein Verlagsbuchhandlung, 111 pp.

FitzGerald, D.M., 1988: Shoreline erosional-depositional processes associated with tidal inlets. In: Aubrey, D.G. and Weishar, L. (Eds): Hydrodynamics and sediment dynamics of tidal inlets, Springer Verlag, New York, Inc., 186-225.

FitzGerald, D.M., Penland, S. & Nummedal, D., 1984: Control of barrier island shape by inlet sediment bypassing: east frisian islands, West Germany, Mar. Geol., 355-376.

FitzGerald, D.M. & Penland, S., 1987: Backbarrier dynamics of the east Friesian Islands. J. Sed. Pet., 57, 746-754.

Flemming, B.W., 1991: Holozäne Entwicklung, Morphologie und fazielle Gliederung der Ostfrieschischen Insel Spiekeroog (südliche Nordsee). Senckenberg-am-Meer, Bericht 91/3, 51 pp.

Flemming, B.W. & Davis, Jr. R.A., 1992: Holocene evolution, morphodynamics and sedimentology of the Spiekeroog barrier island system (southern North Sea). - Tidal Clastics 92, Wilhelmshaven, Germany, Field Guide-Book, 32 pp.

Fockema Andreae, I.S.J., 1950: Studiën over waterschapsgeschiedenis, VI Oostelijk Groningen, Brill, Leyden., 43 pp.

Formsma, W.J. 1954: Een reis naar Schiermonnikoog in 1556. Groningse Volksalmanak 1954, 65-71.

Formsma, W.J. 1956: De heerlijkheid Schiermonnikoog. Groningse Volksalmanak 1956, 81-106.

Formsma, W.J., 1966: De oorlog tussen de Ernerenses en de Uthusenses in de dertiende eeuw. Groningse Volksalmanak 1965-1966, 1-16.

Franken, A.F., 1987: Rekonstruktie van het paleo-getijklimaat in de Noordzee, Delft Int. Rep., X, 0029-00, 74 pp.

Franz, G., 1967: Quellen zur Geschichte des deutschen Bauernstandes im Mittelalter, gesammelt und herausgegeben *(p. 326: Bartholomaeus Anglicus, written 1240 A.D.*. Wissenschaftliche Buchgesellschaft, Darmstadt, 626 pp.

Friedlaender, E., 1881: Ostfriesisches Urkundenbuch. Zweiter Band, 1471-1500, Emden, W.Haynel, p. 785 and onwards.

Geudeke, P.W. & Zandvliet, L., 1990: Grote Historische Atlas van Nederland, 1:50,000, 2 Noord-Nederland, 1851-1855, (facsimile), Wolters-Noordhoff, Groningen 110 pp.

Gosses, I.H., 1923: Deensche heerschappijen in Friesland gedurende den Noormannentijd. Mededeelingen der Koninklijke Akademie van Wetenschappen, Afdeeling Letterkunde, Deel 56, Serie B, No 4.

Gottschalk, M.K.E., 1971: Stormvloeden en rivieroverstromingen in Nederland (Stormsurges and river floods in the Netherlands) deel 1, de periode voor 1400. Sociaal Geografische Studies, 10, De Vries, A.C. (Ed.), Van Gorcum Assen, Amsterdam, 581 pp.

Gottschalk, M.K.E., 1975: Stormvloeden en rivieroverstromingen in Nederland (Stormsurges and river floods in the Netherlands) deel 2, de periode 1400-1600. Sociaal Geografische Studies, 13, De Vries, A.C., (Ed.), Van Gorcum Assen, Amsterdam, 896 pp.

Gottschalk, M.K.E., 1977: Stormvloeden en rivieroverstromingen in Nederland (Stormsurges and river floods in the Netherlands) deel 3, de periode 1600-1700. Sociaal Geografische Studies, 14, De Vries, A.C. (Ed.), Van Gorcum Assen, Amsterdam, 474 pp.

Gottschalk, M.K.E., 1980: Subatlantische transgressiefasen en stormvloeden. In: Verhulst, A. & Gottschalk, M.K.E. (Eds): Transgressies en occupatiegeschiedenis in de kustgebieden van Nederland en België. Belgisch centrum voor landelijke geschiedenis, publ. 66, 21-27.

Griede, J.W., 1978: Het ontstaan van Frieslands Noordhoek, Thesis, Vrije Univ. Amsterdam, 186 pp.

Griede, J.W. & Roeleveld, W., 1982: De geologische en paleogeografische ontwikkeling van het Noordelijk Zeekleigebied, K.N.A.G. Geog. Tijdschrift, 16, 439-543.

Halbertsma, H., 1969: Esonstad, bijdrage tot de ontstaansgeschiedenis der Friese steden. Waddenbulletin, 4, 1, 4-8.

Harley, J.B. & Woodward, D., 1987: The history of cartography. Volume one: Cartography in prehistoric, ancient, and medieval Europe and the Mediterranean. Chicago, University of Chicago Press, 600 pp.

Hayes, M.O., 1975: Morphology of sand accumulations in estuaries. In: Cronin, L.E. (Ed.): Estuarine Research vol. 2 Geology and Engineering. Acad. Press, N.Y., 3-22.

Hayes, M.O., 1979: Barrier Island morphology as a function of tidal and wave regime. In: S. Leatherman (Ed.): Proceedings of the coastal symposium on Barrier islands. Academic Press, N.Y., 1-28.

Hayes, M.O., 1980: General morphology and sediment patterns in tidal inlets. Sed. Geol. 26, 139-156.

Hayes, M.O. & Kana, T.W., 1976: Terrigenous clastic depositional environment. Tech. Rpt. No. CRD-11, Dept. of Geology, Univ. of South Carolina, Columbia, SC, 364 pp.

Heslinga, M.W., De Klerk, A.P., Schmal, H., Stol, T. & Thurkow, A.J., 1985: Nederland in kaarten: verandering van stad en land in vier eeuwen cartografie. Zomer en Keuning, Ede, 216 pp.

Hofstede, J.L.A., 1991: Sea level rise in the inner German Bight (Germany) since AD 600 and its implications upon tidal flats geomorphology. In: Brückner, H. & Radtke, U. (Eds): Von der Nordsee bis zum Indischen Ozean. Ergebnisse der 8. Jahrestagung des Arbeitskreises" Geographie der Meere und Küsten", 13-15 Juni 1990, Düsseldorf. Franz Steiner Verlag, Stuttgart, 11-27.

Hohlbaum, K., 1879: Hansisches Urkundenbuch, Band II, Halle a. S., 395 pp.

Hooykaas, J., 1969: Van Amsterdam tot de stadt Hamborch toe. Waddenbulletin, 9-13.

Huber, H.H., 1879: Duinen Ameland. Advies Mr. H.H. Huber, advocaat te Leeuwarden, over de aanspraken van het Rijk op duingronden en kwelders op het eiland Ameland, maart 1879., 24 pp.

Huussen, A.H., 1979: Het Fries-Groningse geschil voor de Geheime Raad te Brussel over het leggen der tonnen in de Scholbalg (1550-1556), met enige aantekeningen over de rol van de Hollandse landmeter Jasper Adriaensz. Groningse Volksalmanak historisch jaarboek voor Groningen, 1978-1979, 87-115.

Isbary, G., 1936: Das Inselgebiet von Ameland bis Rottumeroog. Morphologische und hydrographische Beiträge zur Entwicklungsgeschichte der friesischen Inseln. Archiv der deutschen Seewarte, 56, 3, 55 pp.

IJnsen, F., 1974: Onderzoek naar de basispeilen in het gebied van de Westelijke Waddenzee, RWS, Studiedienst Hoorn, Nota 74.10, 25 pp.

IJnsen, F., 1977: Over de invloed van troposferische circulatiepatronen op het ontstaan van stormvloeden in de Noordzee. Rijkswaterstaat, Dir. Friesland, Afd. ANP, rapport 77.01, 27 pp.

IJnsen, F., 1987: Getijverlaging door afwaaiing. Harlingen, Terschelling, Griend Steekproef 1940-1985. Rijkswaterstaat, Directie Friesland, Notitie ANW 87-20.

IJnsen, F., 1988: Het karakteriseren van winters. Zenit, 18, 55-58.

IJnsen, F., 1991a: Karaktergetallen voor de winters vanaf 1705. Zenit, 18, 55-58.

IJnsen, F., 1991b: Karaktergetallen voor de zomers vanaf 1706. Zenit, 18, 313-315.

IJnsen, F., 1992: Beschouwing over de algemene circulatie in verband met het optreden van stormvloeden. RWS Memono: ANW 92.06.03.FIJ.

IJnsen, F., 1994: Karaktergetallen en temperaturen voor winters en zomers, 1591-1705/06. Stiens, 17 pp.

IJnsen, F., 1995a: Jaar- en halfjaartemperaturen en maritimiteitsinidices, 761-1990, Memo S 95.01.29.FY, 1 pp.
IJnsen, F., 1995b: Detectie tijdvak Kleine IJstijd. Memo 95.02.19.FY, 3 pp.
IJnsen, F. & Buisman, J., 1993: Winters en zomers in het tijdvak 1591-1706. Meteorologica, 2, 36-39.
Jansen, P.Ph., 1941: Groninger Wadden. Rijkswaterstaat, Studiedienst Hoorn, no. C 110, 132 pp.
Jarrett, J.T., 1976: Tidal prism-inlet area relationships. GITI, Rep. 3, US Army Eng. Waterways Exp. Stat. Vicksburg.
Jelgersma, S. 1979: Sea-level changes in the North Sea basin. In Oele, E., Schüttenhelm, R.T.E. & Wiggers, A.J. (Eds), The Quaternary History of the North Sea, 233-248.
Jensen, J., Hofstede, J.L.A., Kunz, H., De Ronde, J.G., Heinen, P.F. & Siefert, W., 1993: Long term water level observations and variations. In: American Society of Civil Engineers (Eds), Coastlines of the southern North Sea, 115-136.
Joustra, D. Sj., 1971: Geulbeweging in de buitendeltas van de Waddenzee. Rijkswaterstaat, studiedienst Hoorn, Studierapport WWK 71-14, 1-27.
Kalma, J.J., 1978: Christianus Schotanus, Franeker, 1664, Beschrijvinge van de Heerlyckheydt van Frieslandt tussen 't Flie end de Lauwers met nieuwe Caerten van 't Landschap in 't algemeen soo Oud als Nieuw end De drie Gedeelten in 't besonder als mede De Grietenien end Steden Op kosten van 't Landt afgemeten In't net gebracht, geteeckent en gesne den op Copere platen. Toegelicht door. Facsimile (280 pp.), Theatrum Orbis Terrarum, Amsterdam, 11 pp.
Keuchenius Driessen, R., 1824: Monumenta Groningana of verzameling van onuitgegevene oude charters en stukken betreffende de provincie Groningen aanvang nemende met de vroegste tijden en eindigende met het laatste van de veertiende eeuw, 2e stuk, Oomkens, Groningen.
Keuchenius Driessen, R., 1827: Monumenta Groningana veteris aevi inedita, of verzameling van onuitgevende oude charters en stukken, betreffende de provincie Groningen; aanvang nemende met de vroegste tijden, en eindigende met het laatste van de veertiende eeuw. 321 pp.
Keuning, J., 1950a: Cornelis Anthonisz. Imago Mundi, VII, 51-65.
Keuning, J., 1950b: Cornelis Anthonisz, zijn Caerte van Oostlant, zijn Onderwijsinge vander zee en zijn Caerte van die Oosterse See. Tijdschrift K.N.A.G., 2e reeks, 67, 6, 687-714.
Kiden, P., 1989: The Late Holocene evolution of the perimarine part of the river Scheldt. In: Henriet, J.P. & De Moor, G. (Eds): The Quarternary and Tertiary geology of the Southern Bight, North Sea, Min. Ec. Aff. Belgian Geol. Survey, 173-184.
Knol, E., 1991: Het kustland in de donkere eeuwen. Jaarverslagen van de Vereniging voor Terpenonderzoek, 75, 118-132.
Knol, E., 1993: De Noordnederlandse kustlanden in de Vroege Middeleeuwen. Thesis, Groningen, 272 pp.
Knudsen, J., 1914: De kaert vander Zee van Jan Seuerszoon (1532), reprint. Het oudste gedrukte Nederlandsche leeskaartboek.'s-Gravenhage, Nijhoff, 99 pp.
Knudsen, J., 1920: Het leeskaartboek van Wisbuy. Gedrukt te Antwerpen by Jan Roelants en te koop te Amsterdam by Hendrick Albertsz (1566), reprint. Met inleiding door Dr. C.P. Burger Jr. Nijhoff, 's-Gravenhage,221 pp.
Koeman, C., 1967-85: Atlantes Neerlandici: bibliography of terrestrial, maritime and celestical atlases and pilotbooks, published in the Netherlands up to 1880. Theatrum Orbis Terrarum, 6 dln.
Koeman, C., 1973: The Sea-beacon by Willem Jansz. Blaeu (1643), facsimile, with an introduction by. Theatrum Orbis Terrarum Ltd. Amsterdam.
Koeman, C., 1985: Geschiedenis van de kartografie van Nederland. Zes eeuwen land- en zeekaarten en stadsplattegronden. Canaletto, Alphen aan den Rijn, 2nd pr., 293 pp.
Koeman, C. & Kalma, J.J., 1970: Uitbeelding der Heerlijheit Friesland; zoo in 't algemeen als in haare XXX bijzondere Grietenijen door Bern. Schotanus à Sterringa, nu niewelijks met bijgevoegde aangrenzingen, en veel vermeerderingen, nevens d'afteekening van Oud Firesland in VII ver-

scheidene landkaarten; door den heere Menso Alting, 1718. Met inleiding en bio-bibliografische beschrijving door. Facsimile, Theatrum Orbis Terrarum, Amsterdam, 16 pp.

Koenders, M., 1986: Het kaartbeeld van de Friese Waddeneilanden, 1540-1860. Een historisch-kartografisch onderzoek naar de ontwikkeling van de nauwkeurigheid van het kaartbeeld van de tegenwoordige Friese Waddeneilanden tussen ca. 1540 en 1860. Rijksuniversiteit Utrecht, Internal Report Hist. Kartog., 78 pp. (Also published in Caert-thresoor, tijdsch. Ges. kartografie in Nederland, 1986, 4, 65-72.)

Koenders, M., 1987: Oude kaarten van de Friese waddeneilanden. Waddenbulletin, 22, 3, 135-141.

Kolk, B. & Schalkers, K.M., 1980: De veranderingen in de Waddenzee ten gevolge van de afsluiting van de Zuiderzee, Rijkswaterstaat, Directie waterhuishouding en waterbeweging, Notitie 78.H238, 13 pp.

Kooper, J., 1939: Het waterstaatsverleden van de provincie Groningen. Bijdrage tot de kennis van de provincie Groningen en omgelegen streken.

Koopstra, R., Lenselink, G. & Menke, U., 1993: Geologische kaart IJsselmeer, Dienst IJsselmeerpolders, 8 pp.

Koppmann, K., 1876: Das Seebuch. Mit einer nautischen Einleitung von Arthur Breusing. Mit Glossar von Christoph Walther. Niederdeutsche Denkmaler, herausgegeben vom Verein fur niederdeutsche Sprachforschung. J. Kuhtmann's Buchhandlung, Bremen, 120 pp.

Kros, M., 1848: Memorie wegens den vroegeren en lateren staat der beide Vriesche Eilanden Ameland en Schiermonnikoog. RAF, Archief Gedeputeerde Staten, no. 11201.

Kunze, K., 1896: Hansisches Urkundenbuch, vierter Band Hanze-reces in 353 p. 148-149, Halle a. S., 522 pp.

Kusternig, A., 1982: Die vier Bücher der Chroniken des sogenannten Fredegar. Quellen zur Geschichte des 7. und 8.Jahrhunderts, Wissenschaftliche Buchgesellschaft, Darmstadt, 567 pp.

Labrijn, A., 1945: The climate of the Netherlands during the last two and a half centuries. Meded. Verhand. KNMI, 102/49, 114 pp.

Lamb, H.H., 1977: Climate, present, past and future, vol. 2, Climate history and the future, 835 pp. Methuen & Co Ltd London.

Lamb, H.H., 1980: Climatic fluctuations in historical times and their connexion with transgressions of the sea, storm floods and other coastal changes. In: Verhulst, A. & Gottschalk, M.K.E. (Eds): Transgressies en occupatiegeschiedenis in de kustgebieden van Nederland en België. Belgisch centrum vor landelijke geschiedenis, publ. 66, 251-284.

Lang, A.W., 1955: Traces of lost North European sea charts of the 15th century. Imago Mundi, 12, 31-44.

Lang, A.W., 1957: Die älteste gedruckte Seekarte der Ems. Nordseeküste, 1, 5-19, Juist, Die Bake.

Lang, A.W., 1958: Gestaltungswandel des Emsmundungstrichters. Untersuchungen zur Entwicklung der Emsmundung von der Mitte des 16. bis zum Beginn des 20. Jahrhunderts. Veroffentlichungen des Niedersachisischen Amtes fur Landesplanung und Statistik, Reihe A: Forschungen zur Landes- und Volkskunde. 1. Natur, Wirtschaft, Siedlung und Planung (Schriften der Wirtschaftswissenschaflichen Gesellschaft zum Studium Niedersachsens e.V.) Bd.58, Walter Dorn Verlag, Bremen, 1-153.

Lang, A.W., 1959: II: Textband. In: Mortensen, H. & Lang, A. (Eds): Die Karten deutscher Länder im Brüsseler Atlas des Christian s'Grooten (1573). Abhandlungen der Akademie der Wissenschaften in Göttingen philologisch-historische Klasse, Dritte Folge, 44, 80 pp.

Lang, A.W., 1968: Seekarten der südlichen Nord- und Ostsee. Deutschen Hydrographischen Zeitschrift (Ergänzungsheft), Reihe B (4), 10, 105 pp. Borntraeger, Berlin.

Lang, A.W., 1986: Die "Caerte van oostlant" des Cornelis Anthonisz. 1543, die älteste gedruckte Seekarte Nordeuropas und ihre Segelanweisung. Schriften des Deutschen Schiffahrtsmuseums, 8, Ernst Kabel, Hamburg, 167 pp.

Lelewel, J., 1966: Géographie du Moyen Age, tome III and IV, (Edrisi). Meridian Publ. Co., Amsterdam, 220 pp.

Libby, L.M., 1987: Evalution of historic climate and prediction of near-future climate form stable-isotope variations in tree rings. 81-89.

Ligtendag, W.A., 1990: Van IJzer tot Jade, Een reconstructie van de zuidelijke Noordzeekust in de jaren 1600 en 1750, Rijkswaterstaat, 's-Gravenhage, 55 pp. and bijlagen.

Ligtendag, W.A., 1991: Oude kaarten als kenbron voor verleden en toekomst. Historisch-Geografisch Tijdschrift, 3, 77-87.

Luck, G., 1975: Der Einfluss der Schutzwerke der Ostfriesischen Inseln auf die morphologische Vorgänge im bereich der Seegaten und ihrer Einzugsgebiete, Mitteliungen des Leichtweiss-Instituts für Wasserbau der Techn. Univ. Braunschweig, Heft 47, 122 pp.

Mansholt D.R., 1929: Kort geschiedkundig overzicht van het Amelander Wad, 38 pp.

Mars, S. & Cordia, P., 1901: Gids voor de Zuiderzee en de Noordelijke Zeegaten. Handleiding bij het gebruik van Seyffardt's Zuiderzeekaart. Bom, Amsterdam, 178-194.

Mazure, J.P. et al., 1974: Rapport van de Waddenzeecommissie; Advies inzake de principiële mogelijkheden en de voor- een nadelen van inpolderingen in de Waddenzee, Rijkswaterstaat - Algemene Zaken, Den Haag, 328 pp.

Mellema, L., 1981: Schiermonnikoog, lytje pole. Knoop, Haren, 2e druk, 280 pp.

Meurer, P.H., 1984: Mappae germaniae: die schönsten und bedeutendsten Deutschlandkarten von 1482 bis 1803. Pfaehler, Bad Neustadt a.d. Saale, 72 pp.

Midderham, H., 1979: Ontwikkeling Boschplaat-Boschgat tot 1978. Rijkswaterstaat, Directie Waterhuishouding en waterbeweging, district kust en zee, studiedienst Hoorn, Notitie WWKZ-79.H249, 4 pp.

Milikowski, E., 1983: Voormalige landindeling op Schiermonnikoog. Waddenbulletin, 18, 4, 149-153.

Mörner, N.A., 1995: Recorded sea level variability in the Holocene and expected futere changes. In: Eisma, D. (Ed.): Climate change impact on coastal habitation, Lewis Publishers, London, 17-28.

Mortensen, H. & Lang, A., 1959: Die Karten Deutscher Länder im Brüsseler Atlas des Christian S'Grooten (1573). Abhandlungen der Akademie der wissenschaften in Göttingen Philologisch-Historische Klasse, Dritte Folge, 44, 80 pp.

Niemeijer, J.A., 1975: Land van terpen en dijken. Boekencentrum bv, the Hague, 2nd print, 118 pp.

Oost, A.P., 1995: Sedimentological implications of morphodynamic changes in the ebb-tidal delta, the inlet, and the drainage basin of the Zoutkamperlaag tidal inlet (Dutch Wadden Sea), induced by a sudden decrease in the tidal prism. In: Flemming, B.W. & Bartholomä, A. (Eds.): Tidal signatures in modern and ancient sediments. Int. Assoc. Sedimentol., Spec. Publ., 24: 101-119.

Oost, A.P. & De Haas, H., 1992: Het Friesche Zeegat. Morfologisch-Sedimentologische veranderingen in de periode 1970-1987; een getijde inlet systeem uit evenwicht. Rapport in het kader van Kustgenese, Werkgroep 1 voor Rijkswaterstaat, Deel 1 (68 pp.) and 2 (Fig.).

Oost, A.P. & K.S. Dijkema, 1993: Effecten van bodemdaling door gaswinning in de Waddenzee, IBN-rapport 025, 134 pp.

Oost, A.P. & De Haas, H., 1993: Het Friesche Zeegat. Morfologisch-Sedimentologische veranderingen in de periode 1927-1970. Rapport in het kader van Kustgenese, Werkgroep 1 voor Rijkswaterstaat, Deel 1 (83 pp.) and 2 (Fig.).

Otto, L., Zimmermann, J.T.F., Furnes, G.K., Mork, M., Saetre, R. & Becker, G., 1990: Review of the physical oceanography of the North Sea. Neth. J. Sea Res., 26, 161-238.

Overdiep, G., 1958: De grens tussen de provincies Groningen en Friesland in de Wadden. Groningse Volksalmanak 1958, 27-42.

Overdiep, G., 1964: Toelichting bij de kaart van Ameland uit 1731. 23 pp.

Overdiep, G., 1968: Ameland kort vóór en in de Franse tijd. Waddenbulletin, 3, 2, 2-4.

Platner, C., 1867: Die grossen kölnischen Jahrbücher, nach der Ausgabe der Monumenta Germaniae übersetszt *(p. 82: 1170 A.D. contemporanous perhaps by Burchard.* Die Geschichtschreiber der deutschen Vorzeit, 13th Cent., 1. Band, Verlag von Franz Duncker, Berlin, 272 pp.

Pons, L.J. & Wiggers, A.J., 1960: De holocene wordingsgeschiedenis van Noordholland en het Zuiderzeegebied, deel II. Tijdschrift KNAG, XX, 3-57.

Poortinga, Y., 1956: Wierumer Wad- en Noardse-Toponymen, Fryske Plaknammen VIII, 30-36.

Postma, H. & Dijkema, K.S., 1982: Hydrography of the Dutch Wadden Sea: Movements and properties of water and particulate matter. - Final report on 'Hydrography' of the Wadden Sea Working Group, Report 2, Balkema, Rotterdam, 75 pp.

Rau, R., 1955: Die Reichsannalen, Einhards Leben Karls des Grossen, Zwei "Leben' Ludwigs, Nithard Geschichten, unter Benützung der Übersetzungen von O. Abel und J. v. Jasmund neu bearbeitet *(p. 95: Ann. Regni Francorum, contemporanous; p. 189: vita Caroli Magni by Einhard, written between 814-820 A.D.).* Quellen zur Karolingischen Reichsgeschichte, 1st part, Ausgewählte Quellen zur deutschen Geschichte des Mittelalters, 5th vol., Wissenschaftliche Buchgesellschaft E.V., Darmstadt, 484 pp.

Rau, R., 1958: Jahrbücher von St. Bertin, Jahrbücher von St. Vaast, Xantener Jahrbücher, unter Benützung der Übersetzungen von J. v. Jasmund und C. Rehdantz neu bearbeitet *(p. 41: Ann. Bertiniani, period 835/37-861 contemporanous by Prudentius of Troyes; p. 343 and 361: Ann. Xantenses, contemporanous by Gerward).* Quellen zur Karolingischen Reichsgeschichte, 2nd part, Ausgewählte Quellen zur deutschen Geschichte des Mittelalters, 6th vol., Wissenschaftliche Buchgesellschaft E.V., Darmstadt, 400 pp.

Rau, R., 1960: Jahrbücher von Fulda, Regino Chronik, Notker Taten Karls, unter Benützung der Übersetzungen von C. Rehdantz, E. Dümmler und W. Wattenbach neu bearbeitet *(p. 73: Ann. Fuldensis, written in the 9th century).* Quellen zur Karolingischen Reichsgeschichte, 3rd part, Ausgewählte Quellen zur deutschen Geschichte des Mittelalters, 7th vol., Wissenschaftliche Buchgesellschaft E.V., Darmstadt, 448 pp.

Reenders, R., 1986: Veranderingen Rottumeroog, periode 1539-1965 (historisch overzicht). Rijkswaterstaat, directie Groningen, Afdeling Advies en Onderzoek, nota ANA 86.17, 11 pp.

Reitsma, D.Th., 1974: Betekenis van de kaarten. Int. publ. RWS, 5 pp.

Reitsma, D.Th., 1979: Heffesant, int. rep., 2 pp.

Reitsma, D.Th., 1984: Waterstaatkundige historie Ameland. Rijkswaterstaat, Directie Friesland, 10 p.

Reitsma, D.Th., 1988: Pathmos: de Stachouwerboerderij op Schiermonnikoog. Hallum, AT-SEE, 63 pp.

Reitsma, D.Th., 1991a: Historie oostelijk waddengebied. Waddenbulletin, 26, 4, 176-178.

Reitsma, D.Th., 1991b: Simonszand. Manuscript, 8 pp.

Revier, H., 1991: Rottum, ja natuurlijk! Waddenbulletin, 26, 2, 71.

Rienks, K.A. & Walther, G.L., 1955: Zur Geschichte der Binnendeiche und Schlafdeiche im westerlauwerschen Friesland. Friesisches Jahrbuch 1955, 123 pp.

Roeleveld, W., 1974: The Groningen coastal area: A study in Holocene geology and low-land physical geography, Reprint: Berichten van de Rijksdienst voor Oudheidkundig bodemonderzoek, Vol. 20-21, 1970-'71 and Vol. 24, 1974, Supplement.

Roelse, P., 1988: Zandsuppleties in Nederland. In: Roelse, P. (Ed.): Handboek zandsuppleties. RWS, The Hague, 33-46.

Roep, Th.B. & Beets, D.J., 1988: Sea level rise and paleotidal levels from sedimentary structures in the coastal barriers in the western Netherlands since 5600 B.P.. Geol. Mijnb., 67, 1, 53-60.

Rogge, H.C., 1885: Dit is die Caerte van der zee: om Oost en West te zeylen. Réimpression, Leide, Brill, 80 pp.

RWS, 1948: Beschrijving van de provincie Friesland, behorende bij de Waterstaatskaart, bijgewerkt tot 1 januari 1948. Staatsdrukkerij en uitgeverijbedrijf 's-Gravenhage, 686 pp.

RWS, 1959: Beschrijving van de provincie Groningen, behorende bij de Waterstaatskaart, bijgewerkt tot 1 januari 1959. Staatsdrukkerij en uitgeverijbedrijf 's-Gravenhage, 255 pp.

RWS, 1994: Verslag van de stormvloed van 13 en 14 maart 1994 (SR72). Min. van Verkeer en Waterstaat RIKZ, 18 pp.

Saxo Grammaticus, 1145-1185: The first nine books of the Danish History of Saxo Grammaticus, translated by Oliver Elton, B.A., London David Nutt 270-271, Strand, 1894.

Schilder, G., 1991: A Dutch manuscript rutter: an unique portrait of the European coasts in the late sixteenth century. Imago Mundi, 43, 59-71.

Schmeidler, B. & Lappenberg, J.M., 1910: Helmolds Chronik der Slaven, nach der Ausgabe der Monumenta Germainiae übersetszt von J.C.M. Laurent und W. Wattenbach *(97, p. 227: Chronica Slavorum, written 1164 A.D. by Helmold, contemporanous)*. Die Geschichtschreiber der deutschen Vorzeit, Band 44, Verlag der Dutschen Buchhandlung, Leipzig, 3rd Ed. 271 pp.

Schönfeld, M., 1951: Scholbalg. Fryske Plaknammen IV, 82-84.

Schoorl, H., 1973: Zeshonderd jaar water en land. Bijdrage tot de historische Geo- en Hydrografie van de Kop van Noord-Holland in de periode 1150-1750. Verhandeling van het Koninklijk Nederlands Aardrijkskundig Genootschap, Nr.2, Wolters-Noordhoff, Groningen, 534 pp.

Schoorl, H., in prep.: De Convexe Kustboog, Deel I (Alg. Overzicht) & II (Terschelling).

Schortinghuis, D.H., 1975: Cleyn eilant Rottum. 2nd pr., Knoop & Niemeyer, Haren, 120 pp.

Schweingruber, F.H., Bartholin, Th., Schär, E. & Brifa, K.R., 1988: Radiodensitometric-dendroclimatological conifer chronologies from Lapland (Scandinavia) and the Alps (Switzerland). Boreas, 18, 559-566.

Sha, L.P., 1989a: Cyclic morphologic changes of the ebb-tidal delta, Texel Inlet, The Netherlands. Geol. Mijnb., 68, 35-48.

Sha, L.P., 1989b: Variation in ebb-delta morphologies along the West and East Frisian Islands, the Netherlands and Germany. Marine Geol. 89, 11-28.

Sha, L.P., 1990: Sedimentological studies of the ebb-tidal deltas along the West Frisian Islands, the Netherlands. Phd-Thesis. Geologica Ultraiectina. Mededelingen van het Instituut voor Aardwetenschappen der Rijksuniversiteit te Utrecht, 64, 159 pp.

Sha, L.P., 1992: Geological Research in the Ebb-Tidal Delta of 'Het Friesche Zeegat', Wadden Sea, The Netherlands, Rapport R.G.D.-Projekt 40010, 20 pp.

Sha, L.P. & de Boer, P.L., 1991: Ebb-tidal delta deposits along the West Frisian Islands (The Netherlands): processes, facies architecture and preservation. In: Smith, D.G., Reinson, G.E., Zaitlin, B.A. & Rahmani, R.A. (Eds), Canadian Soc. Petrol. Geol. Mem. 16, 199-218.

Sha, L.P. & Van den Berg, J.H. 1993: Variation in ebb-tidal delta geometry along the coast of the Netherlands and the German Bight. J. Coastal Res., 9, 730-746.

Sipma, P., 1927: Oudfriesche oorkonden, eerste deel, Nr.326, 227-228.

Skelton, R.A., 1964: Spieghel der zeevaerdt, by Lucas Jansz Waghenaer, 1584-1585, facsimile, with an introduction by. Theatrum Orbis Terrarum Ltd. Amsterdam.

Skelton, R.A., 1965: Thresoor der zeevaert by Lucas Jansz Waghenaar, Leyden 1592, facsimile, with an introduction by. Theatrum Orbis Terrarum Ltd. Amsterdam.

Steijn, R.C., 1991: Some Considerations on Tidal Inlets. A literature survey on hydrodynamic and morfodynamic characteristics of tidal inlets with special attention to 'Het Friesche Zeegat', Literature Survey H 840.45., 109 pp.

Steijn, R.C. & Louters, T., 1992: Hydro- and morphodynamics of a mesotidal inlet in the Dutch Wadden Sea. Oceanology International 92 Conference, 9 pp.

Steijn, R.C., Hartsuiker, G. & Van Banning, G.K.F.M., 1992: Morphodynamic behaviour of "Het Friesche Zeegat", numerical simulation of hydro- and morphodynamics of a mesotidal inlet in the Dutch Wadden Sea, 94 pp.

Steinberg, S. & Schmeidler, B., 1926: Adam von Bremen hamburgische Kirchengeschichte, nach der Ausgabe der Scriptores reum Germanicarum *(p. 209, Scriptores, B.4,3: Adam von Bremen, written 1066-1075 A.D., contemporanous)*. Die Geschichtschreiber der deutschen Vorzeit, Band 44, Verlag der Dutschen Buchhandlung, Leipzig, 3rd Ed. 271 pp.

Stiboka, 1973: Bodemkaart van Nederland, Blad 7 West Groningen. Wageningen, Stichting voor Bodemkartering, 149 pp.

Stiboka, 1976: Bodemkaart van Nederland, Blad 5 West Harlingen, Blad 5 Oost Harlingen. Wageningen, Stichting voor Bodemkartering, 95 pp.

Stiboka, 1981: Bodemkaart van Nederland, Blad 6 West Leeuwarden, Blad 6 Oost Leeuwarden. Wageningen, Stichting voor Bodemkartering, 181 pp.

Stoob, H., 1963: Helmold von Bosau, Slawenchronik, neu übertragen und erläutert *(p. 339: Chronica Slavorum, written 1164 A.D. by Helmold, contemporanous)*. Ausgewählte Quellen zur deutschen Geschichte des Mittelalters, 19th vol., Wissenschaftliche Buchgesellschaft E.V., Darmstadt, 399 pp.

Streif, H., 1987: Barrier islands, tidal flats, and coastal marshes resulting form a relative rise of sea level in East Frisia on the German North Sea coast. In: Van der Linden, W.J.M., Cloetingh, S.A.P.L., Kaasschieter, J.P.K., Van de Graaff, W.J.E., Vandenberghe, J. & Van der Gun, J.A.M. (Eds): Proc. KNGMG Symposium 'Coastal Lowlands, Geology and Geotechnology', 213-223.

Stuart, P., 1991: De Tabula Peutingeriana, I, map and II, comment. Museumstukken, Gerhartl Witteveen, A.M. & Stuart, P. (Eds), 31 and 20 pp.

Tangl, M., 1920: Leben des h. Bonifazius von Willibald bis Otloh, der h. Leoba von Rudolf von Fulda, des Abtes Sturmi von Eigil, nach den Ausgaben der Monumenta Germaniae übersetzt *(1st vita, written 754-768 A.D.; 2nd vita, written 800-850 A.D)*. Die Geschichtschreiber der deutschen Vorzeit, 13, 8th Cent., 2nd vol., 3rd Ed., Verlag der Dutschen Buchhandlung, Leipzig, 144 pp.

Ufkes, T., 1988: Stranden op Bosch in de 17de eeuw. Waddenbulletin, 23, 2, 80-81.

Ufkes, T., 1989a: Rottumeroog in de 17de eeuw. Waddenbulletin, 24, 1, 39-41.

Ufkes, T., 1989b: Rottumeroog in de 17de eeuw, deel 2. Waddenbulletin, 24, 4, 207-209.

Van Buijtenen, M., 1954: De Fries-Groningse grens in Lauwerszee en Wadden. Studie op verzoek van Ged. Staten van Friesland, Laverman N.V. Drachten, 60 pp.

Van de Plassche, O., 1982: Sea-level change and water-level movements in the Netherlands during the Holocene, Phd-thesis, V.U. Amsterdam, 138 pp.

Van de Plassche, O., in press: Evolution of the intra-coastal tidal range in the Rhine-Meuse delta and Flevo lagoon, 5700-3000 yrs cal B.C. Marine Geology, 20 pp.

Van de Plassche, O. & Roep, T.B., 1989: Sea-level changes in the Netherlands during the last 6500 years: basal peat vs. coastal barrier data. In: Scott, D.B., Pirazzoli, P.A. & Honig, C.A. (Eds), Late Quaternary sea-level correlation and application, NATO ASI Series, Series C: Mathematical and Physical Sciences, vol. 256, 41-56.

Van Es, W.A., 1981: De Romeinen in Nederland. Fibula- Van Dishoeck, Haarlem, 271 pp.

Van de Kreeke, J. & Robaczewska, R., 1993: Tide-induced residual transport of coarse sediment; application to the Ems estuary. Neth. Journ. Sea Research, 31, 3, 209-220.

Van de Ven, G.P., (Ed.), 1993: Leefbaar laagland. Geschiedenis van de waterbeheersing en landaanwinning in Nederland, Matrijs, Utrecht, 303 pp.

Van der Molen, S.J., 1968: De afslag van Ameland en Schiermonnikoog na de stormvloed van 1825. Waddenbulletin, 3, 15-16.

Van der Spek, A.J. & Beets, D.J., 1992: Mid-Holocene evolution of a tidal basin in the western Netherlands: a model for future changes in the norhtern Netherlands under conditions of accelerated sea-level rise? Sed. Geol., 80: 185-197.

Van der Spek, A.J., 1994: Large-scale evolution of Holocene tidal basins in the Netherlands. PhD-thesis, Univ, of Utrecht, 191 pp.

Van der Top, A.M.I., 1992: De cirkelmethode; nauwkeurigheidsonderzoek van een reeks kaarten uit de periode 1559-1820. Doktoraalscriptie, Rijksuniv. Utrecht, faculteit Ruimtelijke Wetenschappen, vakgroep Kartografie, 49 pp. + bijlagen.

Van der Wal, J., 1988: Dokkum: stad bij het wad. Waddenbulletin, 23, 1, 39-42.

Van Giffen, A.E., 1964: De ouderdom onzer dijken. Tijdschrift K.N.A.G., tweede reeks, deel LXXXI, No. 3, 273-286.

Van Leeuwen, J., 1826: Geschiedkundig tafereel van de watervloed en de overstroomingen in de provincie Vriesland, voorgevallen in de Sprokkelmaand 1825.

Van Oosten, M.F., 1986: Toelichting bij de kaarten van de Waddeneilanden Vlieland, Terschelling, Ameland en Schiermonnikoog. Stichting voor Bodemkartering, Wageningen, bodemkaart van Nederland Schaal 1:50 000, 29 pp.

Van Staalduinen, C.J., 1977: Geologisch onderzoek van het Nederlandse Waddengebied. State Geol. Survey, Haarlem, 77 pp.

Van Straaten, L.M.J.U., 1964: De Bodem der Waddenzee. In: Het Waddenboek. Nederl. Geol. Ver., Thieme, Zutphen, 75-151.

Van Veen, J., 1945: Bestaat er een geologische bodemdaling te Amsterdam sedert 1700? Tijdschrift K.N.A.G., 2, 62, 2-36.

Verhoeff, J.M., 1983: De oude Nederlandse maten en gewichten. Publikaties van het P.J. Meertens-Instituut voor Dialectologie, Volkskunde en Naamkunde van de Koninklijke Nederlandse Akademie van Wetenschappen, deel 3, 2e druk, 132 pp.

Verwijs, E. & Verdam, J., 1889: Middelnederlandsch Woordenboek. Martinus Nijhoff, 'sGravenhage, tweede deel, 2212 pp.

Vredenberg-Alink, J.J., 1974: De kaarten van Groningerland, de ontwikkeling van het kaartbeeld van de tegenwoordige provincie Groningen. Bakker, Uithuizen, 151 pp.

Vroom, M.G., Steyaert, F.H.I.M., Misdorp, R., Venema, H., Elgershuizen, J.H.B.W., Rakhorst, H.D. & Damoiseaux, M.A., 1989: Wadatlas, Min. Verkeer en Waterstaat, 95 pp.

Wang, Z.B., 1991: A morphodynamic model for a tidal inlet, 'Het Friesch Zeegat', Delft Hydraulics.

Warrick, R. & Oerlemans, J. 1990: Sea level rise. In: Houghton, J.T., Jenkins, G.J. & Ephraums, J.J. (Eds).In: Climate change, the IPCC scientific assessment, Cambridge University Press, 259-261.

Waterbolk, H.J., Waterbolk, H.T., 1992: Waddenbulletin, 27, 2, 70-77.

Wattenbach, W., Grandaur, G. & Laurent, M., 1888: Die Lebensbeschreibungen des hl. Willibrord, Gregors vun Utrecht, Liudgers und Willehads von Bremen, nach den Ausgaben der Monumenta Germaniae übersetzt *(p. 75 and 80: vita Liudgeri, written 839-849 A.D.; p. 96: vita of Willehad).* Die Geschichtschreiber der deutschen Vorzeit, 8th Cent., 3rd vol., Verlag der Dutschen Buchhandlung, Leipzig, 128 pp.

Winkler Prins, A., 1867: Geschiedenis en beschrijving van het eiland Schiermonnikoog. Loman & Verster, Amsterdam, 109 pp.

Winsemius, P., 1622: Chronique ofte historische geschiedenisse van Friesland, beginnende van den jaere na des werelts scheppinge 3635 ende loopende tot den haere nae de geboerte Christi 1622. Franeker, 611 pp.

Winter, H., 1954: Catalan Portolan maps and their place in the total view of cartographic development. Imago Mundi, XI, 1-12.

Wolff, W.J., 1986: De Waddenzee: eigenschappen van een dynamisch kustgebied. Flevobericht 252, RIJP, 11-22.

Wumkes, G.A., 1900: Tussen Flie en Borne. Schetsen uit de geschiedenis van Schellingerland. Manuscript, 156 pp.

Zagwijn, W.H., 1985: De paleogeographische ontwikkeling van Nederland in de laatste drie miljoen jaar. Geog. Tijdsch. 9, 181-201.

Zagwijn, W.H., 1986: Nederland in het Holoceen, Geologie van Nederland, Deel I, State Geol. Surv., Staatsdrukkerij, The Hague, 46 pp.

Zagwijn, W.H., Beets, D.J., van den Berg, M., Van Montfrans, H.M. & Van Rooijen, P., 1985: Atlas van Nederland in 20 delen, Deel 13 Geologie, 23 pp.

Zuidema, W. & Douma, J., 1972: Kronieken van de abdij Bloemhof te Wittewierum, loopende over de dertiende eeuw, door de Abten Emo, Menko en een Ongenoemde. In opdracht van het Hoofdbestuur der Algemeene Vereeniging "Groningen" uit het Latijn vertaald en uitgegeven met medewerking van Dr. C.D.J. Brandt. Broekhoff N.V. v/h Kemink en zoon over den Dom te Utrecht, 268 pp.

APPENDIX 1: CHRONOLOGICAL[90] OVERVIEW OF THE CHARTS, MAPS, AND SAILING DIRECTIONS USED

General

Terminology used

Chart: map used for navigation, such as the maps of Waghenaer.
Map: Land map.
Portulanos: charts drawn mostly on vellum.
Sailing direction: written navigation direction (rutter), sometimes with side-views of the coast, or with charts.

List of abbreviations of collections

ABGP: Archives du Ministre de la Guerre, Chateau de Vincennes, Paris (Archives Min. of War, Chateau de Vincennes, Paris)
ARA: Algemeen Rijks Archief (Central State Archives The Hague)
BLL: Bodelean Library London
BNM: Bibliotheca Nacional, Madrid (National Lib. Madrid)
BNP: Bibliothèque National, Paris (National Lib. Paris)
BVL: Bibliothèque de Ville, Lyon, (Municipal Lib. Lyon)
GBA: Gemeente Bibliotheek Antwerpen (Municipal Lib. Antwerp)
GAG: Gemeente Archief Groningen (Municipal Archives Groningen)
HABW: Herzog August Bibliothek, Wolffenbüttel (Duke Augustus Lib., Wolffenbüttel)
KBB: Koninklijke Bibliotheek Brussel (Royal Lib. Brussels)
KBH: Koninklijke Bibliotheek Den Haag (Royal Lib. The Hague)
NBW: National Bibliothek Wien (National Lib. Vienna)
RAF: Rijks Archief Friesland (State Archives Friesland)
RAG: Rijks Archief Groningen (State Archives Groningen)
RANH: Rijks Archief Noord-Holland (State Archives Noord-Holland)
RWS: Rijkswaterstaat, mainly Dir. Noord, Leeuwarden
SB: Staatsarchiv Marburg (State Archives, Marburg)
SMA: Scheepvaart Museum Amsterdam (Sailing Museum Amsterdam)
UBA: Universiteits Bibliotheek Amsterdam (University Lib. Amsterdam)
UBG: Universiteits Bibliotheek Groningen (University Lib, Groningen)
UBL: Universiteits Bibliotheek Leiden (University Lib. Leyden)
UBL-BN: Universiteits Bibliotheek Leiden, Bodel Nijenhuis Coll. (University Lib. Leyden, collection Bodel Nijenhuis)
ULR: Universitäts Bibliothek Rostock (University Lib. Rostock)
YAL: Yale University Library

[90] Based on the (estimated) date of publication, or if not available, the (estimated) date of production.

Sources used

At the moment there is no systematic and complete overview of the available maps of the eastern Wadden Sea. The choice of maps, charts and sailing-directions used in this study is for the large part based on the overviews given by Isbary (1936), Vredenberg-Alink (1974), Lang (1958, 1968), Donkersloot-de Vrij (1981), Koeman (1985), Koenders (1986), Ligtendag (1990), and Van der Top, (1992). These overviews were supplemented by other studies (among others Bodel Nyenhuis & Eekhoff (1846), Koeman (1967-1971), De la Ronciere & Mollat du Jourdin (1984), and Lang (1986). Furthermore, charts and maps of RWS Friesland were used for the reconstructions after 1800.

The charts, maps, and sailing directions have been listed as follows: Roman Numeral (used in the text). Maker, year of publication or production, place of print, printer: Name or short description of it, place where it is kept, number, approximate scale. Publication in which it was depicted. Short remarks.

Overview of the charts, maps and sailing directions

I. Anonymous, 12th to early 13th century: "Tabula Peutingeriana", NBW. Dutch part depicted in Vredenberg-Alink (1974) & Stuart (1991). Copy of a Late Roman map (4th century) which was based on the world map made during the rule of Emperor Augustus (27 B.C.-14 A.D.; Stuart, 1991). The map gives an overview of the travelling distances of the Roman Empire and the surrounding world. For the Dutch area only some details are given for the area S of the Rhine. Remarkable are the large estuaries drawn at the end of the Rhine and Meuse.

II. Petrus Vesconte, 1321: Portulano atlas, page 8: The coast of France, England and Ireland, BVL, Ms. 175, Depicted in De la Roncière & Mollat du Jourdin (1984). Hanseatic influence and the establishment of a staple in Bruges in 1323 probably strongly restricted the possibility for frequent direct observations of the North Sea coast by mediterranean sailors (Harley & Woodward, 1987). As shown in Table I the descriptions remained relatively static until approximately 1439 (Lang, 1955; Harley & Woodward, 1987; Koeman, 1985).

III. Abraham Cresques, 1375: "Atlas Catalan", Portulano atlas, BNP, Ms Espagnol 30. Depicted in De la Roncière & Mollat du Jourdin (1984).

IV. Mecia de Viladestes, 1413: Balthic to Niger, Portulano, 1 page, BNP, Rés. Ge AA 566. Depicted in De la Roncière & Mollat du Jourdin (1984).

V. Anonymous, c. 1450: "Das Seebuch", manuscript sailing direction, two manuscripts. Reprinted by Koppmann (1876).

VI. Petrus Roselli, 1462: Atlantic, Mediterranean and Black Sea, manuscript Portulano, 1 page, BNP, Res. Ge C.5090. Depicted in Lang (1955, 1958) & Koeman (1985). The chart is one of the first known, giving a fairly accurate depiction of the Dutch coast and the Wadden Sea.

VII. Nicolaus of Cusa, 1491, Eichstädt: "Cusanus map", copper engraving, BNP and BLL. Depicted in Koeman (1985). Map of Middle Europe, original drawing probably made between 1450-1460 (Koeman, 1985).

VIII. Anonymous, 1492(?): so-called "Chart of Christ. Columbus", Portulano. BNP, Cartes et Plans, Rés Ge AA 562. Depicted in De La Roncière & Mollat du Jourdin (1984). Perhaps made by Ch. Columbus (De La Roncière & Mollat du Jourdin, 1984).

IX. Erhard Etzlaub, <1500, Nuremberg: "Dass ist der Romweg von meylen zu meylen mit puncten verzeychnet... ("Romweg map"), woodcut, BLL. Depicted in Lang (1955) & Koeman (1985). The map was probably made to serve as a pilgrimage-map for the Holy Year 1500 (Krüger, 1967).

X. Anonymous, 1524: Fragments showing Friesland, woodcut, UBL. Depicted in Koeman (1985). The map gives the various waterways from the mainland through the Wadden Sea. Their highly contorted position gives the impression that they were extracted from a list. In the Wadden Sea can be read: elue (Elbe), yaed (Jade), wesel (Weser), oesteremse (eastern Ems), westereems (western Ems), borkêmendiep (Borkumerbalg), lauwerse (Lauwers), Scolbalch (Zoutkamperlaag), nyeamerlât (Nieuwe Amelandse Gat), paedse (river? Paesens), borne (Ameland Inlet), bornryf (Bornrif-ebb tidal delta).

XI. Jan van Hoirne, 1526, Antwerp: "Caerte van de Oosterscher zee", 'chart', woodcut, RAG. Depicted in Lang (1968).

XII. Anonymous, 1532, Amsterdam, printed by Jan Zeuerszoon: "De kaert vander Zee", sailing direction, UBA, 1021 D 20 (original in KBB). Reprinted by Knudsen (1914).

XIII. Anonymous, 1541, Amsterdam, printed by Jan Jacobszoon: "Dit is die Caerte van der Zee: om Oost en West te zeylen", sailing direction. Reprinted by Rogge (1885). Mentions the Lauwers Inlet (with its ebb-tidal delta) and the Zoutkamperlaag (with its ebb-tidal delta and 2 outer inlet channels).

XIV. Cornelis Anthonisz, 1543: Amsterdam: "Caerte van oostlant", chart, woodcut, first print. No exemplars known, only copies of M. Tramezzino: "Septentrionalium regionum ... descriptio", 1558 and 1562 (see below). For a large part the chart has been based on the sailing direction of Jan Jacobszoon, 1541; other sources are probably the province-maps of northern Holland (1508), Friesland (1515 and 1523) by Pieter Gerritsz., Pieter Jansz. Tyebaut, Jan de Pape & Cornelisz and perhaps own observations (probably before 1527; Keunig, 1950; Lang, 1986). Furthermore the chart shows close resemblance to the map of Zell (1560).

XV. Cornelis Anthonisz, 1543, Amsterdam: "Caerte van die Oosterse See", sailing direction with side views, first print. No exemplars known, the book must have been of smaller format than the third edition. Own observations and the sailing direction of 1541. Remaining are the side views of the Frisian islands (woodcuts); they were used for Danish and Dutch sailing directions of 1561, 1566 (Burger, 1920; Lang, 1986).

XVI. Anonymous (possibly Jacob van Deventer or a helper, because his method of encircled survey points was applied (Avis, 1934). However, the map does not resemble the map of Friesland by Van Deventer of 1545 (Ligtendag, 1990, 1991), 1545: Northwest Friesland with Griend, Terschelling and Ameland, manuscript, scale circa 1:38,000, ARA, VTH 3044. Depicted in Avis (1934), Koeman, (1985). Made for plans to polder outerdyke parts of De Bildt area.

XVII. Jacob Heeres (Avis, 1934), 1556: Northwest Friesland with Terschelling and Ameland, partly coloured manuscript, scale circa 1:35,000, ARA, VTH 3043.

XVIII. Cornelis Anthonisz, 1558, Amsterdam: "Caerte van die Oosterse See", sailing direction with side-views, third print, YAL (Lib., Henry Taylor). Photo-copy in SMA, A III-2 245 (a-b). Detailed description of the area, printed after his death (1553/57; Keunig, 1950; Koeman, 1985). Sources: New observations over the period 1543 to 1553/57, and the sailing direction of 1541 (Lang, 1986).

XIX. Michael Tramezzino, 1558, Venice: "Septentrionalium regionum svetiae gothiae norvegiae daniae et terrarum adiacentium recens exactaqs descriptio", chart, Vossius Collection, Leyden, copper engraving. Facsimile by Muller & Co, Amsterdam. The chart is a very detailed and precise copy of the original first print of the chart of C. Anthonisz (1543) as is evident from comparison of the unchanged parts of the third print of Anthonisz (1558).

XX. Jacob van Deventer, 1559, Antwerp, by Bernaerd van den Putte: "Frieslandt:, woodcut, 2nd print. Depicted in Vredenberg-Alink (1974). Only known as facsimile of Martinus Nijhoff (1941). Map of Friesland and Groningen based on land survey. The manuscript and the first print were made in 1545 (Vredenberg-Alink, 1974; Koeman, 1985), for that time extraordinary accurate field-mapping (for which land survey was used; Koenders, 1986) was done in the period 1536-1545 (Van der Top, 1992).

XXI. Cornelis Anthonisz, 1560, Antwerp, by A. Nicolai: "Caerte van oostlant", chart, woodcut, third print, HABW. Depicted in Lang (1986). Sources: The Netherlands derived from province-maps by Van Deventer (1536-1546) and new observations over the period 1543 to 1553/57 (Lang, 1986). Strongly changed with regard to the first and second print (1550), printed after his death (1553/57; Keunig, 1950; Koeman, 1985; Lang, 1986). This is the first time that the triangle of tidal flats in the Lauwerszee is depicted. The outflow direction of the Lauwerszee is both along the Lauwers and along the Zoutkamperlaag.

XXII. Heinrich Zell, 1560, Strassbourg: "Ein neuwe und eygentliche Beschreibung des Teutschen Lands...", woodcut. Depicted in Meurer (1984). Of the map of Germany the earlier versions may have existed as early as perhaps 1532-36, probably around 1542, or at least in 1552 (copy by Gastaldi). In all cases data have been collected from about 1532 onwards (Meurer, 1984). There is a close resemblance to the copies of the first seamap of C. Anthonisz (1543) and his chart of 1560. A remarkable coincidence is that Zell also worked on a chart (lost) around 1542 in Danzig.

XXIII. Michael Tramezzino, 1562, Venice: "Septentrionalium regionum ... descriptio", ULR, copper engraving. Depicted in Lang (1986). The chart is a copy of slightly lesser quality than the one of 1558 of the first (lost) print of C. Anthonisz's "Caerte van oostlant", of 1543 (Lang, 1986).

XXIV. Anonymous, copy of Dirck Zael, probably between 1565-1585 (Denucé & Gernez, 1936): manuscript sailing direction, KBB, 21758. Depicted in Lang (1958). The chart referred to is no. 13: Ems Estuary. It was compiled between 1550-1580, since former courses of 1549 were mentioned (cf. Denucé & Gernez, 1936; Koeman, 1985). Based on the 7 buoys in the Ems (in 1564 there were 5 buoys) and the absence of a high tower on Borkum (built in 1576) the chart of the Ems Estuary must have been mapped between 1565 and 1575, after the mapping by Anonymous (1586), and changed with respect to Anonymous (1572-1580), indicating that the island Bosch had become a strand plain, which happened probably largely in 1570 (Lang, 1958).

XXV. Anonymous, 1566, Amsterdam, by Jan Roelants: "Dit is die Caerte vander See om oost ende west te seylen...", sailing direction with side views. Reprinted by Knudsen (1920). The sailing direction contains the original woodcuts of the first or second taken from Cornelis Anthonisz, 1544: "Caerte van die Oosterse See" (Lang, 1986).

XXVI. Anonymous, perhaps Dirck Zael?, probably between 1572-1580 (Denucé & Gernez, 1936): "Zeeboek", manuscript sailing direction with charts, GBA, B.29166. Depicted in Denucé & Gernez (1936). The chart referred to is p. 4 and 5: Ems-estuary. Like the other copy in Brussels, the work was compiled between 1550-1580 (Koeman, 1985). Based on the 7 buoys in the Ems (in 1564 there were 5 buoys and before that 4 to 5) and the absence of a high tower on Borkum (built in 1576) the chart of the Ems-estuary must have been mapped between 1565 and 1575, after the mapping by Anonymous (1586), before the changes of Anonymous (1565-1585) and before the island Bosch with its high dunes changed into a strandplain, which happened probably largely in 1570 (Lang, 1958).

XXVII. Christiaan sGrooten, c. 1573: "Brussels Atlas", manuscript, scale circa 1:230,000, KBB, Nr. M 21596. The map referred to is the map of Groningen and Oost Friesland. Depicted in Vredenberg-Alink (1974). Compiled from older maps of own making and other authors, such as Van Deventer (Lang, 1958; Koeman, 1985; Van der Top, 1992).

XXVIII. Christiaan sGrooten, 1579-92: "Madrilenean Atlas", manuscript, BNM. The maps referred to are: I, map 11 (drawn >1582; (Vredenberg-Alink, 1974), land partially mapped 1572-1592 (Reitsma, 1974) and II, map 29, dated 1579 (Vredenberg-Alink, 1974), land mapped 1567-1570 (Reitsma, 1974)). Depicted in Vredenberg-Alink (1974). The terrestrial geography was partly based on maps by Van Deventer, Sibrandus Leo and Braun & Hogenberg and supplemented with own data. The Wadden Sea on these maps has been compiled from several (lost) charts, which in some cases must have been quite old (Schoorl, 1973; Vredenberg-Alink, 1974; Koeman, 1985). For the area W of the watershed of Ameland the map of Waghenaer (1584/85) has been used. For the area E of Ameland up to the E-Ems Inlet a map has been used, showing the western Ems in the period before 1563 (showing a main route out through the eastern outer channel), perhaps just after 1541 (cf. Lang, 1958). From the presence of the island Heffesant, and dunes on Bosch, which both disappeared around 1570 and a Lauwers/Zoutkamperlaag configuration, which probably still existed around 1540, but was no longer, or only as a shallow subtidal channel, existent in 1550 (Formsma, 1954; Van Buijtenen, 1954; Huussen, 1979), it can be concluded that this part has been copied from charts of the first half of the 16th century or earlier (The red dots on the map in the area E of Ameland up to the E-Ems Inlet do not indicate buoys, but sounding points!). The cloister-owned island Rottumeroog may have been derived from a detail map of the early 16th century or earlier.

XXIX. Lucas Jansz Waghenaer, 1584/85, Leyden: "Spieghel der Zeevaerdt", sailing direction with charts and side views. The chart referred to is no. 46 from part II (1585): 'Beschrijvinghe van de zee Custen van Oost Vriesslandt...', SMA, Cat. AIII20, scale circa 1:380,000. Facsimile by: Theatrum Orbis Terrerum Ltd, 1965. Field mapping must have been done before 1579, when Waghenaer got his first octroi, probably between 1570-1579 (Lang, 1957, 1958, (Skelton, 1964). Arguments are the reworking of the original copper-engraving to change the islands Bosch (the dunes were removed!) and Bant, the needle form of the tower on Rottum (replaced in 1576 by a large blunt tower) and the number of 9 buoys, which had become already 10 somewhere between 1580-1584 (cf. Haeyen, 1585; cf. Lang, 1957, 1958).

XXX. Albert Haeyen, 1585, Leyden: "Amstelredamsche Zeecaerten, niet sonder excessive costen der selver stede, met grooter nersticheit ende moiten, den zeevarenden ten besten nieuwelijck bij een vergadert", sailing direction with charts, SMA, A III 2-3, scale 1:500,000. The chart referred to is: 'Beschrijvinghe van de Ooster ende Wester Eemsen'. Depicted in Lang (1968). Field mapping for these fairly accurate charts, for the city of Amsterdam, has been done between 1580-1584 (Koeman, 1985).

XXXI. Anonymous (attributed to Haeyen (Schilder, 1991); however, the style, names and presumable date of field-mapping make this unlikely), approximately 1586: "Recueil et pourtraict daulcunes villes maritimes et plus memorables portz et leurs advenues et marcqves servantes a la navigation en la mer oceane", a Dutch manuscript sailing direction with charts, BNM, Ms. Res 237, scales circa 1:185,000-640,000. This sailing direction gives a wealth of details, especially on the Wadden Sea configuration. The charts referred to are: chart l (coast between the mouth of the Elbe and the Groningerdiep) and chart m (West Frisian islands with part of the Zuiderzee; Schilder, 1991, pers. comm.). They were (at least partly) mapped earlier: chart l probably between 1565 and 1575, chart m probably between 1568-1579. This can be concluded from: 1) the absence of a high tower on Rottum, built in 1576 (Lang, 1968) the presence of dunes on Bosch (absent in 1580/84), Cornasant, with a beacon! and Heffesant (likely destroyed during the storm surge of 1570) and the six buoys in the Wester Ems (>1564-<1579; cf. Lang, 1958), 2) the situation at Nieuwe Zijl differs from the chart of Waghenaer (1584/85), mapped between 1570-1579, 3) the absence of a NE-SW oriented inlet through Bornrif (cf. Haeyen (1585), mapped between 1580-1584), 4) the tidal flats in front of the Bildt-polders which represent give a situation which probably existed in the decennia around 1556 (cf. Van Deventer(?) (1545), ARA VTH 3044 and Heeres (1556), ARA VTH 3043), and 5) the closure at Kommerzijl (1568) which is already shown. The 6 buoys indicate that chart l was mapped before Anonymous (1565-1585) and Anonymous (1572-1580) were mapped, which count 7 buoys.

XXXII. Lucas Jansz Waghenaer, around 1588: manuscript, drafts of "Thresoor der Zeevaert", BNP, NAL 2313. The chart referred to is chart 1 of the "Thresoor".

XXXIII. Lucas Jansz Waghenaer, 1592, Leyden: "Thresoor der Zeevaert", sailing direction with charts. Facsimile by: Theatrum Orbis Terrerum Ltd, 1965. The chart referred to is chart 1, scale circa 1:1,000,000. For the Wadden Sea the sailing direction is partly a copy of Haeyens' work of 1585. The drafts were done around 1588. Skelton (1964) points out that the preparatory compilation started in 1586 or earlier.

XXXIV. Sybrandt Hansz. Cardinael, 1602 or slightly later (pers. comm. Donkersloot-de Vrij): "Beschryvinghe van Schellingherlant ende van de nieuwe bedyckte landen", manuscript map, ARA, VTH 2730, scale circa 1:20,000. Extremely detailed large scale manuscript map of the island Terschelling (Koenders, 1986; Ligtendag, 1990).

XXXV. Willem Jansz. (Blaeu), 1608: "Licht der Zeevaert", UBA, standaardnr., OL-80, scale circa 1:500,000. The chart referred to is no. 20. Made with help of Waghenaer, whose influence is marked.

XXXVI. Albert Haeyen, 1613, Amsterdam: "Amstelredamsche Zeecaerten, niet sonder excessive costen der selver stede, met grooter neerstigheyt ende moeyten, den Zeevarenden ten besten nieuwelijck by een vergadert", sailing direction with charts, KBH, page 26-29. Text is strongly updated with respect to the sailing direction of 1585.

XXXVII. Barthold Wicheringe, 1616, Amsterdam: "En tibi lector regio Frisiae, quae inter Lavicam et Amasum flumina Dullartumque sinum porrigitur, ... Bartoldi Wicheringij Groningani iam recens accuratissime descripta...", scale circa 1:170,000, UBG, BB VIII 13; D.N. Depicted in Vredenberg-Alink (1974). Although the map is the most accurate province map for the Groningen area before 1800 and probably partly based on Van Deventer (1559) (Van der Top, 1992), it is rather poor in the depiction of the Wadden islands, when compared to, for instance, the chart of Faber (1642); a conclusion also drawn by Koenders (1986).

XXXVIII. W.J. Blaeu, 1623: "De Zeespiegel", sailing direction with charts, UBA, 1802 D5. The charts referred to are: nr. 5: 'Het Vlie en Amelander gat', scale circa 1:125,000 and nr. 6: 'De Wester ende Ooster Eemsen met de andere gaten der Zee tusschen Amelandt en Langeroogh', scale circa 1:350,000.

XXXIX. Anonymous, 1625-1650 (Donkersloot-de Vrij (1981) mentions 1600 as year, but this is debatable. Although found in a trunk with papers of ≤1600 (Van Keulen, pers. comm.) the handwriting dates from 1625-1650 (Bruggeman, Roos & Koen, pers. comm.)): Schiermonnikoog and a part of the Lauwerszee, manuscript, coloured, scale circa 1:50,000, strongly contorted (Koenders, 1986). RAG, no. 1071., cat Noordhoff, 24. Depicted in Abrahamse & Koning (1969, p. 27). Perhaps used for the sale of Schiermonnikoog in 1638 (to G. Botma & H. Van Marssum), in 1639 (to P. Bauckes Hauckema) or in 1640? (to Stachouwer).

XL. Johannes Sems, 1632: "Generale caerte van het buitenlandt", manuscript, UBG. Depicted in Lang (1958). Supratidal marshes along the coast of Groningen.

XLI. Martin Faber, 1642, Amsterdam(?): "Pascaerte en Beschrijvinghe van de seer vermaerde Riviere Ooster ende Wester Eems. aenwijsende alle Sanden, Bancken, Droochten en Ondiepten daer in ghelegen, hoemen die Schouwen en Voorsichtelyck myden sal", print, SB. Depicted in Lang (1958, 1968). The chart, made by the city-architect of Emden, Faber, a pupil of map maker and engineer Sems, is one of the most accurate of the 17th century in the depiction of a large part of the Wadden Sea (Lang, 1968).

XLII. Willem Jansz. Blaeu, 1643: "The Sea-Beacon", UBA, standaardnr., OL-80, scale circa 1:500,000. The chart referred to is no. 20. Made with help of Waghenaer, whose influence is marked. Facsimile by Koeman (1973).

XLIII. Frederik de Wit, Amsterdam, 1657(?): "Tabulae Dominii Groeningae Quae Et Complectitur Maximam Partem Drentiae Emendata", GAG, 1289. Depicted in Vredenberg-Alink (1974).

XLIV. Sjoerd Ates Haacma, Systse Gravius & Bernardus Schotanus a Sterringa, before 1664: "De Grietenie van Dongerdeel, Oostzijde der Paessens, ...". Depicted in Christianus Schotanus, Franeker, 1664: 'Beschryvinge van de Heerlyckheydt van Frieslandt'. Facsimile by: Theatrum Orbis Terrarum, Amsterdam and De Tille bv., Leeuwarden, 1978. 280 pp, with an appendix by J.J. Kalma, 11 pp.

XLV. Anonymous, circa 1665: Ameland, manuscript, scale circa 1:26,000, including 7 drafts. ARA, VTH 3052. Highly accurate depiction. The mapping is characteristic of circa 1640-1660 (Koen, pers. comm.) in a note on the map the date of 1665 is given. At the North Sea coast the low water line is depicted and at the Wadden Sea side the HW line and part of the low water line.

XLVI. Sanson, H. Jaillot 1673 (1681): "La Seigneurie De Groningue subdivisée en toutes ses Iuridictions. Dressée sur les Memoires les plus nouveaux", copper-engraving, GAG 1326. Depicted in Vredenberg-Alink (1974). Partly after Wicheringe (1616), first version printed in 1673 (Vredenberg-Alink, 1974). Depiction of the channel Reitdiep.

XLVII. Ludolf Tjarda van Starckenborch & Nicolaas (I) Visscher, second half 17th century, before 1684, most likely before 1679: "Groningae Et Omlandiae" Dominium vulgo De Provincie van Stadt En Lande...', copper-engraving, scale circa 1:128,000, UBL-BN. Depicted in Vredenberg-Alink (1974). This rather unprecise map is likely partly based on older material (Van der Top, 1992). It gives new details of the mainland in the second half of the 17th century.

XLVIII. Jan van Petersom Hansz. & Adriaan Hermansz. Wentel, 1695: Kaart van het oosteinde van Terschelling (RANH. A. 492.718) & report RANH Aanw. 301-22. Copy in reconstruction by H. Schoorl (in prep.).

XLIX. Mathurin Guitet, 1708/10, Amsterdam: "Wad en Buytenkaart van het Vlie tot Hamburg...", scale circa 1:300,000. UBL, IV-15-4. Depicted in Lang (1968). The chart was made for the Frisian Admiralty to help to sail through the Wadden Sea (in order to escape pirates; Hooykaas, 1968) and to sail over the North Sea along the barrier islands (Lang, 1968). Although largely new (partly going back on charts of Witsen, 1708) and detailed work (Lang, 1968; Koenders, 1986), the chart is not precise in the depiction of the backbarrier areas. Strong distortions occur due to the enlarging of the inlets and shortening of the islands (which are also too broad; Hooykaas, 1968; Koenders, 1986).

L. Bernardus Schotanus a Sterringa, 1718: Map of Oost Dongeradeel. Depicted in Halma, F., 1718: Uitbeelding der Heerlijkheit Friesland; zoo in 't algemeen als in haare XXX bijzondere Grietenijen door Bern. Schotanus a Sterringa, nu nieweliijks met bijgevoegde aangrenzingen, en veel vermeerderingen, nevens d'afteekening van Oud Friesland in VII verscheidene landkaarten; door den heere Menso Alting. Facsimile by: Theatrum Orbis Terrarum, Amsterdam, 1970, with an introduction by Koeman, C. & Kalma, J.J., 16 pp.

LI. v(an) d(e) P. O. Gravius, 1724: Anjumer and Liousemer Polder, manuscript, RAF, 10634. Map shows that the Bant is still outerdyke supratidal land in 1724.

LII. P.W. Donama, 1732: "Map of Schiermonnikoog with the decayed cape and the new one, which should be made", manuscript, N. to the bottom, RAF, no. 13.212. P.W. Donama was the local guardian of Schiermonnikoog on behalf of the owners (Winkler Prins, 1867). Depicted in Mellema (1981, p. 53) and Abrahamse & Koning (1969, p. 33).

LIII. Anonymous (probably P.W. Donama), 1730-1735(?): "The pirate map", manuscript, coloured, scale circa 1:50,000, N. to the bottom, RAF, Stamno. 758. Depicted in Mellema (1981, p. 73). Low water situation (Koenders, 1986) after a storm (Donkersloot-de Vrij (1981), showing Schiermonnikoog with a large swash bar attached onto the beach. The map may originally have been annexed with the message of Donama to the Staten Generaal of Friesland (mentioned in a resolution concerning beach- and sea-law of 1747).

LIV. Bernardus Schotanus a Sterringa, 1739: "Nieuwe Caert van Frieslant". Facsimile by: Canaletto, Alpen aan den Rijn, 1983, with introduction by De Vries, D., 24 pp.

LV. Theodore van der Haven, circa 1740: Lauwerszee, manuscript, N. to the top. RAG 1064. Depicted in Abrahamse (1967).

LVI. Anonymous, 1745: De zeestromen tussen Ameland en Terschelling, manuscript, N. to the bottom, RAF, no. 13.065. Annexed to a request of the poor-guardians of Harlingen to the Staten of Friesland, concerning the shift of the Amelander Inlet.

LVII. A.T.B. (probably Ass. Theodorus Beckeringh, possibly A.T. Bontekoe (Vredenberg-Alink, 1974)), 1745: (Map of the Northcoast of the province Groningen from Zoutkamp to Enkhuizen and the adjacent Wadden Sea), manuscript, scale circa 1:76,000, N to the top, RAG no. 1071. Later copies depicted in Vredenberg-Alink (1974). N is to the top left. See also Beckeringh (1781).

LVIII. D.W.C. Hattinga in 1749: "Kaarte van het westelyk gedeelte van Amelandt" and "Kaarte van het oostelyk gedeelte van Amelandt", print, scale circa 1:20,000, N to the top, two pages, RAG no. 1048/42, 43. Copy of the originals of Pieter de la Rive, 1731 (lost), for the Atlas Hattinga (1754). Depicted in Heslinga et al. (1985, p. 130) and Overdiep (1964; facsimile).

LIX. Anonymous (P. Portier), <(?)1754: Ships on the Lauwerszee, views. manuscript, N to the top. RAF, no. 13,326. Depicted in Lang, 1976.

LX. Anonymous, 1757: Chart of the Lauwerszee and the adjacent Wadden Sea and the waterway along Schiermonnikoog to the North Sea, manuscript coloured, RAG, 1072 (Noordhoff, nr. 25). Close copy of the original in the GAG, no. 525g.

LXI. Anonymous, 1762: "Kaarte van de Zeegaten en het leggen der tonnen in deselve tussen Schiermonkoog en Amelandt en Amelandt en der Schellingh als mede naar de Wadden", manuscript, N to the top, RAF, no. 13,214. Depicted in De Haan et al., 1983.

LXII. Pieter Idsertdts Portier, 1763: "Caerte van de banken en plaatsen waar de tonnen leggen tusschen de Eijlanden Ameland en Schiermonnikoog", manuscript, N to the top, RAF, no. 13.018.

LXIII. Sluyterman, 1768: seadike of Oostdongeradeel, manuscript, N to the top, RAF, no. 10.632.

LXIV. Anonymous (CLS, probably Coppen Lambert Stachouwer), 1769: Die Heerlijkheid Schiermonnik-Oog. RAF, no. 12,900. In contrast to the conclusion of Ligtendag (1990) the map turns out to be accurate when compared to the original dykes and polders (cf. Milikowski, 1983).

LXV. Theodorus Beckeringh, 1781: "Kaart of Land tafereel der Provincie van Groningen en ommelanden verdeelt in Deszelfs byzondere Quartieren, Districten en voornaamste Iurisdictien. Beneffens de Heerlyckheid Westerwolde, van Nieuws Opgenomen, Verbetert en Vermeerdert...", copper-engraving, GAG, 863. Depicted in Vredenberg-Alink (1974). Probably the only original map of the Groningen area of the 18th century, although parts may have been copied from military surveys (Koeman, 1985; Van der Top, 1992). In the inset map with the Wadden Sea the N is to the upper left, instead of up as in the main map.

LXVI. A.D.G. de Gross, Amsterdam, 1792: "Kaart Der Provintie Van Groningen En Ommelanden Mitsgaders de Heerlijkheid Westerwolde, met een gedeelte van Vriesland, Oostvriesland, Munsterland en Drenthe, de nieuw ingedijkte Polders en Limietsscheidingen and....", copper-engraving, scale circa 1:200,000, copy RWS of original RAG, 26. Depicted in Vredenberg-Alink (1974). The mapping-work of the Wadden Sea is new (source unknown) and shows an intermediate stage between the downdrift orientation of the Lauwers Inlet and the a Kapersplaat W of it (Beckeringh, 1745) and the updrift shift of the Lauwers Inlet with the Kapersplaat amalgamated with Koeplaat and Bosch (Athallin, 1811).

LXVII. J. Peereboom, 1794 (drawn), 1802 (printed): "Terschelling", print, scale circa 1:72,000. UBL-BN, port. 32, 89^{X} (drawn) and ARA, VTH 2737 (print).

LXVIII. A.A. Buyskes, 1798: "Schets van het Amelander Gat, opgenomen op ordre van den agent der marine, door Capt. Lieut. ter Zee", hydrographical chart, copy by Beckering Vinckers (1943), scale 1:100,000. Depicted in Beckering Vinckers (1943).

LXIX. Rabe van Weezel, 1806: "Kaart van het Friesche Gat, benevens de vaarwaters na het Groninger wad, Zoutkamp, Dokkumr en Ezemer-zijl, en het Friesche wad. Nautical chart of the Frisian Inlet. One of the first detailed nautical charts of the Dutch Navy. Archives of RWS, Dir. Noord. Scale: 1:57,000.

LXX. P. van Diggelen, 1806: "Kaarte van het oostelyk gedeelte van Amelandt", manuscript, scale circa 1:14,000, N to the top. RAF, 13254. Copy of the manuscript map (lost) of Hattinga, 1749, which was a copy of the map of P. de la Rive (1731) (Overdiep, 1964).

LXXI. L. den Berger, 1809: "Kaart van het Eiland Schiermonnikoog, meetkundig opgenomen en gekaarteerd op order van de Minister van Binnenlandsche Zaaken", print, scale 1:30,000. UBL-BN, port 43, no. 223/224.

LXXII. L. den Berger, 1809: "Kaart van het Eiland Rottum, meetkundig opgenomen en gekaarteerd op order van de Minister van Binnenlandsche Zaaken", printed, scale 1:30,000. ARA, VTH 3098. Depicted in Lang (1968).

LXXIII. P.A. Overduyn, 1809: "Kaart van het Eiland Ameland meetkundig opgenomen en gekaarteerd op order van de Minister van Binnenlandsche Zaaken", print, scale 1:30,000. UBL-BN, port 43, no. 223/224. Depicted in Bakker (1970). Detailed depiction of the island in the year 1809 (Koenders, 1986).

LXXIV. Atthalin, 1811: "Reconnoissance de l'ile d'Ameland", ABGP, no. 1471 and 1472. Parts of the chart have been derived from the map of P.A. Overduyn, 1809. Furthermore, Atthalin made many own observations. Depicted in Overdiep (1968).

LXXV. Atthalin, 1811: "Reconnoissance des Iles Schiermonnikoog et Rottum", ABGP, no. 1471 and 1472. Parts of the chart have been derived from the maps of L. den Berger, 1809. Furthermore, Atthalin made many own observations. Depicted in: Overdiep (1968), Abrahamse & Koning (1969, p. 35), Lang (1976).

LXXVI. C.R.T. Krayenhoff, 1809-1823: "Choro-Topografische Kaart der noordelijke provinciën van het Koninkrijk der Nederlanden". Copy, original in: RANH Atl. (492), no. 8 (I-IX). The most accurate map until then (Van der Top, 1992). Facsimile Topografische Dienst 1981. Although accurately positioned, the barrier islands and especially the Wadden Sea, were not depicted very well (cf. Koenders, 1982).

LXXVII. S.J. Keuchenius, 1831: "Hydrografische kaart der Zeegaten van Vlieland, Terschelling en Ameland, de vaarwaters naar Harlingen en verder de Zuiderzee opgaande tot aan de middelgronden", trigronometrisch opgenomen en in plan gebracht, hydrographical chart, scale 1:50,000. Copy by Beckering Vinckers (1943), scale 1:100,000.

LXXVIII. S.J. Keuchenius, 1832: "Hydrografische kaart van het Vriesche Zeegat en Groninger Wadden", trigronometrisch opgenomen en in plan gebracht, hydrographical chart, scale 1:50,000.

LXXIX. Sj. Keuchenius, 1833: "Hydrografische kaart der monden van de Eems", trigronometrisch opgenomen en in plan gebracht, hydrographical chart, scale 1:50,000. Copy by Jansen (1941), scale 1:100,000.

LXXX. Anonymous, 1833: Map of Rottumeroog, scale 1:25,000. In: Jansen (1941). Detailed map with the position of the house of the island-keeper from 1743-1799.

LXXXI. Anonymous, 1843: Kadastrale Gemeente Schiermonnikoog. Archives RWS, Dir. Noord, copy, scale 1:20,000.

LXXXII. Boltz, 1843: Situatie van het Eiland Schiermonnik-oog in November 1843 opgenomen bij gewoon laagwater, met aanwijzing der afneming aan de Westzijde sedert November 1842. Ingezonden bij missive van den 31 December 1843 n 233/1 door den ondergeteekenden Ingenieur. Manuscript, Archives RWS, Dir. Noord, Regno. 26001, scale 1:10,000.

LXXXIII. Van Rhijn, 1850: "Hydrografische kaart der monden van de Eems", trigronometrisch opgenomen en in plan gebracht, hydrographical chart, scale 1:50,000. Copy by Jansen (1941), scale 1:100,000.

LXXXIV. Van Rhijn, 1850: "Hydrografische kaart van het Vriesche Zeegat met een gedeelte der Vriesche en Groninger Wadden", trigronometrisch opgenomen en in plan gebracht, hydrographical chart, scale 1:50,000.

LXXXV. Van Rhijn, 1853: Hydrografische kaart der Zeegaten van Vlieland, Terschelling en Ameland, de vaarwaters naar Harlingen en verder de Zuiderzee opgaande tot aan de Middelgronden", trigronometrisch opgenomen en in plan gebracht, hydrographical chart, scale 1:50,000. Partially after map of 1831. Copy by Beckering Vinckers (1943), scale 1:100,000.

LXXXVI. Blommendal, 1859 and 1866: "Hydrografische kaart der monden van de Eems", trigronometrisch opgenomen en in plan gebracht, hydrographical chart, scale 1:50,000. Copy by Jansen (1941), scale 1:100,000.

LXXXVII. Eekhoff, 1854: Map of Ameland. In: Jappé, H., Van Leeuwen, J. & Eekhoff, W., 1859: "Nieuwe Atlas van de Provincie Friesland", part 8. Scale 1:25,000. Partly based on land registration measurements between 1811 and 1843.

LXXXVIII. Ministry of War, 1854: Topographische en Militaire Kaart van het Koningrijk der Nederlanden, Scale 1:50,000. Facsimile by: Geudeke & Zandvliet (1990).

LXXXIX. Blommendal, 1859: "Hydrografische kaart van het Vriesche Zeegat met een gedeelte der Vriesche en Groninger Wadden", trigronometrisch opgenomen en in plan gebracht, hydrographical chart, scale 1:50,000.

XC. Blommendal, 1859: "Hydrografische kaart der Zeegaten van Vlieland, Terschelling en Ameland, de vaarwaters naar Harlingen en verder de Zuiderzee opgaande tot aan de Middelgronden", trigronometrisch opgenomen en in plan gebracht, hydrographical chart, scale 1:50,000. Copy by Beckering Vinckers (1943), scale 1:100,000. Partially after map of 1831.

XCI. Anonymous, 1861: Map of Rottumeroog, scale 1:25,000. Depicted in Jansen (1941).

XCII. Anonymous, 1866: map of Ameland Inlet. RWS, hydrographical chart, with reference to DOL, scale 1:100,000. Copy by Beckering Vinckers (1943), scale 1:100,000.

XCIII. P.J. Buijskes & T.E. de Brauw, 1873/74: "Friesche Zeegat en gedeelte der Wadden". Uitgegeven door het Ministerie van Marine, Afdeeling Hydrographie. Print, scale 1:50,000, Archives RWS, Dir. Noord. Hydrographical chart.

XCIV. P.J. Buijskes & T.E. de Brauw (?), 1873/74: "Hydrografische kaart der monden van de Eems", trigronometrisch opgenomen en in plan gebracht, hydrographical chart, scale 1:50,000. Copy by Jansen (1941), scale 1:100,000.

XCV. C.J. de Jong & R.O.J. Verschoor, 1891: "Friesche Zeegat en omliggende Wadden". Uitgegeven door het Ministerie van Marine, Afdeeling Hydrographie. Print, scale 1:50,000, Archives of RWS, Dir. Noord. Hydrographical chart.

XCVI. C.J. de Jong & R.O.J. Verschoor(?), 1891: "Monden van de Eems". Hydrographical chart, scale 1:50,000. Copy by Jansen (1941), scale 1:100,000.

XCVII. C.J. de Jong & R.O.J. Verschoor(?), 1892: "Hydrografische kaart der Zeegaten van Vlieland, Terschelling en Ameland.." Hydrographical chart, scale 1:50,000. Copy by Beckering Vinckers (1943), scale 1:100,000.

XCVIII. Anonymous, 1899: Strandmetingen en oeverloodingen Boschplaat en Rottumeroog. Raaienstelsel. Scale 1:25,000. Archives RWS, Dir. Noord, Regno 629, no 81302.

XCIX. Anonymous, 1899: Map of Rottumeroog, scale 1:25,000. Depicted in Jansen (1941).

C. C.J. de Jong & R.O.J. Verschoor, 1891, revised 1903: Friesche Zeegat en omliggende Wadden. Uitgegeven door het Ministerie van Marine, Afdeeling Hydrographie. Print, scale 1:50,000, Archives of RWS, Dir. Noord. Hydrographical chart.

CI. C.J. de Jong & R.O.J. Verschoor (?), 1892, revised 1903/1904: "Hydrografische kaart der Zeegaten van Vlieland, Terschelling en Ameland.." Hydrographical chart, scale 1:50,000. Copy by Beckering Vinckers (1943), scale 1:100,000.

CII. Anonymous, 1908: Map of Rottumeroog, scale 1:25,000. Depicted in Jansen (1941).

CIII. Anonymous, 1915: Map of Rottumeroog, scale 1:25,000. Depicted in Jansen (1941).

CIV. Anonymous, 1920, surveyed 1891, partly revised 1920: Friesche Zeegat en omliggende Wadden. Ministerie van Marine, print, scale 1:50,000, Archives of RWS, Dir. Noord. Hydrographical chart.

CV. Anonymous, 1926: Map of Ameland Inlet in 1926. RWS, scale 1:100,000, hydrographical chart, with reference to DOL. In: Beckering Vinckers (1943)

CVI. Anonymous, 1927: Friesche Zeegat. Hydrographisch Bureau, print, scale 1:50,000, Archives of RWS, Dir. Noord. Hydrographical chart.

CVII. Anonymous, 1926/27: "Monden van de Eems". Hydrographical chart, scale 1:50,000. Copy by Jansen (1941), scale 1:100,000.

CVIII. Anonymous, 1928: Map of Rottumeroog, scale 1:25,000. In: Jansen (1941).

CIX. Anonymous, 1950, surveyed 1927, largely revised 1949: "Friesche Zeegat". Hydrographisch Bureau, print, scale 1:50,000, hydrographical map, with reference to DOL. Archives of RWS, Dir. Noord.

CX. Anonymous, 1950: Noordblad Waddenzee, 1950. RWS, Dir. N-Holland, Arrondissement Hoorn, scale 1:100,000. Study area surveyed between 1927-1950.

CXI. Anonymous, 1958: Map of Ameland Inlet in 1958, scale 1:100,000, hydrographical map, with reference to DOL. Archives of RWS, Dir. Noord.

CXII. Anonymous, 1979: Oostelijke Waddenzee en Eems-Dollard. RWS, Dir. Groningen, Meet- en Adviesdienst, print, scale 1:100,000. Surveyed between 1971-1976.

APPENDIX 2

Sources used for reconstructions

In all reconstructions use was made of the following sources: Isbary (1936); Kooper (1939); Rienks & Walther (1955); Rijkswaterstaat (1948, 1961) and Stiboka (1973, 1976, 1981).

Reconstruction 1300 (Figure A1)
Roeleveld (1974); Van Staalduinen (1977); Griede (1978); Knol (1991); Bosch & Vos (1991); Sha (1992); Van der Spek (1994).
Maps, charts and sailing directions: II to X and the reconstruction 1500.

Reconstruction 1500 (Figure A2)
Maps, charts and sailing directions: X to XVI, XVIII, XX, XXII, XXVII and XXVIII and the reconstructions 1300 and 1550.

Reconstruction 1550 (Figure A3)
Lang (1963): Morphologischer Zustand der Emsmündung, 1580.
Van Oosten (1986).
Maps, charts and sailing directions: XIV to XXIV, XXIX, XXX, XXXI and the reconstructions 1500 and 1600.

Reconstruction 1600 (Figure A4)
Lang (1963): Morphologischer Zustand der Emsmündung, 1580 and 1650.
Ligtendag (1990): Reconstruction Ameland, Schiermonnikoog & Rottumeroog in 1600.
Maps, charts and sailing directions: XXX to XXXIX and the reconstructions 1550 and 1650.

Reconstruction 1650 (Figure A5)
Lang (1963): Morphologischer Zustand der Emsmündung, 1650.
Maps, charts and sailing directions: XXXIX to XLVI and the reconstructions 1600 and 1700.

Reconstruction 1700 (Figure A6)
Maps, charts and sailing directions: XLVI to LII and the reconstructions 1650 and 1750.

Reconstruction 1750 (Figure A7)
Hayes (1762), Winkler Prins (1867).
Lang (1963): Morphologischer Zustand der Emsmündung, 1720/1790.
Ligtendag (1990): Reconstruction Ameland, Schiermonnikoog and Rottumeroog in 1750.
Maps, charts and sailing directions: LII to LXV and the reconstructions 1700 and 1800.

Reconstruction 1800 (Figure A8)
Lang (1963): Morphologischer Zustand der Emsmündung, 1790. Maps, charts and sailing directions: LXVI to LXXVI and the reconstructions 1750 and 1832/34.

Reconstruction 1831-34 (Figure A9)
Winkler Prins (1867), Van der Molen, (1968).
Charts: LXXVII to LXXXI

Reconstruction 1866-74 (Figure A10)
Charts: XCII-XCV

Reconstruction 1888-1992 (Figure A11)
Charts: XCVI-XCVIII

Reconstruction 1927-1950; Figure A12)
Chart: CX

Reconstruction 1971-1976; Figure A13)
Chart: CXIII

CHAPTER 3

THE CYCLIC DEVELOPMENT OF THE PINKEGAT INLET SYSTEM AND THE ENGELSMANPLAAT/SMERIGGAT, DUTCH WADDEN SEA, IN THE PERIOD 1832-1991

ABSTRACT

The developments of the Pinkegat Inlet system and the shoal complex east of it, Engelsmanplaat, have been studied for the period 1809-1991. A cyclic development of the morphology of the Pinkegat Inlet is observed, from a single inlet phase to a multiple inlet phase and back, with periods of 20 to, at maximum, 54 (but probably 41) years. All inlets migrate downdrift, due to the strong sediment supply from the updrift (west) side, outer bend erosion, and shoal migration. A multiple inlet configuration is always an unstable configuration. The downdrift inlets tend to migrate more slowly than updrift ones and thus inlets merge. Also, other inlets are abandoned. During downdrift migration of the channels, a shoal starts to build up at the downdrift side of the island. These processes result in the establishment of a single inlet configuration. The single inlet continues to migrate downdrift, which results in the further growth of the shoal at the downdrift side of the island Ameland. When the transport path to the drainage basin becomes too long, new inlets are formed more updrift, starting as spill-overs, which, in their turn, probably are generated from flood chutes.

The developments in the drainage basin follow the developments in the ebb-tidal delta. Upon downdrift shift of the inlets, the main backbarrier channels of the Pinkegat are forced to orient ENE-WSW, as they form the connection between the drainage area and the inlet(s). The channels erode the shoal at the downdrift end of Ameland at its southern side, thus enhancing the formation of new inlets at the updrift side. A NNE-SSW-orientation can only develop when the shoal is cut and new inlets develop to the west. When the transport path along the downdrift end of Ameland becomes longer/slower, the watershed shifts to the E, due to the local decrease of tidal forces from that side.

The cyclic development in the Pinkegat is reflected in the sediments. They show many internal erosion features and other indications of local, strong changes in energy conditions. Extensive lateral accretion surfaces are formed during downdrift migration of the inlets. The erosion depth become deeper during the development of the single inlet phase. In the fossil record such accretion surfaces may form important indicators of the residual current in the open sea basin. Due to the net expansion of the eastern part of Ameland during subsequent cycles, a sedimentary sequence is formed. It consists of a deep channel lag formed during a single inlet phase and one or several shallower channel lags formed during the subsequent multiple inlet phase. The outer channels of the ebb-tidal delta show a sedimentary development comparable to that of the inlet, but at the western side of the ebb-tidal delta they may

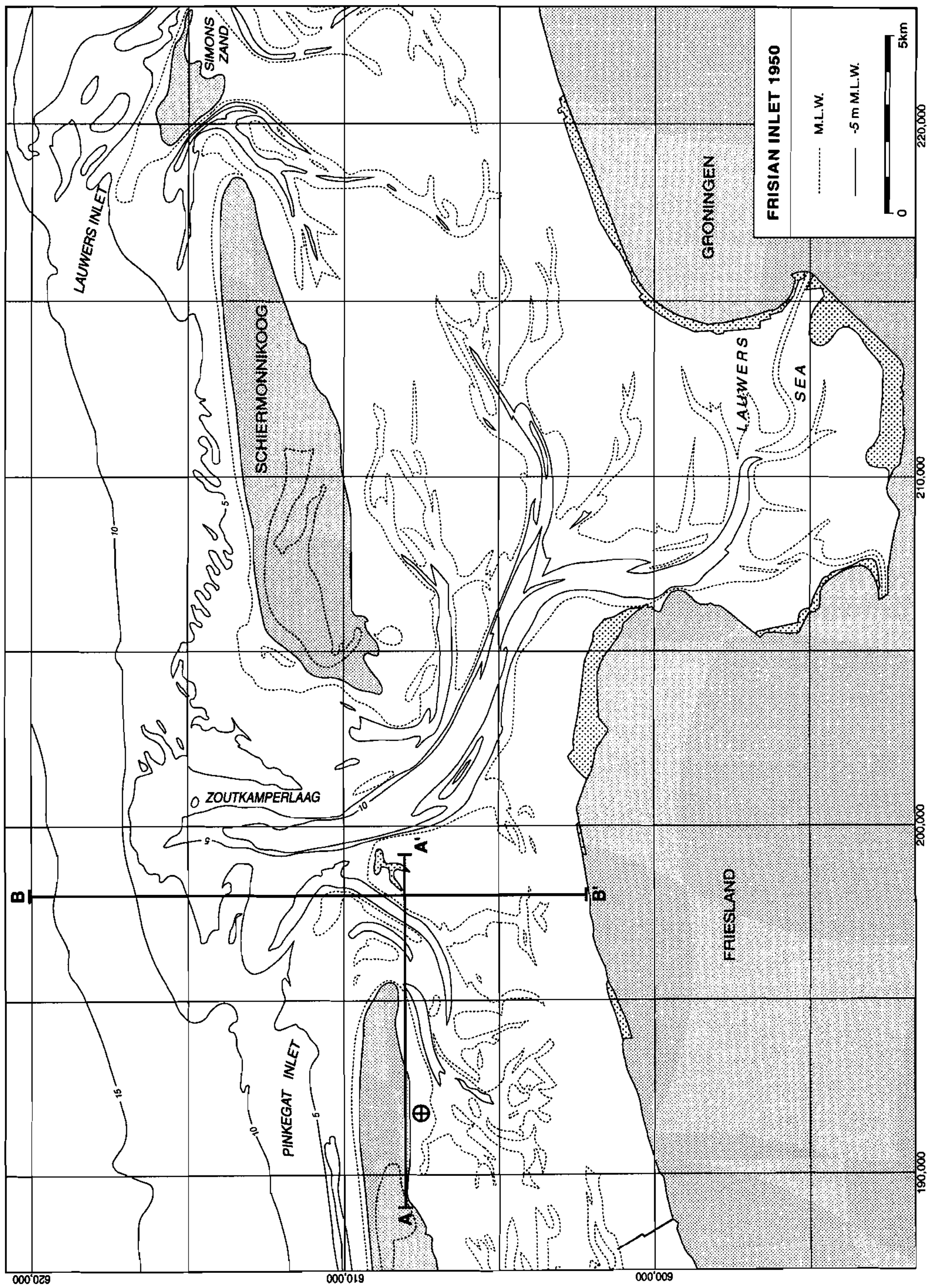

Figure 1: Overview of the area and location of the profiles (Figs 7 and 10); cross marks location of core 91.032 (Fig. 9).

have lateral accretion surfaces dipping to the NNE. When the inlet and main channel orientation changes, larger channels in the drainage basin may become abandoned quite abruptly. This results in the deposition of thick, clay-rich infills in these channels. The net eastward displacement of the watershed in combination with the relatively restricted possibilities to migrate results in a laterally and vertically rather discontinuous, inhomogeneous sedimentary sequence with frequent channel lags and clay(-rich) infills formed near the watershed area.

The northern part of the Engelsmanplaat cyclically (54-86 yrs) grows and wanes due to alternating erosion by channel migration and the formation of shoals by waves and the flood current. The channels are, at least partly, formed as the easternmost channels of the Pinkegat Inlet system, and shift clockwise to form the connection between the Pinkegat and the Zoutkamperlaag Inlet system. Mergers of the shoals N of the channels with the Engelsmanplaat result in the formation of abandoned channel deposits, which are more or less coast-parallel. The western side of Engelsmanplaat has eroded between 1832/34-1991 from 7.1 to 2.2 km. If erosion continues the shoal will disappear in the near future.

INTRODUCTION

Cyclic changes in ebb-tidal deltas are relatively common (FitzGerald, 1988). Data are often insufficient to follow the morphological development and to link the development with the controlling parameters and the sedimentary sequence deposited. It is therefore difficult to understand similar fossil sedimentary deposits in terms of morphodynamic development. The development of the Dutch Wadden Sea area during the last two centuries can be reconstructed in great detail, because many observations were made.

The Wadden Sea area consists of the backbarrier area, and a series of inlets and barrier islands in the southern North Sea. It stretches from the Northern Netherlands to Denmark. The Dutch barrier islands are oriented SW-NE in the west and WSW-ENE in the east. They have a 'downdrift offset', meaning that an island at the downdrift side of an inlet (i.e., in the direction of the residual coastal current, here towards the E) extends further seaward than the island updrift of it (Edelman, 1961; Sha, 1989b). The barrier islands shelter the backbarrier area against waves from the North Sea. The backbarrier area, the Wadden Sea proper, consists of a series of partially intertidal drainage basins. These are more or less separated from each other by tidal watersheds (south of the islands) and are drained through individual inlets. River influence is limited and the backbarrier area can be classified as lagoonal (Reinson, 1992). In front of the inlets (seawards) large ebb-tidal deltas are situated.

Several studies have shown that the island coasts adjacent to the ebb-tidal delta form an interactive part of the ebb delta (cf. Dean, 1988; FitzGerald, 1988; Sha, 1989a,b, 1990; Oost & De Haas, 1992, 1993; Huijs, 1993). For most natural tidal systems in unconsolidated sediments the dimensions of the ebb-tidal delta, of the inlets, and of the backbarrier channels are defined by the tidal prism of the drainage basin[1]. The volume of the tidal prism is determined by the surface area of the drainage basin, the average height of the intertidal area, the tidal amplitude and the celerity of the tidal wave. The surface area of the drainage basin is determined by the distance from the islands to the shore and the distance from the inlets to the watersheds at either side. In the Dutch Wadden Sea, the latter distance depends for a large part on the length of the barrier islands. This is because the watershed is situated, as a rule-of-thumb, at a distance of about 2/3 downdrift of the updrift ends of the islands, due to the phase difference of the flood waves entering through the inlets E and W of the islands and wind effects. The length of the island in its turn depends on the tidal amplitude: the higher the tidal amplitude, the shorter the islands (Hayes, 1975; Wolff, 1986). The average water depth of the Wadden Sea depends for an important part on the erosional force exerted by the waves (Eysink, 1979), the water level[2], local hydrodynamic conditions, and perhaps the size of the basin (Eysink & Biegel, 1992). Under conditions of a constant wave climate and tidal amplitude, all parts of a drainage system (ebb-tidal delta with the adjacent island ends, the inlets, the backbarrier channels, and the drainage basin itself) in the Wadden Sea can be considered to be coupled and to be in dynamical equilibrium[3]; they form a so-called "Sand Sharing System" (cf. O'Brien, 1969; Dean & Walton, 1975; Jarrett, 1976; Walton & Adams, 1976; Van den Berg, 1986; Dean, 1988; Dieckmann et al., 1988; Stive & Eysink, 1989; Gerritsen, 1990; Hume & Herdendorf, 1990; Misdorp et al., 1990; Niemeyer, 1990; Sha, 1990; Biegel, 1991b; Flemming, 1991; Steijn, 1991; Van Kleef, 1991; Eysink & Biegel, 1992; Bilse, 1993; De Vriend & Bakker, 1993; Eysink, 1993). The dynamical equilibrium is an important factor when disturbances occur within the system.

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[1] In estuarine systems also the river discharge is of importance.

[2] According to Eysink and Biegel (1992) this is the high water level; measurements of net sedimentation (De Boer et al., 1991; Oost & De Haas, 1992, 1993) suggest that sedimentation follows MSL rise. In an investigation of 22 United States lagoons Nichols (1989) showed that most of them show a near-balance between relative sea-level rise and accretion. That balancing is a common feature in lagoons is also suggested by the observation that barrier systems can exist for many thousands of years although the filling of a backbarrier area might theoretically take a few centuries.

[3] Actually the system will continuously be lagging slightly behind the real equilibrium conditions, because always some time is needed to adapt to the continuous changes. Since the morphological developments in a system are often quite rapid, and because almost all relations between the parts of the system and the tidal prism are (semi-)empirical, this lag is neglected in this study. At present the Zoutkamperlaag (chapter 5) and probably several of the drainage basins in the western Wadden Sea (see chapter 1) have not yet reached their equilibrium conditions, so that sedimentation rates are high (see chapter 1). Apart from tidal prism also the influence of waves has to be considered; this has not yet been fully quantified. Also, it should be realized that, depending on the local external conditions, drainage basins may drown (in case of insufficient sand supply) or be filled up (FitzGerald, 1988; Steijn, 1991).

When temporary disturbances occur (such as dredging, erosion by storms) other parts of the system will temporarily deliver or store sand to compensate for such disturbances. Over a longer space of time the system will tend to regain its old dynamic equilibrium. Where the disturbances are permanent (for instance a change in wave climate, the closure of a part of the drainage basin; Chapter 5), the whole system shifts towards a new dynamical equilibrium. In both cases, the sediment, needed to regain dynamical equilibrium, has to be exported to, or imported from outside. To emphasize the interaction between the different elements which together form a Sand Sharing System the term "inlet system" is frequently used throughout the text. The term comprises an inlet, its ebb-tidal delta, the adjacent island ends, and its complete backbarrier (drainage) area.

The barrier island Ameland is 25 km long and 3 km wide (Fig. 1). The island to the east, Schiermonnikoog, is 15 km long and 3 km wide. The, about 11 km wide, Frisian Inlet is situated between the islands. It consists of two inlet systems (two sand sharing systems), which are separated by the supratidal shoal Engelsmanplaat. The western inlet is called Pinkegat Inlet and the eastern one is called Zoutkamperlaag Inlet. Their, mainly intertidal, drainage basins are bordered by the already mentioned Engelsmanplaat, the tidal watersheds, and by the mainland coast (Fig. 1). The Lauwerszee embayment also was part of the drainage basin of the Zoutkamperlaag before the closure in 1969 (Chapter 5). Two ebb-tidal deltas are present at the North Sea side of the inlets. The ebb-tidal delta of the Zoutkamperlaag Inlet is the largest and influences the morphology of the coastal zone to at least -12 m DOL (DOL = Dutch Ordnance Level = NAP = ± MSL). The main channels of the ebb-tidal delta of the Zoutkamperlaag Inlet are at present the ebb-defined[4] Plaatgat and the flood-defined Westgat (Fig. 1). Coastward they merge into the Zoutkamperlaag proper, an approximately 1 km wide backbarrier channel. This main channel branches into several lower order channels. The Pinkegat is the smaller westernmost inlet system. It drains the largely updrift drainage basin S of the eastern part of Ameland. Its small ebb-tidal delta is largely surrounded by the ebb-tidal delta of the Zoutkamperlaag Inlet system.

The tidal regime in the area is semi-diurnal and mesotidal, with a 2.3 m tidal range at Schiermonnikoog, (cf. Hayes, 1979; Postma & Dijkema, 1982). In winter waves are highest, with a wave height of up to two metres at 20 metres water depth. The coast is a mixed energy, tide-dominated shoreline, influenced by both tides and waves (cf. Hayes, 1975).

The developments of the Pinkegat Inlet system during the past two centuries show, in great detail, the strong morphodynamics of a inlet system in a mixed energy shoreline setting and the interaction between the various parts of the system. The study provides insights into

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[4] This distinction is mainly based on the form: if the channel becomes shallower towards the drainage basin it is considered flood-defined; ebb-defined is used for the reverse morphology. Model calculations (Steijn et al., 1992) as well as direct observations (RWS, unpublished data) indicate that ebb-defined and flood-defined channels roughly coincide with ebb-dominated, respectively flood-dominated channels. In the flood-defined channels the flood currents tend to be concentrated south of the channel axis.

the formation of barrier and barrier-related deposits, and gives clues where to look in the fossil reach in order to reconstruct barrier-related deposits. The sedimentological/morphodynamical changes in the drainage basin and ebb-tidal delta of the Pinkegat over the period 1832-1991 are described, with special attention for the period 1927-1991. This paper is a further elaboration of two studies for the Project COASTAL GENESIS (Oost & De Haas, 1992, 1993) coordinated by the Directorate General for Public Works and Water Management (further referred to as RWS = Rijkswaterstaat), National Institute for Coastal and Marine Management/RIKZ (further referred to as RIKZ).

METHODS

Several data sets were collected, depth soundings being the most important. For the period 1832-1934 these depth-sounding data have been redrawn from maps of RWS and the Navy (Figs B1 to B10)[5]. For the period 1927-1991 computerized depth maps (Figs C1 to C10) were compiled from available depth soundings and the yearly coastal observations of the RWS (JARKUS). The data from the period before 1982 were hand digitized from the original data maps (minutes, appendix 1, next chapter). Where needed, detailed observations of tidal marshes (Dijkema, pers. comm.) and detailed topography maps were added (less than 1% of the observations). For the years 1927 en 1949 the scale of the minutes was mainly 1:25,000, whereas for the other years it was usually 1:10,000. For the observations from 1970 to 1979 hand-digitized maps of RWS, which were originally stored in a 200*200 m grid were used. This makes these maps somewhat less detailed. Data after 1979 were directly taken from depth soundings. Most maps were compiled from depth soundings in a few subsequent years (appendix 1, next chapter). The year for which the most depth soundings were available is used in the text and marked on the relative maps. For detailed discussions the very (or most important) year of that specific observation is used.

All sounding data were calibrated to Dutch Ordnance Level (DOL) and real bottom depth (Oost & De Haas, 1993). The data were gridded into a 90 * 90 m grid, which is identical for all observation years, in order to allow comparisons. Depth values for empty cells were interpolated using the CONLOD-program, which was developed by the RWS, Friesland (ANW) for the handling of depth-sounding data (Van den Boogert & Noordstra 1988; Van den Boogert, 1991). The interpolation is based on inverse distance interpolation, but also applies several other weighting factors, which take into account the local topography (Van den Boogert, 1991).

From the interpolated grid, the depth-sounding maps were made (Figs C1 to C10). Also sedimentation-erosion maps were made by subtracting the grids of different years. Such

[5] For the figures B1 to B10 and C1 to C10 the reader is referred to appendix A of this thesis.

maps have been made for 1949-1927, 1958-1949, 1967-1958, 1970-1967, 1967-1927, 1970-1967, and 1991-1987 (Fig. D1). On the basis of the grids (synthetic) profiles were computed (see below). They show those parts of the various sedimentary profiles which have not been "eroded" in later profiles. They thus give an indication of what has been preserved of older sedimentary deposits and they enable the prediction of the sedimentary sequence within the subsurface.

For the various calculations the Frisian Inlet was subdivided into seven areas (Fig. 2): the ebb-tidal deltas of the Pinkegat and Zoutkamperlaag Inlet, the drainage basins of the Pinkegat and Zoutkamperlaag, the channel Smeriggat, the shoal Engelsmanplaat, and the Lauwerszee embayment. All sub-areas match together (with exception of the Smeriggat/Engelsmanplaat, which form part of the other areas). A change in the configuration of the areas had to be taken into account for the period 1970-1987, due to the strongly changed configuration of the whole area and the limitations of the available maps.

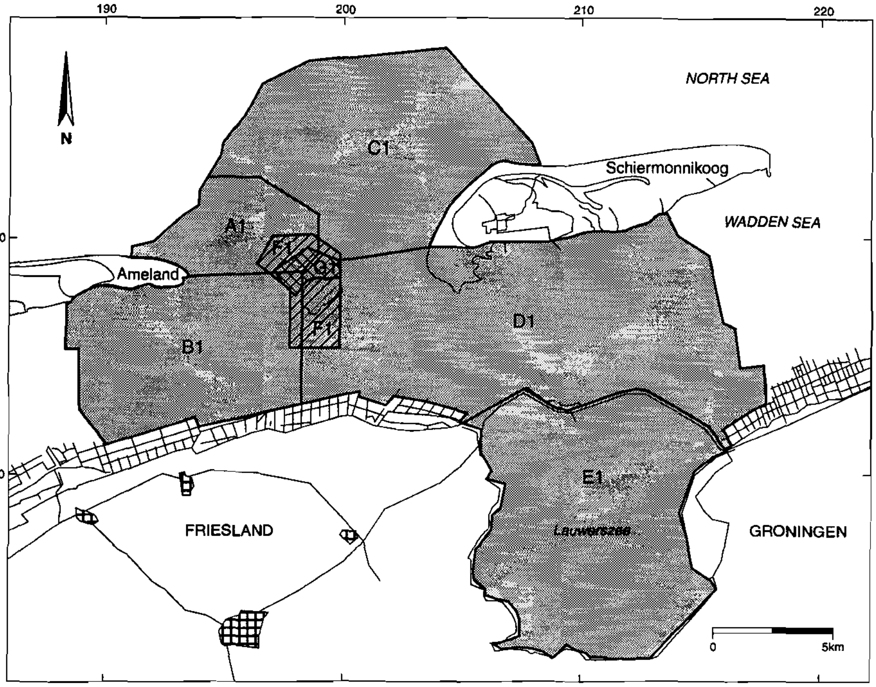
Accuracy

All depth-soundings were made with various ships applying self-registering echo-sounding (except in 1927, when sounding by hand was used). After 1955 the distance between observations was about 5 m, earlier it was up to 100 m. The track-to-track distance was usually 800 m before 1958 and 200 m after 1958. In general the observations have been made parallel to the strongest topographical gradient. All observations were made by experienced crews who applied the state-of-the-art methods of their time. It was always attempted to minimize the measurement error, both during the field observation (cross-checks of measurements, compensation of systematic errors) and during the processing (removal of unrealistic data, cross-checks, compensation of systematic errors). Even for the oldest measurements of 1927 a fair degree of accuracy was reached.

The depth-soundings constitute a sample of the population of the depths of the bottom. This implies that the final values always contain a sampling error. In addition, there are also the errors in the observations themselves. The causes of these errors can be subdivided into 5 main groups:

1) water and bottom conditions,
2) registration equipment,
3) ship movements,
4) tides, and
5) the distance between the observations; the larger the distance the larger the interpolation errors (De Looff, 1975; Van Dam, 1988).

The errors differ with the years, because of changes of instrumentation and methods. For an extensive discussion of errors, which can occur during the soundings the reader is referred to


190
200
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N
NORTH SEA
C1
Schiermonnikoog
WADDEN SEA
A1
F1
G1
Ameland
D1
F1
B1
E1
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Lauwerszee
GRONINGEN
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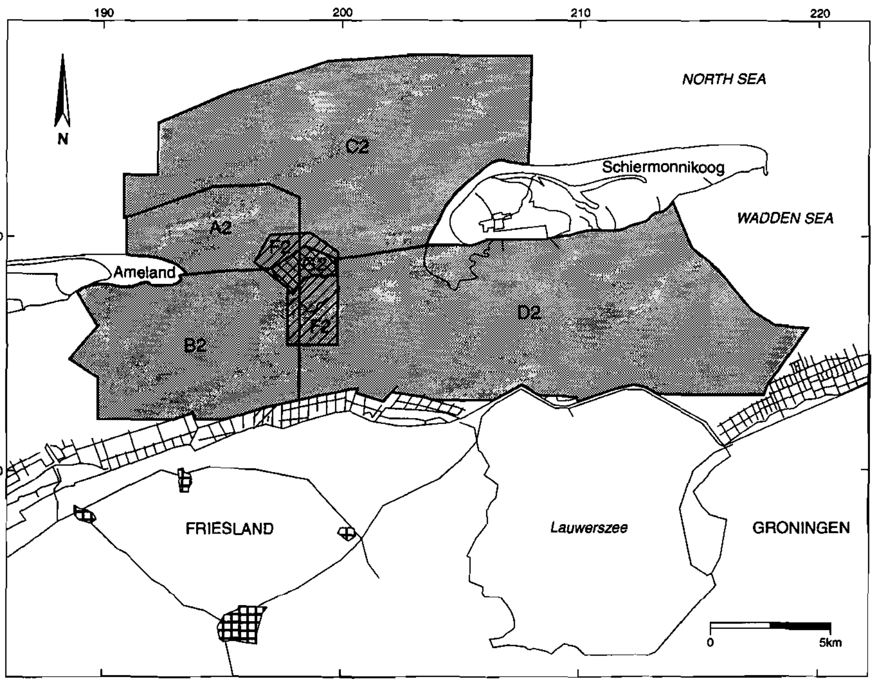
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WADDEN SEA
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G2
Ameland
D2
F2
B2
FRIESLAND
Lauwerszee
GRONINGEN
0
5km

Table I: Estimated depth errors over larger areas for the various gridded data sets.

Year	Error (1σ) over a larger area
1927	± 30 cm
1949	± 30 cm
1958	± 25 cm
1967	± 20 cm
After 1967	± 3 cm

De Boer et al., (1991a) and Oost & De Haas (1992, 1993). In addition to errors during sounding also errors are made during the conversion into a grid structure because of digitizing, the grid size, the bottom structure, and the interpolation method applied. Based on an extensive literature study (Anonymous, 1952, 1974; De Looff, 1975; Schaik, 1979; Hartman & Pastoor, 1985; Nanninga, 1985; Anonymous, 1991; Biegel, 1991a; De Boer et al., 1991a; Oost & De Haas, 1992, 1993; Navy pers. comm.) the systematic error over larger areas (tens of km^2) was estimated for the various years (Table I).

OBSERVATIONS

Throughout the presentation and the discussion the following sub-areas are used (Fig. 2): the ebb-tidal delta of the Pinkegat, the drainage basin of the Pinkegat, Smeriggat, and Engelsmanplaat. Special attention is given to the results of the computer studies. For the morphological changes the reader is referred to Figs B1 to C10 and C1 to C10. In each subchapter the qualitative changes are discussed, followed by a discussion of the quantitative changes (see Tables II and III and Oost & De Haas, 1992, 1993). For the ebb-tidal deltas channels are normally defined as being -5 m DOL and deeper, unless explicitly stated otherwise.

Figure 2 (opposite page): Overview of the various sub-areas for which calculations were made: A = the ebb-tidal delta of the Pinkegat Inlet, B = the drainage basin of the Pinkegat Inlet, C = the ebb-tidal delta of the Zoutkamperlaag Inlet, D = the drainage basin of the Zoutkamperlaag Inlet, E = Lauwerszee embayment, F = 'Engelsmanplaat', G = 'Smeriggat'. Prefixes 1 and 2 indicate the area before, respectively after 1970.

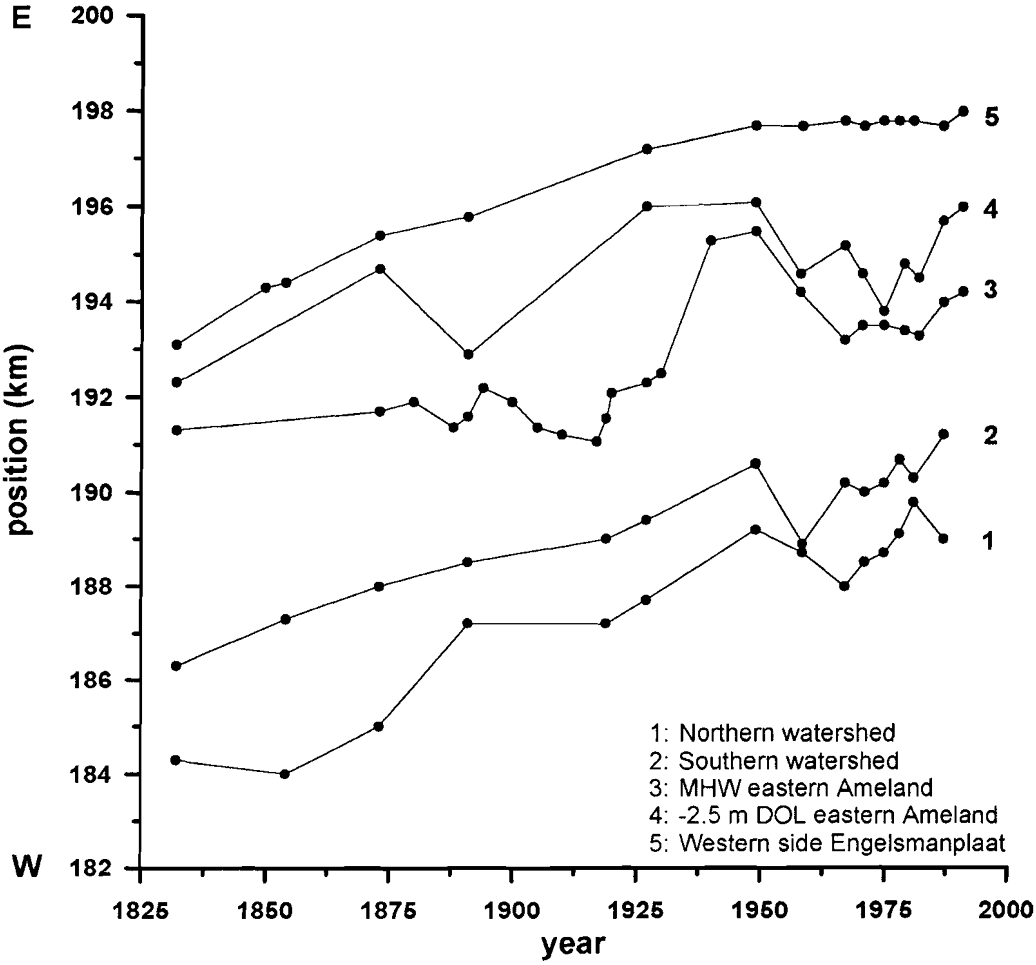

Figure 3: Migration of the watershed, of the downdrift end of Ameland, and of the Engelsmanplaat. Note the close relation between the growth and decrease of the easternmost end of Ameland and the general pattern of shift of the watershed. Before about 1900 the eastward displacement of Ameland Inlet (west of Ameland) also enhanced the downdrift shift of the watershed. The migration of the HW-line of Ameland between 1880-1917 suggests the presence of another single inlet phase around 1894. Also, note the strong erosion of the western side of the Engelsmanplaat (measured along E-W-line 608, see Figs C).

The Pinkegat system drains the area SE of Ameland. The watershed runs at two-third E from the western end of Ameland (Fig. 1), so that the Pinkegat system has only a small drainage area, a small tidal prism, and also a small ebb-tidal delta (cf. Dean & Walton, 1975; Walton & Adams, 1976). The ebb-tidal delta is surrounded almost entirely by the large ebb-tidal delta of the Zoutkamperlaag Inlet system (Fig. 1).

The Pinkegat Inlet system (HW-line) has shifted over a net distance of approximately 3 km to the E during the period 1832-1987 (Fig. 3). The watershed shifted about 4.8 km to the E in the period 1832-1987 (Fig. 3; De Boer, 1991; Oost & De Haas, 1993). However, the tidal prism has not changed much, because of the erosion of the Engelsmanplaat at its western side (Fig. 3).

Pinkegat ebb-tidal delta

1832-1873/74

Periods with one inlet alternate with periods with several inlets in the Pinkegat system. In 1832 (Fig. B1) the Pinkegat had one broad inlet, which was oriented NW-SE. In 1850 (Fig. B2) only the easternmost part of the Pinkegat was sounded; it shows that the Engelsmanplaat (E of the Pinkegat) was eroded over about 1 km in the period 1832-1850 (Fig. 3). The names given by Van Rhijn (1851) indicate that in 1850 the westernmost inlet had been newly formed, whereas the eastern inlet was the original one. In 1854 (Fig. B3) two outer channels had formed seaward of the easternmost inlet.

1873/74-1894

In 1873/74 (Fig. B5) the easternmost inlet of the Pinkegat had become the dominant one (Anonymous, 1877). It had both a flood- and an ebb-defined outer channel, located closely to each other. W of it only a small channel was present. After an eastward expansion from about 1873/74 onwards, the easternmost area above the HW-line of Ameland retreated more than 0.5 km to the W in the period 1881-1888 (data RWS). The beachmark at the eastern end (no. 24) was eroded in 1887 (Isbary, 1936). In 1891 (Fig. B6) the Pinkegat consisted of several inlets (2 or 3 depending on whether the middle channel is considered as an outer channel or as an individual inlet). The easternmost inlet had a large ebb-defined outer channel and a large flood-defined one. In the period 1888-1894 the easternmost HW-line of Ameland shifted eastward over 0.8 km (Fig. 3; Isbary, 1936; data RWS).

1894-1927

From 1894 to 1917 the HW-line of Ameland shifted westward again over about 1.1 km (Fig. 3; Isbary, 1936; Postma, 1982). This was likely related to channel erosion. Coastal observations around 1900 state that at that time a multiple inlet configuration was present (Mars & Cordia, 1901). In 1903 (Fig. B7) only the northeasternmost part of the ebb-tidal delta was sounded. The sounding shows that the shoals in front of the ebb-defined channel had disappeared. In the period from 1917 to 1950, the eastern HW-line was displaced to the E over 4.4 km (Fig. 3; cf. Postma, 1982), and also the inlets must have shifted eastward. No depth-soundings were made in 1921 (Fig. B8), except of the easternmost inlet. The shoal in front of the eastern ebb-defined channel, above -5m DOL, had disappeared, suggesting an eastward shift of the updrift flood-defined channel.

1927-1967

In 1927 (Figs B9 and C1) the Pinkegat consisted of one inlet. This inlet bifurcated in the seaward direction into a deep ebb-defined channel to the east and a somewhat shallower flood-defined channel to the west. A large subtidal sandy shoal was situated west of the inlet. It formed the eastern end of Ameland. In the period 1927-1949 (especially until 1940; Postma, 1982) the eastern end of Ameland became higher, from -2.5/-1 m DOL to 0/+1 m DOL, and it slightly shifted to the E (Figs C1, C2 and 3). By 1949 the inlet of 1927 had been replaced by several, more or less separate inlets, with a smaller maximum inlet depth (the westernmost one being less than -5 m deep). The new westernmost inlet(s) probably formed by deepening of small channels W of the main inlet, which were already present around 1941 (Jansen, 1941). They had already become large inlets in 1944, when they were clearly visible on an aerial photograph (Van Straaten, 1964). Shoals developed between the inlets, in particular W of the easternmost channel (from at least 1944 onwards; Van Straaten, 1964), close to the shoal Engelsmanplaat. The ebb-defined channel (Smeriggat) became connected with the western flood-defined outer channel in the Zoutkamperlaag ebb-tidal delta. The workmaps of RWS show that this happened after 1938, probably around 1941/42 (Reitsma, pers. comm.), but in any case before or in 1945 (notes on workmaps RWS indicate a channel in/or before 1947, in which ammunition was dumped in 1945; Van der Geest, pers. comm.).

In 1949-1958 (Figs C2, C3, and 3) strong erosion occurred at the eastern end of Ameland (the HW-line retreated westward over 1.3 km). A new western outer channel developed into a separate inlet. In 1958 the Pinkegat system consisted of three clearly separate inlets. The shoals in between them had also become larger. The middle inlet bifurcated into two outer channels. In the period 1949-1958 the eastern end of the easternmost inlet (Smeriggat) shifted southward. The channel was bordered by sandy shoals by 1958. In the period 1958-1967 (Figs C3, C4, and 3) the westernmost inlet shifted to the east (over 1 km). It merged with the western outer channel of the middle inlet, which had hardly migrated. The eastern outer channel of the middle inlet separated, and became an inlet for a short time (around 1960), but was soon abandoned. In the period 1958-1967 the eastern end of Ameland decreased in height. A subtidal shoal formed ENE of Ameland in the same period. The ebb-tidal delta as a whole expanded to the WNW.

1967-1991

By 1967 (Fig. C4), the easternmost inlet of 1958 had become a relatively shallow channel, which only had a function as a drainage channel through the Zoutkamperlaag Inlet for a small part of the Pinkegat system. It could no longer be considered a tidal inlet. Like in 1927, the Pinkegat system consisted thus again of a single inlet.

A new cycle from a single inlet to a multiple inlet and back again occurred between 1967 and 1987 (Oost & De Haas, 1992). In the period 1967-1970/71 (probably as early as 1968/69; Fig. C5) two partly separate channels developed by incision of the shoal ENE of Ameland. A rapidly migrating middle channel existed also for a while, but was abandoned in the period 1970/71-1974 (Huijs, 1993). In the period 1970/71-1975 (Fig. C6) the eastern outer channel rotated clockwise, whereas the seaward part of the western channel was formed by incision (around 1972; Huijs, 1993) near the eastern end of Ameland. The western outer channel branched on the ebb-tidal delta into two channels which were surrounded by sub- to intertidal shoals. In 1975 two, clearly separate inlets were present in the Pinkegat. In the period 1972-1979 the eastern end of Ameland expanded, because the westernmost inlet shifted rapidly to the east (with a maximum velocity of 225 $m.y^{-1}$; Oost & De Haas, 1992; Huijs, 1993). In 1979 (Fig. C7) several outer channels of the western inlet had developed. The throat of the eastern inlet migrated only over a short distance to the E (370 m in the period 1972-1979, with maximum migration velocities of 120 $m.y^{-1}$ in the period 1974-1976; Huijs, 1993), so that the distance between the two inlets was reduced.

This development continued in the period 1979-1987 (Figs C7 to C9). The shoal at the eastern end of Ameland became higher (mainly intertidal), while the two inlets merged (Fig. 3). The eastern inlet decreased in depth and cross-sectional area over the period 1970-1979, whereas its remaining outer channel (<-2.5 m) was filled up and abandoned in the period 1979-1991 (Figs C7 to C10; Oost & De Haas, 1992; Huijs, 1993). In the period 1972-1982 the depth and the cross-sectional area of the western inlet increased as it moved eastward (Oost & De Haas, 1992; Huijs, 1993). In the period 1982-1987 a pronounced single inlet configuration developed. It took over the complete drainage and increased in cross-sectional area until 1987; afterwards (1987-1991) it decreased in size. This single inlet migrated also to the E; the throat of the inlet moved with velocities of up to 125 $m.y^{-1}$ (Oost & De Haas, 1992; Huijs, 1993). In 1979 the inlet system still had three outer channels (of which the westernmost one was <-5 m deep). In 1982 (Fig. C8) two were connected with the western inlet, and one was connected with the eastern inlet. The middle outer channel became the most important one, while in the period 1982-1987 (Fig. C9) the western and eastern channels became abandoned. In 1987 the middle channel became some -14 m deep, and drained the whole drainage basin, so that once more a clear-cut single inlet configuration was present. From the map of 1991 (Fig. C10) it appears that small new channels (up to -2.5 m DOL) had already started to form near the eastern end of Ameland. These channels could already be sailed during high tide in 1993 (Reitsma, pers. comm.).

Summarizing: The evolution of the Pinkegat Inlet follows a cyclic pattern, from a multiple inlet to a single inlet and back. One cycle lasts several decennia.

Quantitative changes based on the digitized data (Table II)

The ebb-tidal delta of the Pinkegat is relatively small (approximately 22 km^2). Sedimentation and erosion are quite strong as a result of the intensive morphologic developments. Only a slight net sedimentation occurred over the period 1927-1970/71, in particular at depths below -3 m DOL. The overall net erosion/sedimentation pattern of 1927-1949 (Figs C1 and C2: a deep single inlet gives way to several shallower inlets) is partly the opposite of that of 1958-1967 (Figs C3 and C4; several shallower inlets merge and get abandoned to give way to a single inlet). Over the period 1975-1987 net sedimentation occurred mainly above -6 m DOL. Net erosion was mainly concentrated below that level, due to the formation of a deep single inlet channel.

Table II: sedimentation over various periods; the sedimentation and erosion in the area over various periods has been calculated based on the gridded topography.

Net sedimentation on the Pinkegat ebb-tidal delta; 1927-1970: 22,145,400 m^2, 1970-1987: 22,396,500 m^2 (corrected for extraction of sand)			
Period	Total (10^6 m^3)	Annual (10^6 $m^3.yr^{-1}$)	Annual ($cm.yr^{-1}$)
1927-1949	+2.7	+0.12	+0.55
1949-1958	-3.7	-0.41	-1.85
1958-1967	+4.2	+0.47	+2.11
1967-1970/71	-2.8	-0.80	-3.61
1970/71-1975	+2.4	+0.53	+2.38
1975-1979	+3.6	+0.90	+4.02
1979-1982	+1,8	+0.60	+2.68
1982-1987	-1.8	-0.36	-1.61

Pinkegat, drainage basin

1832-1873/74

In 1832 the channel pattern in the drainage basin of the Pinkegat (Fig. B1) consisted of a deep ENE-WSW-oriented main channel, with N-S and NE-SW-oriented 2nd order[6] branches. The western watershed was situated 4.8 km further to the west than at present

[6] The 1st order channel is the main channel, 2nd order channels are its branches, 3rd order channels are the branches of the 2nd order channels, etc.

(Fig. 3). In 1850 (Fig. B2) only limited observations in the Pinkegat area indicated an increasing eastward directed drainage of the easternmost 2nd order branch (N-S-oriented). The development continued so that by 1854 (Fig. B3) the N-S-oriented channel may have been (one of) the main drainage channel(s); unfortunately the rest of the drainage area was not sounded.

In the years 1871 and 1872 an intertidal earth dam was constructed between Ameland and the mainland over the watershed of that time (Fig. 3). The dam suffered badly from overflowing flood water from the E and storms from especially the W. After repeated repairs (1873, 1875, 1876, 1877) and new damage, the dam was strengthened and heightened to MHW (1879-1881). After repeated repairs and new damage in 1882, new breaches remained unrepaired. The dam was abandoned after a critical review by Lely in 1888 (officially in 1903; Anonymous, 1930).

1873/74-1927

In 1873/74 (Fig. B5) the main drainage channel curved from the NNE-SSW-oriented inlet towards an ENE-WSW-orientation into the drainage basin. To the south 2nd order branches were oriented mainly NE-SW. By then the western watershed had a more eastern position than before (Fig. 3). In 1891 (Fig. B6) the easternmost and westernmost inlets were both connected to the drainage basin by their own main channel. These main channels both were oriented NE-SW. Several channels of the Ameland Inlet system had shifted eastward, thereby breaching the above mentioned dam around 1882. In 1919 the channel system of the Ameland Inlet had partly expanded to the E and the breaches had become larger. Until 1927 only minor changes were made in the map of 1891.

1927-1967

In 1927 (Fig. B9) the channel pattern in the drainage basin of the Pinkegat consisted of an E-W-oriented main channel, with N-S-oriented 2nd order branches. The southern watershed was situated almost 1 km further to the E than in 1891 (Fig. 3). In the period 1927-1949 (Figs C2 and 3) the channels near the watershed became shallower and retreated, while the watershed shifted over more than 1 km to the east.

Between 1927 and 1949 the main channel split up and it rotated counterclockwise in the period 1927-1967. With the exception of the easternmost channel (Smeriggat) all main channels merged during the period 1958/59-1967. The watershed shifted westward in the period 1949-1958/59 (Figs C3 and 3). The southern part shifted to the E in the period 1958/59-1967 (Figs C4 and 3).

1967-1987

During 1967-1971 the main backbarrier channels remained more or less in place. After 1971 strong changes in the orientation of these channels occurred. By 1975, two separate inlets had formed and the main channels became NNE-SSW-oriented connections with the North

Sea. The watershed shifted mainly to the E in the period 1967-1987, interrupted by periods of westward shift (Figs C4 to C10 and 3). In the period 1975-1987 the main channels merged and rotated clockwise. By 1987 a series of smaller 2nd order channels, oriented perpendicular to the main channel, drained the drainage basin. The ENE-WSW-oriented main channel passed seaward into the inlet, while curving around the eastern end of Ameland. The situation resembles the situation of 1927 (Fig. C1) and, to a lesser extent, that of 1967 (Fig. C4). After 1971 the drainage of the area SW of the Engelsmanplaat by the Zoutkamperlaag (via the Smeriggat channel) was taken over by the Pinkegat Inlet, in particular in the period 1978-1981 (Fig. 3).

Summarizing: also here a cyclic behaviour is observed, with the main channel(s) changing in orientation by alternating clockwise and counterclockwise shifts.

Quantitative changes based on the digitized data (Table III)

For the period 1927-1971 calculations for the drainage basin of the Pinkegat refer to an area of almost 57 km^2. Strong sedimentation occurred in the period 1927-1971 (approx. $1*10^6$ $m^3.y^{-1}$). Especially in 1927-1949 sedimentation prevailed on all depths. Sedimentation was low in the period 1958/59-1967.

Table III: sedimentation over various periods; the sedimentation and erosion in the area over various periods has been calculated based on the gridded topography.

Net sedimentation in the Pinkegat drainage basin; 1927-1971: 56,983,500 m^2, 1971-1987: 53,119,800 m^2 (corrected for extraction of sand)			
Period	Total (10^6 m^3)	Annual (10^6 $m^3.yr^{-1}$)	Annual ($cm.yr^{-1}$)
1927-1949	+24.5	+1.12	+1.96
1949-1958/59	+9.1	+0.96	+1.68
1958/59-1967	+2.8	+0.33	+0.58
1967-1971	+5.4	+1.35	+2.37
1971-1975	+3.7	+0.93	+1.75
1975-1978	-0.3	-0.11	-0.21
1978-1981	-1.2	-0.40	-0.75
1981-1987	+3.3	+0.55	+1.04

The calculations for the drainage basin of the Pinkegat were reduced to an area of some 53 km^2 in the period 1971-1987, due to restrictions of the depth soundings. Strong sedimentation occurred in the period 1971-1975, while in 1975-1981 net erosion occurred. In 1981-1987 net erosion below -2.6 m DOL was more than compensated by sedimentation above that level, resulting in a net sedimentation. Over the whole period 1971-1987 net sedimentation occurred above -3 m and erosion below it, the net result being the deposition of $4.5 * 10^6$ m^3 sediment.

Engelsmanplaat/Smeriggat

The Engelsmanplaat is an inter- to supratidal shoal which separates the Zoutkamperlaag Inlet system from the Pinkegat Inlet system. Seismics (Oost & De Haas, 1992) and drillings (Core 91.031; Sha, 1992) show that the shoal sits on top of a massive early Holocene clay deposit at depths of -5 to -10 m DOL. On top of it a sandy shoal is present. New inter- to supratidal sandy shoals are continuously formed at the North Sea side. These are at present separated from the Engelsmanplaat by a channel, the Smeriggat. In the period 1832-1927 the E-W dimensions of the Engelsmanplaat decreased from 7.1 to 2.7 km, mainly because of erosion at the western side (Fig. 3). The erosion continued at a decreasing rate until 1987, after which it accelerated again. In 1991 the Engelsmanplaat was only 2.2 km wide.

1786-1891

Around 1786 a channel was present N of Engelsmanplaat (De Haan et al., 1983). The channel was abandoned in the period 1806-1832 (Figs A8 & B1). In 1832 (Fig. B1) the Engelsmanplaat was a large, mainly intertidal shoal. In the period 1832-1854 lateral migration of several channels of the Pinkegat System eroded the shoal at the western side over 1.3 km. By 1854 a supratidal shoal was present at the northern side of the Engelsmanplaat. In the period 1854-1873/74 (Figs B3 to B5) erosion at the E-side continued both in the drainage basin and along the inlet over another km (Fig. 3). In 1873/74 the supratidal part of the shoal was located at the NE-side of Engelsmanplaat.

1891-1941/45

Erosion at the western side continued in the period 1873/74-1927 (Figs B5, B6, and 3). The erosion by migration of the inlets of the Pinkegat also resulted in a considerable reduction of the northern part of the Engelsmanplaat. By 1891 (Fig. B6) the supratidal part of the shoal had shifted to the NW-side. A small (rest?)channel, with a depth of -2.5 m DOL was present N of the intertidal Engelsmanplaat. By 1903 (Fig. B7) the supratidal shoal of the Engelsmanplaat had become elongated and E-W-oriented. The presence of a small channel in front of the Engelsmanplaat is still suggested by the E-W orientation of the supratidal area at that time. By 1921 (Fig. B8) the -2.5 m zone N of the Engelsmanplaat extended northward and several small intertidal shoals had formed on it.

In 1927 (Fig. B9) the channel 'Smeriggat', north of the Engelsmanplaat, was absent. West of Engelsmanplaat a channel was present, locally known between 1900 and 1950 as the 'Darchgatsje' (= Peat Channel) or, probably derived from it via Smarggat, the 'Smeriggat' (= Dirty Channel; Reitsma, pers. comm.). These names refer to the boulders and remnants of peat, which were frequently found in that channel. With the northeastward shift of the channel in combination with the formation of the present-day Smeriggat channel, the name also migrated northeastwards (Reitsma, pers. comm.). In 1927 the intertidal part of the Engelsmanplaat extended further to the N again, with in front of it a sub- to intertidal shoal which reached far to the north (Fig. B9).

1941/45-1991

In the period 1927-1949 the shoal Engelsmanplaat was eroded over 0.5 km by the easternmost outer channel of the Pinkegat which migrated clockwise and shifted to the south (Figs C1, C2, and 3). Probably around 1941/42 (Reitsma, pers. comm.), but at latest in 1945, the Smeriggat channel became connected with the western outer channel of the Zoutkamperlaag Inlet, and became bordered to the NW by a sub- to intertidal sandy shoal. The shoal increased in size, and shifted to the SE in the period 1949-1958 (Figs C2 and C3), thereby enclosing the Smeriggat channel. The development continued (Fig. C4), so that by 1970/71 (Fig. C5) the Smeriggat had become a completely isolated channel. The shoal N of it had become intertidal by that time. After reduction of the supratidal area of the Engelsmanplaat in the period 1927-1949 the intertidal shoal gradually increased in size until 1970.

In the period 1970/71-1991 (Figs C5 to C10) the Engelsmanplaat became lower: from supratidal to intertidal. At the same time the mainly wave-built shoal north of the Smeriggat became higher (from intertidal to supratidal). In the period 1970/71-1979 the shoal shifted southward and got a sickle form. In the period 1979-1982 especially the northern side of the shoal continued to shift to the S. N of it a new shoal formed. After 1982 the whole shoal continued to migrate southward, and it became oriented ENE-WSW. In the period 1982-1991 the height of the shoal decreased from supratidal to mainly intertidal as a result of strong erosion at its western side. In the same period the new shoal N of it had become higher (mainly intertidal), and migrated to the south as well.

In 1970/71 the Smeriggat still formed a shallow connection between the Zoutkamperlaag and the drainage area SW of the supratidal Engelsmanplaat (Fig. C5). In connection with the southward shift of the shoal north of it, the Smeriggat was gradually abandoned, and became shallower over the period 1970/71-1991 (Figs C5 to C10). By 1991 the overall depth had decreased to less than -2.5 m and an intertidal ridge had formed across the channel (Oost & De Haas, 1992; 1993). During the abandonment, the Smeriggat channel was filled rapidly (Fig. 10) with two main types of sediments (Van den Heuvel, 1993). One type shows little or no variation in grainsize over many dm. Sedimentary structures consist of strongly bioturbated, small-scale cross bedding, erosive surfaces and upper stage plane beds. The other type shows a rhythmic, likely annual, alternation of cm-thick layers of fine-grained sediment (clay/sand mixture) alternating with sands. The fine-grained sediments show a stronger bioturbation than do the sandy layers, which show ripples, parallel laminae, loadcasts, and contorted bedding. Both types of sediments point to relatively quiet conditions, enabling deposition of fines and bioturbation, alternating with higher energetic events, promoting erosion and the fast deposition of sands.

Quantitative changes, based on the digitized data

The Engelsmanplaat area used for calculations includes the Engelsmanplaat proper, the channel Smeriggat, and the shoals N of it (total 11.9 km2). Erosion dominated the area in the period 1927-1949 with a maximum at the -1 m level (erosion of subtidal and lower intertidal shoals during the formation of the channel). In the period 1949-1970/71 sedimentation was dominant at greater depths. The maximum of sedimentation shifted from -5 m to -2 m DOL, while the Smeriggat channel was filled up. In the supra-tidal zone above the +1 m level slight sedimentation occurred as the supratidal part of the Engelsmanplaat increased in size. In the period 1970/71-1979 slight net sedimentation occurred ($5.3 * 10^6$ m^3). Afterwards erosion became dominant. Erosion occurred in particular above DOL, due to the erosion of the Engelsmanplaat. Sedimentation occurred below DOL, because the infill of the Smeriggat channel continued.

Although the size of the Smeriggat area is only 3 km^2, the sedimentary changes are quite intense. During the formation of the Smeriggat in the period 1927-1949, erosion prevailed at all depths. In all succeeding periods between 1949-1970/71 net sedimentation occurred at greater depths, whereas erosion dominated in the shallower parts (above -3 m). During the whole period 1927-1967 net erosion occurred between -8.5 to +0.9 m (max. $19*10^3$ m^3). Above +0.9 m sedimentation occurred (max. $4.5*10^3$ m^3). In the period 1967-1970/71 sedimentation occurred below -0.5 m and erosion above it. After 1970/71 sedimentation dominated at greater depths, because the channel became gradually shallower. The net effect was an average sedimentation of approximately 1 m in the period 1970/71-1987 (6 $cm.yr^{-1}$!). Net sedimentation over that period was $2.3 * 10^6$ m^3.

THE PINKEGAT INLET SYSTEM AND THE ENGELSMANPLAAT/SMERIGGAT: DISCUSSION

In general inlets tend to migrate in the direction of the residual current (FitzGerald, 1988; Steijn, 1991). The same accounts for the smaller outer channels (Oost & De Haas, 1992, 1993). At the same time there is a tendency of the prominent tidal channels on the outer delta to be located at the updrift side of the ebb-tidal delta (Van Veen, 1936; Sha & De Boer, 1991).

Pinkegat, ebb-tidal delta

The configuration of the ebb-tidal delta in the years 1832, (1873), 1927, and 1967 resembles the situation of 1987: a single inlet, which curved around the E end of Ameland (Figs B1, B5, B9, C1, C4, and C9). Between these years more than one inlet was present. Clearly the system shows a cyclic development pattern from a single inlet to a multiple inlet configuration and back (Oost & De Haas, 1992, 1993). Based on observations and internal reports (Isbary, 1936; Jansen, 1941; Postema, 1956; Van Straaten, 1964; Postma, 1982; Reitsma, pers. comm.; Van der Geest, pers. comm.) it is concluded that from 1927 to 1987[7, 8] the cycle took 40 and subsequently 20 years (from single inlet configuration to single inlet configuration). Single inlet configurations also existed around 1832 and 1873 and likely also in 1894. The latter is concluded from the eastward expansion of the HW-line of Ameland in that year (Isbary, 1936). Multiple inlets can be observed on maps of 1850-1854 and 1881/91, they and likely also occurred during a part of the period 1894-1927. The latter is concluded from the westward retreat of the easternmost HW-line of Ameland in the period 1894-1917 and the presence of a multiple channel configuration around 1900 (Mars & Cordia, 1901; Isbary, 1936; see description of the cycle below). Thus, from maps of 1832 onwards and the observations it is concluded that the cycle of development of the Pinkegat Inlet repeats with a varying period of 20 to 40, at maximum about 41 years (or 54, if the changes between 1891 and 1915 are not related to multiple and single inlet configurations). Judging from the formation of shallow channels across the eastern end of Ameland in the period 1987-1991 (Fig. C10), which could be sailed during high tide in 1993 (Reitsma, pers. comm.), the present-day Pinkegat develops again towards a multiple inlet.

In detail the cycle develops as follows (cf. Oost & De Haas, 1992, 1993; cf. Huijs, 1993; Fig. 4):

A) a phase with one inlet, with a maximal cross-sectional area. The inlet shifts to the east (downdrift). A sandy shoal develops west of it and forms the eastern end of the adjacent barrier island Ameland. The shoal is gradually eroded at the southern side by channels from the drainage basin.

B) a phase, in which several inlets are formed, mainly by cutting through the western supra- to subtidal sandy shoal. These inlets develop from spill-overs which in their turn probably evolve from small flood chutes. During this phase the cross-sectional area of the eastern inlet becomes smaller.

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[7] The development of the easternmost outer channel in the period 1927 to 1970 will be discussed in the chapter on the Engelsmanplaat/Smeriggat.

[8] No indications could be found that the closure of the Lauwerszee embayment (downdrift of the Pinkegat) influenced the development of the Pinkegat system.

C) a phase during which part of the inlets is abandoned and others merge (the same happens to the outer channels), to give way to a new single inlet. The merger occurs mainly, because the westernmost inlet shifts fastest to the E. In the process the westernmost inlet increases in depth and cross-sectional area as it moves eastward and becomes the main inlet. During the phase a new shoal forms at the eastern side of the updrift island (Ameland).

Cyclical changes in the geometry of tidal inlets, as described above, are not unusual (Oertel, 1973; Oertel, 1977; Tye, 1984; FitzGerald, 1988; Sha, 1989a, 1990). In a tidal system with several inlets, the growth of shoals, and shifts of currents may lead to the formation and widening of certain channels and the abandonment of others. The exact rate of change is de-

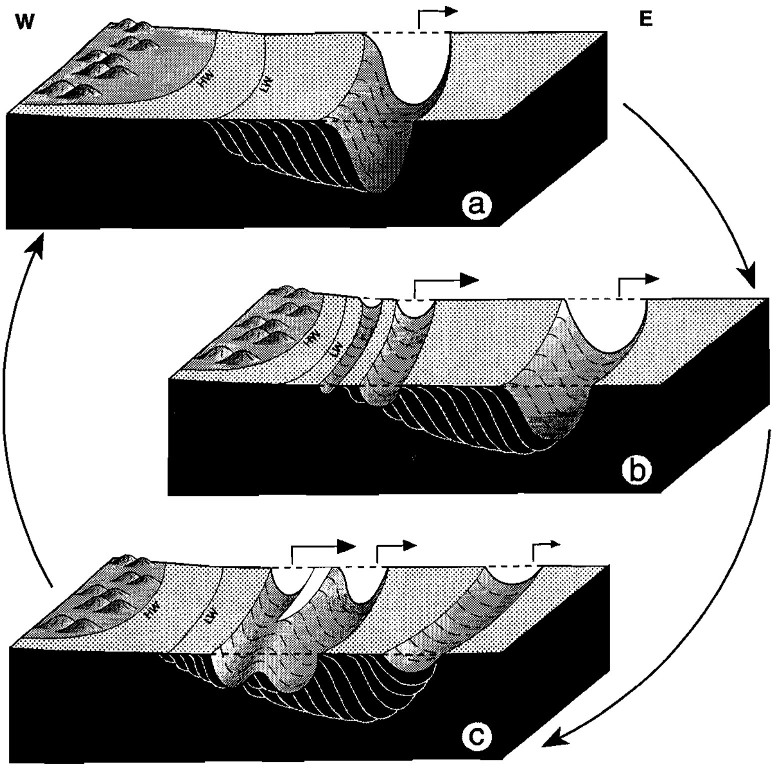
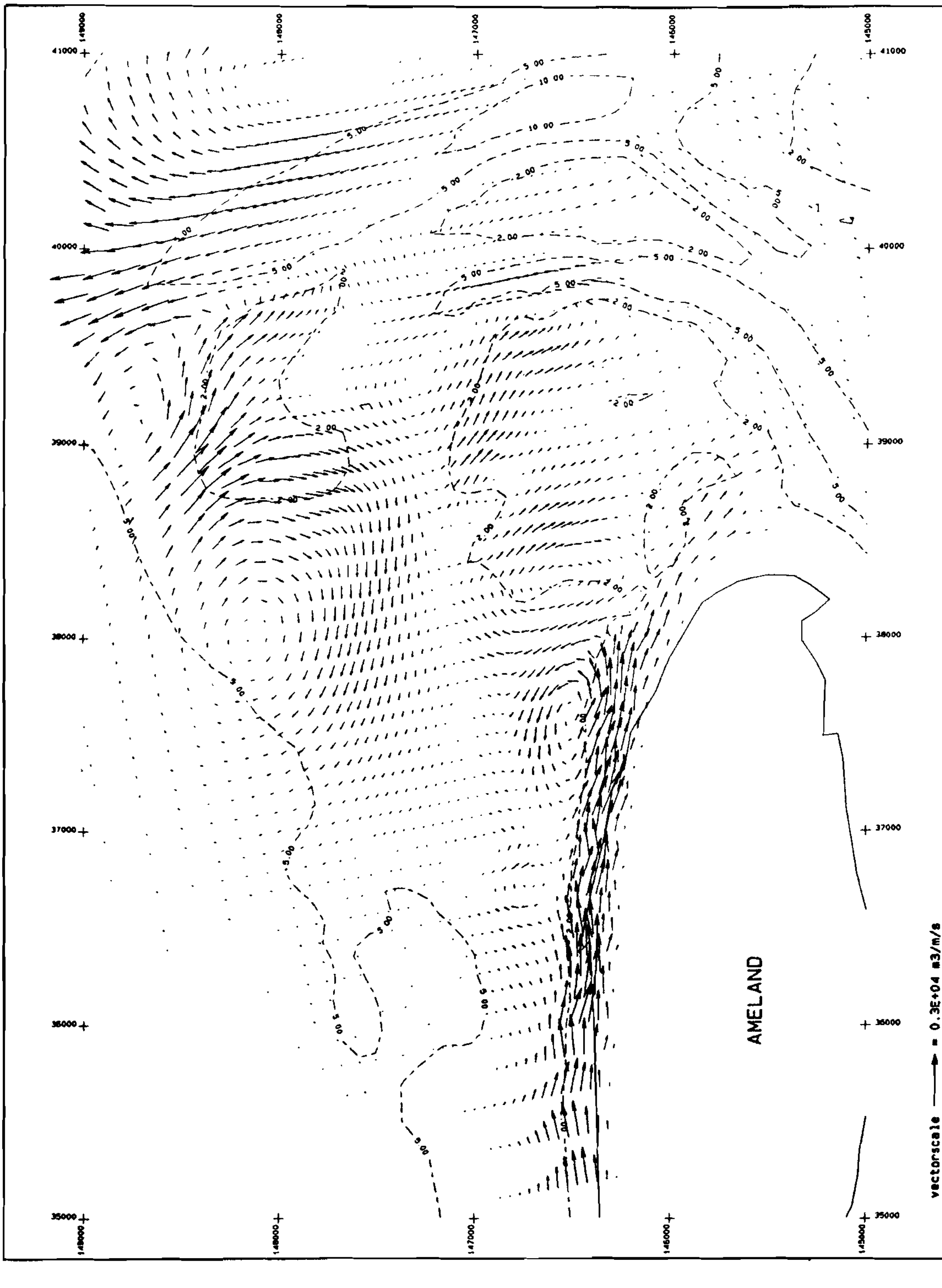


Figure 4: The cyclic development of the inlets in the ebb-tidal delta of the Pinkegat:

a) A single inlet shifting to the E and development of a sandy shoal at the downdrift end of Ameland. Erosion at the backbarrier side of the shoal not depicted.

b) Formation of a multiple inlet configuration; erosion of the sandy shoal.

c) Abandonment and merger of inlets; formation of a new single inlet and the development of a sandy shoal. Small arrows indicate relative migration velocity of channels.


AMELAND
vectorscale = 0.3E+04 m3/m/s

pending on the local relative importance of waves and tidal energy (Hubbard, Oertel & Nummedal, 1979; Steijn, 1991; Steijn et al., 1992; Huijs, 1993). For the ebb-tidal delta of the Pinkegat, model calculations indicate that tide- and wave-induced transport are both important (Fig. 5; Steijn et al., 1992). This is supported on one hand by tidal current data from ebb-tidal deltas, and on the other hand by observations (aerial photographs and visual) of wave-refraction around shallower areas, and frequent swash-bars (Oost & De Haas, 1992). The latter observations indicate a strong influence of waves in shallower areas. Direct measurements in channels show that during storms also the current velocities in the inlet channels may change dramatically (Rakhorst, 1981). Breaker bars (WSW-ENE-oriented, slightly oblique to the coast) and model calculations moreover indicate that in the shallow offshore along the beach of Ameland wave-induced eastward sediment transport is important and that tidal transport is far less influential (Steijn et al., 1992; Oost & De Haas, 1992). Model calculations furthermore indicate that during storms from the west the wave-induced transport zone extends to deeper water (Steijn et al., 1992). These modelling results are supported by measurements along Texel and along the coast of Holland. They show that, as a general trend, currents increase with increasing wind velocities, especially when the tidal-current and wind direction are identical, and also that the zone of higher current velocities extends to deeper water (Rakhorst, 1981). Under fair-weather conditions tidal-current velocities in somewhat deeper water are nearly symmetrical and hence result in only a small residual eastward current.

Offshore bar migration and the displacement of depressions indicate a general eastward sediment transport (Ehlers, 1988). The generally decreasing grainsize from W to E along the foreshores of the Dutch barrier islands indicates that the coarser grains are trapped by the inlet systems and that only finer grains pass the ebb-tidal delta (Veenstra & Winkelmolen, 1976; Ehlers, 1988). The commonly eastward decreasing grainsize of the beaches and dunes of individual islands of the Dutch and German Wadden Sea (Endema, 1979; Veenstra & Winkelmolen, 1976; Veenstra, 1982) also confirms the eastward transport and selection. Local overprints due to e.g., erosion of offshore early Holocene and Pleistocene deposits cause local deviation from this general trend of longshore eastward transport of fine-grained sediment (cf. Veenstra & Winkelmolen, 1976; Veenstra, 1982; Ehlers, 1988).

Figure 5 (opposite page): Computed annual net sediment transport by water through the Pinkegat ebb-tidal delta based on the bottom topography of 1970/71, using sediment transport formulas of Bijker (1967, 1971) and the following weighted occurrence of tides and waves: 125 days.yr^{-1} tides without waves, 112 days.yr^{-1} tides with waves from the north, 128 days.yr^{-1} tides with waves from the northwest (Steijn et al., 1992; courtesy Steijn). Note the strong eastward coast-parallel sediment transport along Ameland and the S- to SE-directed transport over shallow shoals.

What causes the cyclic development of the Pinkegat?

Downdrift migration

An important element in the morphodynamic development of the area is that all the inlets and outer channels tend to migrate downdrift. This is brought about by a the following factors:

1) The eastward sediment transport by the coast-parallel littoral drift (Fig. 5; Steijn et al., 1992; Steijn & Louters, 1992; Huijs, 1993; Oost & De Haas, 1993), in combination with aeolian transport over the beach under the influence of westerly winds (Eysink, 1979). Due to the sand supply, the eastern end of Ameland can expand rapidly and consequently inlets are forced to migrate downdrift.

2) The shoals on the ebb-tidal delta are forced to migrate to the SE to S by the combined influence of flood currents and waves[9], (e.g., Fig. 5; Steijn et al., 1992; Steijn & Louters, 1992; Oost & De Haas 1992, 1993; Allersma, 1993). The shift of these shoals forces the clockwise migration of the channels east of them.

3) The lateral accretion at the inner bend of the inlet channel(s), which drain the drainage basin. During both flood and ebb a secondary flow is generated due to the inertia of the water (centrifugal acceleration). It has a velocity of about 5% of the main flow (Allersma, 1993). The Coriolis force also generates a force proportional to the flow velocity, which is of the same size as the centrifugal force (Allersma, 1993). On the northern hemisphere, the Coriolis force is directed to the right with reference to the direction of flow. Thus, in the Pinkegat it counteracts the centrifugal force during the flood, but strengthens it during the ebb. The forces result (in particular during the ebb) in erosion slightly downstream of the outer bend of the channel. At the same time sediment is deposited slightly downstream of the inner bend (lateral accretion). The dynamical equilibrium is thus maintained, because the cross-sectional area of channels remains constant when tidal prism is constant (Chapter 1; Johnson, 1919; Gaye & Walther, 1935; Nummedal & Penland, 1981; FitzGerald, 1988; Gerritsen, 1990). In the Pinkegat Inlet erosion of the outer bends in combination with lateral migration is quite influential, because of the strong curvature of the inlet and its outer channels (Huijs, 1993).

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[9] The waves are important, because they can generate currents and form breaker bars in the shallow subtidal and intertidal zones, as is obvious from aerial photographs (Oost & De Haas, 1992). Most important, however, is that they stir sediment (Steijn et al., 1992). In the Frisian Inlet this sediment is subsequently easily transported by (tidal) currents, because the small grainsize allows transport as saltation load (Winkelmolen & Veenstra, 1974). Model calculations showed that sediment transport may thus increase by an order of magnitude, in comparison with only the effect of the tidal movement (Steijn et al., 1992).

The strong curvature is determined by both the position of the drainage basin and the propagation of the tidal wave. The drainage basin is positioned updrift with respect to the inlet. This implies that during ebb and flood the water current is deflected almost 180 degrees. Strong curvature also occurs in other Dutch inlets which predominantly drain updrift located backbarrier basins, e.g., the Eijerlandse Gat, Vliestroom, and Eijlander Balg. The propagation of the tidal wave contributes to the curvature of the inlet, because it forces a NW-SE directed inlet. The propagation direction and velocity of the tidal wave determine the phase differences between locations. When tidal amplitudes are identical throughout the area, the tidal current will prefer the path with the strongest gradient in phase difference, which is for the Pinkegat NW-SE (Fig. 6; Van Veen, 1936; Sha, 1989b, 1990; Huijs, 1993).

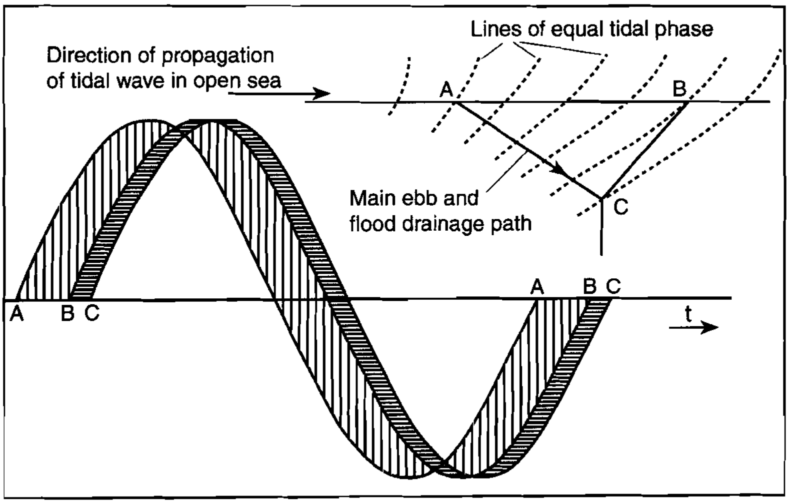

Figure 6: The 'motoric capacity' (Van Veen, 1936): The lines AC and BC are two possible inlet orientations. The tidal wave propagates from W to E and first arrives in A (at t=0), then in B (at t=4) and then in C (at t=5). The tidal curves (here having the same amplitude) of the stations are drawn along a time axis. The surface between the curves for stations A and C is much larger, both for the ebb and for the flood, than between the curves for stations B and C. These surfaces, being determined by the tidal amplitudes ànd by the phase difference between stations, have been called 'motoric surfaces' by Van Veen (1936). The tidal current will prefer the path of the greatest motoric surface, i.e., gradient. In the figure, where tidal amplitudes are identical, it is determined by the strongest gradient in phase difference, i.e., AC, instead of BC. Huijs (1993) showed that a strong gradient in the flood and ebb tides occurs along the eastern end of Ameland in the NW-SE direction. Thus, the preferred transport path, i.e., the stretch along which channel development occurs is oriented in that direction.

The single inlet configuration

From the total balance of tidal forces, wave forces, and sand supply, a balance which also depends on the dimensions of the drainage basin, an optimum position results for the inlet to drain the drainage basin (cf. Sha, 1990). The optimum position is reached fully during a single inlet configuration (cf. Huijs, 1993). Then, the single inlet transports maximum amounts of water and therefore has maximum dimensions. Also, its position is relatively stable, because of the large tidal prism flowing through it (cf. Sha, 1990).

However, the downdrift migration of the single inlet continues, albeit at a slow rate. As a result the position of the inlet becomes less optimal (if only tidal forces are considered the inlet moves more and more towards point B in Fig. 6). At the same time a (large) shoal is formed in the inner bend of the inlet at the eastern end of Ameland. Over the shoal flood chutes are formed, which are scoured by the flood current when the higher subtidal to intertidal shoal is flooded.

Formation of a multiple inlet configuration

At some point the eastern position of the inlet reaches a critical limit, and it becomes so inefficient (too long and slow transport path to the drainage basin) for the drainage of the backbarrier basin, that new inlets form updrift of it along a more preferred, faster pathway (Figs 4, 6, and 7; cf. Van Veen, 1936; Oertel, 1977; FitzGerald, 1988; Huijs, 1993). A multiple inlet configuration is reached by the formation of spill-overs mainly over the shoal updrift of the inlet (= eastern end of Ameland)[10]. These spill-overs may develop during storms from flood chutes. In time they can develop into completely new inlets. Their development through the shoal is enhanced by the main backbarrier channel. The channel becomes mostly ENE-WSW-oriented during a single inlet situation and erodes the shoal at the backbarrier side, so that the distance over the shoal E of Ameland to the open sea becomes smaller. Nevertheless, depending on the extent of the shoal E of Ameland, erosion and updrift generation of inlets can take a considerable period (e.g., from 1927 to 1958).

Instability of a multiple inlet configuration

New inlets which are formed, in addition to an existing main inlet, lead to a change of the current pattern in the area. As a result, the original inlet is out of equilibrium. It cannot

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[10] During the last centuries the eastern side of the barrier island Ameland has expanded net to the E. A similar growth has also been observed for the German barrier islands by Luck (1975). His observations suggest that parts of the E-point which become supratidal dunes, are not easily changed back into inlets. This is probably because flood chutes cannot form easily in such high areas. This, and perhaps the eastward shift of the watersheds, which temporarily reduces the tidal prism and thus the tidal energy through the inlets, enhances the eastward expansion of Ameland. As stated in chapter 2 a net growth at the downdrift side is only possible when it is more or less compensated by a net erosion at the updrift side of the island, or when a double inlet of an estuary gives way to a single inlet (cf. Van der Spek, 1994).

return to its original dimensions, due to the strongly changed hydrodynamical conditions. Adding to the instability is the local, relative increase of wave action, because the tidal prism moving through each individual inlet in a multiple inlet configuration is smaller than in a single inlet configuration.

On a theoretical basis Van de Kreeke (1990) stated that the stability of an inlet depends on the wet cross-sectional surface area and the hydrodynamics in the inlet. The most important parameters are the shear stress and the equilibrium shear stress. The latter is the bottom shear stress needed to erode the sediments from the inlet. The inlet is in equilibrium with the hydrodynamical conditions when the maximum bottom shear stress, defined by the sediment characteristics, tides and wave action, is equal to the equilibrium shear stress. The notion equilibrium refers to a dynamic equilibrium, such that after a slight change (for instance sedimentation during a storm) the cross-sectional area of the channel returns to its original dimensions. With the help of a lumped parameter model Van der Kreeke (1990) showed that in a tidal system with two inlets neither inlet is stable. A slight change in the wet cross-sectional surface area of one of the channels will lead to disturbance of the unstable equilibrium, resulting in the infill of one, or both (Van der Kreeke, 1990). The same holds for a system with more than two inlets. This means in practice that any tidal inlet system with more than one inlet will tend to give way to a system with only one inlet.

Change from a multiple inlet to a single inlet

In the Pinkegat, updrift inlets are observed to migrate downdrift more rapidly than do downdrift inlets (Huijs, 1993), because:

1) the supply of sand to inlets is dominantly from the updrift island coast (Fig. 5; Eysink, 1979; Steijn et al., 1992; Steijn & Louters, 1992; Huijs, 1993), and

2) during a multiple inlet configuration, the inlets with smallest dimensions are the more western ones (Figs 4 and 7). For these inlets the effects of storms and local wave-climate (mainly directed to the SE to S) are relatively more important, because of their small tidal prism (cf. Sha, 1990), than for the larger eastern inlets.

During eastward migration the dimensions of updrift inlets increase with the increasing tidal prism through it, until a single inlet configuration is reached by merger of adjacent channels (Figs 4 and 7). During the process (Fig. 4) outer channels and other inlets merge with the downdrift migrating updrift inlet (e.g., 1975-1982), or get abandoned (e.g., Fig. C2, Smeriggat after 1949; Figs C8 and C9, outer channels 1982-1987). Afterwards the new inlet decreases in size (Fig. 7; cf. Oost & De Haas, 1992; 1993; Huijs, 1993), because in time again a multiple inlet system with new inlets in the updrift part of the system develops. Complete abandonment of the old inlet can follow when it does not merge with updrift inlets and continues to migrate downdrift. Thus, it becomes more and more inefficient for drainage

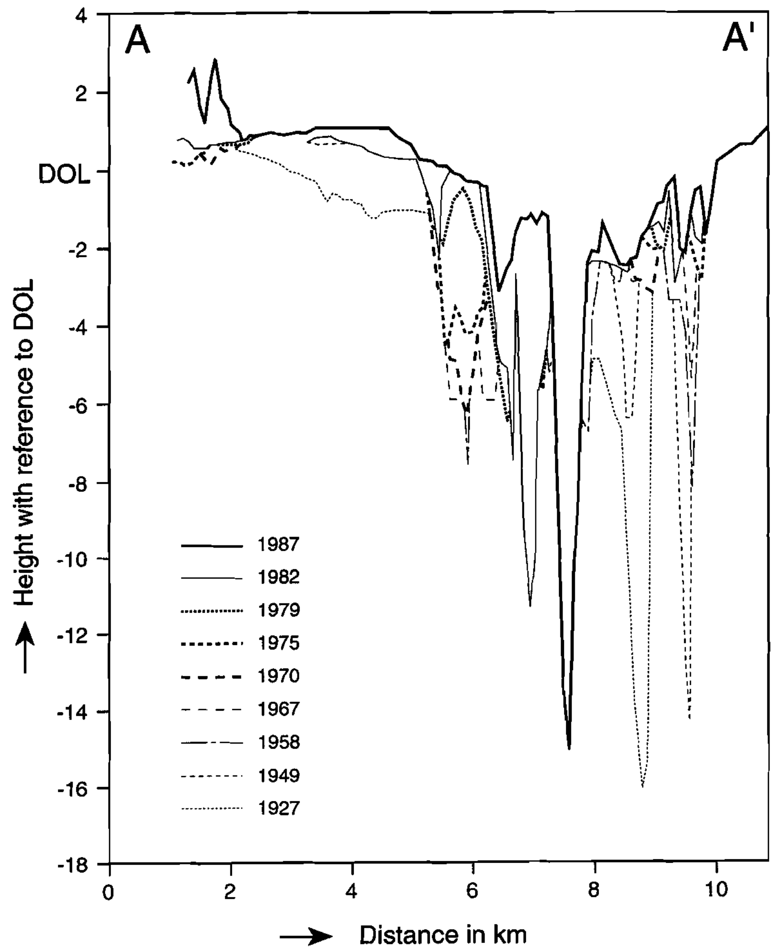

Figure 7: W-E cross-section A-A' (for location see Fig. 2) showing those parts of the profiles for the period 1927-1987, which have not been 'eroded' by later profiles. Note the deepening and subsequent shallowing of the channels of the Pinkegat Inlet system to the E (i.e., downdrift).

of the backbarrier basin (cf. Tye, 1984). The old inlet or parts of it (outer channels) are filled in with sediments from the sandy shoals, which approach from the updrift direction or from the N. In the mean time the other, or one of the other inlets develops into a new main inlet. The above historical data indicate that in the Pinkegat system this is mostly the westernmost one, which may, additionally, merge with one or more other inlets.

Pinkegat, drainage basin

Changes in the channel pattern in the drainage basin of the Pinkegat are closely linked to the developments in the ebb-tidal delta. Comparison of the maps from 1832 onwards shows that the main backbarrier channel had often (but not always) a NNE-SSW orientation when a double inlet configuration existed. The main channel usually had a more ENE-WSW orientation during periods of a single inlet, which was positioned far to the east.

Most of the lower order channels (in particular those at some distance from the inlet) have a relatively more stable position than the main channel (Flemming & Davis, 1994). Their ideal position is in the centre of their drainage area. If a channel shifts and enters the drainage area of another channel, one of both will generally be abandoned. At the original place of the channel a new drainage channel will develop (Fig. 8). This and the continuous formation of chutes within the same channel (Fig. 8; Van Straaten, 1958) restricts the lateral migration of smaller channels.

The main channel(s) form the connection between these relatively stable smaller channels, and the continuously shifting inlet. The main channel is NNE-SSW-oriented when the inlet has a western position. It becomes increasingly more ENE-WSW-oriented when the inlet, or inlets, shift to the E (Oost & De Haas, 1992). Such EWE-WSW orientation is enhanced by the inertia of the currents. The inertia of the currents also leads to the deepening of those lower order channels that become oriented in the direction of the main backbarrier channels (or even merge with them), and in the shallowing of those that become oriented at an angle to them. Up to 8 m thick clay infills of channels (cf. Kievits, 1994) show that such changes may sometimes be very abrupt. The cyclic changes are illustrated by the developments in the Pinkegat Inlet:

In 1927 there was a single inlet configuration with an ENE-WSW-oriented main channel. In the period 1927-1941/45 a multiple inlet configuration developed with the main backbarrier channel still having an ENE-WSW orientation. As late as 1949-1958/59 the main channel became strongly NNE-SSW-oriented, as a new inlet formed more to the west. This illustrates that the orientation of the main channel only changes if there is a need for it. This "need" is strongly determined by the position of the inlet. A NNE-SSW orientation of the main channel can only form when short-cuts through the eastern end of the island Ameland are formed leading to multiple inlet configurations. An ENE-WSW orientation of the main channel is forced by the build-up of a shoal at the eastern end of the island, which occurs in particular during the development of a single inlet configuration, when the inlets shift to the E. The backbarrier channels themselves also influence the cyclic development, because they erode the shoal at the backbarrier side (cf. Isbary, 1936). Thus, the scene is set for an easier break-through of the shoal and the onset of a multiple inlet configuration.

Also, the position of the watershed is influenced by the developments in the inlet (Fig. 3). This position is mainly determined by the place where the flood waves, which enter

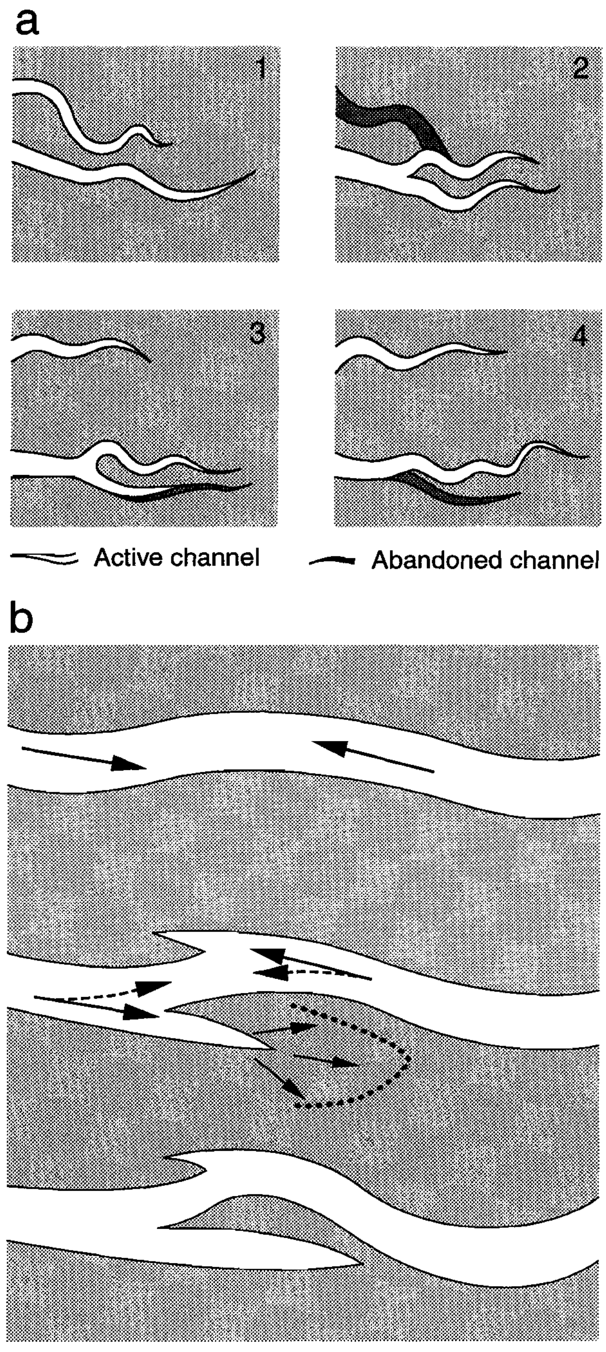

Figure 8: Processes which limit the extent of lateral shift of lower order channels: a) take-over of drainage and abandonment, based on observations of the Pinkegat drainage basin, and b) chute formation (after Van Straaten, 1964).

along the western and eastern end of the island, meet each other[11]. Since the tidal wave in the North Sea moves from W to E, the flood waves meet each other east of the centre of the island. Measurements over a four-month period, showed that a pure astronomical tide is only active during 20% of the time. Usually winds and waves generate an additional flux of water and sediment to the E and thus generate a continuous tendency of the watershed to shift to the E (De Boer, 1979; De Boer et al., 1991a; cf. FitzGerald & Penland, 1987). Normal astronomical tidal currents counteract such development. The balance between these forces determines the position of the watershed. When at the existing watershed a phase difference occurs between the two flood waves, the watershed will shift. In total the watershed of Ameland shifted over about 4.8 km to the E in the period 1832-1987. It appears that eastward shift of the watershed S of Ameland coincides with the development of situations in which:

1) the water transport path becomes longer due to a more eastern position of the inlet(s) in the Pinkegat system, as is illustrated by the developments 1891-1949, 1958-1967, (S watershed), 1975-1987. The reverse is observed in the period 1949-1958 (Fig. 3),

2) the water transport along the eastern end of Ameland to the watershed becomes more difficult and slower, due to the existence of several smaller inlets or main channels, as illustrated by the changes from 1873/74 to 1891 and 1927 to 1949[12], or

3) the orientation of the major backbarrier channels only allows a slow transport of the tidal water to a part of the watershed. The reverse process is illustrated by the westward shift of the northern watershed in the period 1981-1987, when the main channel of the Pinkegat Inlet became directed to that location, although the transport distance became larger.

It should be noted that during the formation of a multiple inlet configuration these effects can counteract each other. Short-cuts through the shoal favour a westward displacement of the watershed, but the smaller inlets favour an eastward displacement. Furthermore, the same time the reorientation of the main backbarrier channel from NNE-SSW to ENE-WSW leads to a westward shift of the southern watershed and an eastward shift of the northern watershed. Depending on their relative influence these forces will together determine the shift of the watershed. The relatively close link between the position of the watershed and the position of the eastern end of Ameland suggests that the position of the island ends is quite

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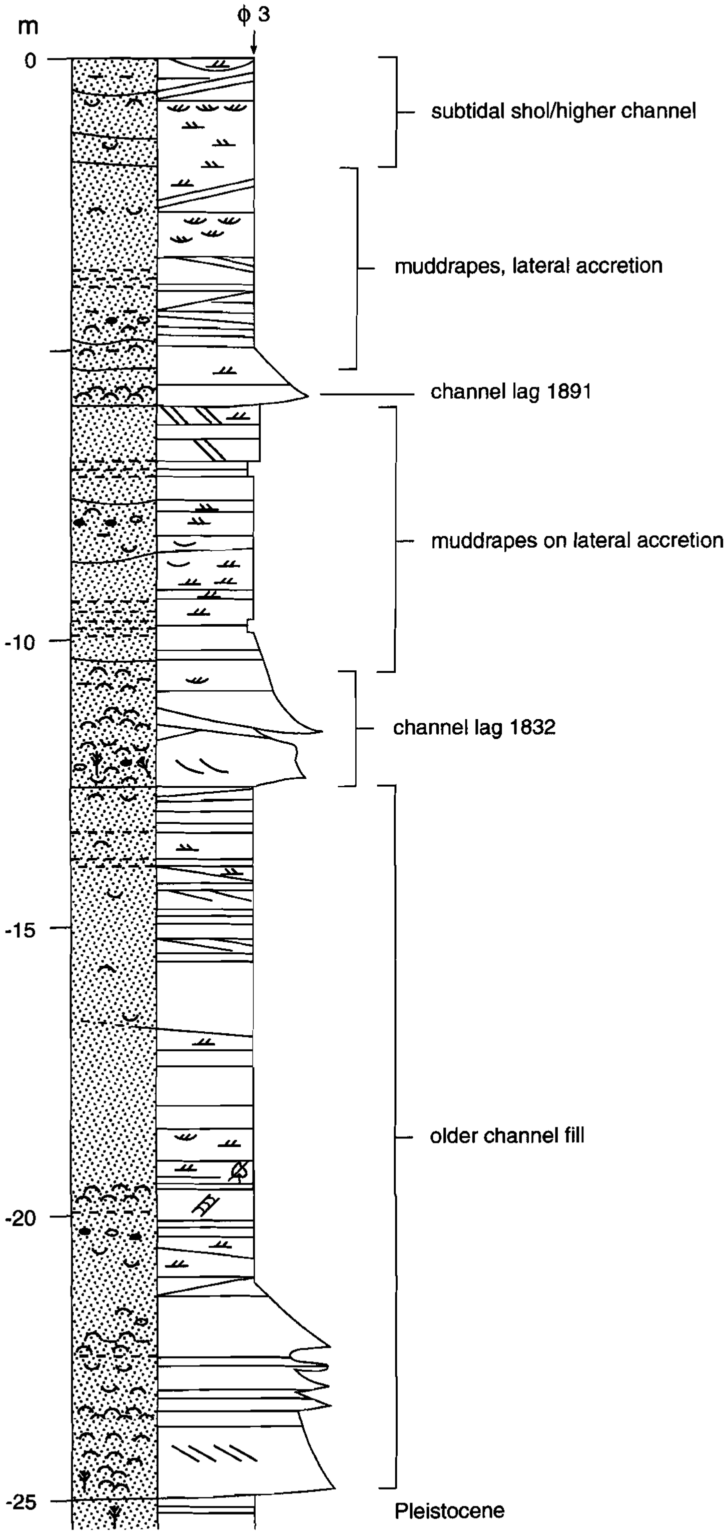
[11] Also, the velocity asymmetry of the tidal currents (higher flood velocity) due to contortion of the tidal wave is of importance (FitzGerald & Penland, 1987).

[12] The development over the period 1927-1949 was also due to the eastward shift of the HW-line in that period and perhaps also to some time-lag between the eastward shift of the shoal which forms the eastern end of Ameland (the LW-line shifted 3 km eastward in the period 1915-1927) and the eastward shift of the watershed.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channel lag 1832
older channel fill
Pleistocene

important (Fig. 3). The quick shift of the watershed in the 19th century is, for a large part, attributed to the reorientation of Ameland Inlet and the coinciding erosion of the western end of Ameland. Later shifts of the watershed are mainly attributed to the shifts of the eastern end of Ameland. A strong determination of the position of the watershed by the position of the island ends, in particular the eastern end of the barrier island, is also suggested by a study of the German Wadden Sea (Luck, 1975).

Although the changes in the inlet and the ebb-tidal delta strongly determine the developments in the drainage basin, also a reverse forcing occurs. This has already been illustrated by the influence of the updrift position of the drainage basin on the curvature of the inlet and the role of the erosion by backbarrier channels in the development of the multiple inlet phase (see above). Furthermore, the tendency of the watershed to shift to the E may reduce the tidal prism from time to time[13]. Such a temporal reduction of the tidal prism of the drainage basin leads to a temporal reduction of the tidal forces, the related cross-sectional area of the inlet and the size of the ebb-tidal delta (see chapter 1).

Geological relevance

The observed cyclical development of the Pinkegat Inlet generates several phenomena:

1) During the strong morphodynamic changes of the Pinkegat Inlet system millions of m^3 of sediment (especially fine sand) are transported annually between the various parts of the Sand Sharing System (tables II and III). The system thus adjusts constantly to the ever-changing hydrodynamic conditions. In general the resulting deposits will show many internal erosion features and other indications of local, strong changes in energy conditions, especially of the tidal current energy (for instance abandonment facies).

2) During the migration of the western inlet in the downdrift direction (i.e., to the E) extensive lateral accretion deposits are formed along its inner bend (Fig. 7). In the period 1972 to 1991, the throat of the (western) inlet migrated over a distance of 2,280 m (average 120 $m.yr^{-1}$; Huijs, 1991), leaving extensive lateral accretion deposits, formed at the inner bend of the channels. In the downdrift direction these deposits first increase in maximum depth (single inlet configuration) and then they decrease, proportional to the changes in the tidal prism flowing through it, during the development of new inlets updrift. From such inlet

Figure 9 (opposite page): Sedimentary log of core 91.032 (after Sha, 1992). For interpretation see remarks along the log and in the text. Position indicated by cross-mark in Fig. 1.

[13] For the Pinkegat this reduction is roughly compensated by the retreat of the western side of the shoal Engelsmanplaat.

deposits in fossil examples the residual current direction through the adjacent open sea basin can be estimated, because channel migration in mixed energy inlets is usually downdrift (cf. FitzGerald, 1988). Furthermore, the maximum depth of the resulting inlet deposits can be used to estimate the tidal prism of the total drainage basin (Sha, 1990)[14]. The sedimentary development of Smeriggat shows that an abandoned inlet may be filled by partly fine-grained sediments when it becomes sheltered seaward by shoals (see Engelsmanplaat).

3) The eastern end of Ameland has expanded net to the E. Therefore the ideal resulting sedimentary sequence below Ameland can be inferred to be (from bottom to top): a) a deep channel lag and some channel deposits (single inlet phase with maximum cross-sectional area), b) mainly lateral accretion deposits), c) at least one shallower channel lag, covered by shallowing channel deposits (multiple inlet phase, mainly lateral accretion deposits), d) sub- to intertidal shoal deposits, possibly alternating with flood chute deposits, e) intertidal to supratidal beach deposits, and f) supratidal marsh deposits and/or dune deposits.

Indeed, double channel lags resulting from the shift of the inlet and its cyclic behaviour have been observed at approximately -15 m DOL and -9 m DOL with in between lateral accretion deposits (Sha, 1992; Fig. 9). Comparison of core 91.032 (Sha, 1992) with the historical development shows that the channel lag at -15 m DOL was most likely formed during the single inlet phase around 1832 (Fig. B1) when the site was located in the centre of the inlet throat. Postma (1982) reports a depth of -14 m DOL. The channel lag at -9 m DOL may have been deposited by the westernmost inlet channel during the multiple inlet configuration around 1891 (Fig. B6), or around 1917 when there may also have been a multiple inlet configuration (Fig. 9). Afterwards no inlet channel has been present at that place. The sedimentary sequence is comparable to those described by Kumar & Sanders (1974), Hubbard (1977), and Reinson (1984). It therefore is concluded that, alternatively to a deep channel formed by the main ebb channel and shallower marginal flood channels (Kumar & Sanders, 1974; Hubbard, 1977), a cyclic development of the inlet system in combination with a net downdrift shift may result in an identical sequence. It should be noted that during a multiple inlet configuration the most updrift inlet may be flood-dominated.

4) In general, the outer channels of the ebb-tidal delta will show a similar sedimentary development as the inlet. Seismic profiles at the western side of the ebb-tidal delta (Sha, 1992) show lateral accretion surfaces dipping to the NNE. This is the result of the clockwise migration of these channels. Repeated shell layers in cores (Sha, 1992) suggest a systematic decrease in channel size, comparable to that observed in the inlets (see previous point).

[14] Using the maximal depth of the inlet to establish the tidal prism was criticized by Eysink & Biegel (1992). Nevertheless, although a considerable data scatter is present, which may be attributed to wind, local flow conditions, subsurface conditions, and grainsize, the depth increases with increasing tidal prism (cf. Sha, 1990; Fig. 15, chapter 1).

Occasionally isolated channel fills, showing both channel walls, can be observed on the seismic profiles. They indicate a sudden stop of lateral migration and an abandonment of the inlets or outer channels, in which the local decrease in current velocity resulted in a strong, more or less vertical sedimentation. The infills consist mostly of sand, because of the relatively high wave and current energies along the ebb-tidal delta and because flood- and wave-driven shoals tend to move over the abandoned channels.

5) In the drainage basin the channels may become partially abandoned when the inlet/main channel orientation changes. Cores, 2 to 3 km S of the inlet, show channel fills consisting of thick layers of clays and clay-rich sands (thicknesses of up to 8 m!; cf. Kievits, 1994). This indicates a locally rapid reduction of current velocity and shows that reorientation of the main backbarrier channel can be an abrupt process.

6) The discontinuous eastward shift of the watershed of the Pinkegat[15] has resulted in a shallowing of the channels on a fixed location and an increase in the height of the shoals. Thus, in time, a sedimentary sequence is formed with several channel lags as the channels become shallower[16]. The deepest channel deposits are formed close to the inlet and are commonly well interconnected, due to the continuous cyclic movement of the main channels and inlets, although clay-plugs, resulting from abandonment may occur (see above). The channel facies formed on the spot by the fill of the shallower, lower order channels, upon migration of the watershed, is less continuous and less homogenous than those of the inlets, because the extent over which the shallower 'end-channels' can migrate depends on their drainage area (see above). Abandonment deposits, i.e., infill of the channel with clay or sand, but often with an alternation of both, with a thickness of up to 2-3 m (cf. Kievits, 1994), are frequently formed. Laterally, these deposits may alternate with sub- and intertidal shoal deposits or (especially near the mainland) older Holocene/Pleistocene deposits (Kievits, 1994). Higher up the sedimentary sequence becomes increasingly finer grained, because both

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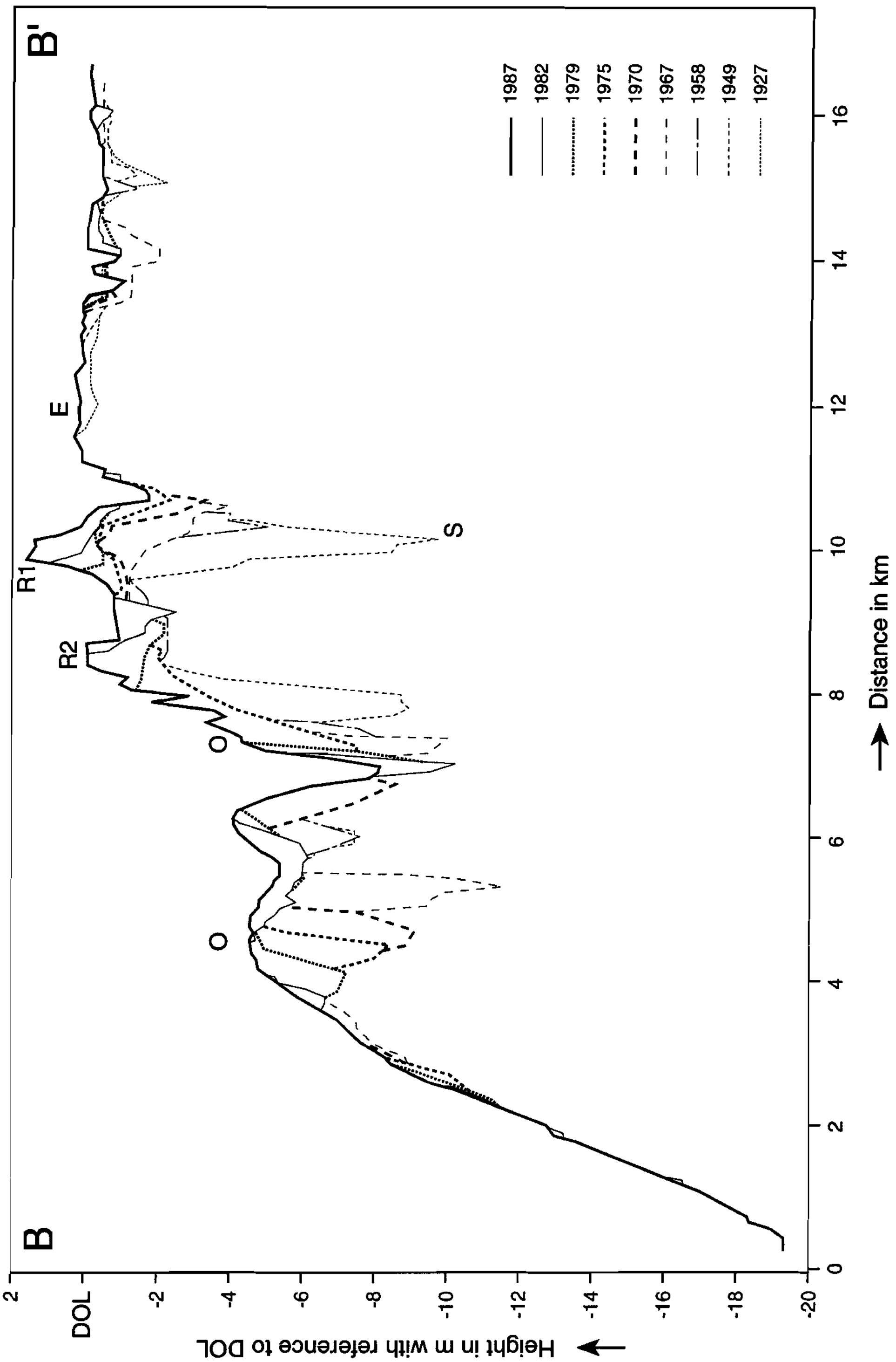
[15] This shift has not always been to the E, as historical reconstructions show (chapter 2). Although in general the eastern ends of the barrier islands tend to migrate eastward, the early Holocene deposits show that the position of many of the islands is rather stable (cf. Van Staalduinen, 1977; cf. de Jong, 1984; Sha, 1992). One of the main factors is probably the location of depressions, which are or are not filled in with easily erodible materials such as peats. These depressions, which are largely defined by the Pleistocene subsurface topography (Griede & Roeleveld, 1982; Bosch & Vos, 1992), largely determine the position of drainage basins, and hence the position of the inlets (Sha, 1992). Since the length of an island is determined by the tidal amplitude, a net eastward shift of the eastern end has to be compensated by erosion of the western end of the island, as happened with Ameland after about 1800. Another possibility is that a shift occurs from a double-inlet estuarine configuration to a single-inlet lagoonal configuration.

[16] Here channel deposits are considered to be confined to the somewhat deeper channels (more then 2.5 m below MLW), because above this level channels merge into subtidal shoals and can become quite broad.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Distance in km
0
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6
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12
14
16

the channel facies and the shoal deposits are formed closer to the relatively muddy tidal watershed. When the watershed migrates over such spot and becomes positioned E of it (compare 1832, Fig. B1 with 1987, Fig. C9), the height of the tidal flat decreases and the higher lying deposits are eroded. Thus, the resulting deposits mainly consist of channel facies. In the resulting sedimentary sequence the higher deposits are rather inhomogeneous and not very continuous and they show frequent channel lags and frequent clay(-rich) infills (cf. Kievits, 1994).

7) In general, a greater lateral extent of the inlet deposits also leads to a greater lateral extent of a more or less homogenous sand body consisting of the lower part of the channel facies, formed near the inlet, because the shallower backbarrier channels (which have a less homogeneous infill) cannot erode the deeper inlet deposits.

Engelsmanplaat/Smeriggat

Analysis of the historical data since 1786 shows that the development of Engelsmanplaat is also cyclic, closely linked to, but different from, the cyclic behaviour of the Pinkegat. At the beginning of such a cycle the Engelsmanplaat is extending far seaward. It becomes higher by deposition of sand (e.g., 1854-1873/74). Then channel erosion occurs at the northern side (e.g., around 1786; De Haan et al., 1983; before 1891, around 1941/45). At least a part of these channels (e.g., 1934-1945) are inlets/outer channels of the Pinkegat system, which shift downdrift and become W-E-oriented and thus form a connection between the Zoutkamperlaag and the Pinkegat Inlet. They develop during a multiple inlet stage. A shoal is formed at the North Sea side (e.g., 1806, 1891-1921, after 1941/45) mainly by wave action (Oost & De Haas, 1992), in combination with flood currents (Steijn et al., 1992). The shoal, N of the channel, becomes usually higher as it catches the sand coming in from the North Sea. The sand is no longer available for the Engelsmanplaat, especially when the northern shoal grows to supratidal heights. When this happens[17] erosion by waves results in a gradual lowering of

Figure 10 (opposite page): N-S cross-section B-B' (for location see Fig. 2) showing those parts of the profiles for the period 1927-1987, which have not been 'eroded' by later profiles. Clearly visible, from S to N are 1) the limited height of Engelsmanplaat (E), which was eroded after 1970, 2) the southward shift and gradual infill of the Smeriggat channel (S), 3) the growth of the northern shoal (R1) and its southward shift, 4) the formation of a second shoal N of it (R2), and 5) the mainly northward lateral migration of the outer channels of the Zoutkamperlaag Inlet system (O).

[17] In 1891, when the channel, separating the shoal and the Engelsmanplaat, was rather shallow, the northern shoal only reached intertidal heights before it merged with the Engelsmanplaat.

the Engelsmanplaat from supratidal to intertidal. The channel separating the two shoals is gradually abandoned and filled in by the settling of muds and by the deposition of sands, partly derived from the southward moving shoal north of it. The two shoals merge (e.g., 1806-1832; 1891-1927) and consequently the Engelsmanplaat becomes larger again at its northern side. In an earlier study, based on changes of names of channels and shoals N of Engelsmanplaat, De Haan et al. (1983) concluded that over the last five centuries the cycle has taken 90-100 y. If, however, the developments of 1786-1891 and 1891-1941/45 are considered to represent cycles, as well as 1873-1927, 1927-±2000, it would take a period of 50-100 y for a complete cycle to occur in the last two centuries (Oost & De Haas, 1992).

The developments since 1921 show a part of the cycle in great detail (Figs B8 to B10 and C1 to C10): in the period 1921-1927 the Engelsmanplaat expanded seaward. In the period 1937-1941/45 the easternmost inlet of the Pinkegat shifted eastward, causing strong erosion at the western side of the Engelsmanplaat. This ebb-dominated channel (Smeriggat) contacted the SW-channel in the ebb-tidal delta of the Zoutkamperlaag Inlet system in 1941/45 (Fig. 10). This resulted in the formation of a ridge at that point at the eastern side of the Smeriggat in the period 1949-1958 (Figs C2 and C3). Due to the shoal N of the Smeriggat, which built up and migrated to the south, the channel became gradually abandoned and shifted somewhat to the south (Fig. 10). The northern shoal became higher, because it caught the sand brought in by waves and flood currents from the North Sea; after 1967 it had become, for a large part, intertidal and after 1970 even supratidal (Figs C4 to C10). In particular after 1970 this affected the Engelsmanplaat, which became lower (Figs C5 to C10 and 10). The shoal did no longer receive substantial amounts of sediment, because waves broke on the inter- and supratidal shoals N of Engelsmanplaat and deposited the larger part of their sediment there. From 1970 until 1993 the Engelsmanplaat has become almost totally intertidal. The strong effect of the formation of a northern shoal is also illustrated by the decrease in height of the northern shoal after 1982, when a second shoal formed N of it during 1979-1982 (Figs C7, C8, and 10). The abandonment and infill of the Smeriggat was in 1991 almost total (Figs C10 and 10). As stated, next to this cyclic development, the western side of the Engelsmanplaat was eroded by the channels of the Pinkegat system in the last 160 years over 5 km.

Future of the Engelsmanplaat

It is to be expected that in the near future the shoal(s) N of the Smeriggat will merge with the Engelsmanplaat. Thus the Engelsmanplaat will receive sand brought in by waves and flood currents from the North Sea, and it will become larger and probably higher again. With this, a new cycle will start. New erosion of the Engelsmanplaat by the Pinkegat-system during the cycle (Fig. 3) may eventually lead to the total disappearance of the Engelsmanplaat and to a merger of the Pinkegat and Zoutkamperlaag Inlet systems (compare Figs B1 to B10 and C1 to C10). The developments may then be similar to those which occurred during the merger of the inlet systems between the East Frisian barrier islands Juist and

Norderney, after the disappearance of the island Buise in between (in the period 1650-1860). There, the change led to a downdrift expansion of the eastern end of the updrift island Juist (Luck, 1975). If such an expansion of Ameland would occur, a considerable shift of sediment (of the order of 10^7 m^3) would be needed (Oost & De Haas, 1993).

Geological relevance

The southward shift of the shoal N of Smeriggat leads to the abandonment and infill of the channel (Fig. 10). These are the only deposits formed in the course of the cycle, which have some preservation potential. The decrease in current velocity in combination with its sheltered position leads to the deposition of sands, clays, and clay-rich sands or an alternation of both (cf. Tye, 1984). The finer sediments are preferentially deposited in summer, whereas the sands are mainly deposited in winter (cf. Oost, 1995). The sedimentary structures in the sands indicate that they have been deposited at high current velocities (e.g., during storms over the N shoal) to low current velocities (slow tidal flow). In general, the facies shows strong bioturbation. Such deposits can easily be mistaken for tidal flat or lagoonal facies (cf. Van den Heuvel, 1993). Their great thickness (here approximately 5 m) and their lateral dimensions (channel form) may prevent such an erroneous interpretation, if sufficient lateral control is available. Thick abandoned channel/inlet facies can be considered indicative for migrating inlets (cf. Tye, 1984; Sha, 1989a). This is due to the fact that normally the strongest sedimentary, and related hydraulic changes within the inlet system occur at the inlet throat. There, two of the important formative forces, tidal currents and waves have a maximal impact. This leads to strong shifts of sediment and to strong, sometimes swift changes in the hydraulic patterns enabling sudden abandonment.

CONCLUSIONS

The Pinkegat System shows a clear cyclic development, from a single inlet to a multiple one and back, covering a period of 20 years to at maximum 54 (but probably 41) years, in the period since 1832. When the position of a single inlet becomes hydraulically less efficient, because it becomes positioned too far downdrift, new inlets evolve from spill-overs, which tend to breach through the sandy shoal that forms the eastern end of the updrift island (Ameland). These new, small western inlets migrate rapidly downdrift (to the E) and increase in dimensions. At the same time the eastern inlets decrease in dimensions and migrate more slowly. By merging of inlets and abandonment of others, a single inlet with maximum dimensions remains. The process is driven by the continuous downdrift supply of sediment, the tidal forces, and by the local wave climate (which builds up shoals). The strong curvature of the inlet, partly due to the fact that it drains an updrift drainage basin and the erosion in the outer bend further enhance strong lateral migration of the inlets.

The developments in the drainage basin of the Pinkegat are largely dictated by the developments in the ebb-tidal delta. The main channels form the connection between the continuously shifting channels in the ebb-tidal delta and the relatively stable lower order channels along the borders of the drainage basin. The main channels in the drainage basin shift from a NNE-SSW orientation to an ENE-WSW orientation, the more the eastern end of the barrier island Ameland becomes positioned to the east. In general, the shoal at the eastern end evolves when the western inlets shift downdrift and the system develops into a single inlet, which in its turn also shifts to the east. The backbarrier channels erode the shoal at its S side and enhance the formation of new channels through it. When a more western pathway develops during the formation of a multiple inlet configuration, the main channel once again becomes oriented NNE-SSW. The watershed south of Ameland shifts to the E when the transport route along the eastern end of the island to the watershed behind it becomes longer and slower, due to changes in the inlet configuration in the ebb-tidal delta.

The strong cyclical development of the Pinkegat Inlet is illustrated by the strong changes in energy conditions, in particular the tidal current energy, and the frequent internal erosion features which can be observed in the sedimentary succession:

1) the cyclical migration of the inlets, in combination with the net downdrift shift of Ameland, results in characteristic downdrift-oriented lateral accretion deposits, which show at least a double channel lag, because updrift channels are in general shallower,

2) the outer channel facies in the seaward part of the ebb-tidal delta also shows lateral accretion dipping downdrift and, at the updrift side of the ebb-tidal delta, dipping seaward. In addition sandy abandoned channel facies are present,

3) the backbarrier facies near the inlet is characterized by extensive lateral migration in clock- and anticlockwise directions and abrupt channel abandonment facies with clay-rich fills of several metres thick, and

4) the facies formed E of the net eastward migrating watershed will be characterized by a shallowing up and an increase in clay content in combination with a laterally inhomogeneous character.

The Engelsmanplaat also experiences cyclic changes (54-86 yrs). The originally large and high northern side of the shoal is eroded by a channel. In the 20th century the channel was an eastward shifting inlet channel of the Pinkegat Inlet. It became connected to the Zoutkamperlaag Inlet system. Wave action and flood currents produce a secondary shoal at the North Sea side, separated from the Engelsmanplaat by the channel. The shoal becomes higher, because it catches sand coming in from the North Sea. The sand thus is not available for the Engelsmanplaat. Therefore, when the secondary shoal reaches supratidal heights, the

Engelsmanplaat gradually decreases in height from supratidal to intertidal. As soon as the channel is surrounded by shoals it is gradually abandoned and filled in by fine-grained sediments and by sands, which are partly derived from the southward moving secondary shoal north of it. The two shoals merge and the cycle starts again. The cyclic development thus produces channel abandonment deposits.

The net eastward shift of the Pinkegat system (which has at least partially been attributed to the eastward shift of Ameland Inlet, see chapter 2) has resulted in an erosion of the Engelsmanplaat by channels of the Pinkegat in the last two centuries. If erosion continues the Engelsmanplaat will disappear and Pinkegat Inlet will likely merge with the Zoutkamperlaag Inlet.

ACKNOWLEDGEMENTS & REFERENCES

See next chapter.

CHAPTER 4

THE CYCLIC DEVELOPMENT OF THE ZOUTKAMPERLAAG INLET SYSTEM, DUTCH WADDEN SEA, IN THE PERIOD 1832-1991, AND A COMPARISON WITH THE PINKEGAT INLET SYSTEM

ABSTRACT

The development of the Zoutkamperlaag Inlet system has been studied for the period 1806-1991. The outer channels of the ebb-tidal delta of the Zoutkamperlaag migrate roughly downdrift. At the updrift side of the ebb-tidal delta they migrate mainly to the N to NE. At the N to NE-side they migrate to the E to SE. New outer channels are generated at the SW-side and the NE-side of the ebb-tidal delta. The latter, flood-defined channels particularly develop when the tidal flow is concentrated along the W-side of the ebb-tidal delta. Eventually, all outer channels which arrive at the downdrift, NE-side of the ebb-tidal delta are abandoned. The main inlet is positioned at the W-side of the ebb-tidal delta. During the 19th century a double inlet system developed. The most eastern inlet migrated downdrift and was abandoned. The downdrift migration of the outer channels and inlet results in extensive lateral accretion deposits.

Before the closure of the Lauwerszee embayment the geomorphological developments in the drainage basin of the Zoutkamperlaag Inlet were also cyclic. The main inlet developed from a single channel to a double channel and back. To a large extent the development was controlled by the position of the outer channels and of the inlet in the ebb-tidal delta. Upon a concentration of the tidal current along the eastern side of the main channel the channel bifurcated into two channels. Between these channels an intertidal shoal formed and the western channel was finally abandoned. The apparent stop of the cyclic development after the closure of the Lauwerszee embayment suggests that this embayment has been crucial for the maintenance of the cyclic development.

In the Zoutkamperlaag Inlet system seaward to downdrift dipping, lateral accretion surfaces are mainly formed within the ebb-tidal delta, in particular by lateral migration of the outer channels. In the drainage basin lateral accretion and channel abandonment is strongly restricted to a narrow zone along the channel. Upon gradual infill of the drainage basin the main channel would fill up to form a channel-fill body with mainly sand and be flanked by clay-rich sediments. In the Lauwerszee embayment the channel-fill body would also be flanked ànd capped by clay-rich sediments. Such channel-fills are indicative of a strong influence of the subsurface relief, keeping the position of the inlet and of the main channels within narrow limits.

In general, observations of any part of a barrier system may provide valuable clues to all other parts of the system, because many parts of barrier-related deposits are in dynamical

equilibrium with the tidal prism. Both in the Zoutkamperlaag and Pinkegat Inlet system the developments in the drainage basin are strongly determined by the developments in the ebb-tidal delta and, to a minor extent, vice versa, because they are mutually dependent. This, and the dependence of all parts of the system on the tidal prism allows the reconstruction of large parts of the tidal system, based on observations of other parts. To understand and reconstruct the sedimentary conditions the better information is to be obtained from the inlet and the ebb-tidal delta deposits. In vertical sections (cores) and often also in coast-parallel sections the deposits primarily reflect the prevailing, external conditions during a short period, because in a coast-parallel direction, laterally extensive deposits are deposited almost instantaneously, compared to the geologic timescale.

INTRODUCTION

Barrier and barrier-related facies are formed at the border of land and sea. They provide valuable information about basin dynamics and paleo-hydrodynamics. Important amounts of fossil hydrocarbons can be stored in barrier-related sandstone bodies, such as channel fills and sandy shoals. For an optimum prediction of the 3-dimensional extent of sandstone bodies, reliable reconstructions of the depositional conditions are crucial. Detailed reconstructions, however, are often difficult, since the information is commonly available from a few cores or sections only. Interpretations are further hampered, because geologists are inclined to consider the time needed for the formation of sedimentary deposits in many thousands of years, whereas recent barrier-related deposits are formed in time spans of centuries or even decades. Therefore, the study of the sedimentary development and morphodynamics of recent barrier deposits is a major key for understanding their fossil equivalents.

The developments of the Zoutkamperlaag Inlet system and the Pinkegat Inlet system (Fig. 1) during the past two centuries show in detail the strong morphodynamics of two inlet systems in a mixed energy shoreline setting and the interaction of the various parts of the systems. This study provides insights into the formation of barrier and barrier-related deposits and gives clues where to look in the fossil reach to reconstruct barrier-related deposits. This paper is a further elaboration of two studies for the Project COASTAL GENESIS (Oost & De Haas, 1992, 1993) coordinated by the Directorate General for Public Works and Water Management (further referred to as RWS = Rijkswaterstaat), National Institute for Coastal and Marine Management/RIKZ (further referred to as RIKZ). The sedimentological/ morphodynamical changes in the backbarrier and ebb-tidal delta of the Zoutkamperlaag over the period 1806-1991 are described, with special attention for the period 1927-1991 (for the effects of the closure of the Lauwerszee embayment the reader is referred to Chapters 1 and 4 (Oost & De Boer, 1994; Oost, 1995).

METHODS

For a discussion of material and methods the reader is referred to the previous chapter.

OBSERVATIONS

In 1969 the Lauwerszee embayment (Fig. 1) was dyked. Before, it formed part of the drainage area of the Zoutkamperlaag. The semi-enclosed Lauwerszee had its own sedimentary character, and is therefore discussed separately. The closure of the Lauwerszee embayment resulted in a decrease in tidal prism with 1/3, from $305*10^6$ m^3 to $200*10^6$ m^3 (Van Sijp, 1989). Due to the reduction, the system was out of equilibrium. During the transition to a new equilibrium, strong changes in the morphological, hydraulic, and sedimentary characteristics of the drainage basin, the ebb-tidal delta, and the inlet occurred. These changes are dealt with extensively in Chapter 5. The changes in the Zoutkamperlaag Inlet after 1969 will only be mentioned as far as needed for the present discussion.

Zoutkamperlaag, ebb-tidal delta

1806-1873/74

In 1832 (Fig. B1[1]) the Zoutkamperlaag Inlet consisted of one large channel, oriented to the North. Towards the North Sea it bifurcated into two outer channels. Of these two the westernmost one was most important. The other was oriented to the NNE. To the SW a flood-defined channel (Middelgat, marked M in Fig. B1) was present.

In 1850 (Fig. B2) a flood-defined outer channel was present at the SW side (Oude Middelgat, marked OM). The inlet itself had begun to bifurcate into two more or less separate channels, directed to the N. The parts which formed the connection to open sea were directed WNW-ESE (the southernmost one called Middelgat, marked M). The western channel was the most important one in 1850, and still was in 1854 (Fig. B3). By 1854 the part which formed the connection to the open sea had shifted to the N (Northwest Gat, marked NW). The eastern channel was obstructed by several shoals, and the connection to the open sea became oriented to the NNE (Northeast Gat, marked NO). West of the eastern channel a shoal built up, which separated the channel from the more western channel. Moreover, the eastern channel had merged with the SW-NE-defined channel along Schiermonnikoog (Oostgat, marked O). The gradual separation of the two channels in the inlet continued and by 1859 (Fig. B4) they were partly separated by an elongated, mainly subtidal shoal. The western channel had three outer channels: two flood-defined (NW-SE) ones (unnamed and Oude Middelgat, marked OM) of which the most northern was a remnant of the old connection to

[1] Figures indicated with letters are to be found in appendix A of this thesis.

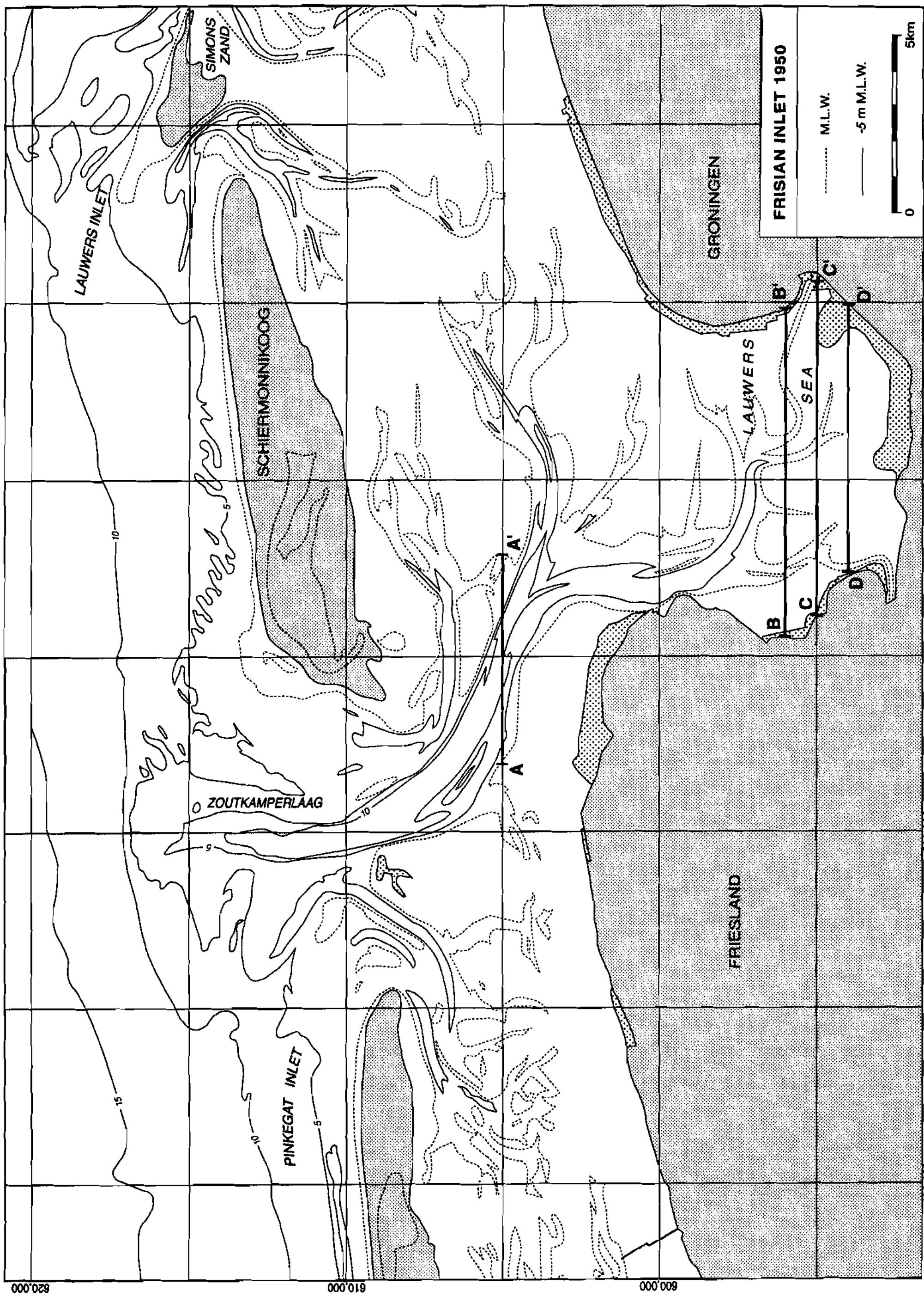


Figure 1: Overview of the area and location of the profiles (Figs 2 and 3).

the open sea of 1854 (Fig. B3, marked NW), and a northward-oriented ebb channel (Rifgat, marked R). More to the east the outer channel Northeast Gat (marked NO) was situated, which had shifted eastward. The channel Oostgat was now largely abandoned (Oude Oostgat, marked OO). In the period 1850-1859 the outer channels migrated northward at the W-side of the ebb-tidal delta and downdrift at the N-side.

1873/74-1921

By 1873/74 (Fig. B5) the two channels were largely separated by a wedge-shaped sandy shoal at the seaward side and another, elongate shoal with flood chutes more landward. The westernmost inlet channel was still the most important one, and it had a good connection with the flood-defined westernmost outer channel (called Plaatgat; marked P), which actually consisted of two closely connected outer channels. The ridge between the westernmost inlet channel and the now SE-NW-oriented, ebb-defined outer channel led to a somewhat poor lateral connection in that direction. Because of this obstruction, the eastern inlet channel had to transport more water and became deeper. The northern part of the eastern inlet channel became increasingly more oriented to the NE. By 1891 (Fig. B6) the sandy shoal, which had separated the two inlet channels shifted towards the SW.

Within the drainage basin two separate channels were present, which merged into the western inlet channel. After 1873/74 the western inlet channel became longer in a northward direction and became slightly NNE-oriented. Also, the outer flood-defined channel (Plaatgat, marked P) and the NW-SE-oriented ebb channel (Noordgat, marked N) had both shifted to the north. The eastern inlet channel had become almost completely separate. It was less significant than in 1873/74, and had largely lost its connection to the main backbarrier channel. Further to the E another shallow channel had developed, with N of it an intertidal shoal. Together with the eastern inlet, the channel drained the backbarrier channel directly S of Schiermonnikoog.

By 1903 (Fig. B7) the most eastern part of the main backbarrier channel had become somewhat more important than the western part. Seawards it merged with the western backbarrier channel to form the western inlet channel. It had several outer channels. The most western one was new. It was abandoned again in the period 1903-1921. The flood-defined outer channel (Plaatgat, marked P), N of it, had rotated somewhat clockwise. The third channel was oriented to the NNE (it was called Noordwestgat (marked NW) indicative of an originally more NW-orientation) and merged with the eastern inlet[2]. This eastern channel had again established a connection with the main backbarrier channel. It provided also the

[2] On the original map traces at the eastern side of this channel are visible, which might be the remnants of the NNE-SSW oriented outer channel(s) of the eastern inlet channel of 1891. This indicates a downdrift (to the E) shift of these channels. Subsequent merging of shoals after complete abandonment seems to be indicated by the sedimentation at the NW side of Schiermonnikoog in 1910-1920, observed by Isbary (1936).

drainage of the backbarrier channel S of Schiermonnikoog, because the more eastern channel of 1891 had been abandoned. Shoals west of it had merged with Schiermonnikoog around 1896 (Isbary, 1936).

1921-1939

By 1921 (Fig. B8) the main channel of the drainage basin was drained by the western inlet channel, which had extended seaward, and had two outer channels: one to the W (Plaatgat, marked P) which had slightly shifted to the N, and a less well developed one to the N. By then the eastern inlet (with its outer channel 'Noordwestgat', marked NW) only served to drain the backbarrier channel S of Schiermonnikoog, and was separated from the main western inlet by a series of shoals.

By 1927 (Fig. B9) the Zoutkamperlaag ebb-tidal delta had a N-S-oriented inlet, which bifurcated seaward into a small flood-defined channel with a W-E-orientation (at the location of the Plaatgat in 1921) and an important ebb-defined outer channel with a small branch to the NNE (together named Plaatgat, marked P). The eastern inlet was largely abandoned, and probably merged with the channel E of it by clockwise rotation (until 1937; workmaps RWS). In 1927 the small remaining channel flowed along a recurved bar at the southwestern end of Schiermonnikoog and partly drained the channel S of Schiermonnikoog. In the period 1927-1970 this remnant was gradually abandoned.

The many relocations of buoys in the 'Plaatgat' channel in the period 1927-1934 (Figs B9 and B10) show that the channel rotated clockwise from a N-S orientation to an ENE-WSW orientation (by 1934; workmaps RWS). Then the outer channel became too shallow for sailing. In 1935 (Fig. B10) buoys were placed in a *new* NW-SE-oriented Plaatgat, which had evolved from the flood-defined channel S of it, which had migrated (at least partly[3]) to the N (Postema, 1956; workmaps RWS).

By 1934 the Zoutkamperlaag Inlet had five separate outer channels. The most eastern, ENE-WSW-oriented, outer channel in 1934 (Fig. B10) was the *old* Plaatgat of 1927 (marked OP). In the period 1934-1939 the *old* Plaatgat was gradually abandoned, and sandy shoals formed at that location (workmaps RWS). In the same period the new Plaatgat (marked NP) rotated clockwise, but its connection to the open sea remained largely in place (workmaps RWS).

1939-1970

In the period 1939-1950 (Fig. C2) the new Plaatgat outer channel continued to rotate, and became directed to the N; the shallow connection to the open sea shifted northward and became oriented WNW. In the period 1934-1950 the flood-defined NW-SE trending outer

[3] This flood-dominated channel may have been separated in two parts: one part migrating N and one part forming the new flood-dominated channel SW of the ebb-tidal delta. Alternatively, a new flood-dominated channel may have been formed, like in 1903 and 1949.

channel (Noordwestgat) became more important, in particular in the period 1927-1934. Between 1939-1949 (Fig. C2) another outer channel formed SW of it. A large flood-defined funnel-shaped outer channel formed at the eastern side of the inlet. The recurved bar at the SW end of Schiermonnikoog became lower and extended to the SW. The bar on the NW end of the island became also recurved. The area N of it, between -2.5 m and -5 m DOL, was strongly eroded.

By 1958 (Fig. C3) the new outer channel which had formed S of the Noordwestgat had become deeper and took over the function of the Noordwestgat of 1950. It became flood-defined. The Plaatgat had now two ebb-defined branches: the original Plaatgat (NNE-oriented) and the old Noordwestgat (NW-oriented) of 1950. In the period 1950-1958 the throat of the inlet became deeper. The funnel-shaped, flood-defined, SSW-NNE-oriented outer channel of 1949 had become broader and had shifted slightly and rotated clockwise. North of it a shoal had formed above -5 m and locally above -2.5 m DOL. The shoal at the NW coast of Schiermonnikoog disappeared in the period 1949-1958, while the N-S-oriented channel SW of Schiermonnikoog had become quite shallow.

In the period 1958-1965 (Figs C3 and C4) the flood-defined channel at the SW-side of the ebb-tidal delta and the inlet throat became deeper. The NW-SE-oriented outer channel (the former Noordwestgat) of the Plaatgat became deeper, while the extension to the NNE became shallower. The ebb-tidal delta as a whole extended further to the NW (-5 m line in Figs C3 and C4). The funnel-shaped flood-defined, SSW-NNE-oriented outer channel between the inlet and Schiermonnikoog was squeezed between the shoal N of it, which migrated to the SE, and the coast, and became gradually abandoned. The N-S-oriented channel SW of Schiermonnikoog had become even shallower in the period 1958-1965.

In the period 1965-1970 (Figs C4 and C5) the westernmost outer channel became slightly shallower and migrated northwards (near the inlet). The outer NW channel of Plaatgat rotated clockwise. The NNE-outer branch of Plaatgat became abandoned and the shoals above -5 m DOL at either side merged. The funnel-shaped flood-defined SSW-NNE-oriented outer channel between the inlet and Schiermonnikoog was almost completely abandoned.

1970-1991

In the period 1970-1991 (Figs C5 to C10) the -10 meter contour of the ebb-tidal delta retreated somewhat landward, while the -15 metre contour kept its position. Also, over the period 1970-1991 the main inlet became shallower. The throat of the inlet migrated over 1 km to the east, with a maximum velocity of 72 $m.yr^{-1}$ (cf. Oost & De Haas, 1992, 1993; Huijs, 1993). After an initial counterclockwise movement, the inlet rotated clockwise from 1976 onwards (Huijs, 1993).

In 1970 (Fig. C5) the westernmost outer channel (Westgat) had a WNW-ESE orientation, and the Plaatgat had a NW-SE orientation. They were separated by a subtidal shoal (-2.5 to -5 m DOL) of 2 km^2. The Westgat rotated clockwise during the period 1970-1982 (Figs C5 to C8), especially in 1970-1975. In 1982-1987 (Figs C8 and C9) the Westgat got a more

E-W orientation, but clockwise rotation continued in the period 1987-1991 (Figs C9 and C10). In 1987-1991 the western flood-defined outer channel became shallower and wider.

Table I: Sedimentation and erosion in the area over various periods, calculations based on the gridded topography (Fig. 2 in Chapter 3).

Net sedimentation on the Zoutkamperlaag ebb-tidal delta (1927-1970: 73,450,800 m^2; 1970-1987: 94,324,504 m^2); corrected for the extraction of sand by dredging)

Period	Total (10^6 m^3)	Annual (10^6 $m^3.yr^{-1}$)	Annual ($cm.yr^{-1}$)
1927-1950	+15.1	+0.66	+0.89
1950-1958	-13.7	-1.71	-2.33
1958-1965	-11.6	-1.66	-2.26
1965-1970	-5.9	-1.18	-1.61
1970-1975	-9.8	-1.96	-2.08
1975-1979	+0.4	+0.10	+0.11
1979-1982	-10.0	-3.33	-3.53
1982-1987	-6.6	-1.32	-1.40

The northward directed ebb-defined outer channel (Plaatgat) shifted to the E during the period 1970-1991 (Figs C5 to C10), except for the period 1974-1976, when the axis shifted to the W (Huijs, 1993), and 1982-1987, when the outer channel was partly abandoned (Oost & De Haas, 1992). In the period 1970-1975 (Figs C5 and C6) the Plaatgat branched into two outer channels, close to each other and became oriented to the NNW. Thereafter the easternmost channel was gradually abandoned, while the western one expanded seaward. As can be concluded from the remainders of the eastern outer-channel of the Plaatgat in 1987, this eastern part continued to rotate clockwise in the period 1982-1987 (Figs C8 and C9), while it became largely abandoned at the same time. Over the period 1982-1991 the western part shifted slightly downdrift and became curved at the seaward end.
In the period 1970-1987 a bar developed W of Schiermonnikoog by the action of waves, and became supratidal (Oost & De Haas, 1992). As discussed in Chapter 5, the extremely large size has been attributed to the enclosure of the Lauwerszee embayment and the consequent release of sand from the shrinking ebb-tidal delta. After 1987 the western arm of the recurved bar merged with Schiermonnikoog, closing off the embayment which it surrounded. In the same period the bar decreased in height, and a channel of approximately 5 m deep cut through its western arm (Oost, 1995).

Quantitative changes based on digitized data (Table I)

The large ebb-tidal delta of the Zoutkamperlaag Inlet experienced dominantly erosion after 1927, except for the periods 1927-1950 and 1975-1979. Before the closure of the Lauwerszee embayment, the loci of net sedimentation and erosion shifted continuously.

As a result of the closure of the Lauwerszee in 1969, the higher parts (above -4 to -5 m DOL) were characterized by net sedimentation due to formation of wave-built shoals, and, in the deepest part of the ebb-tidal delta (mainly below -12 m) due to the infill of the inlet. Due to erosion at other sites, the overall net effect was the erosion of $26*10^6$ m^3 of sand in the period 1970-1987.

Zoutkamperlaag, drainage basin

During the years 1806 to 1987, the main backbarrier channel(s) followed roughly the same course. The channel ran from the inlet along the eastern side of the Engelsmanplaat with a slight curve towards the entrance of the Lauwerszee embayment. From that point there was always a 2nd order branch into the Lauwerszee and one towards the E. Furthermore, from 1850 (Figs B2 to B10 and C1 to C10) onwards a channel system was present S of Schiermonnikoog.

1806-1873/74

In 1806 there were two main channels, separated by a shoal (Fig. A8). In 1832 (Fig. B1) there was only one main channel. By 1850 (Fig. B2) the channel had bifurcated seaward into two channels separated by a shoal. In 1854 (Fig. B3) no observations were made. By 1859 (Fig. B4) the main backbarrier channel had become more and more separated into two channels (SE of Engelsmanplaat). By 1873/74 (Fig. B5) the main backbarrier channel consisted of two channel beds, separated by a subtidal ridge. At the western side, a flood chute had formed, which transported the flood water towards the Lauwerszee embayment. At the eastern side another channel was present which transported mainly the ebb water from the Lauwerszee and from the southeastern drainage basin, N of Groningen (Fig. 1).

1873/74-1949

In 1891 the shoal which originally separated the inlets of the Zoutkamperlaag had shifted to the SW, making the separation between the two channel beds even more complete (Fig. B6). By 1903 (Fig. B7) the western channel bed within the main channel had been gradually abandoned. A large elongated shoal formed E of it, and built up gradually. The process continued, and by 1921 (Fig. B8) the most eastern channel had become the main backbarrier channel. Thus, in 1921 the main backbarrier channel consisted of two separate channels, a shallow flood chute ('Noordwestrak') in the SW and the main channel more to the NE. These were separated from each other by a sub- to intertidal shoal. The shoal had formed by the merger of two shoals, Eilanderbalg and a part of Schulprug, in the period 1903-1921

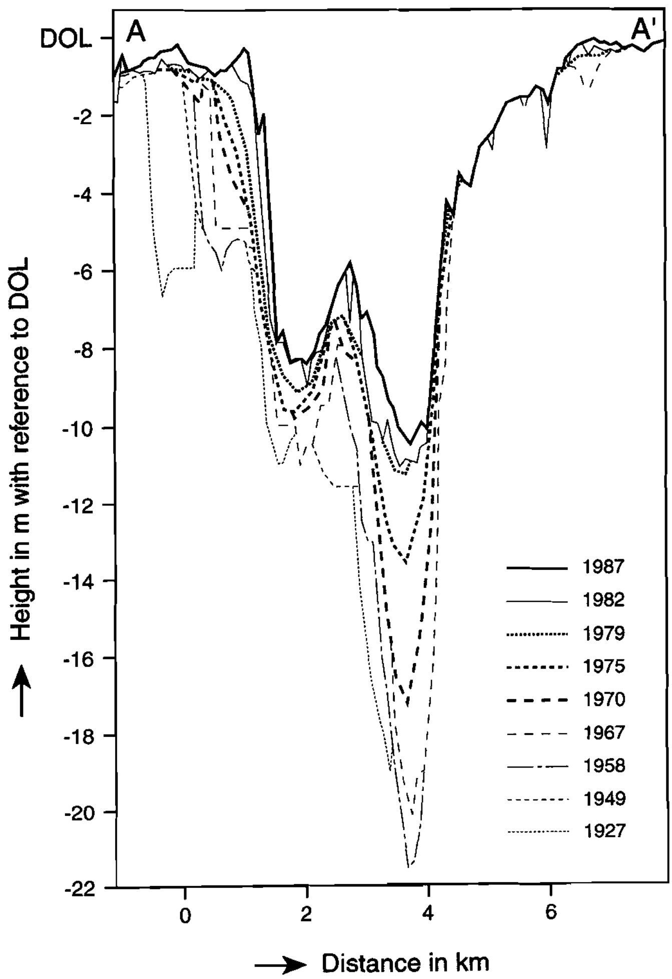


Figure 2: W-E cross-section A-A' through the backbarrier-area (for location see Fig. 1) showing the parts of the profiles which have not been eroded by later 'profiles' over the period 1927-1967. Note the ridge in the main channel, and left the old channel of 1927.

(working maps RWS). Another part of Schulprug formed a large separate shoal. By 1927 (Figs B9, 1 and 2) the main channel already had started to develop a new flood chute and a channel NE of it. The old southwestern flood chute had become unimportant.

In the period 1927-1934 the ridges at the end of the old flood chute had become so high that the channel could transport only little water (Postema, 1956). In this process the eastern main channel became increasingly more important for the transport of water. In the period

1934-1949 (Figs B10 and C2) the western channel was abandoned almost completely, and filled up to intertidal heights; the shoal E of the western channel thus merged with the tidal flats. To the N remnants of the western channel and parts of the new flood chute east of it ('Westrak') became shallower and were gradually abandoned and covered by a shoal (period 1927-1970). By 1949 the flood chute, which had formed in the period 1927-1934 was less well developed. It became more pronounced again in the period 1949-1958 and once more there were two channel beds within the main channel, separated by a ridge.

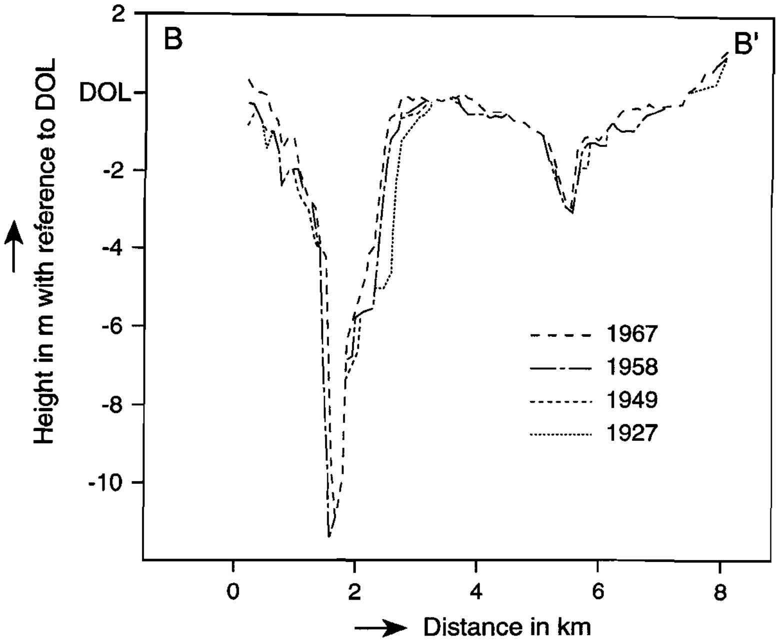
1949-1987

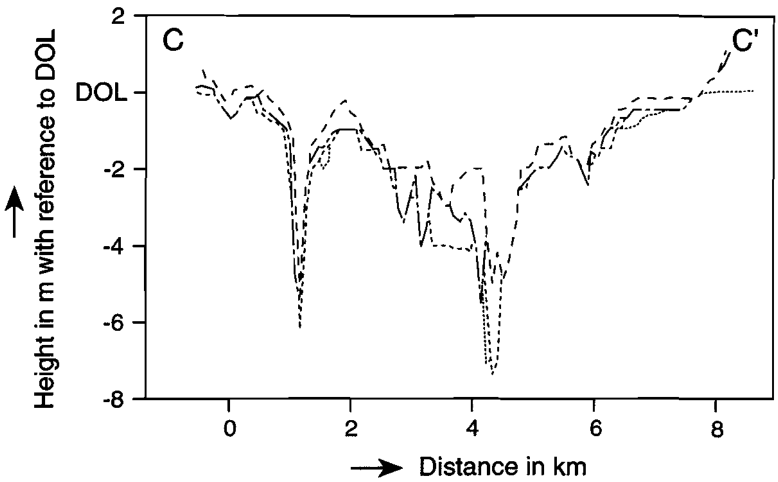
The main channel became deeper and broader in the period 1949-1966 (Figs C2 to C4). The most important change in the period 1966-1970 (Figs C4 and C5, Table I) was strong sedimentation in the axis of the Zoutkamperlaag inlet, in particular in the area N of the closure dam. At the same time the southern channels near the watershed extended over approximately 0.5 km to the E. For the changes in the backbarrier after 1970 the reader is referred to Chapter 5. Notwithstanding the strong sedimentation in the period 1970-1987, separate channel beds persisted until 1987 (Fig. 2). However, the axis which connected the inlet with the main backbarrier channel (N of the closure dyke) had shifted to the side of the western channel bed in the period 1970-1987 (cf. Oost & De Haas, 1992). Sedimentation has been strongest in the eastern chute.

Quantitative changes based on digitized data (Table II)

For the period 1927-1971 the calculations cover the major part of the drainage basin of the Zoutkamperlaag, some 122 km^2 in surface area, not including the Lauwerszee area (Table II, see below). Sedimentation dominated in the successive periods, except 1949-1957/58. The net result over the period 1927-1966, the period before the closure of the Lauwerszee embayment, was erosion between -25 m and -9 m, sedimentation between -9 m and -2.5 m, erosion between -2.5 m and -2 m and sedimentation above -2 m. The net result over the whole period 1927-1966 was a slight sedimentation of $0.4 * 10^6$ m^3. In the period 1966-1970 sedimentation dominated, perhaps already resulting from the closure of Lauwerszee in 1969.

In contrast to the previous periods, sedimentation was strong in the succeeding period 1970-1987, due to the enclosure of the Lauwerszee (Chapter 5). Over the period 1970-1987 it amounted $33.6 * 10^6$ m^3 (over an area of 126 km^2). Sediments were, for an important part, deposited in the channels, in which the locus of sedimentation shifted upwards through time (1970-1975, 1975-1979, 1982-1987) because they were filled up.


B
B'
DOL
Height in m with reference to DOL
Distance in km
1967
1958
1949
1927


C
C'
DOL
Height in m with reference to DOL
Distance in km

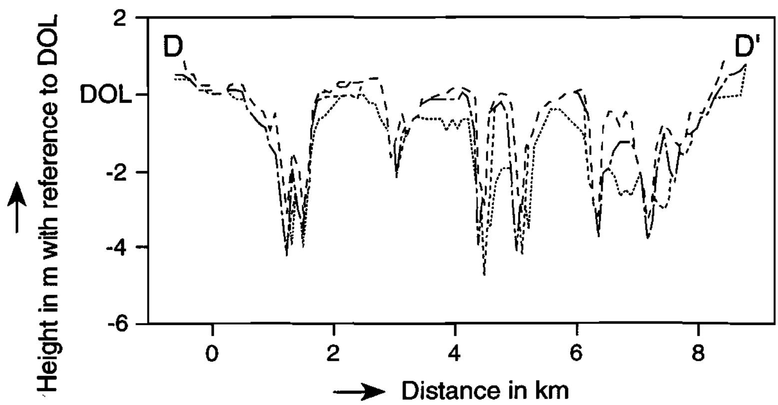

D
D'
DOL
Height in m with reference to DOL
Distance in km

Table II: Sedimentation and erosion in the area over various periods, calculated on the basis of the gridded topography (Fig. 2 in Chapter 3).

Net sedimentation in the Zoutkamperlaag drainage basin (1927-1970: 122,058,904 m^2; 1970-1987: 126,416,704 m^2); corrected for the extraction of sand by dredging.			
Period	Total (10^6 m^3)	Annual (10^6 $m^3.yr^{-1}$)	Annual ($cm.yr^{-1}$)
1927-1949	+12.3	+0.56	+0.46
1949-1957/58	-13.7	-1.61	-1.32
1957/58-1966	+1.7	+0.20	+0.16
1966-1970	+10.0	+2.50	+2,05
1970-1975	+14.8	+2.95	+2.33
1975-1979	+10.7	+2.68	+2.12
1979-1982	+4.3	+1.43	+1.13
1982-1987	+3.8	+0.76	+0.60

The Lauwerszee embayment

By 1832 (Fig. B1) the Lauwerszee area was the last remnant part of a much larger embayment, which had formed in the 8-9th century. Comparison of the situations in 1832, 1850, and 1873/74 shows a stable picture, the area being dissected by one 2nd order channel. At the junction of the channel system with the main backbarrier channel one or several flood chutes were present. In the embayment the 2nd order channel bifurcated into two branches: one to the S and one to the SE (to Zoutkamp). The latter branch cut the mainly intertidal area into two parts, N and S of it. These branches showed a tendency to meander slowly.

In the period 1874-1877 the eastern river mouth, Reitdiep, was closed off (already indicated on the map of 1873/74; Anonymous, 1948). As a result the connecting branch became shallower (Figs B7 and B8). The meander in the western branch was cut off and abandoned in the period 1891-1903. Some meandering occurred, and a 3rd order E-W-oriented channel disappeared after 1903.

Figure 3 (opposite page): W-E cross-sections B-B', C-C' and D-D' through the Lauwerszee area (for location see Fig. 1) showing the parts of the profiles which have not been eroded by later 'profiles' over the period 1927-1967. Note the extremely constant position of the channels in this area.

Subsequently, at the same spot, a flood chute changed into a 3rd order channel with 4th order branches (1903-1967). However, the channel system as a whole remained largely stable. The sediments were finer than in the major part of the drainage basin of the Zoutkamperlaag Inlet system. Analyses show that in 1890 the larger part of the area consisted of clay with less than 60% sand to clay-rich sand (Van Elzelingen, 1904). Observations around the time of closure of the Lauwerszee embayment show that the sediment grainsize had not changed significantly in the period 1890-1969.

In the period 1927-1949 net erosion of shoals occurred, especially in the NW, whereas sedimentation was important in the S and E of the embayment. In the period 1949-1958 erosion prevailed on the shoals, with exception of the southern part. In the period 1959-1967 net sedimentation occurred on the shoal. The position of most of the channels was rather stable in the period 1927-1967 (Fig. 3). In 1969 the Lauwerszee embayment was closed, and it now forms a fresh-water lake.

Quantitative changes based on digitized data (Table III)

In 1967 the Lauwerszee embayment was some 94 km^2 in size. The semi-enclosed character of the embayment resulted in sedimentation patterns which are not fully comparable with those in the rest of the drainage basin (Tables II and III). In particular after 1959 the pattern differed: the Lauwerszee was, much more than the rest of drainage basin of the Zoutkamperlaag, an area with net sedimentation. Except for 1949-1959 net sedimentation prevailed over the period 1927-1967 ($14.9 * 10^6$ m^3). Over the whole period net erosion dominated between -14.5 m to -5 m, i.e., in the channels, mainly near the entrance of the Lauwerszee embayment. Slight sedimentation occurred between -5 m and -2 m, and strong sedimentation on the shoals above -2 m.

Table III: Sedimentation and erosion in the area over various periods, calculations based on the gridded topography (Fig. 2 in Chapter 3).

Net sedimentation in the Lauwerszee embayment; 1927-1967: 93,781,800 m^2 (corrected for the extraction of sand by dredging)			
Period	Total (10^6 m^3)	Annual (10^6 $m^3.yr^{-1}$)	Annual ($cm.yr^{-1}$)
1927-1949	+10.0	+0.45	+0.48
1949-1959	-6.8	-0.68	-0.73
1959-1967	+11.8	+1.48	+1.57

DISCUSSION

Ebb-tidal deltas are largely the result of the ebb current, the flood current, waves, and the coast-parallel current (Oertel 1973; Hayes, 1979; FitzGerald, 1988; Sha, 1990; Steijn et al., 1992; Steijn & Louters, 1992). Together with the location and dimensions of the drainage basin these processes determine to a large extent the development and dynamics of the ebb-tidal and the channels in it.

Zoutkamperlaag, ebb-tidal delta

The most commonly observed behaviour of outer channels in the Dutch Wadden Sea is a migration which is clockwise and roughly downdrift (Joustra, 1971; Sha, 1989a; Oost & De Haas, 1992, 1993; Huijs, 1993). This is also true for the outer channels of the Zoutkamperlaag Inlet. New outer channels are mainly generated at the western side of the ebb-tidal delta. A part is formed as westward directed truncations of existing outer channels at the seaward side (e.g., 1832/34-1850, 1927-1934). Completely new outer channels are formed mainly at the SW-side of the ebb-tidal delta (e.g., 1891-1903, 1927-1934(?), 1934-1949). They may develop fast when after storms huge amounts of ebb water leave the drainage basin, forcing strong sedimentation/erosion. The truncation of the ebb-delta shoals may also be a gradual process, as is illustrated by the beautiful development of a new ebb-defined outer channel, W of the original one, over the period 1970-1991.

At the updrift (W) side of the ebb-tidal delta, outer channels migrate mainly to the N to NE. For the mostly flood-defined channel at the SW-side of the ebb-tidal delta this is mainly accomplished by growth of sandy shoals close to the inlet throat (e.g., 1873-1891, 1927-1934, 1934-1950, 1958-1970, 1982-1987). For the mostly ebb-defined channel, N of the flood-defined channel, the migration is mainly brought about by northward shift (e.g., 1850-1854, 1873-1891) and/or clockwise rotation (e.g., 1921-1927, 1934-1939, 1958-1982). At the N and NE-side of the ebb-tidal delta the outer channels migrate to the E to SE, mainly by clockwise rotation (e.g., 1854-1859, NE-channel; 1891-1921/36(?), eastern inlet channel; 1921-1934, old Plaatgat; 1949-1967, northern channel; 1950-1967, eastern flood-defined channel).

Remnants of old seaward parts of outer channels and sudden changes in geographical names (e.g., 1854-1859, NW-channel; 1921-1927, westernmost channel) indicate that sometimes the part of the outer channel, which forms the connection to the open sea becomes suddenly abandoned during migration[4]. A new connection can then form slightly more downdrift.

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[4] The official guide of the Navy of 1877 states: "*There are examples that one heavy storm totally changed the course of the sailing water, so that pilots coming from the sea could not find their way back*" (Anonymous, 1877).

A part of the channels at the eastern side of the ebb-tidal delta is newly formed as marginal flood channels. All channels (also those which arrive due to lateral migration) at the NE-side of the ebb-tidal delta are abandoned after some time (e.g., 1850-1859, eastern most channel; 1891-1921, easternmost channel; 1891-1970, eastern inlet and channel E of it; 1958-1970, eastern flood-defined channel). After abandonment the sandy shoal W of the channel merges with the barrier island downdrift (Schiermonnikoog), and forms an often inter- to supratidal, sandy, often recurved bar (1965-1987; cf. Tye, 1984). Behind such a bar a (tidal) embayment may form, characterized by alternating fine-grained and coarse-grained sedimentation and often relatively strong bioturbation. Where a bar encloses parts of the abandoned channel the clay deposited in the channel forms a potential 'weak spot'. The clay is easily scoured after disappearance of the bar, so that a new channel can form relatively quickly (cf. Tye, 1984). The migration of outer channels is a cyclic process. Two cycles could be followed fully, from generation at the SW-side until abandonment at the NE-side, namely 1927-1970 and 1934-1987.

Also, the eastern inlet experienced clockwise migration, especially after 1891, when it was forced gradually downdrift, and became abandoned.

The following factors determine the development of channels in the ebb-tidal delta of the Zoutkamperlaag:

1) The gradient of the lines of equal tidal phase is less steep than in the Pinkegat Inlet, and according to Huijs (1993) of little influence. However, the preferential orientation of (parts of) the outer channels to the W to NW indicates that this effect (cf. Van Veen, 1936) is probably important for the orientation of especially the outer channels. Flood dominance induced by wave-pumping under the influence of the dominant waves (Nummedal & Penland, 1981) may further enhance the formation of the E-W-oriented marginal flood-channels at the SW side of the ebb-tidal delta.

Moreover, the low gradient of the lines of equal tidal phase in the Zoutkamperlaag area will be an important condition for enabling the formation of a flood-defined channel at the eastern side of the ebb-delta (see point 5).

2) The position of the main backbarrier channel of the Zoutkamperlaag Inlet system at the western side of the drainage basin is relatively stable. Model calculations for the configuration of 1970 indicate a low residual sediment transport through the Zoutkamperlaag main channel (Steijn et al., 1992; Steijn & Louters, 1992), which is in agreement with the observed stable position. Moreover, the large tidal prism of the Zoutkamperlaag forces an updrift asymmetry (cf. Sha, 1989b; Sha & Van den Berg, 1993). Together with point 1 the relatively stable westward position of the main channel favours the formation of new outer channels at the western, updrift side of the ebb-tidal delta.

3) The inertia of the ebb current, which leaves the main, N-S-oriented backbarrier channel (E of Engelsmanplaat) tends to direct the main ebb channel to the north. The formation of shoals at the southern side of the flood-defined channels near the main channel and the related northward shift of these outer channels are likely brought about by the eastward shift of the main ebb channel, when, after a NNW orientation, it becomes oriented to the N. This allows a northward shift of the zone of low net sediment transport capacity S of the flood-defined channels (Fig. 4; cf. Steijn et al., 1992). The other zones of low sediment transport capacity N of the flood-defined channel (Fig. 4) will also shift northwards and force channels N of them to migrate. The tendency of the main ebb channel to be directed to the N induces also reorientation of the outer channels at the NW side, from NW to N.

Model studies show that the zone updrift of the inlet throat is influenced by complicated, mainly circular, patterns of ebb-, flood-, and wave-driven currents, resulting in a zig-zag seaward movement of the sand (Steijn et al., 1992). A comparable zig-zag movement has been observed in the East German Inlets (Hanisch, 1981; Nummedal & Penland, 1981; FitzGerald et al., 1984). The difference, however, is that those inlets are somewhat more dominated by bypassing of many clearly separated bars along a pronounced reefbow (mainly a 'bar bypasser'; Hanisch, 1981; Nummedal & Penland, 1981; FitzGerald et al., 1984; Oertel, 1988; Steijn, 1991). Due to its relatively large tidal-prism-to-littoral-drift ratio, the Zoutkamperlaag Inlet is more of the 'tidal flow bypasser' type, where sediment is transported by tidal flow through channels and by tidal channels and bars which move across the inlet throat (Steijn, 1991). Detailed observations show that the bars between the channels at the updrift side of the main inlet can, but not always do, consist of several smaller, more or less E-W-oriented ridges, although this is not pronounced (cf. Oost & De Haas, 1992).

4) Like in the Pinkegat, lateral shift is enhanced by the curvature of the channels. In the Zoutkamperlaag system this is mainly important for strongly curved outer channels (mainly at the western side), and to a lesser extent for the inlet itself (see point 3; Huijs, 1993).

5) Tidal currents through the inlets disturb the shore-parallel tidal currents, as can clearly be observed on aerial photos. Updrift of the inlet this results in an enhancement of shore-parallel currents, while more downdrift refraction and deflection occur during the ebb and flood (Fig. 5). The large tidal prism of the Zoutkamperlaag Inlet causes a strong disturbance of the shore-parallel flow. The disturbing effect of the outflowing ebb current is maximal, when the ebb channel(s) are oriented to the NNW-N. Then, the ebb current flows perpendicular to the coast-parallel tidal flow and the ebb-jet flows out from the drainage basin in a straight course (Postema, 1956). Downdrift of the inlet channel, weak rotational currents are generated (Sha, 1989b), which enhance sedimentation. Indeed, models, based on the morphology in 1970, suggest the presence of a rotational current NW of Schiermonnikoog. Models indicate that its influence on sedimentation becomes especially pronounced if some wave action occurs (Steijn et al., 1992; Steijn & Louters, 1992).

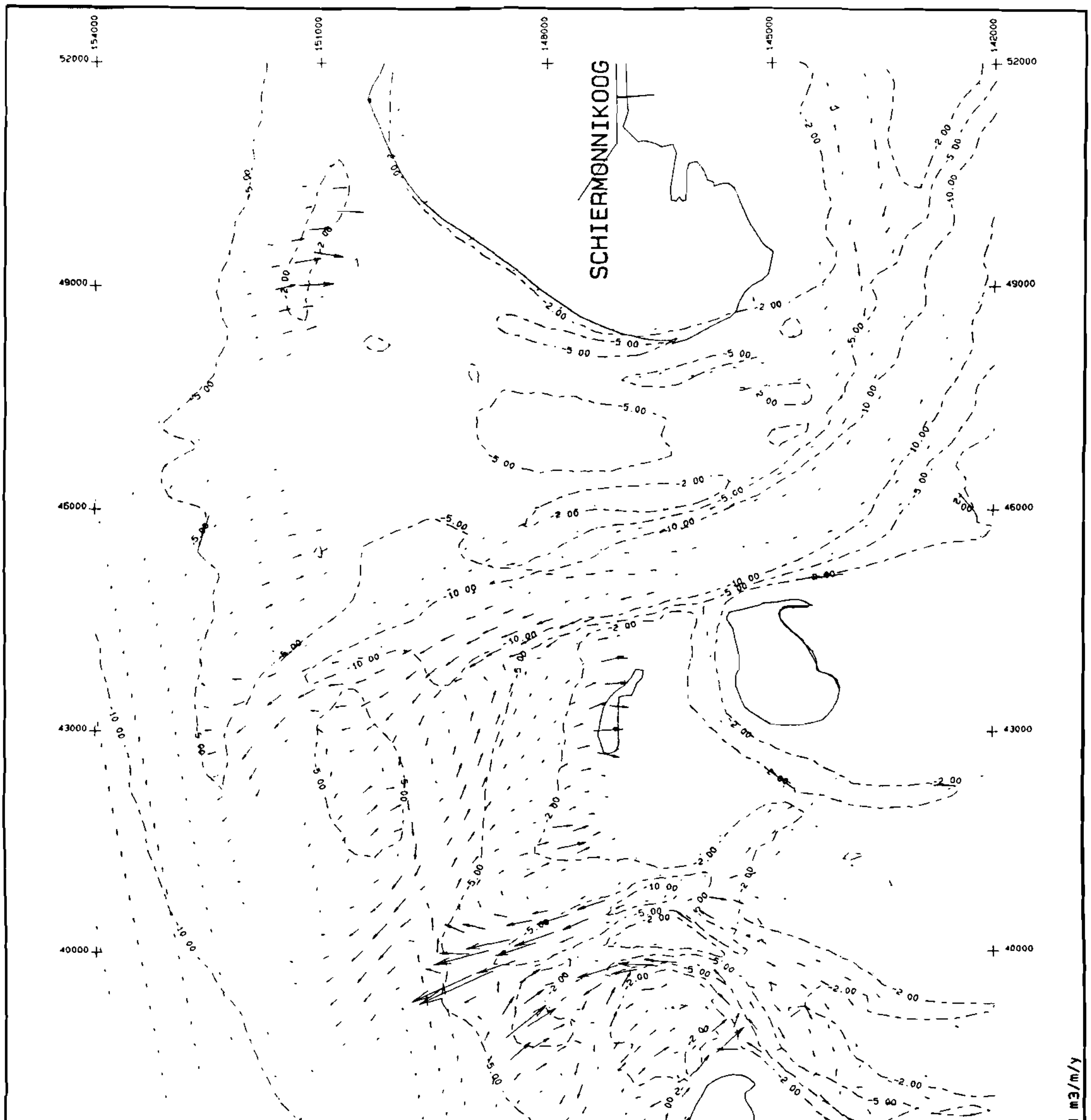
SCHIERMONNIKOOG
AMELAND
NORTH SEA
vectorscale ——▷ = 0.5E+04 m3/m/y

Table IV. The position of the inlet/outer channel and the occurrence of marginal flood channels at the eastern side of the ebb-tidal delta

Year	Main position inlet/outer channels	Marginal flood channel at E-side?
1832	north side	no, shoal off NW Schiermonnikoog
1850	west side	yes
1854	intermediate	yes, smaller than in 1850
1859	north side	no, shoal off NW Schiermonnikoog
1873	west side	yes, but eastern inlet channel served also as flood way
1891	intermediate	yes, in abandonment stage
1903	intermediate	no, only remnant present
1921	west side	no, remnant eastern inlet channel served as flood way
1927	north side	no, remnant eastern inlet smaller than in 1921
1934	intermediate	yes, but flood-defined old Plaatgat was abandoned when the inlets oriented more to the N
1950	west side	yes, but smaller than in 1958
1958	west side	yes, orientation of inlet more to W
1965	intermediate	yes, in abandonment stage
1970 1982	intermediate, increasingly to the N	gradual abandonment, development of large bar
1982 1991	intermediate, more to the west side	no, no channel possible due to large bar

Figure 4 (opposite page): Computed annual net sediment transports in the ebb-tidal deltas of the Frisian Inlet based on the bottom topography of 1970/71, sediment transport formulas of Baillard (1981) and the following weighted occurrence of tides and waves: 125 days/y tides without waves, 112 days/y tides with N-waves, 128 days/y tides with NW-waves (from Steijn et al., 1992; courtesy Steijn).

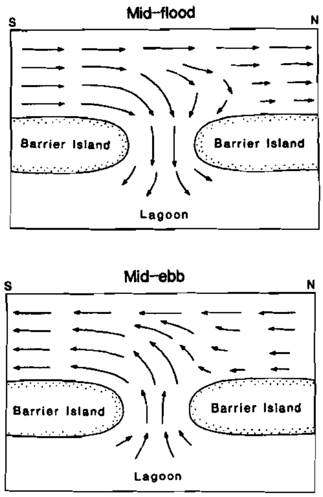

Figure 5. Schematic current patterns at mid-flood and mid-ebb showing the effect of the interaction of tidal currents parallel to the coast with the tidal currents through the inlet. Strong, bidirectional currents are produced at the left hand (SW) side of the ebb-tidal deltas, and a weak, rotational tidal current pattern occurs at the right (N/NE) side (Sha, 1989b).

Furthermore, Sha (1989b) pointed out that the ebb-tidal delta sediment body works as a groyne, which further enhances the above process. The more the ebb-tidal delta protrudes seaward, the stronger the effect. The seaward extension of the ebb-tidal delta is maximal when the inlet and its outer channels are directed to the N (compare 1859 with 1873/74; Figs B4 and B5); the same configuration which enhances the ebb-jet. When the channels are concentrated at the updrift side of the ebb-tidal delta, the ebb-jet effect is less influential, and the ebb-tidal delta then protrudes less to the N.

The above effects are important for the development of the downdrift side of the ebb-tidal delta (Table IV). Flood-defined channels at the eastern side of the ebb delta are generally absent when the inlet/outer channels are located largely at the N-side, because sedimentation and the deceleration of the current hamper the inflow of flood water along the downdrift side. Shoals can build up more easily, and enhance channel migration and the abandonment of channels at the eastern side of the ebb-tidal delta (see point 6; cf. Sha, 1989a). The need for additional marginal flood channels at the NE-side of the ebb-tidal delta is further reduced, because the northwards oriented ebb-dominated channels will be used also for the flood. On the other hand, marginal flood channels do occur in the east when the inlet/outer channels are located mainly at the W-side. Then sedimentation at the downdrift side of the ebb-tidal delta is less effective and the flood can scour a marginal flood channel there. The process is enhanced by the small difference in tidal phase at either side of the inlet (see point 1).

An additional factor for the formation of a marginal flood channel at the eastern side may be the clogging of the marginal flood channel at the westernmost side by sediment supplied through the most eastern outer channel of the Pinkegat (Huijs, 1993).

6) Observations of the developments show that, like in the Pinkegat, also in the ebb-tidal delta of the Zoutkamperlaag large shoals (above -5 m DOL) built up (amongst others enhanced by the effects discussed in point 5). In the zone of -5 m DOL W of the inlet, they tend to migrate mainly to the north and east (see point 3).

East of the main inlet shoals migrate mainly to the E and S, and thus influence the position of the downdrift channels[5] (cf. FitzGerald et al., 1984; cf. Steijn, 1991). In the area, fair-weather waves have an erosive influence on the bottom down to -5 m DOL (data RWS). This, and model research show that a large part of the shoal dynamics at the east side must be wave-driven, with in addition the influence of flood currents (Oost & De Haas, 1992; Steijn et al., 1992; Fig. 4). The highest areas of the shoals may consist of several smaller, up to intertidal, bars (Figs C1 to C10). On detailed maps many even smaller, SW-NE-oriented, subtidal linear wave-built bars[6] are shown (e.g., Fig C10; cf. Oost & De Haas, 1992). Aerial photographs show that the large shoals influence the local wave and tidal climate (cf. Sha, 1989a; Oost & De Haas, 1992; Oost, 1995). Also here, like everywhere in the tidal system, hydrodynamical conditions and morphological developments are mutually dependent. Shoals in the northern part of the ebb-tidal delta often show a tendency to expand and migrate to the S to SSE (e.g., 1854-1873; 1903-1921). Models (Steijn et al., 1992; Steijn & Louters, 1992) and direct observations (measurements RWS) indicate that this is also due to flood dominance in that area, in combination with wave action.

The shoals play an important role in the abandonment processes (cf. 1850-1859). Downdrift of the main inlet, the shoals build up to intertidal heights at the upstream side of the channels, and force them downdrift. When a channel becomes situated too far downdrift, the flow through it, to and from the drainage basin, takes too much time: it becomes hydraulically inefficient (cf. FitzGerald et al., 1984; FitzGerald, 1988). In combination with the downdrift shift of the adjacent outer channel W of it, which by and by takes over its function, this results in abandonment of the channel. The shoal then merges with the downdrift island and forms a bar.

7) Models indicate that along the periphery of the ebb-tidal delta westerly storms may generate a considerable eastward sediment transport E of the main inlet (Steijn et al., 1992; Steijn & Louters, 1992). The observed sudden changes in the ebb-tidal delta during storms

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[5] In the Zoutkamperlaag sand transport by littoral drift (probably) and by aeolian transport is less important, in contrast to the situation in the Pinkegat Inlet (Steijn et al., 1992; Huijs, 1993).

[6] They strongly resemble the swash bars observed in the East Frisian ebb-tidal deltas (Hanisch, 1981; FitzGerald et al., 1984). Since they are mainly below the swash-zone they are better referred to as wave-built bars, or breaker bars.

(Anonymous, 1877; Veenstra & Winkelmolen, 1980) show that storms indeed contribute strongly to the observed changes in the Zoutkamperlaag ebb-tidal delta.

To conclude: Most outer channels in the ebb-tidal delta of the Zoutkamperlaag show a cyclic development. New channels are formed from time to time at the SW side of the ebb-tidal delta. The outer channels migrate clockwise in the downdrift direction, by a combination of lateral shifts and rotational movements, sometimes in combination with downdrift short-cuts. The channel migration W of the main inlet, to the N to NE, results from the tendency of the main inlet channel to orient to the N, in combination with the formation of shoals and lateral accretion along the inner bends of the outer channels. East of the main inlet, in particular the migration of sandy shoals by wind and tides results in a downdrift shift. The migration process can temporarily be reversed when channels form updrift short-cuts, and thus deviate the course of an outer channel. The formation of new channels occurs mainly at the NW-side of the ebb-tidal delta. Moreover, completely new channels are formed at the SW-side, and can be the start of a new cycle (e.g., 1949). The preferential formation of new channels at the western side of the ebb-tidal delta results from the effect of the gradient of lines of equal tidal phase (Fig. 6 of Chapter 3; Van Veen, 1936; Sha, 1989b; Huijs, 1993), the dominant westerly waves, as well as from the western location of the main backbarrier channel. When the migration is not interrupted, a channel can migrate over large parts of the ebb-tidal delta (e.g., 1927-1970 and 1949-1982), thereby even changing from flood- into ebb-defined. The channels thus may arrive at the E-side of the ebb-tidal delta. The mainly wave-induced shift of sandy shoals forces them downdrift. Finally the channels are abandoned, because they become hydraulically inefficient and are covered by migrating shoals.

From time to time marginal flood channels are formed at the eastern side of the ebb-tidal delta. This occurs especially when the inlet/outer channels are located far to the west, perhaps in combination with an eastward position of the Pinkegat inlet channels. These marginal flood channels become abandoned when the position of the inlet/outer channels is more towards the north. These outer channels cause a decrease of the downdrift currents and take over part of the drainage of the flood channel.

Zoutkamperlaag, drainage basin

The position of the Zoutkamperlaag main backbarrier channel has been relatively stable at the western side of the drainage basin. The stable position must, for an important part, be brought about by the form of the drainage basin, which consists of a southern area (Lauwerszee embayment) and an eastern area (N of the Groningen mainland). The main channel simply forms the shortest (fastest) possible way between these areas and the inlet. The curvature of the main channel is caused by the momentum of the water going and coming from the east. Moreover, the substratum likely restricts lateral movements of the main channel: to the W early Holocene clay-rich deposits beneath the Engelsmanplaat and to the E a Pleistocene high and clay-rich early Holocene deposits (cf. Kievits, 1994) form the substratum.

Before the closure of the Lauwerszee embayment, the developments in the backbarrier were also cyclic. During short periods only one main channel provided the drainage (e.g., 1832, 1949). The channel became separated in a flood chute at the western side and a channel bed, used by both the ebb and the flood current (Postema, 1956; e.g., between 1832-1859, 1927-1934). Between these channel beds an intertidal shoal was formed which became more and more elongate (e.g., 1806, 1859-1927). Finally the western channel was abandoned and was filled up (e.g., 1806-1832, 1927-1949). This led again to a single main channel configuration. The formation of new flood chutes sometimes partly overlapped with the abandonment of the old channel (e.g., 1927-1934).

The maps from 1806 onward suggest that the developments in the drainage basin were closely related to the developments in the ebb-tidal delta. When the outer channels and inlet configuration induced a concentration of part of the tidal currents along the eastern side of the main backbarrier channel, the latter separated in a flood chute at the western side and a channel bed which was used by both the ebb and the flood current at the eastern side (e.g., 1832-1859; 1927-1934, main channel). The flood chute mainly served for the drainage of the Lauwerszee embayment. The bed at the eastern side of the channel was used by both the ebb and the flood, and usually served to drain both the Lauwerszee and the drainage basin to the E. The flow concentration along the eastern side of the main channel was mainly brought about by a positioning of the outer channels at the northern side of the ebb-tidal delta in combination with an eastern position of (part of) the inlet channel(s).

A gradual erosion at the outer bends of the eastern channel bed, in combination with a flood-driven displacement of the shoals from the inlet into the drainage basin (e.g., 1873) resulted in increasingly better connections with all the outer channels of the ebb-tidal delta. Thus, the eastern channel bed took over the drainage more and more. Between the two channel beds a shoal built up and the western channel bed was gradually abandoned (e.g., 1806-1832; 1921-1949). The eastern channel bed became the new main channel, and the whole process started again.

The cyclic development of the main channels in the drainage basin would probably not have occurred in the absence of the Lauwerszee. The embayment enhanced the formation of the flood chute. This is illustrated by the changes after the closure of the Lauwerszee. The separation of the main drainage channel into two channel beds occurred in 1927 and the dimensions of the eastern channel bed increased (especially 1949-1967). However, after the closure, the main drainage channel partly switched from the eastern channel bed to the western channel bed, thereby terminating the separation into two subchannels. Thus, the cyclic development of the backbarrier channel likely was ended by the enclosure of the Lauwerszee embayment.

The Lauwerszee embayment

Historical observations over the period 1550-1800 and, in more detail, over the period 1832-1967, show that the channel pattern did not change much (Figs A3 to C4). Over the period 1832-1927 some channel cut-off and the generation of several 3rd and 4th order channels can be observed, in particular at the entrance of the Lauwerszee embayment. Overall, the morphodynamical development was relatively slow (Fig. 3). In this aspect the area differs from the more dynamic drainage basins in the rest of the Wadden Sea. The slow dynamics of the channels has several causes:

1) The Lauwerszee formed a semi-enclosed embayment, only open at one side. Only along that side water transport was possible, in contrast to the rest of the drainage basin where water could flow also over the watersheds. In both cases the position of the smaller order channels was strongly determined by the drainage area. The ideal position for the channels was in the centre of such an area. If a channel shifted and entered the drainage area of another channel, one of both had to be abandoned and at the original place of the migrated channel a new drainage channel formed. This, and the formation of chutes (e.g., 1891-1903; Van Straaten, 1954) strongly restricted the possibilities for lateral migration of the smaller channels (Fig. 8 of Chapter 3). Near the watershed such "end"-channels could become longer or shorter with the expansion or retreat of the drainage basin. The Lauwerszee embayment has no watershed, but was surrounded by supratidal land, restricting lateral as well as longitudinal movements.

2) The main channel of the Lauwerszee was cut into the Pleistocene subsurface. The Pleistocene clay mass was resistant to erosion and hampered lateral migration of the main channel in the embayment.

3) The influence of the wind was weak, because the area was surrounded by land at three sides, thus allowing the settling of fine-grained matter. The larger part of the area consequently consisted of clays to clay-rich sands, which were difficult to erode. Also, the sheltered position in combination with the clay-rich sediments favoured a stable position of shoals and channels, as is still the case in the present-day Dollard area.

The dominance of fine-grained sedimentation in the Lauwerszee embayment was favoured by:

1) The relatively high percentage of intertidal shoal area (79%). At low tide the water was concentrated in the channels and the average water depth was large (in the areas covered by water). The average water depth was much less during high tide. Suspended sediment could therefore settle more easily during high tide (Van Straaten & Kuenen, 1958).

2) In the inner parts of the Wadden Sea the tidal wave becomes asymmetrical in such way that around low water the tide turns much more quickly than around high tide. The time span during which low current velocities, allowing suspended matter to settle, are active, is therefore much greater when the water covers the flats during high water, than during low water when it turns quickly, while only filling the channels (Postma, 1961).

3) Due to settling-lag and scourlag effects (Chapter 1) higher flow velocities are needed to resuspend particles than to deposit them. The net result is a landward transport of these particles (Hjulström, 1935; Van Straaten & Kuenen 1958).

4) Suspended matter is transported by waves in the direction of wave propagation, which is most of the time into the Wadden Sea (Ehlers & Kunz, 1993). Suspended matter, deposited during calm periods is resuspended and transported by waves. This usually does not happen in the more sheltered areas, where fine-grained sediments thus preferentially accumulate.

5) Organisms (invertebrates, diatoms, plants) bind the sediment (Chapter 1 and 6; cf. Vos et al., 1988). The vegetated salt marshes around the embayment strongly reduce current velocities and enhance the suspended sediment to settle. Sediment is later fixed by the root systems of the plants. This has especially been an important factor for the strong sedimentation between -2 m and +1.5 m DOL. Man artificially encouraged the natural vertical growth, and thus was an important factor for the sedimentation in the area.

Geological relevance

At the North Sea side, the Zoutkamperlaag Inlet system shows a strong clockwise and downdrift migration of the outer channels, and, over the period 1891-1934, of one of the inlets. The downdrift migration of the eastern inlet channel in the period 1891-1949, away from the optimal position, caused a decrease in hydraulic efficiency. This led to a gradual abandonment and shallowing of the eastern channel. Thus eastward dipping lateral accretion deposits formed at its western margin.

The shift of the outer channels produced also lateral accretion surfaces, of which the orientation changed clockwise in the downdrift direction. It implies that they dip roughly to the N to NE at the western side of the ebb-tidal delta and change in dip to E to SE at the northern side of the delta (Fig. 10 of Chapter 3). In general, the outer channels, which are newly formed at the SW-side of the ebb-tidal delta are deep (more than -10 m DOL). Upon northward migration, they decrease in depth and become flood-defined. When they migrate further to the N they increase in depth and, in general, they become ebb-defined. As soon as the outer channels become oriented NNE to E, they generally decrease in depth. These eastern channels can be ebb- or flood-defined. The occasional marginal flood channel at the downdrift side normally shows a strong shallowing. Shoal deposits are, in general, flood-dominated and may be expected to show frequent indications of wave action. Notwithstanding this general picture of the Zoutkamperlaag ebb-tidal delta, the depth of the outer channels shows irregular, strong changes in time.

Like in the Pinkegat, the clockwise and downdrift migration of the outer channels is a good indicator of the residual current in the basin. Seismic sections through older ebb-tidal delta deposits (from 5,400 B.P. onwards), seaward of the present ebb delta, indeed show a dominance of lateral accretion surfaces dipping to the N to E (Sha, 1992). Since such deposits have a fair preservation potential (Sha, 1990), they provide a useful tool for the reconstruction of paleo-basin dynamics.

The main channel system of the Zoutkamperlaag Inlet is far less active than that of the Pinkegat Inlet system. The difference is due to the relatively low sand transport by littoral drift and aeolian transport towards the Zoutkamperlaag Inlet, the position of the large Lauwerszee embayment, the presence of (subsurface) highs of Pleistocene and Early Holocene sediments and, in case of the Lauwerszee, due to the abundance of recent clay deposits. Moreover, the large tidal prism forces a stable updrift position of the main channel system (Sha, 1989b), so that also the main channels in the drainage basin remain stable. The erosion base thus will be a reliable indication of the maximum channel depth and of the tidal prism.

The drainage basin of the Zoutkamperlaag Inlet system has gradually decreased in volume since at least 1300 A.D. Due to the decrease, channels became shallower through time. The infill of the channels is mainly sandy (Kievits, 1994), because the decrease in tidal volume and tidal currents has not been abrupt (Chapter 4), but gradual. Upon continued decrease of the drainage basin the sandy sediment body deposited, would eventually become a long channel form, with several branches, characterized by many lateral accretion and internal erosion surfaces. Laterally and vertically the channel form will mainly consist of sand and shell lags, with sometimes clay-rich layers on the lateral accretion surfaces and clay drapes on megaripples (Van den Heuvel, 1993). The channel fill may be flanked by deposits with large amounts of clays formed during abandonment of the flood chute. Observations of the fill of the main channel in the Lauwerszee embayment do not indicate a much higher clay content than in the rest of the drainage basin (cf. Bosch & Vos, 1992; Kievits, 1994). However, the sediments deposited in the shallower parts of the channels and in the smaller channels, may contain large amounts of clay, because they can be influenced by the clay sedimentation on the shoals.

In general, such channel fills, as formed in the Zoutkamperlaag Inlet system, indicate a relatively stable position of the main channel(s). The stable position is determined by the position of embayments, which have to be drained and, probably to a lesser extent, by older substratum, difficult to erode. In addition, the position of embayments, such as the Lauwerszee, can be determined by the paleo-relief of the substratum (cf. Bosch & Vos, 1992). Thus, in areas with an irregular substratum, laterally confined, well developed channel fills can be expected. The deposits described here resemble incised valley fills (cf. Reinson, 1992). It is emphasized that their genesis is different. In incised valley fills the tidal channel deposits mainly fill up the paleo-valley which developed during a lowstand. Here tidal channels have incised deeply, to drain low-lying drainage basins, formed during gradual rise of relative sea level and flooding of the mainland.

COMPARISON OF THE PINKEGAT AND ZOUTKAMPERLAAG; GEOLOGICAL RELEVANCE

The inlet systems of the Zoutkamperlaag and in particular the Pinkegat show a strong cyclic development over periods of the order of decennia to a century. The differences between both systems are determined by the differences in shape and position of the drainage basin, by the size of the tidal prism relative to the influence of wave action, and by the supply of sand (Eysink, 1979; Sha, 1989b; Steijn et al., 1992; Oost & De Haas, 1992, 1993; Huijs, 1993). In the Pinkegat the cyclic shift of the inlet is registered by extensive, seaward to downdrift dipping, lateral accretion deposits in the ebb-tidal delta and, throughout the basin, by many internal erosion features and other indications of strong changes in local energy conditions. In the ebb-tidal delta the channel deposits increase in thickness (depth) in a downdrift direction to a certain point, after which they decrease. Due to the net downdrift shift of the Pinkegat inlet, two channel lags are deposited on top of each other. In the Zoutkamperlaag, seaward to downdrift dipping, lateral accretion surfaces are formed within the ebb-tidal delta by lateral migration of the outer channels. These deposits show no clear trend in thickness development. In the drainage basin of the Zoutkamperlaag lateral accretion is strongly restricted to a narrow zone along the channel, so that the cyclic development results in a constant reworking of older lateral accretion deposits. Both in the Zoutkamperlaag Inlet system and in the Pinkegat Inlet system, the developments in the ebb-tidal deltas and inlets are closely followed by those in the drainage basin[7]. Thus, especially the ebb-tidal delta and inlet deposits provide clues to understand the formative conditions and to predict the character of the backbarrier deposits.

The reconstruction of the formative conditions and the 3-D distribution of barrier-related facies in the fossil record is often difficult. However, the above described relationships (Chapter 1) allow predictions more easily than was previously thought. The dimensions of many parts of barrier-related deposits strongly depend on the size of the tidal prism. In the eastern part of the Dutch Wadden Sea the various parts of the inlet systems (inlet, ebb-tidal delta, adjacent island ends, and the drainage basin) are coupled and in dynamical equilibrium with the prevailing hydrodynamic conditions (in particular the tidal prism)[8]. More in

[7] To a lesser extent also reverse, i.e., influences of the drainage basin on the ebb-tidal delta development can be observed.

[8] At present the Zoutkamperlaag is not in equilibrium, due to the closure of the Lauwerszee embayment.

general: except for rapid changes in the drainage basin, (due to infill, increase/reduction of tidal prism, etc.), the ebb-tidal delta, the inlets and the backbarrier channels of most inlet systems can be considered to be coupled, and to be in dynamical equilibrium with tidal prism over periods of the order of decennia to centuries, if the hydrodynamic conditions (in particular tides and wave climate) and the availability of sediment are stable. Thus, observations of any part of the system provide clues to other parts of the system at that time. For instance, observed changes in channel size can be related to changes of the related drainage area (when tidal range is constant) and to the related sand volume of the ebb-tidal delta.

To which extent a certain equilibrium is registered in the subsurface depends in the first place on the intensity of lateral shifting of especially the largest channels during such an equilibrium situation. For instance, the Pinkegat system experienced strong lateral shifts, but in the Zoutkamperlaag Inlet system the shift was relatively slow over the period 1832-1991. FitzGerald (1988) recognized three modes of lateral shift for mixed energy coasts, namely:

1) Inlet migration (usually downdrift) in combination with spit breaching. When water exchange through the inlet becomes inefficient, because the spit becomes too long, the latter is breached (during storms) and a new inlet is formed. A extreme example was discussed by Giese (1988).

2) Stable inlet processes: the throat of the inlet has a stable position and the main ebb channel does not migrate. Waves rework the coast-parallel transported sand to swash bars, which accumulate along the downdrift shoreline.

3) Ebb-tidal delta breaching: the throat of the inlet is stable, but the outer ebb channel migrates. Sediment accumulation at the updrift side of the ebb-tidal delta results in the deflection of the outer ebb channel. During the migration of the channel it looses its efficiency, and is replaced by another new, more updrift channel (cf. Tye, 1984).

In particular in model 2, lateral migration of the inlet throat and of the outer ebb channels, which have the highest preservation potential, is limited. However, all three mechanisms can be active at some time in an inlet system (FitzGerald, 1988), generating a fair change that inlet-related facies are deposited over a considerable lateral extent (up to several km; cf. Sha, 1989a, 1990). Such deposits generally consist of lateral accretion surfaces which roughly dip downdrift. At the most updrift side of the ebb-tidal delta the accretion surfaces of the outer channels may dip seaward (Fig. 10 of Chapter 3). The preservation potential depends also on the tidal prism. From the relations between the tidal prism and the dimensions of the various parts of the tidal system it follows that deposits formed during conditions of maximum tidal prism, resulting from a large tidal amplitude or a large drainage basin, will have a high preservation potential, because this leads to large ebb-tidal deltas, deep channels, and deep inlets.

The observations of the Pinkegat system (mainly model 1 of FitzGerald (1988), in combination with 3; Huijs, 1993) and the Zoutkamperlaag Inlet system (mainly model 3, in combination with 1) show that in both ebb-tidal deltas the channel development is cyclic. During each cycle there is a configuration in which the inlet reaches maximum dimensions. This happens preferentially when it is the only or almost only transport pathway of water; in all other inlet configurations channels have lesser dimensions. Because in a single inlet configuration the cross-sectional area and the depth of an inlet are normally maximal (Sha, 1990; Gerritsen, 1990; Niermeyer, 1990; Biegel, 1991b), the preservation potential of the single inlet deposits is large. When the cross-sectional area or, to a lesser extent (Sha, 1990; Eysink & Biegel, 1992), the maximum inlet depth are known, this allows the calculation of the tidal prism through the inlet[9]. In general, in all tidal inlet systems with a cyclic development there will be a phase during which maximum dimensions are reached. All other inlet configurations have smaller dimensions. Thus, a cyclic sedimentary sequence with multiple channel lags will form, comparable to the inlet migration models of Kumar & Sanders (1974) and Reinson (1984). Based on the maximum dimensions of the tidal inlet, estimates can be made of the dimensions of other parts of the system, which are closely related to the tidal prism (Chapters 1 and 5).

In geology, the time needed for the formation of sedimentary deposits is often measured in thousands or many of thousands of years. However, the formation of thick, wide-spread sandy deposits in tidal systems such as the Pinkegat and Zoutkamperlaag is almost instantaneous. Therefore, a vertical sequence of barrier-related deposits without major erosional breaks should primarily be considered to be a snap-shot, which reflects the prevailing, external conditions over a short period. The deposits reflect the dynamical equilibrium conditions during that very period. Sudden, often erosional breaks within the vertical sequence may be generated by (cyclic) changes in the tidal-inlet system (e.g., Pinkegat) or may result from erosion and deposition in a much later stage. To a large extent coast-parallel profiles of barrier-related deposits, which show no great breaks in sedimentation, will have been formed in a short span of time, as is illustrated by the downdrift lateral accretion surfaces in the ebb-deltas of the Pinkegat and Zoutkamperlaag. Profiles of barrier-related deposits perpendicular to the coast, which are part of a transgressive systems tract (even a fifth order cycle succession is commonly considered to represent an interval of 10^4-10^5 yr), can be expected to have formed over a long time span, during which external conditions changed (cf. Thom, 1983; Beets et al., 1992; Chapter 1). Thus such units may show a considerable change in sedimentary characteristics. The same accounts for regressive systems, if preserved.

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[9] Not considered in this study are partially estuarine settings (with often separate channels for ebb currents and flood currents) or complicated situations, where resonance plays a dominant role.

CONCLUSIONS

The sedimentary/geomorphological developments in the Zoutkamperlaag Inlet system are cyclic. From time to time, new flood-defined channels are formed at the western margin of the ebb-tidal delta, due to the tidal gradient and the western position of the inlet. The outer channels mainly migrate clockwise. At the western side of the main inlet, channels migrate to the N, resulting from the tendency of the main inlet channel to orient to the N, in combination with the formation of shoals and outer bend erosion of the outer channels. During migration flood-defined channels at the western side evolve into ebb-defined channels. The latter get a N-S-orientation, due to the inertia of the ebb current. East of the main inlet, the migration of sandy shoals by wind and tides enhances a further downdrift shift of the outer channels and inlet channels. During a further downdrift shift to a NE-SW-orientation the channels become abandoned. This facilitates the merger of shoals west of the abandoned channel with the downdrift island. New flood-defined outer channels are formed from time to time at the eastern side of the ebb-tidal delta of the Zoutkamperlaag. It is inferred that these result from a mainly westward position of the other outer channels and the inlet, perhaps in combination with an eastward orientation of the Pinkegat Inlet channels.

The cyclic development (before the closure of the Lauwerszee embayment in 1969) in the main backbarrier channel of the Zoutkamperlaag was, to a large extent, controlled by developments in the ebb-tidal delta. During short periods only one main channel drained the drainage basin. The channel split up in a flood chute at the western side and a channel bed used by both the ebb and the flood current at the eastern side. This happened when the outer channels and the inlet configuration concentrated part of the tidal currents along the eastern side of the main backbarrier channel. Successively the eastern channel became increasingly more important for the drainage of the system. Between both channels an elongated intertidal shoal was formed. Finally the most western channel became abandoned and was filled up. This again led to a single main channel configuration. The formation of new flood chutes could partly overlap with the abandonment of the old one. The stop of the cyclic developments and the change in the orientation of the channel axis after closure of the Lauwerszee suggest that this embayment, i.e., the original dimensions of the total drainage basin, was crucial for the maintenance of the cycle. Over the period of observation (1832-1967) the morphology of the Lauwerszee changed only slightly. It showed a relatively slow morphodynamic development, because the Lauwerszee forms a wind-sheltered, semi-enclosed embayment, with characteristic fine-grained sedimentation. Moreover, the erosion resistant Pleistocene to early Holocene subsurface restricted lateral shifts.

The dynamics of the Zoutkamperlaag system can be recognized in the sediments deposited. Seaward to downdrift dipping lateral accretion surfaces are formed mainly within the outer ebb-tidal delta. In the drainage basin, lateral migration of channels is restricted to a narrow zone, so that the cyclic development results in a constant reworking of older deposits. Upon gradual infill of the drainage basin the main channel would fill up and form a channel body mainly consisting of sands and flanked by a more clay-rich facies. In the Lauwerszee embayment the body would be flanked and capped by clay-rich sediments. Such types of fills are considered to indicate an influence of the relief of the subsurface on the dimensions of the drainage basin and on the position of the main channels.

Observations of any part of a barrier system may provide valuable clues to all other parts of the system, because the dimensions of many parts of barrier-related deposits are in dynamical equilibrium with the tidal prism. The intensity of lateral shifts of especially the channels and the size of the tidal prism, primarily determine the volume of sediment deposited during an equilibrium situation and hence the chance that the equilibrium is registered in the sedimentary record. Within almost the same setting large differences in the rate of lateral shift may occur between inlets (compare Zoutkamperlaag with Pinkegat). This depends largely on their size and shape.

In both the Zoutkamperlaag and Pinkegat Inlet systems the developments in the drainage basin are strongly determined by the developments in the ebb-tidal delta, although also a reversed influence is observed. To understand and reconstruct the sedimentary conditions the most useful information is to obtained from the inlet and the ebb-tidal delta deposits. The deposits which are formed in tidal systems such as the Pinkegat and Zoutkamperlaag are deposited almost instantaneously, when compared to the time scales to which a geologist is used to. In vertical sections (cores) and, to a large extent, also in coast-parallel sections the deposits primarily reflect the prevailing, external conditions over short periods. Long sections perpendicular to the (paleo-)coastline will in general represent longer periods. Such perpendicular sections thus may show variations in sedimentary development due to the changing external conditions with time.

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REFERENCES

Allersma, E., 1993: Geulmigratie in buitendelta's. Rapport WL, H. 840.55, 64 pp.

Anonymous, 1877: Beschrijving van het Friesche Zeegat. Issued by the Min. of Navy, Hydrographical Dep., The Hague, 58 pp.

Anonymous, 1930: Kort geschiedkundig overzicht van het amelander wad. RWS Hoorn, D 53 (internal report), 33 pp.

Anonymous, 1948: Beschrijving van de provincie Friesland, behorende bij de Waterstaatskaart. Staatsdrukkerij, The Hague, 686 pp.

Anonymous, 1952: Hydrografisch opnemen. Min. van Defensie, Algemene Landsdrukkerij, 2nd ed., 344 pp.

Anonymous, 1974: Met lood en lijn, catalogus van de tentoonstelling, Honderd-jarig bestaan van de afdeling Hydrografie van het Ministerie van Defensie (Marine), ed. Chef der Hydrografie, Chevalier, Rotterdam, 135 pp.

Anonymous, 1981: Zandwinning in de Waddenzee. Resultaten van een hydrografisch-sedimentologisch onderzoek. Werkgroep 1, 48 pp.

Anonymous, 1991: Evaluatie van interpolatiemethoden voor het verwerken van vaklodingen op de Noordzee. Icim, rapport CSO, 1991-475.

Baillard, J.A., 1981: An energetics bedload model for a plane sloping beach: local transport. J. of Geophys. Res., 86, C3.

Beets, D.J., Van der Valk, L. & Stive, M.J., 1992: Holocene evolution of the coast of Holland Mar. Geol., 103, 423-443.

Biegel, E.J., 1991a: De ontwikkeling van de ebgetijde delta en het kombergingsgebied van het Friesche Zeegat in relatie tot de sluiting van de Lauwerszee. GEOPRO, 1991.07, Fac. Geogr. Sci., Utrecht University.

Biegel, E.J., 1991b: Equilibrium relations in the ebb tidal delta, inlet and backbarrier area of the Frisian inlet system (concept), Fac. of Geogr. Sci., Univ. of Utrecht.

Bilse, D.P., 1993: A large scale model of the morphological behaviour of the outer delta of tidal inlets. Delft Hydraulics, Report H 1215: 42 pp.

Bosch, A. & Vos, P.C., 1992: Paleogeografische reconstructie van het Lauwersmeergebied. Concept RGD-Project 40009, 22 pp.

De Boer, M., 1979: Morphologisch onderzoek Ameland. Verslag van het onderzoek op het Amelander Wantij in 1973. Rijkswaterstaat, Directie waterhuishouding en waterbeweging, district kust en zee, studiedienst Hoorn, Nota WWKZ-79.H005, 67 pp.

De Boer, M., Kool, G. & Lieshout, M.F., 1991a: Erosie en sedimentatie in de binnendelta van het Zeegat van Ameland, 1926-1984, Deelonderzoek no.4. RWS Directie Noord-Holland, Rapport ANVX-91.H202, 42 pp.

De Boer, P.L., Oost, A.P. & Visser, M.J., 1989: The diurnal inequality of the tide as a parameter for recognizing tidal influences. J. Sedim. Petrol., 59, 912-921.

De Haan, H., IJbema, R. & Reitsma, D.T., 1983: Engelsmanplaat: geschiedenis van en gebeurtenissen rond een zandbank. Moddergat: Stichting 't Fiskershúske, Necumdruk, Drachten, 168 pp.

De Haas, H., 1991: Morphology of tidal channels in the Ameland-Schiermonnikoog area. Comp. Sedimentology Division, University Utrecht, internal report, 25 pp.

De Jong, J., 1984: Age and vegetational history of the coastal dunes in the Frisian Islands, The Netherlands. Geol. & Mijnbouw 63, 269-276.

De Looff, M., 1975: Enkele opmerkingen over de nauwkeurigheid van de bij de inhoudsberekeningen voor de Westerschelde toegepaste methode. Rijkswaterstaat, Directie Waterhuishouding en waterbeweging, Studiedienst Vlissingen, Memo 75.1., 25 pp.

De Vriend, H.J. & Bakker, W.T., 1993: Sedimentary processes and morphological behaviour models for mixed-energy tidal inlets. Netherl. Centre coast. Res. Delft, Rep. H 1887: 25 pp.

Dean, R.G., 1988: Sediment interaction at modified coastal inlets: processes and policies. In Aubrey, D.G. & Weishar, L. (Eds): Hydrodynamics and sediment dynamics of tidal inlets, Lect. Notes Coastal Estuarine Studies, 29, 412-439.

Dean, R.G. & Walton, T.L., 1975: Sediment transport processes in the vicinity of inlets with special reference to sand trapping. - In: Cronin, L.E. (Ed.): Estuarine Research: Vol. II, Geology and Engineering: 129-150; Acad. Press, NY.

Dieckmann, R., Osterthun, M. & Partenscky, H.W., 1988: A comparison between German and North American tidal inlets. Proc. 21st Coast. Engng. Conf., ASCE, 3, 199: 2681-2691.

Edelman, T., 1961: Verstoringen in de kustlijn tengevolge van de aanwezigheid van zeegaten. Rijkswaterstaat, nota WWK 61-3, 12 pp.

Ehlers, J., 1988: Morphodynamics of the Wadden Sea. 397 pp.; Rotterdam (Balkema).

Ehlers, J. & Kunz, H., 1993: Morphology of the Wadden Sea natural processes and human interference. Coastlines of the southern North Sea, Proc. 8th Symp. on Coast. and Ocean Management, 65-84.

Endema, D., 1979: Grootheden / gegevens Waddenzee, waddeneilanden en Noorzeekust. RWS, Hoorn, Nota WWKZ-79.H003.

Eysink, W.D., 1979: Morphologie van de Waddenzee, gevolgen van zand- en schelpenwinning, verslag literatuuronderzoek. Waterl. Lab., Report R-1336, 92 pp.

Eysink, W.D., 1992: Impact of sea level rise on the morphology of the Wadden Sea in the scope of its ecological function. Proposed set-up of a dynamic morphological model for Wadden Sea basins and estuaries based on empirical relations. Vol. 1. Rijkswaterstaat Dienst Getijdewateren, ISOS2 Project, Phase 3, 38 pp.

Eysink, W.D., 1993: Impact of sea-level rise on the morphology of the Wadden Sea within the scope of its ecological function. General considerations on hydraulic conditions, sediment transports, sand balance, bed composition and impact of sea-level rise on tidal flats. Rijkswaterstaat Dienst Getijdewateren, ISOS*2 Proj., Phase 4: 35 pp

Eysink, W.D. & Biegel, E.J., 1992: Impact of sea level rise on the morphology of the Wadden Sea in the scope of its ecological function. Investigations on empirical morphological relations, Rijkswaterstaat Dienst Getijdewateren, ISOS2 Project, Phase 2, 73 pp.

FitzGerald, D.M., 1988: Shoreline erosional-depositional processes associated with tidal inlets. In: Aubrey, D.G. & Weishar, L. (Eds): Hydrodynamics and sediment dynamics of tidal inlets, Springer-Verlag, N.Y., 186-225.

FitzGerald, D.M., Penland, S. & Nummedal, D., 1984: Control of barrier island shape by inlet sediment bypassing: East Frisian Islands, West Germany. Marine Geology, 60, 355-376.

FitzGerald, D.M. & Penland, S., 1987: Backbarrier dynamics of the East Frisian Islands. J. Sedim. Petrol., 57, 746-754.

Flemming, B.W., 1991: Holozäne Entwicklung, Morphologie und fazielle Gliederung der Ostfriesischen Insel Spiekeroog (südliche Nordsee). Senckenberg-am-Meer, Ber. 91/3: 51 pp.

Flemming, B.W. & Davis, R.A. Jr., 1994: Holocene evolution, morphodynamics and sedimentology of the Spiekeroog barrier island system (S. North Sea). Senckenbergiana Maritima, 24, 3-64.

Gaye, J. & Walther, F., 1935: Die Wanderung der Sandriffe vor den ostfriesischen Inseln. Bautechnik 41, 1-13.

Gerritsen, F., 1990: Morphological stability of inlets and channels of the western Wadden Sea. Rijkswaterstaat, Dienst Getijdewateren, Nota GWAO-90.019, 86 pp.

Giese, G.S., 1988: Cyclical behavior of the tidal inlet at Nauset Beach, Chatham, Massachusetts.

Griede J.W. & Roeleveld, W., 1982: De geologische en paleogeografische ontwikkeling van het Noordelijk Zeekleigebied, K.N.A.G. Geog. Tijdschrift, 16, 439-543.

Hanisch, J., 1981: Sand transport in the tidal inlet between Wangerooge and Spiekeroog (Western Germany). In: S.D. Nio, R.T.E. Shüttenhelm & Tj.C.E. van Weering (Eds): Holocene Marine Sedimentation in the North Sea Basin. Int. Assoc. Sediment. Spec. Pub., 5, 175-185.

Hartman, J. & Pastoor, K., 1985: Onderzoek naar de verandering van de bodemligging voor de Oostelijke Waddenzee en buitendeltas voor de perioden, 1975, 1979 en, 1970, 1979. Rijkswaterstaat, directie Groningen, nota 85-24, 23 pp.

Hayes, M.O., 1975: Morphology of sand accumulations in estuaries. In: Cronin, L.E. (Ed.): Estuarine Research vol. 2 Geology and Engineering.Acad. Press, N.Y., 3-22.

Hayes, M.O., 1979: Barrier island morphology as a function of tidal and wave regime. In. Leatherman, S. (Ed.): Proceedings of the coastal symposium on Barrier islands. Acad. Press, N.Y., 1-28.

Heynis, H., Berger, G.W. & Eisma, D., 1987: Accumulation rates of estuarine sediment in the Dollard area: comparison of Pb-210 and pollen influx methods. Neth. J. Sea Res., 21, 295-301.

Hjulström, F., 1935: Studies of the morphological activity of rivers as illustrated by the River Fyris. Bull. Geol. Inst. Univ. Uppsala 25, 221-527.

Hubbard, D.K., 1977: Variations in tidal inlet processes and morphology in the Georgia Embayment: Tech. Rept. 14-CRD, Coastal Res. Div. Geol. Dept. Univ. of South Carolina, 79 pp.

Hubbard, D.K., Oertel, G. & Nummedal, D., 1979: The role of waves and tidal currents in the development of tidal inlet sedimentary structures and sand body geometry: examples from North Carolina, South Carolina and Georgia. J. Sedim. Petrol. 47, 1121-1131.

Huijs, S.W.E., 1993: Geulmigratie op de buitendelta, Het Friesche Zeegat en het Amelander Gat, met bijdragen van J.J.P. Lambeek, Inst. voor Marien en Atmosferisch onderzoek Utrecht, 58 pp.

Hume, T.M. & Herdendorf, C.E., 1990: Morphologic and hydraulic characteristics of tidal inlets on a headland dominated low littoral drift coast, Northeastern New Zealand. J. coast. Res., Spec. Iss: Proc. Skagen Symp.: 527-563.

Isbary, G., 1936: Das Inselgebiet von Ameland bis Rottumeroog. Morphologische und hydrographische Beiträge zur Entwicklungsgeschichte der friesischen Inseln. Archiv der deutschen Seewarte, 56, 3, 55 pp.

Jansen, P.Ph., 1941: Groninger Wadden. Studiedienst Hoorn, internal report C110, 132 pp.

Jarrett, J.T., 1976: Tidal prism-inlet area relationships. GITI, Rep. 3, US Army Eng. Waterways Exp. Stat. Vicksburg.

Johnson, J.W., 1919: Shore processes and Shoreline Development. John Wiley and Sons, 584 pp.

Joustra, D.Sj., 1971: Geulbeweging in de buitendeltas van de Waddenzee. Rijkswaterstaat, Studiedienst Hoorn, Studierapport WWK 71-14, 27 pp.

Kievits, S.S., 1994: The historical evolution of the 'Friesche Zeegat' inlet channel, and the adjacent barrier islands Ameland and Schiermonnikoog, Comp. Sed., 143 pp., int. rep.

Kumar, N. & Sanders, J.E., 1974: Inlet sequence: a vertical succession of sedimentary structures and textures created by the lateral migration of tidal inlets: Sedimentology, 21, 491-532.

Luck, G., 1975: Der Einfluss der Schutzwerke der Ostfriesischen Inseln auf die morphologische Vorgänge im bereich der Seegaten und ihrer Einzugsgebiete, Mitteilungen des Leichtweiss-Instituts für Wasserbau der Techn. Univ. Braunschweig, Heft 47, 122 pp.

Mars, S. & Cordia, P., 1901: Gids voor de Zuiderzee en de Noordelijke Zeegaten. Handleiding bij het gebruik van Seyffardt's Zuiderzeekaart. Bom, Amsterdam, 178-194.

Misdorp, R., Steyaert, F., Hallie, F. & De Ronde, J., 1990: Climate change, sea-level rise and morphological developments in the Dutch Wadden Sea, a marine wetland. In: Beukema, J.J., Wolff, W.J. & Brouns, J.J.W.M. (Eds): Expected Effects of Climatic Change on Marine Coastal Ecosystems: Kluwer, Dordrecht, 123-131.

Nanninga, M., 1985: The accuracy of echo sounding, description in a mathematical model. Nota WWKZ 85.H016, 92 pp.

Niemeyer, H.D., 1990: CM5 Morphodynamics of tidal inlets. In: Coastal Morphology, Stichting Postacademisch Onderwijs, Civ. Technieken en Bouwtechniek, Delft.

Nummedal, D. & Penland, S., 1981: Sediment dispersal in Norderneyer Seegat, West Germany. In: Nio, S.D., Schüttenhelm, R.T.E. & Van Weering, Tj.C.E. (Eds): Holocene Marine Sedimentation in the North Sea Basin. Int. Assoc. Sedimentol. Spec. Pub., 5, 197-210.

O'Brien, M.P., 1969: Equilibrium flow areas ofinlets on sandy coasts. J. Waterways and Harbors Div., ASCE, WWI: 43-52.

Oertel, G.F., 1977: Geomorphic cycles in ebb deltas and related patterns of shore erosion and accretion. J. Sedim. Petrol. 47, 1121-1131.

Oertel, G.F., 1988: Processes of sediment exchange between tidal inlets, ebb deltas and barrier islands. In Aubrey, D.G. & Weishar, L. (Eds): Hydrodynamics and sediment dynamics of tidal inlets, Lect. Notes Coastal Estuarine Studies, 29, 297-318.

Oost, A.P., 1995: Sedimentological implications of morphodynamic changes in the ebb-tidal delta, the inlet, and the drainage basin of the Zoutkamperlaag tidal inlet (Dutch Wadden Sea), induced by a sudden decrease in the tidal prism. In: Flemming, B.W. & Bartholomä, A. (Eds.): Tidal signatures in modern and ancient sediments. Int. Assoc. Sedimentol., Spec. Publ., 24: 101-119.

Oost, A.P. & De Boer, P.L., 1994: Sedimentology and development of barrier islands, ebb-tidal deltas, inlets and backbarrier areas of the Dutch Wadden Sea. Senckenb. Mar. 24, 1/6, 65-115.

Oost, A.P. & De Haas, H., 1992: Het Friesche Zeegat, morphologisch-sedimentologische veranderingen in de periode 1970-1987, een getijde inlet systeem uit evenwicht. Deel 1 and 2, rapport Kustgenese, 68 pp. and 22 Figs.

Oost, A.P. & De Haas, H., 1993: Het Friesche Zeegat, morphologisch-sedimentologische veranderingen in de periode 1927-1970, cyclische veranderingen in een tidal-inlet systeem, 96 pp. and 15 Figs.

Oost, A.P., De Haas, H., IJnsen, F., Van den Boogert, J.M. & De Boer, P.L., 1993: The 18.6 year nodal cycle and its impact on tidal sedimentation, Sedim. Geol., 87, 1-11.

Postema, P.W., 1956: Wijzigingen in geulenstelsel Waddenzee, omgeving Lauwerszee. Int. Rep. RWS Friesland, 10 pp.

Postma, H., 1961: Transport and accumulation of suspended matter in the Dutch Wadden Sea. Netherl. J. Sea Res. 1, 148-190.

Postma, H., 1967: Sediment transport and sedimentation in the estuarine environment. In: Lauff, G.A. (Ed.): Estuaries. Washington, D.C., A.A.A.S, 158-179.

Postma, H. & Dijkema, K.S., 1982: Hydrography of the Dutch Wadden Sea: Movements and properties of water and particulate matter. - Final report on 'Hydrography' of the Wadden Sea Working Group, Report 2, Balkema, Rotterdam, 75 pp.

Postma, J.T., 1982: Ontwikkeling van het Pinkegat voor zover van belang voor de kustontwikkeling van Ameland. Rijkswaterstaat, Meet- en Adviesdienst, nota 82-30, 16 pp.

Rakhorst, H.D., 1981: Stroming veroorzaakt door wind. Deel 2, RWS, District kust en zee, Adviesdienst Hoorn, Nota WWKZ-81.H028, 7 pp.

Reinson, G.E., 1984: Barrier-island and associated strand-plain systems. In: Walker, R.G., 1984: Facies Models, Geol. Ass. of Can, Geos. Rep. Ser. 1, 119-140.

Reinson, G.E., 1992: Transgressive barrier island and estuarine systems. In: Walker, R.G. & James, N.P., 1992: Facies Models: response to sea level change, Geol. Ass. of Can, Love Print. Serv. Ltd., Stittsville, Ontario, 179-194.

Schaik, 1979: Eventueel optredende verschillen bij het verrichten van lodingen. RWS Friesland, int. rep. F162, 4 pp.

Sha, L.P., 1989a: Cyclic morphological changes of the ebb-tidal delta, Texel Inlet, The Netherlands. Geol. Mijnbouw 68, 35-49.

Sha, L.P., 1989b: Variation in ebb-delta morphologies along the West and East Frisian Islands, the Netherlands and Germany. Marine Geol. 89, 11-28.

Sha, L.P., 1990: Sedimentological studies of the ebb-tidal deltas along the West Frisian Islands, the Netherlands. Phd Thesis, University of Utrecht, Geol. Ultraiectana 64, Utrecht. 160 pp.

Sha, L.P., 1992: Geological Research in the Ebb-Tidal Delta of 'Het Friesche Zeegat', Wadden Sea, The Netherlands, Rapport R.G.D., 20 pp.

Sha, L.P. & De Boer, P.L., 1991: Ebb-tidal delta deposits along the West Frisian Islands (The Netherlands): processes, facies architecture and preservation. In: Smith, D.G., Reinson, G.E., Zaitlin, B.A. & Rahmani, R.A. (Eds): Canadian Soc. Petrol. Geol. Mem. 16, 199-218.

Sha, L.P. & Van den Berg, J.H. 1993: Variation in ebb-tidal delta geometry along the coast of the Netherlands and the German Bight. J. Coastal Res., 9, 730-746.

Steijn, R.C., 1991: Some Considerations on Tidal Inlets. A literature survey on hydrodynamic and morphodynamic characteristics of tidal inlets with special attention to 'Het Friesche Zeegat', Literature Survey H 840.45., 109 pp.

Steijn, R.C. & Louters, T., 1992: Hydro- and morphodynamics of a mesotidal inlet in the Dutch Wadden Sea. Oceanology International '92 Conference, Vol. 2.

Steijn, R.C., Hartsuiker, G. & Van Banning, G.K.F.M., 1992: Morphodynamic behaviour of "Het Friesche Zeegat", numerical simulation of hydro- and morphodynamics of a mesotidal inlet in the Dutch Wadden Sea, 94 pp.

Stive, M.J.F. & Eysink, W.D., 1989: Voorspelling ontwikkeling kustlijn 1990-2020, fase 3, deelrapport 3.1: dynamisch model van het Nederlandse kustsysteem. Rapport H825, Waterloopkundig Laboratorium, Delft: 66 pp.

Thom, B.G., 1983: Transgressive and regressive stratigraphies of coastal sand barriers in Southeast Australia. Mar. Geol., 56, 137-158.

Tye, R.S., 1984: Geomorphic evolution and stratigraphy of Price and Capers Inlets, South Carolina. Sedimentology, 31, 655-674.

Van Dam, R., 1988: Een aanzet tot een 'Decision Supportsystem' ten behoeve van hydrografisch opnemen. Int. Rep. T.U. Delft, Afd. Geodesie, 33 pp.

Van de Kreeke, J., 1990: Can multiple tidal inlets be stable? Estuarine, Coastal and Shelf Science 30, 261-273.

Van den Berg, J.H., 1986: Aspects of sediment- and morphodynamics of subtidal deposits of the Oosterschelde (The Netherlands). Rijkswaterstaat Commun. 43, The Hague.

Van den Boogert, J.M., & Noordstra, P., 1988: Beschrijving van de Conlod programmatuur. Rijkswaterstaat, directie Friesland, rapport ANW 88.30, 34 pp.

Van den Boogert, J.M., 1991: Beschrijving van de Conlod interpolatiemethode. Rijkswaterstaat, directie Friesland, rapport ANW 91.056, 107 pp.

Van den Heuvel, P.B.J., 1993: From the ebb-tidal delta to the mainland shore of the Holwerderbalg, Dutch Wadden Sea; a facies description, a project at the University of Utrecht, Int. Report XGS/32, 103 pp.

Van der Spek, A.J., 1994: Large-scale evolution of Holocene tidal basins in the Netherlands. PhD-thesis, Univ, of Utrecht, 191 pp.

Van Elzelingen, 1904: Rapportage over de mogelijkheden tot afsluiting van de Lauwerszee, Int. Report RWS.

Van Kleef, A.W., 1991: Empirical relationships for tidal inlets, basins and deltas. Fac. Geogr. Sci., Univ. Utrecht, Rep. GEOPRO 1991.19: 53 pp.

Van Rhijn, A., 1851: Beschrijving behoorende bij de Hydrographische Kaart van het Friesche-Zeegat met een gedeelte der Friesche- en Groninger Wadden, trigonometrisch opgenomen en in plan gebragt door S.J. Keuchenius, 1831, herzien door A. van Rhijn, Algemeene Lands-drukkerij, The Hague, 30 pp.

Van Sijp, D., 1989: Correcties op gemeten eb- en vloedvolume bij de omrekening naar gemiddeld getij, in het Friese Zeegat. Rijkswaterstaat, directie Friesland, notitie ANW 89-02, 4 pp.

Van Staalduinen, C.J., 1977: Geologisch onderzoek van het Nederlandse Waddengebied. State Geol. Survey, Haarlem, 77 pp.

Van Straaten, L.M.J.U., 1954: Composition and structure of Recent marine sediments in the Netherlands. Leidse Geol. Mededel. 19: 1-110.

Van Straaten, L.M.J.U., 1964: De Bodem der Waddenzee. In: Het Waddenboek. Nederl. Geol. Ver., Thieme, Zutphen, 75-151.

Van Straaten, L.M.J.U. & Kuenen, Ph.H., 1958: Tidal action as cause of clay accumulation. J. Sedim. Petrol. 28, 406-413.

Van Veen, J., 1936: Onderzoekingen in de Hoofden in verband met de gesteldheid der Nederlandsche kust. Nieuwe verhandelingen van het Bataafsch Gen. der Proefondervindelijke Wijsbegeerte te Rotterdam, 2, 11, 252 pp.

Veenstra, H.J., 1982: Size, shape and origin of the sands of the East Frisian Islands (Norht Sea, Germany). Geologie en Mijnbouw, 61, 2, 141-146.

Veenstra, H.J. & Winkelmolen, A.M., 1976: Size, shape and density sorting around two barrier islands along the north coast of Holland. Geol. en Mijnb., 55: 87-104.

Visser, M.J., 1980: Neap-spring cycles reflected in Holocene subtidal large-scale bedform deposits: A preliminary note. Geology, 8, 543-546.

Vos, P.C., De Boer, P.L. & Misdorp, R., 1988: Sediment stabilisation by benthic diatoms. - In: De Boer, P.L., van Gelder, A. & Nio, S.D. (Eds): Tide-influenced sedimentary environments and facies. 511-526; Dordrecht (Reidel).

Walton, T.L. & Adams, W.D., 1976: Capacity of inlet bars to store sand. - Proc. 15th Conf. Coastal Eng., Am. Soc. Civ. Eng., Honolulu, 1919-1938.

Wolff, W.J., 1986: De Waddenzee: eigenschappen van een dynamisch kustgebied. Flevobericht 252, RIJP, 11-22.

APPENDIX 1: OVERVIEW OF DATA USED FOR THE COMPILATIONS 1927-1969

- For 1927, charts (Scale 1:25,000, unless mentioned otherwise): 11A/T21, Ameland, 1:12,000 (with aerial photographs!); 12F/M23, N-coast Schiermonnikoog; 9V-T/60, Oosterom (sounded 1926); 11A/M27, sea N of Ameland to Pinkegat; 13E/M85, Lauwers and Schild ebb-tidal delta, S.E. Lauwers, Schild, Eilanderbalg, Groninger and Uithuizer Wad; 12F/M25, Zoutkamperlaag and Plaatgat; 12F/M26 Plaatgat and Huibertgat; 12F/M27, Friesche wad to Pinkegat, 1:50,000 (sounded 1927). Numbers ARA, The Hague.

- For 1949, charts (Scale 1:25,000, unless mentioned otherwise): 12F/M43, Frisian Inlet and Zoutkamperlaag; 12F/M44, Frisian Inlet and Zoutkamperlaag; Friesche wad to Pinkegat; sea N of Ameland to Pinkegat (sounded 1949); 12F/M47, Frisian Inlet, Plaatgat (sounded 1950). Numbers ARA, The Hague.

- For 1958, charts (Scale 1:10,000): 55.167 (sounded 1955), (part) 58.276; (small part) 56.195 (sounded 1956), (part) 58.276; (part) 58.055; (part) 56.181 (sounded 1957), 58.304; 58.373; 59.255; (part) 58.276; 59.051 (sounded 1958), 59.404; 60.010; 60.007; 60.030 (sounded 1959), (part) 60.386; part 60.353 (sounded 1960). Numbers RWS Groningen. Jarkus 1958.

- For 1967, charts (Scale 1:10,000): 66.076 (sounded 1964), 66.029 (sounded 1965), 67.017; 67.018; 67.018; 67.033; 67.001 (sounded 1966), 67.388; 67.186; 68.026 en 68.027 (sounded 1967), 67.194 to 67.198 (sounded 1968), 69.388; 70.013 to 70.017 (sounded in 1969). Numbers RWS Groningen; Jarkus 1967.

- For 1970; soundings of RWS Groningen (200*200 m grid): Charts (Scale 1:10,000): 70.065; 70.421 to 70.426; 70.428; 70.429; 70.549; 70.586 to 70.588; 71.018; 71.238; 71.381 (sounded 1970), part 72.024; 72.029; 72.033; 72.034; 72.036 (sounded 1971, mainly Pinkegat, backbarrier). Numbers RWS Groningen. Jarkus 1970.

- For 1975; soundings of RWS Groningen (200*200 m grid): Charts (Scale 1:10,000): 75.526 to 75.542; Jarkus Ameland and Schiermonnikoog (sounded 1975). Jarkus 1975.

- For 1979: soundings of RWS Groningen (200*200 m grid): 78.443 to 78.446 (sounded 1978, Pinkegat backbarrier), 79.324 to 79.335; 79.382. Jarkus 1979.

- For 1982: Original sounding data RWS; Zoutkamperlaag sounded 1982; Pinkegat sounded 1981; Jarkus 1982.

- For 1987: Original sounding data RWS. Jarkus 1987.

- For 1991: Original sounding data RWS. Jarkus 1991.

CHAPTER 5

SEDIMENTOLOGICAL IMPLICATIONS OF MORPHODYNAMIC CHANGES IN THE EBB-TIDAL DELTA, THE INLET, AND THE DRAINAGE BASIN OF THE ZOUTKAMPERLAAG TIDAL INLET (DUTCH WADDEN SEA), INDUCED BY A SUDDEN DECREASE IN THE TIDAL PRISM

Spec. Publ. IAS, 24, 101-119

ABSTRACT

In 1969 the Lauwerszee, an embayment of the Wadden Sea, was dyked and the tidal prism of the Zoutkamperlaag tidal inlet decreased from $305{\cdot}10^6$ m^3 to $200{\cdot}10^6$ m^3. The sudden decrease in tidal prism caused significant sedimentary changes. In order to document and investigate these changes, high resolution computer studies were conducted and sedimentary cores were analysed.

Reduction of the tidal prism led to a strong erosion of the ebb-tidal delta, a downdrift shift and partial fill of the inlet gorge, the formation of a large recurved bar in the ebb-tidal delta, the partial fill of the main backbarrier channel, a shift of the eastern watershed, and channel fill east of that watershed. Only the inlet deposits and the abandoned channel fills have a significant preservation potential. Such fills reflect the sudden decrease in tidal prism and consist of clays, sands or alternations of both, deposited under relatively quiet conditions. The alternations reflect seasonal depositional cycles.

Sudden reductions of tidal prisms also occur under natural conditions, both in tectonically active and inactive lagoonal/estuarine settings. Apart from tectonics, several other factors can lead to a reduction of tidal prism and to the formation of abandoned channel fills. Several cases of abandoned channel fills from the fossil record can thus be explained in terms of local reduction of tidal prisms.

INTRODUCTION

Ebb-tidal deltas and backbarrier drainage basins have been investigated extensively both in recent environments (Bruun, 1978; FitzGerald & Nummedal, 1983; FitzGerald et al., 1984a; FitzGerald & Penland, 1987; Aubrey & Weishar, 1988; Sha, 1990a) and in the fossil records (cf. Hamberg, 1991; Sloan & Williams, 1991; Beets et al., 1979; Van der Spek & Beets, 1992). In recent systems, study periods generally span relatively short time intervals (decades). Consequently, apart from short-term cyclic and random variations in the behaviour of channels and shoals, these systems are commonly considered to be in dynamic equilibrium

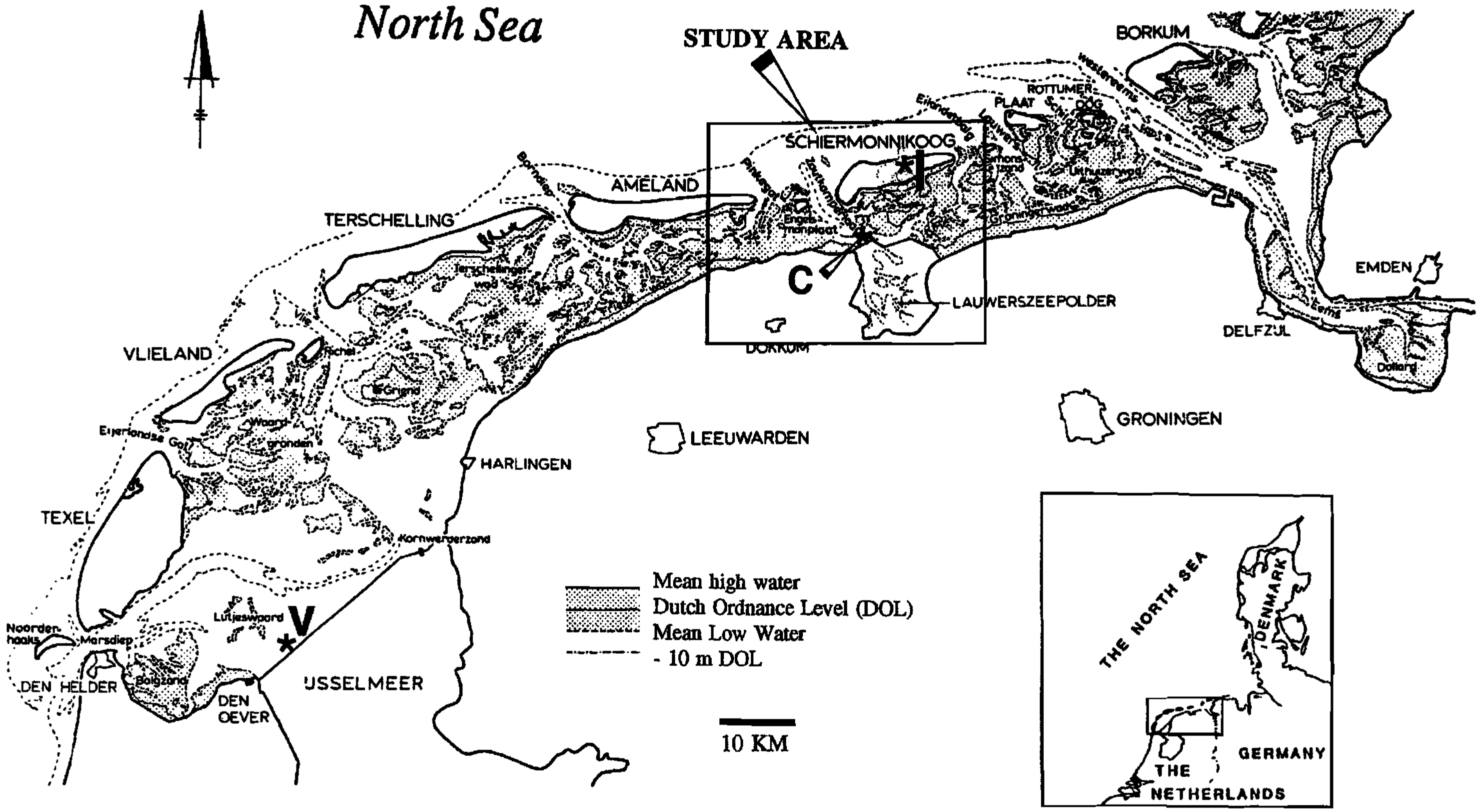

Figure 1: Overview of the area. Asterisk c refers to the location of the core in Fig. 7. Asterisk l indicates thje former position of the Lauwers Inlet in 1300 A.D. Asterisk v indicates the position of the former channel Vlieter (For details see text; after Steijn, 1991). The inset gives an overview of the North Sea.

with prevailing hydrodynamic regimes. Long-term variability is often overlooked also in the case of fossil ebb-tidal delta and backbarrier deposits. Indeed, these are commonly interpreted to represent a specific set of conditions that differs from that of prior and subsequent depositional events. For recent as well as fossil examples, therefore, the response time of barrier systems to changes in external conditions has to date generally not been considered.

In 1969 the tidal prism of the Zoutkamperlaag tidal inlet system was decreased by one-third (Van Sijp, 1989) due to the dyking of the Lauwerszee embayment (Figs 1 and 2). Subsequent sedimentological changes in the ebb-tidal delta, the inlet and the backbarrier drainage basin were analysed with high precision. The purpose of this paper is to discuss the geological relevance of these observations.

This paper is an elaboration of two studies for the Project COASTAL GENESIS (Oost & De Haas, 1992, 1993) coordinated by the Directorate General for Public Works and Water Management (further referred to as RWS = Rijkswaterstaat), National Institute for Coastal and Marine Management/RIKZ (further referred to as RIKZ).

SETTING

The Zoutkamperlaag tidal inlet is situated in the barrier island chain (Fig. 1) along the North Sea coast of The Netherlands, Germany and part of Denmark. The North Sea is a broad and shallow shelf sea on a passive continental margin. The tidal wave propagates from W to E, with a residual current towards the E. The tidal regime is semi-diurnal. In the area under discussion the tidal amplitude is 2.3 m and the tidal regime is therefore mesotidal (Hayes, 1979; Postma, 1982). The wind strength varies seasonally, average wind velocities reaching 15 $m.s^{-1}$ in winter, and 7 $m.s^{-1}$ in summer. Waves are highest in winter, reaching two m at 20 m water depth. The coast can be classified as a mixed-energy shoreline, influenced by both tides and waves (cf. Hayes, 1975). The islands have the typical 'drumstick' form, characteristic for mesotidal coasts (Hayes, 1979).

The Zoutkamperlaag and the Pinkegat Inlet systems (Figs 1 and 2) are situated between the barrier islands of Schiermonnikoog and Ameland. The inlets and their largely intertidal backbarrier flood-basins are separated by a supratidal shoal, which is situated on top of a massive, relatively stable clay body at -5 m with reference to Dutch Ordnance Level (DOL ≈ mean sea level) and its adjacent watershed (Sha, 1992; Oost & De Haas, 1992). In this paper attention is focused on the larger of the two systems, the Zoutkamperlaag Inlet system (Figs 1 and 2).

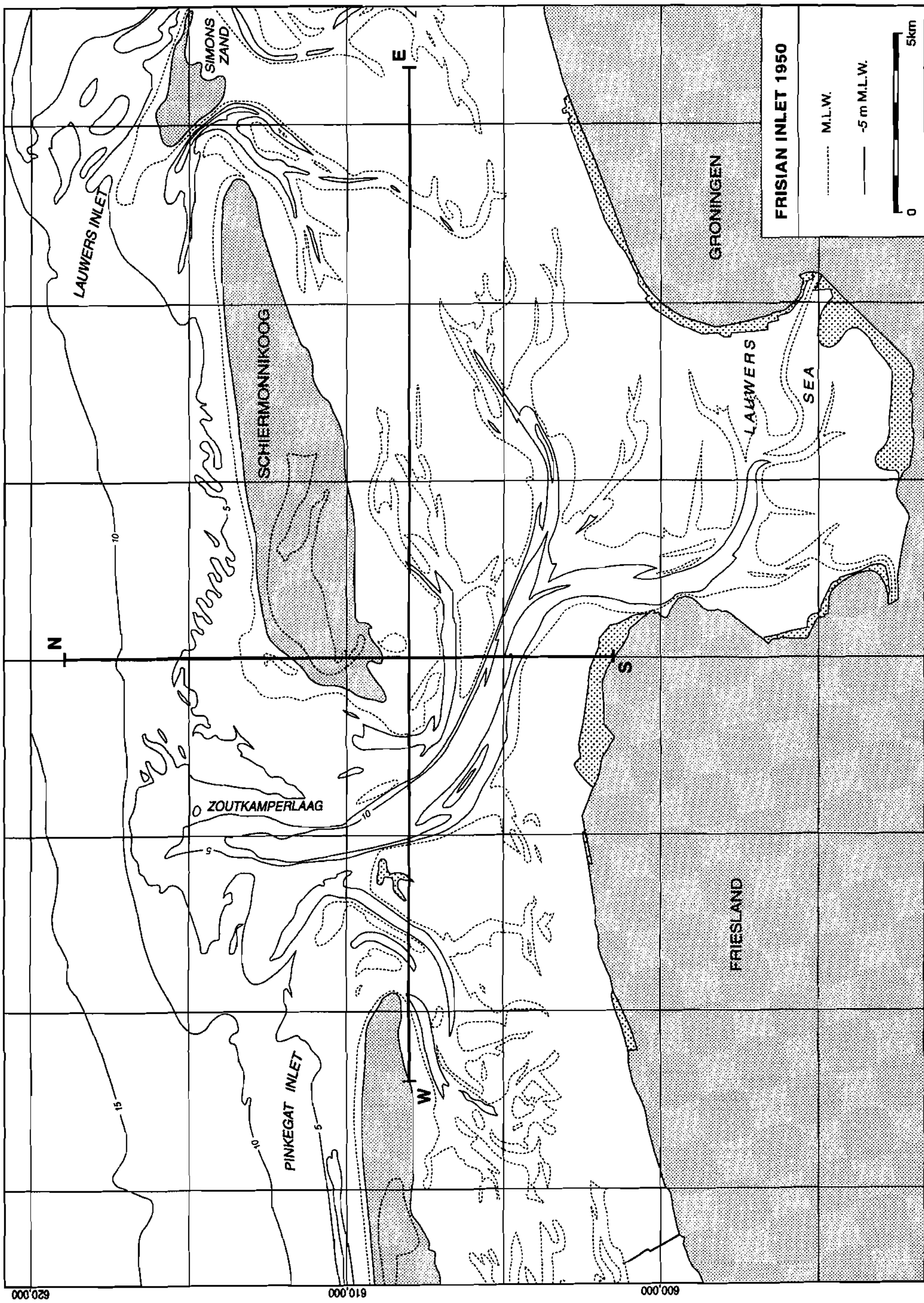

Figure 2: Detailed overview of the area showing the location of the sections W-E (Fig. 5) and N-S (Fig. 6).

METHODS

Depth measurements and soundings, densely spaced in the channels and on the inter- and supratidal flats, were made by the RWS during the years 1967, 1970, 1975, 1979, 1982, 1987, and 1991. The data were condensed into grid of 90*90 m areas. The CONLOD-program especially developed to handle sounding data (Van den Boogert & Noordstra, 1988; Van den Boogert, 1991) was used to fill empty cells with weighted interpolated values. Accuracy of both the original soundings and the interpolation technique is high. Thus, for the difference between two successive years the standard deviation (1σ) is 0.06 m for the interpolated data over larger areas (Oost & De Haas, 1992, 1993). From the filled gridcells, depth maps were generated which show the development of the area between Ameland and Schiermonnikoog over the last 20 years (Figs C4 to C10[1]; Oost & De Haas, 1992, 1993). Net sedimentation maps were produced by subtracting grid values of consecutive years. Figure D1 shows the erosion and sedimentation for 1970-1987, i.e., the period after the closure of the Lauwerszee. In addition, quantitative data of net erosion and sedimentation were calculated for several areas (Figs 3 and 4; Oost & De Haas, 1992, 1993). Owint to the slightly different sizes of the study areas, the net values presented here differ from those of Biegel & Hoekstra (1995); the general picture, however, is the same.

CHANGES IN THE ZOUTKAMPERLAAG SYSTEM

Owing to the closure of the Lauwerszee in 1969 (Figs 1, 2, C4 and C5) the tidal prism of the Zoutkamperlaag Inlet was reduced from $305*10^6$ m^3 to $200*10^6$ m^3 (Van Sijp, 1989). As a result, the system was no longer in equilibrium with existing hydrodynamic conditions and the ebb-tidal delta and the backbarrier drainage basin started to shift towards a new morphodynamic equilibrium (Figs C4 to C10 and 3 to 6; Biegel, 1991a&b; Oost & De Haas, 1992, 1993; Biegel & Hoekstra, 1995). The following changes were observed (for a full description, see Oost & De Haas, 1992, 1993).

The ebb-tidal delta

The ebb-tidal delta of the Zoutkamperlaag measures about 94 km^2. Total net erosion measured $26*10^6$ m^3 in the ebb-tidal delta for the period 1970-1987. Net erosion dominated during the periods 1970-1975, 1979-1982 and 1982-1987. Net erosion was mainly confined to the zones -4 to -12 m (1970-1975, 1982-1987) or -4 to -8.7 m DOL (1975-1979) (Fig. 3). From 1970 to 1991, the -10 m DOL contour on the North Sea side of the ebb-tidal delta retreated shorewards, but the -15 m DOL contour remained largely in place (Figs C4 to C10).

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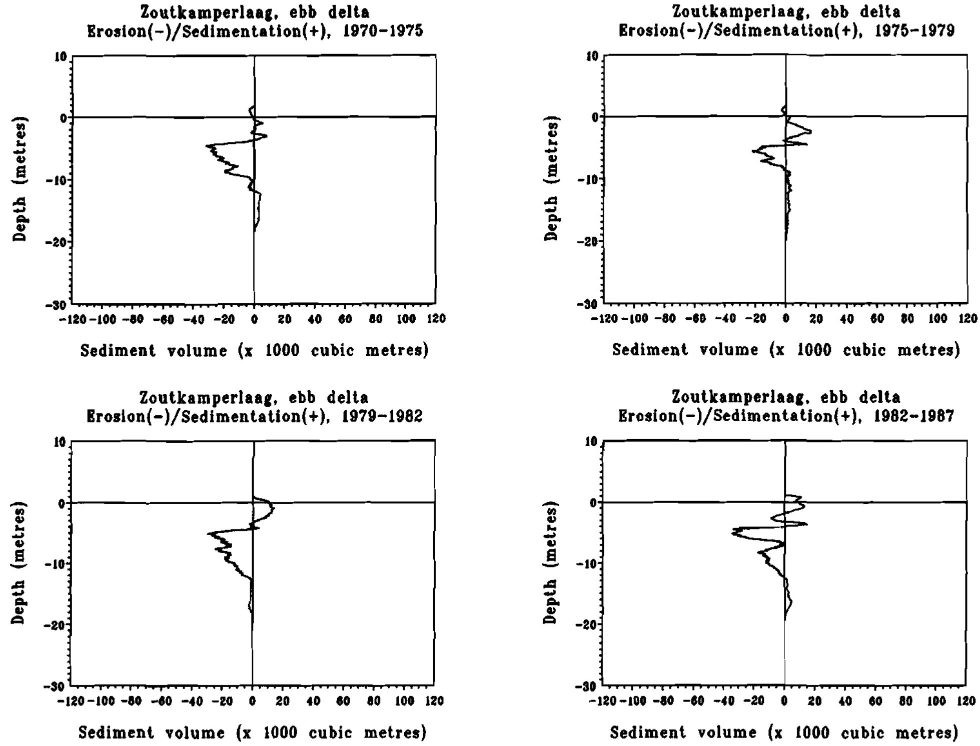
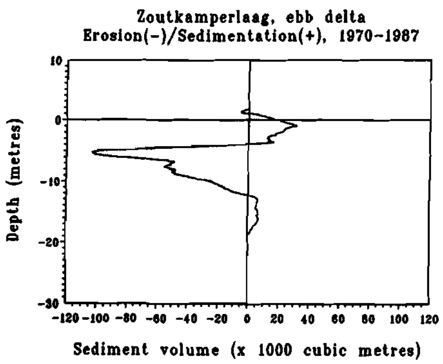
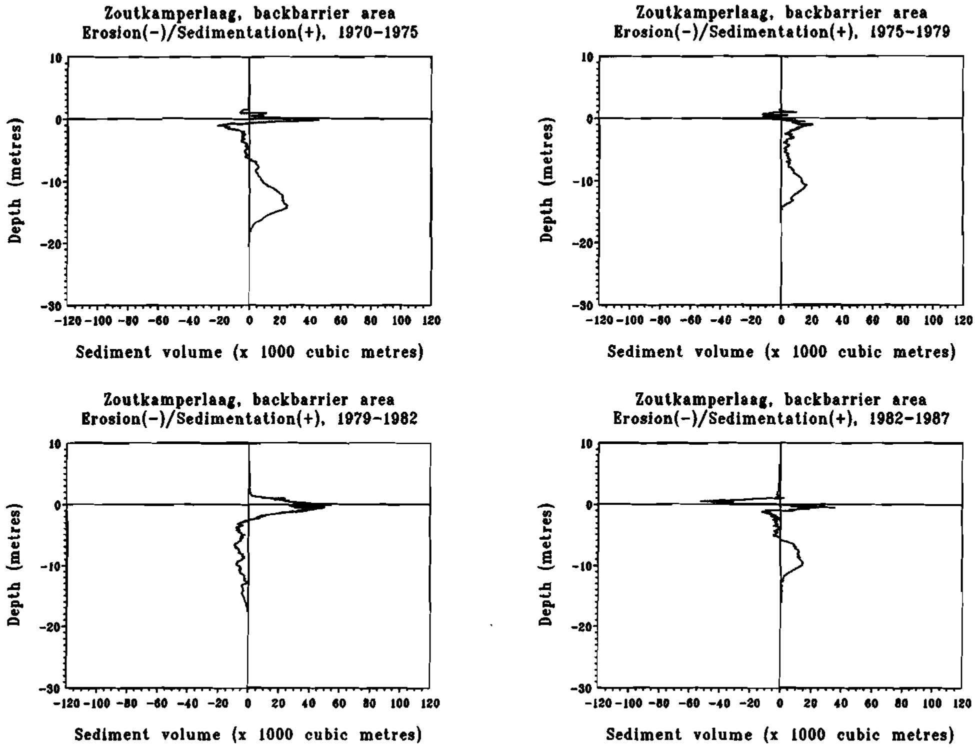
[1] All figures indicated with letters can be found in appendix A in the back of this thesis. For extensive captions the reader is referred to appendix 1: figure captions, at the end of the chapter.

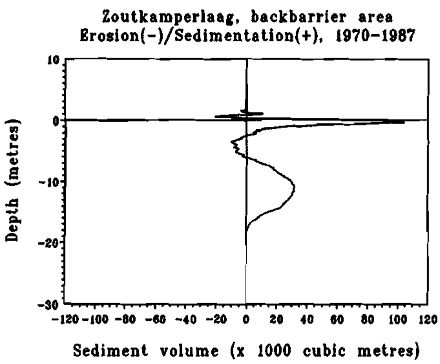




Zoutkamperlaag, ebb-tidal delta	
Period	Net Erosion (-)/ Sedimentation (+)
1970-1975	$-9.8 \cdot 10^6$ m^3
1975-1979	$+0.4 \cdot 10^6$ m^3
1979-1982	$-10.0 \cdot 10^6$ m^3
1982-1987	$-6.6 \cdot 10^6$ m^3
1970-1987	$-26.0 \cdot 10^6$ m^3

Figure 3. Sedimentation (positive) and erosion as a function of depth in 0.01 m increments in the ebb-tidal delta of the Zoutkamperlaag Inlet over the various time intervals, 1970-1975, 1975-1979, 1979-1982, 1982-1987, 1970-1987. Net values are given at the lower right. Note the similarity of trends in all periods. Strong erosion was mainly confined to the zone between -4 m to -12 m DOL, whereas sedimentation dominated above and below this zone.





Zoutkamperlaag, backbarrier area	
Period	Net Erosion (-)/ Sedimentation (+)
1970-1975	$+13.4 \cdot 10^6$ m^3
1975-1979	$+10.3 \cdot 10^6$ m^3
1979-1982	$+2.9 \cdot 10^6$ m^3
1982-1987	$+3.5 \cdot 10^6$ m^3
1970-1987	$+30.1 \cdot 10^6$ m^3

Figure 4. Sedimentation (positive) and erosion as a function of depth in 0.01 m increments in the drainage basin of the Zoutkamperlaag Inlet over the various time intervals 1970-1975, 1975-1979, 1979-1982, 1982-1987, 1970-1987. Net values are given at the lower right. Note the comparable pattern for most periods. Pronounced net sedimentation occurred in the deeper parts due to the fill of the main channel, except in the period 1979-1982, when the formation of drainage channels W of the watershed dominated the sedimentation pattern.

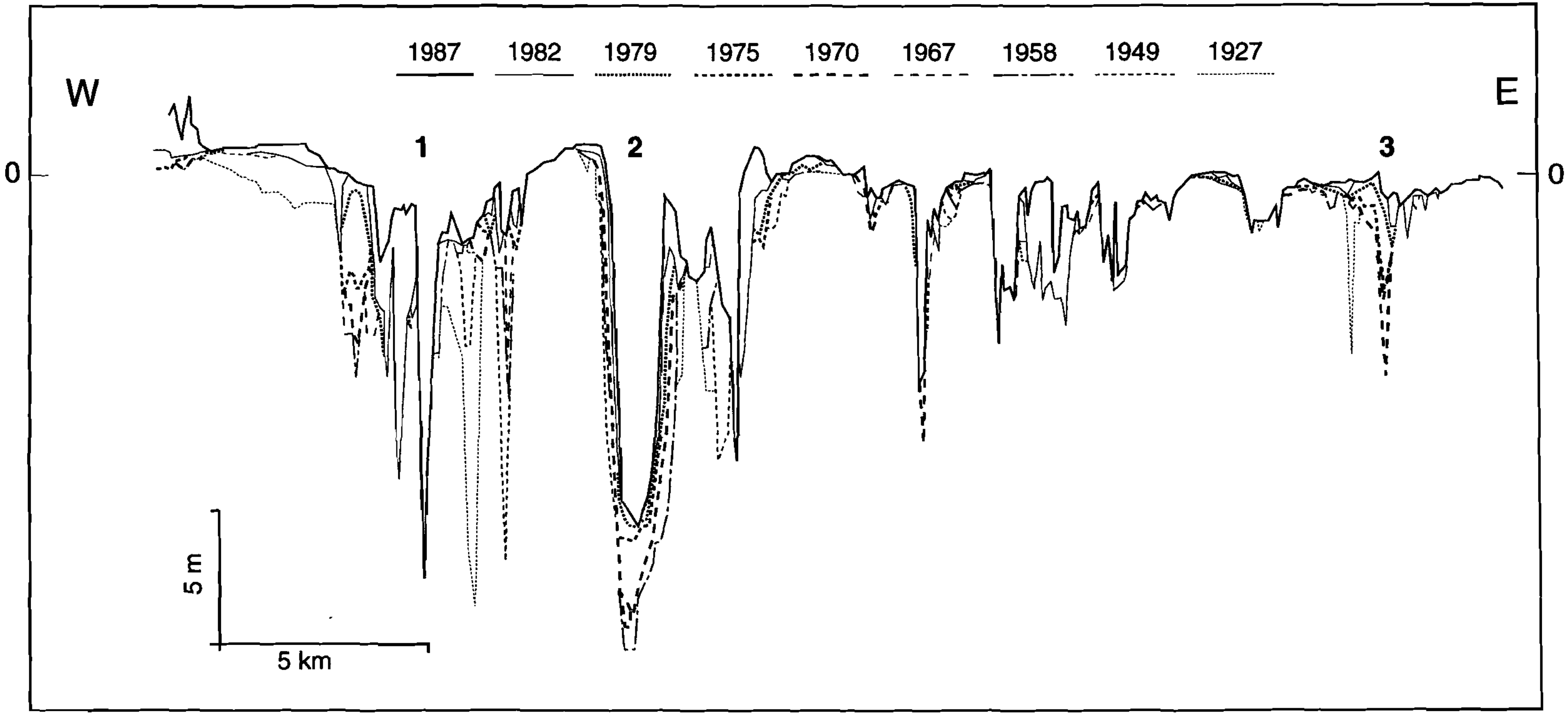


Figure 5: W-E bathymetric transect along line W-E (see Fig. 2), showing the profiles between 1927 and 1987 which have not been eroded by subsequent morphological changes. At the left O indicates the level of DOL (≈ mean sea level). The superimposed profiles illustrate the marked changes in the channel positions in the Pinkegat Inlet (1), the vertical fill of the inlet gorge of the Zoutkamperlaag Inlet (2) and the vertical infill of the inlet system E of Schiermonnikoog (3).

Recurved bar development

In 1970 a triangular swash platform of 26 km^2 was present NW of Schiermonnikoog (Fig. C5). The top of this platform was largely above -5 m DOL (the depth to which fair weather waves can erode the sediment; data RWS) consisting of fine to medium sand, with a median size of 100 to 180 μm (Winkelmolen & Veenstra, 1974). Especially from 1970 to 1982 the platform was eroded strongly (Figs C4 to C8). Southeastward sediment transport by waves and probably by tidal currents (Steijn et al., 1992) is inferred from grain characteristics, the migration of shoals and the refraction of waves (Veenstra & Winkelmolen, 1976).

A large recurved bar developed in the ebb-tidal delta at the northwestern side of Schiermonnikoog. More than $6*10^6$ m^3 sand was stored in this bar above -2 m DOL (Noordstra, 1989). The E-W oriented part of the bar partly evolved as a swash bar from a subtidal shoal (Fig. C4) that migrated towards the SE in the period 1970-1975 (Figs C5 and C6). For the other part it evolved from shoals NE of the channel of 1958 (Fig. C3). Aerial photographs show that structures observed in the foreshore and shallow shore-face region of Schiermonnikoog are largely wave-built. Judging from these structures, the observed development in the period 1970-1982, and the shape characteristics of the sediment, sand was probably transported partly parallel and partly perpendicular to the coast (Winkelmolen, 1969, 1982; Winkelmolen & Veenstra, 1980). During the period 1975-1982 this part of the bar rotated slightly counter-clockwise (Figs C6 to C8). After 1982 it migrated southwards (Figs C8 to C10).

The N-S oriented part of the bar developed after 1979 under the influence of waves and, to a lesser extent, probably also of tidal currents (Figs C7 to C10). Swash bars observed on aerial photographs indicate that the formation of this part of the bar was dominated by fair-weather wave action, with waves approaching mainly from the WNW and refracting around the developing bar (Oost & De Haas, 1992). Quiet conditions prevailed in the embayment behind the bar (Fig. C9), allowing sedimentation of muds alternating with sands. The southward migration of the bar over these deposits resulted in a coarsening upward sequence. By 1987 the recurved bar had become partly supratidal (Figs C9 & 6). Since 1989 the bar has become subject to strong erosion by tides and storms (J. Wiersma, pers. comm.). In 1991 (Fig. C10) the recurved bar was again intertidal and was breached in several places by small tidal channels down to -5 m DOL.

Channel rotation

In 1970 the Zoutkamperlaag Inlet had two outer channels, a flood-defined[2] channel to the W and an ebb-defined channel to the NW (Fig. C5). These channels were separated by a supratidal shoal (-2.5 to -5 m DOL) of about 2 km^2.

Both channels rotated during the period 1970-1982 in a clockwise mode (Figs C5 to C8). Furthermore, they shifted laterally over distances of up to 300 m and 1000 m respectively. Especially during the period 1970-1975, the axes of the flood-defined and ebb-defined chan-

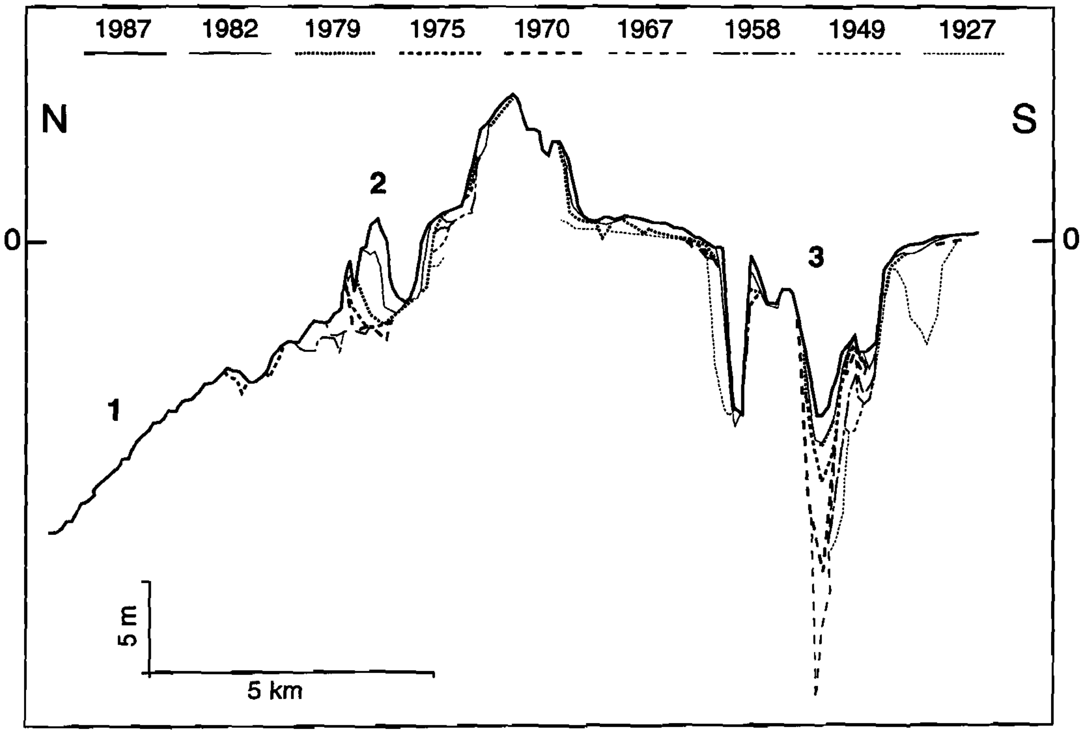


Figure 6. N-S bathymetric transect along line N-S (see Fig. 2), showing the profiles between 1927 and 1987 which have not been eroded by later morphological changes. O (at the left) indicates the level of DOL. Visible are the strong erosion in the ebb-tidal delta, leaving only the 1987 profile line (1), the vertical growth and coastward shift of the recurved bar (2) and the fill of the main channel of the Zoutkamperlaag Inlet system (3).

[2] This distinction is mainly based on the form: if the channel becomes shallower towards the backbarrier drainage basin it is considered flood-defined, whereas ebb-defined is used for the reverse morphology. Model calculations (Steijn et al., 1992) as well as direct observations (data RWS) indicate that ebb-defined and flood-defined channels roughly coincide with ebb-dominated, respectively flood-dominated channels. In the updrift flood-defined channels the flood-currents tend to be concentrated south of the channel axis.

nels experienced strong lateral shifts and a rotation over 10^o and 20^o respectively (Figs C5 and C6). Channel depths decreased by several meters especially that of the ebb-defined channel. At the same time the width of this channel increased, mainly due to strong erosion of the higher parts of the channel wall at the eastern side. From 1975 to 1982 (Figs C6 to C8) the ebb-defined channel split into two north-facing channels, the eastern (original) one being abandoned in the period 1982-1987 (Figs C8 and C9). Concurrently, the flood-defined channel reverted to a more E-W orientation.

Fill and shift of inlet gorge

Below -12 m DOL net deposition of $3.3*10^6$ m^3 of fine sand (Sha, 1992) took place, mainly in the gorge of the inlet (Figs 5 and D1). In 1970 the total length of the part of the gorge lying at -15 m DOL was 5.5 km, the maximum depth reaching -17 m DOL (Fig. C5). During the period 1970-1991 the seaward part of the main gorge shifted downdrift (to the east), and the length of the backbarrier channel region below -15 m DOL decreased to 3.5 km, mainly between 1970 and 1975 (Figs C5 to C7).

The drainage basin

The backbarrier drainage basin of the Zoutkamperlaag tidal system measures about 126 km^2. In the period 1970-1987 sedimentation in this area amounted to $30*10^6$ m^3. Sedimentation rates were highest soon after the reduction of the tidal prism (Fig. 4).

Fill of main channel

The sediments were deposited mainly in the main (1 km wide) backbarrier channel which originally connected the Lauwerszee embayment with the North Sea (Figs C4 to C9, and D1). As a result the cross-sectional area and the depth of the channel were reduced (Fig. 6). Sedimentation rates were high, reaching 1 $m.y^{-1}$ (Fig. 7). The sediments of the fill consist of fine sand or clay or alternations of both. The alternations have a rhythmic appearance; climbing ripple structures, lenticular bedding, loadcasts and bioturbation are common (Fig. 7). Net sedimentation dominated in the deeper parts of the channel (Fig. 4). The locus of sedimentation shifted upwards through time (Figs. 4, 6, and 7), as the channel became shallower (Oost et al., 1993). The deeper parts were dominated by erosion only in the period 1979-1982, when the zone above -3 m DOL experienced strong sedimentation (Fig. 4).

On the tidal flats adjacent to the main channel, erosion occurred in the period 1970-1975 and, to a lesser extent, also in the period 1975-1979 (Fig. 4). Winkelmolen & Veenstra (1974) showed convincingly that the sediments filling up the main channel in the backbarrier drainage basin shortly after closure of the Lauwerszee embayment were mainly derived from adjacent tidal flats. Erosion of these tidal flats did not continue after 1979 (Figs C7 to C9).

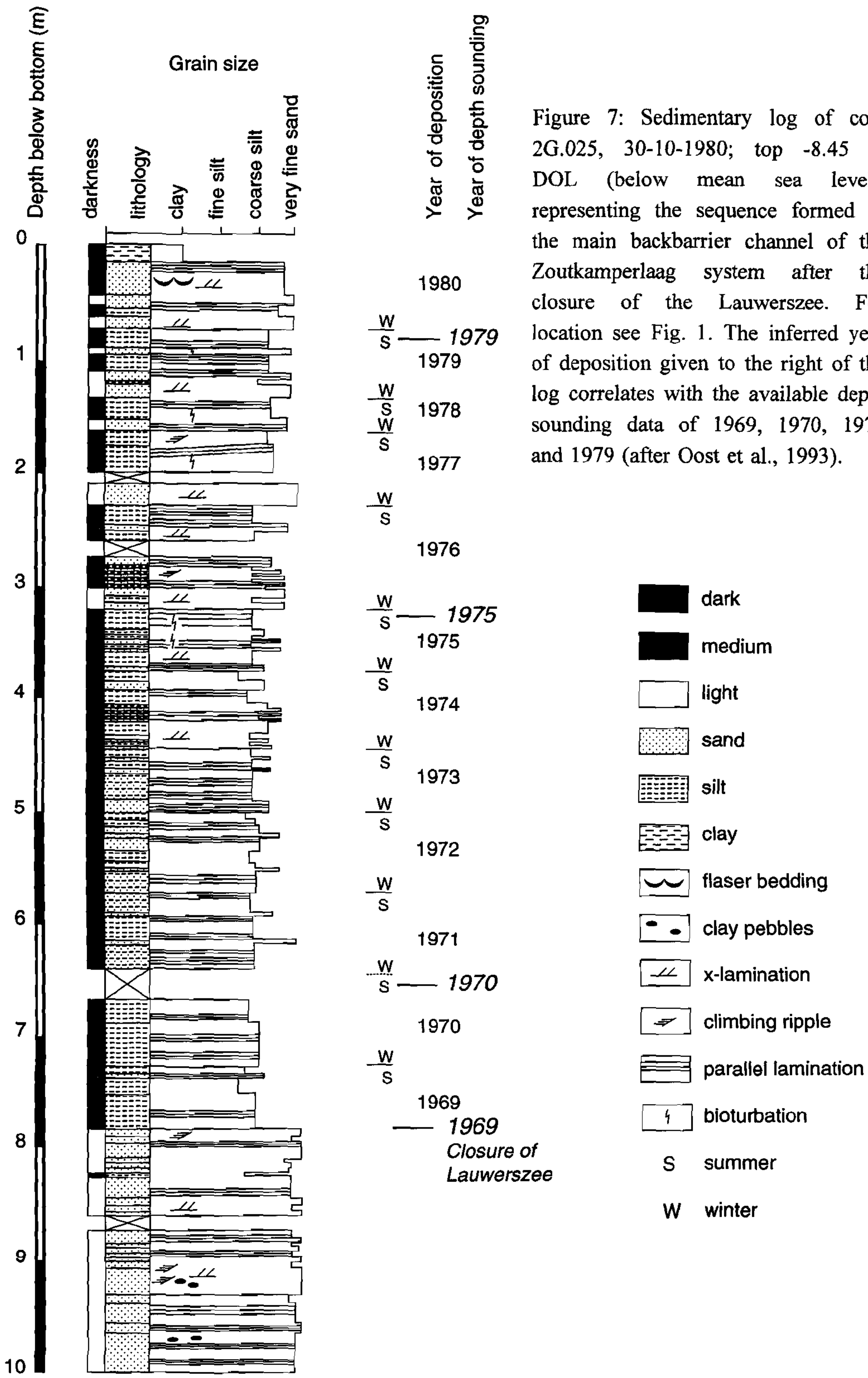


Figure 7: Sedimentary log of core 2G.025, 30-10-1980; top -8.45 m DOL (below mean sea level), representing the sequence formed in the main backbarrier channel of the Zoutkamperlaag system after the closure of the Lauwerszee. For location see Fig. 1. The inferred year of deposition given to the right of the log correlates with the available depth sounding data of 1969, 1970, 1975 and 1979 (after Oost et al., 1993).

Shift of watershed

Historical maps show that the position of the morphological watershed of Schiermonnikoog (S) was quite stable between 1891 (the inlet east of Schiermonnikoog had developed since 1848 at the cost of the more eastern Lauwers Inlet) and 1927. In the period 1927-1965 the watershed shifted 1 km to the west.

In the period 1970-1979 the southern part of the watershed of Schiermonnikoog slowly shifted towards the E (Figs C5 to C7; Van Parreeren, 1980). Soon after the closure of the Lauwerszee in 1969, and at least lasting up to 1983, the backbarrier channels near the watershed became deeper and wider (Postma & Van Parreeren, 1982; Postma, 1983). The main channel of the inlet system E of Schiermonnikoog retreated by rapid infilling in the period 1970-1979 (Figs C5 to C7). After 1979 the eastward migration rate of the southern part of the watershed accelerated, the total migration over the period 1970-1987 amounting to 3-4 km. West of the watershed new small channels were formed (Figs C7 to C9). The northern part of the watershed migrated only c. 1 km over the period 1970-1987.

INTERPRETATION OF THE CHANGES

The ebb-tidal delta

Ebb-tidal deltas are controlled by the tidal currents through the inlets, wave-action and the tidal wave along the coast (Dean & Walton, 1975; Oertel, 1975; Hayes, 1975, 1979, 1980; Walton & Adams, 1976; Hubbard et al., 1977; Nummedal et al., 1977; Nummedal & Fischer, 1978; FitzGerald et al., 1984a; Sha, 1989a-d, 1990a,b; Sha & De Boer, 1991; Steijn; 1991). Before the closure of the Lauwerszee in 1969, the system was in a state of dynamic equilibrium (Oost & De Haas, 1993).

The above data (Figs C4 to C10, and D1) show that the reduction of the tidal prism of the Zoutkamperlaag tidal inlet system resulted in a reduction in the size of the ebb-tidal delta. As the ebb currents decreased with the decreasing tidal prism, the relative importance of wave action increased, leading to net erosion of sand from the ebb-tidal delta (cf. Carter et al., 1987). Indeed, net erosion took place down to the storm erosion base of -12 to -13 m DOL (Fig. 3; Oost & De Haas, 1992).

Transport of sediment within the ebb-tidal delta occurs through the channel system, by migration of channels, by migration of subtidal shoals on the swash platform, and by migration of marginal breaker bars off Schiermonnikoog Island (cf. Hayes, 1979). The fine sand is transported both by saltation (Veenstra & Winkelmolen, 1976) and by bedload as is evident from the abundance of dunes and current ripples observed in box-cores, on sonar records and on aerial photographs (Oost & De Haas, 1992).

The ebb-tidal delta lost $26*10^6$ m^3 of sediment in the period 1970-1987. Sediment budgets for adjacent areas to the W show that no sediment was transported in that direction (Oost & De Haas, 1992). The area E of the ebb-tidal delta, i.e. the coast of Schiermonnikoog, also showed a small sediment surplus over that period. A net sediment transport must therefore have taken place mainly towards the N and/or the S.

Reineck & Singh (1972) and Aigner & Reineck (1982) have demonstrated that seaward sediment transport to below erosional wave base occurs in the German Bight during storms (see also Aigner, 1985). In this way sediment could also have been transferred from the ebb-tidal delta to the offshore region. However, in the present case net offshore transport cannot have been substantial as is indicated by the stable position of the -15 m DOL line at the outer rim of the delta (Figs C4 to C10). Moreover, cores taken at the seaward side of the ebb-tidal delta show that a thin veneer of recent North Sea sands, only a few dm thick, overlies older Holocene to Pleistocene deposits (Sha, 1992). It is therefore concluded that the larger part of the eroded sediment was transported through the inlet into the drainage basin of the Zoutkamperlaag Inlet by the flood current and wave action, resulting in the deposition of $30*10^6$ m^3 of sand (Fig. D1; Oost & De Haas, 1992).

Recurved bar development

Within the ebb-tidal delta, part of the sediment that was eroded from the swash platform contributed to the development of the large recurved bar northwest of Schiermonnikoog (Figs C9 and 6). The formation of such bars is a common phenomenon along the Frisian barrier island coast, having been documented, e.g. off Texel in 1908 (Sha, 1989a, 1990a), Schiermonnikoog in 1927, and Ameland in 1974 (data from nautical maps). Such bars mostly form when marginal (flood) channels are displaced towards the islands where they are abandoned, so that the updrift shoals (with reference to the direction of the residual eastward current) merge with the islands (cf. Moslow & Tye, 1985). Because of the decrease of tidal energy in such marginal channels, the relative influence of waves increases locally, resulting in coastward sediment transport and the formation of linear swash bars parallel to the coast of the island. In many cases there is a subsequent development of bars oriented parallel to the inlet. The nautical map of Schiermonnikoog of 1967 (Fig. C4), i.e. before the closure of the Lauwerzee in 1969, shows that the present bar started to develop in precisely this manner, i.e. after abandonment of a marginal flood channel. Such bars are commonly only a few km long. The unusually large size of the bar off Schiermonnikoog is explained by additional sediment supply from the eroding ebb-tidal delta (Figs C9 and 4; Oost & De Haas, 1992). After 1987 supply of sand from the ebb-tidal delta largely ceased, resulting in rapid erosion of the bar by storms and tides since 1989 (Wiersma, pers. comm.; Fig. C10).

Channel rotation

After closure of the Lauwerszee the outer tidal channels in the ebb-tidal delta of the Zoutkamperlaag Inlet rotated rapidly (Figs C4 to C10). Clockwise rotation and translation of outer channels in ebb-tidal deltas of the Wadden Sea are common phenomena (Joustra, 1971). Longshore sediment supply and interactions of hydrodynamics with morphology force the channels to migrate (Johnson, 1919; FitzGerald, 1988; Sha, 1989a; Chapter 4). Shoals shift towards the adjacent downdrift island, largely driven by wave action and tidal transport. Adjacent channels are forced to rotate and migrate until they become abandoned and are filled up (Sha, 1989a). The abandonment of channels in the downdrift part of the ebb-tidal delta stimulates the creation of new channels and perhaps also the migration and rotation of other channels within the system (Joustra, 1971). The rotations and translations are partly the result of a gradual lateral accretion, enhanced by the above processes. Moreover, massive sediment transport during storms (subtidal to supratidal) probably plays an important role in this respect, especially in triggering the end of a rotational channel movement (abandonment) and in the formation of new channels (Sha, 1989a; Sha & De Boer, 1991).

Historical maps of the area show that the channels of the West Frisian Islands commonly rotate at a rate of about $1^{o}.y^{-1}$ (Joustra, 1971; Sha, 1990a; Oost & De Haas, 1992). In the period 1970-1975, both the flood-defined channel and the ebb-defined channel rotated on average by 2^{o} and $4^{o}.y^{-1}$ respectively. This rapid rotation is ascribed to the sudden reduction in tidal prism and the resulting increase of relative wave influence, which tend to give the ebb-tidal delta a more downdrift, eastward orientation (cf. Sha & De Boer, 1991). Generally, the outer channels maintain their cross-sectional areas because the strength of the ebb current removes sand delivered by littoral drift at the western side of the channels. Moreover, any net deposition along the updrift side of the channels is compensated by erosion along the eastern side, as a result of which channel migration and rotation occur. In the present case, due to the decrease in the tidal prism, wave-generated sediment supply at the updrift side of the channels (Biegel, 1991a) could not be compensated by the weaker tidal currents (Oost & De Haas, 1992). As a result, deposition towards the SE increased, thereby filling up the deeper parts of the channel. This process is particularly well recorded in the ebb-defined channel, which was originally oriented NW-SE.

Along the eastern side of the channels erosion of the upper channel slopes was enhanced by the relative increase of wave energy. After the closure of the Lauwerszee the weaker ebb-current could no longer compensate the wave induced loss of sand loss, especially above -12 m DOL. Sedimentation along the western side, coupled with erosion of the eastern side of the outer channels, resulted in faster than normal rotation (Oost & De Haas, 1992).

Fill and shift of inlet gorge

From Figures 3 and 5 it is evident that net deposition within the ebb-tidal delta below -12 m DOL was largely confined to the main gorge, where c. $3*10^{6}$ m^{3} sand was deposited. The cross-sectional area, depth and width of an inlet are correlated with the tidal volume passing through it (O'Brien, 1969; Walther, 1972; Jarrett, 1976; Dieckmann et al., 1988; Gerritsen, 1990; Hume & Herdendorf, 1990; Niemeyer, 1990; Sha 1990a; Steijn, 1991; Biegel, 1991b; Flemming, 1991; Eysink & Biegel, 1992). Therefore, sedimentation in the main gorge documented by the decrease in length of the -15 m DOL region (Figs C5 to C9 and D1) is logically explained by the decrease in tidal prism and the related strong reduction of the ebb-tidal flow velocities in the main gorge (Biegel and Hoekstra, 1995; Fig. 5). The sediments in the inlet gorge are dominated by fine-grained sands (Oost & De Haas, 1992; Sha, 1992). The downdrift shift of the inlet gorge is attributed to the relative increase of wave forces (cf. Sha, 1990a, Sha & De Boer, 1991).

The drainage basin

Before the closure of the Lauwerszee a dynamic equilibrium existed in the drainage basin, with periods of net erosion alternating with periods of net sedimentation.

Owing to the reduction in tidal prism, a net deposition of $30*10^6$ m^3 occurred in the drainage basin in the period 1970-1987 (Fig. 4). As outlined above, the larger part of this sediment was derived from the ebb-tidal delta, from which $26*10^6$ m^3 were eroded. Erosion in the backbarrier drainage basin of the Pinkegat Inlet and in other, more westerly tidal flats was not substantial (De Boer, 1979; De Boer et al., 1991, Oost & De Haas, 1992). These areas can therefore be disregarded as possible sediment sources. The remaining $4*10^6$ m^3 of sediment probably comprise muds that were carried into the drainage basin in suspension.

Fill of main channel

The rapid vertical accretion of the main channel in the drainage basin (Fig. 6) was caused by the strong decrease of tidal current velocities after the closure, comparable to the development in the main gorge (Biegel 1991b; Biegel & Hoekstra, 1995). Initially these sediments were derived from the adjacent tidal flats (Winkelmolen & Veenstra, 1974), weaker tidal currents being unable to transport them back to the flats (Oost & De Haas, 1992).

The intercalated fine-grained material in the deposits show that current velocities in the channel were at times very low. Analysis of cores from these channel fills (Fig. 7; Oost et al., 1993) suggest that the sedimentation pattern was seasonal, comparable to the example given by Van den Berg (1981), with deposition of fines and bioturbation during the quiet summer season (cf. De Haas & Eisma, 1993) and deposition of sandy sediments in winter, when tides in the northern hemisphere are stronger and storms dominate the North Sea coast. The inferred timing of deposition in the main channel correlates well with available depth sounding data (Fig. 7; Oost & De Haas, 1992). Fine sediments are more prominent in the channel fill of the drainage basin than in the fill of the inlet gorge. This is in accordance with the decrease in current energy and the increasing proportion of fines towards the mainland coast (Van Straaten, 1954; Van Straaten & Kuenen, 1957, 1958; Postma, 1961, 1967; Flemming & Nyandwi, 1994; Oost & De Boer, 1994). Moreover, the formation of pellets of faeces and pseudo-faeces by filterfeeders (especially *Mytilus edulis* and *Cerastoderma edule*) and deposit feeders (for instance *Macoma baltica*) enhances sedimentation of clay and silt in the backbarrier drainage basin, especially during the summer (e.g. Flemming & Delafontaine, 1994; Chapter 6). It should be noted that the abrupt upward decrease in grainsize of sediments (around the level of 1969) in the abandoned channel fill is indicative of a sudden, rather than gradual drop in current velocity. In channels such grain-size 'jumps' can be considered to be indicative of sudden decreases in tidal flow strength.

The observation that the filling of the partly abandoned channel continued for more than a decade, i.e., at least from 1969 until 1987, indicates that the ability of the system to supply sediment largely determines the rate of deposition in such channels. Such constraints on sediment supply in tide-influenced sedimentary systems is also demonstrated by the sedimentary fill of the channel Vlieter. This channel was completely blocked at the landward side by a dyke enclosing the IJsselmeer embayment in 1932 (Fig. 1). Since it almost totally lost its sediment transport capacity after the closure, sedimentation rates in the channel have only been 6-7 $cm.y^{-1}$ (Eisma et al., 1987). This is in contrast to the high sedimentation rates of several $dm.y^{-1}$ in the Zoutkamperlaag Inlet main channel (Oost et al., 1993) and in the partly abandoned channel studied by Van den Berg (1981).

Shift of watershed

Before the closure, the phase difference of the tidal wave between the inlet gorges E and W of Schiermonnikoog was 20 minutes. Also, the resistance of the western channel system was less than that of the eastern channel system. These two effects resulted in an asymmetrical position of the tidal and morphological watersheds south of the eastern part of Schiermonnikoog (Van Parreeren, 1980; pers. com.), which was relatively stable.

The eastward migration of especially the southern part of the morphological watershed S of Schiermonnikoog, starting after the closure of the Lauwerszee embayment in 1969 (Postma, 1983), implies an increase of phase difference between the tidal waves at either side of the morphological watershed. Indeed, although some eastward migration had already occurred, direct measurements in 1980 showed that the tidal wave at the western side of the southern part of the morphological watershed was still up to several tens of minutes earlier than the wave at the eastern side (Van Parreeren, 1980).

The main channel in the backbarrier drainage basin of the Zoutkamperlaag Inlet system drained the Lauwerzee area before its closure (Fig. 2). After closure, the large width of the channel facilitated a faster propagation of both the vertical and horizontal tide towards the southern part of the morphological watershed (Van Parreeren, 1980; Postma & Van Parreeren, 1982). This was further enhanced by the formation of new channels and erosion west of the watershed (Figs C8 and C9; Van Parreeren, 1980; Postma, 1983; Oost & De Haas, 1992). Thus, the tidal (hydrological) watershed and, as a result, also the morphological watershed shifted towards the east. A comparable, but less pronounced, displacement can be observed in the northern part of the watershed, where the relatively wider inlet gorge probably enhanced the propagation of the tide.

The fact that the shift of the southern part of the watershed was rather slow in the period 1970-1979 (Figs C5 to C7) can be explained by analysing the development of the Eilanderbalg Inlet system SE of the barrier island Schiermonnikoog. A rapid shallowing and retreat of the main channel of this system occurred in the period 1970-1979 (Figs C5 to C7 and 5), because part of the drainage was taken over by the Zoutkamperlaag Inlet (Van Parreeren, 1980). The time required for morphological adjustment of the partly abandoned eastern inlet system to the new hydrodynamic conditions is the main reason for the initially slow shift of the watershed. At that time flood waters entering the system through the Zoutkamperlaag Inlet drained through the eastern inlet during the ebb phase (Van Parreeren, 1980). After filling up of the eastern inlet system new channels started to erode at the western side (Zoutkamperlaag) and the shift of the watershed accelerated (Figs C7 to C9). This suggests that, with some delay, the lateral shift of the watershed was indeed brought about by the closure of the Lauwerszee. Wind forced migration of the watershed as proposed by FitzGerald & Penland (1987) and Biegel & Hoekstra (1995) may have played an additional but certainly minor, role, as is apparent from the relatively stable position of the watershed before the closure of the Lauwerszee.

Historical evidence shows that eastward shifts of watersheds and barrier islands are associated with land reclamation schemes and related reductions in tidal prisms, e.g. Terschelling[1], since 1500; Ameland[2], since 1800; Schiermonnikoog[3], since at least 1550; Rottumeroog[2], mainly after 1650; Juist[1,2], 1650-1860; Norderney[1,3], since 1650; Baltrum[3], 1650-1860, Langenoog[3], since 1650; Spiekeroog[1], since 1650 and Wangerooge[3], since 1650 (Homeier & Luck, 1969; FitzGerald & Penland, 1987; FitzGerald, 1988; Ligtendag, 1990; Flemming & Davis, 1994; Oost, unpub. data). For the barrier islands, in the previous sentence marked with 1, the shifts result from the amalgamation of (estuarine) double inlets at the E-side. For those marked with 2 a faster route was formed by channel erosion at the W-side. For those marked with 3 the tidal prism at the updrift side of the watershed was reduced. It seems logical to conclude that in the latter cases the shifts of the watersheds were also brought about by the decreases in tidal prisms, comparable to the shift of the backbarrier system of the Zoutkamperlaag Inlet after the closure of the Lauwerszee. In addition, aeolian forcing may have played a minor role (FitzGerald & Penland, 1987).

Summarizing: a dynamical equilibrium was maintained in the ebb-tidal delta and the drainage basin of the Zoutkamperlaag Inlet, before the closure of the Lauwerszee. After the closure in 1969 the ebb-tidal delta of the Zoutkamperlaag Inlet was affected by erosion, $26*10^6$ m^3 of sand being transferred to the drainage basin in the period 1970-1987. Sediment was also concentrated by wave action in a large inter- to subtidal recurved bar situated NW of Schiermonnikoog. Offshore transport by storms was subordinate or absent.

After the closure a rapid rotation of the outer channels brought about by an increase in sedimentation at the western side and an increase in erosion at the eastern side of the channels, induced by the decrease in tidal prism and the relative increase of wave influence. A substantial deposition of sand in, and a reorientation of the main gorge occurred due to the decrease of the tidal prism.

In the drainage basin the change in tidal prism caused partial abandonment of the main channel resulting in rapid vertical accretion of sands and clays. Furthermore, the rapid propagation of the tides through the wide main channel in the drainage basin of the westerly tidal system (Zoutkamperlaag) was the main cause of an eastward shift of the tidal watershed. Thus the inlet system E of Schiermonnikoog became partially abandoned and, as a result, it was partly filled. Especially after the fill, the southern part of the morphological watershed between the two systems moved rapidly towards the E.

DISCUSSION

Preservation potential of the deposits

The deepest parts of a barrier system, i.e., the channels, the inlet gorge and the lower part of the ebb-tidal delta, have the highest preservation potential, as they suffer least from subsequent erosion by tidal currents and waves (Moslow & Tye, 1985; Sha, 1990a).

Ebb-tidal delta

The strong erosion of the ebb-tidal delta of the Zoutkamperlaag Inlet (Fig. 6) shows that such deposits, if located above the storm-erosion base, have little preservation potential in the wake of a sudden decrease of the tidal volume along a mixed energy coast. The only sediments that have some preservation potential are the deposits formed by the migrating outer channels and the fine sandy deposits in the inlet gorge (Fig. 5).

An accelerated rotation of the outer channels will result in a series of laterally accreting deposits inclined to the E. Unless a very detailed time control (annual basis) is available, these can not be distinguished from the normal deposits formed by eastward shifting inlet channels, as described by Sha (1992). Their preservation depends on the extent of erosion of the ebb-tidal delta. A comparison of the maps from 1967 to 1991 indicates that part of this lateral accretion surface is currently preserved in the ebb-tidal delta of the Zoutkamperlaag.

The sediments in the inlet gorge form nested channel deposits and a vertical channel fill in the non-migrating part (Fig. 5) and shallowing lateral accretion surfaces in the downdrift (eastward) migrating part, features which are commonly observed in the fossil record. Preservation potential of the deposits which fill the original inlet gorge is high, because the newly established inlet channel will be shallower than the preceeding one.

The intertidal to supratidal recurved bar NW of Schiermonnikoog (Fig. 6) has, notwithstanding its size, no significant preservation potential. Ever since sand supply from the ebb-tidal delta has declined, erosion has prevailed and the bar deposits are likely to disappear within the near future, quite comparable to other (recurved) bars in the Wadden Sea.

A general empirical relationship between sand volume (above the normal island shore profile) and tidal prism has been proposed for ebb-tidal deltas (Dean & Walton, 1975; Walton & Adams, 1976; Steijn, 1991):

$$V = 65.6*10^{-4}\ P^{1.23}$$

where V is the sand volume of the ebb-tidal delta and P is the mean tidal prism. At equilibrium, ebb-tidal delta volume decreases with increasing relative wave-influence (Dean, 1988). A lineair empirical relationship has been proposed, between the tidal prism and the cross-sectional area of the main channels in the backbarrier area of the Dutch Wadden Sea (Gerritsen & De Jong, 1985; Gerritsen, 1990). It follows that, depending on the length and geometry of the drainage basin and the size of the ebb-tidal delta, a decrease in tidal prism can generate a surplus of sediment. In general, the larger part of such a surplus would probably be reworked, to be either deposited in the offshore reach as storm deposits (Reineck & Singh, 1972; Aigner & Reineck, 1982; Aigner, 1985) or to be driven onshore and added to existing barrier islands (FitzGerald et al., 1984b; Sha, 1989a; Sha & De Boer, 1991; Flemming & Davis, 1994).

Backbarrier drainage basin

The considerable deposition of sands and fines in the main backbarrier channel of the Zoutkamperlaag tidal system following upon the decrease in tidal prism is comparable in scale to the infill of the gorge. The sediments form nested channel deposits (Fig. 6) reflected in the vertical channel-fill sequences (Fig. 7). Here too, preservation potential is high because the new, main backbarrier channel is shallower than its predecessor.

The observed migration of the watersheds in the drainage basin, caused by changes in tidal prism, will produce significant features in the geological record. The channel fill of the inlet system E of the barrier island Schiermonnikoog is now partly covered by intertidal watershed deposits (Figs C5 to C9, and 5). Although being somewhat smaller in scale, the fill of the channels will be comparable in sedimentary sequence to that observed in the Zoutkamperlaag main channel, since here too the tidal prism has decreased. Obviously these infilled channel deposits, being situated below the new watershed, have a high preservation potential. By contrast the watershed deposits, which were present in 1969, became incised and eroded by new channels (Figs C4 to C10, and D1).

In conclusion, only sediments deposited in the deeper parts of the inlet system have a high preservation potential. These comprise the sedimentary fills of the partially abandoned inlet, the main channel in the backbarrier basin and the main channel of the adjacent (E) inlet system. The fills are characterized by sand in the inlet and sand, mud or an alternation of both in the backbarrier channels.

Natural equivalents

The decrease of the tidal prism in the Zoutkamperlaag Inlet by one third has resulted in rapid and substantial changes in the morphology of both the outer delta and the drainage basin. The change was artificially induced and one might wonder whether equivalent processes occur under natural conditions. In this respect two different settings must be considered: tectonically active and tectonically inactive areas. The changes considered here are so rapid (i.e. occuring within years) that absolute sea-level fluctuations can be ignored.

Tectonically active areas

About 22% of all barrier coasts are located along collision zones (Pilkey et al., 1988). Studies show that average uplift rates in such areas range from several $mm.y^{-1}$ (Hails, 1983; Bishop, 1991; Carter et al., 1991; Fortuin & De Smet, 1991; Flint et al., 1991) to >0.1 $m.y^{-1}$ (Prince et al., 1974). Between 1964 and 1982 a rapid uplift of 0.2-0.3 $m.y^{-1}$ occurred in Portage, Alaska (Atwater et al., 1991). During the 1964 Alaska earthquake sudden localized uplift of the shoreline amounted to 10 m and, during the Yakutat Bay earthquake of 1899, the shoreline rose by 15 m (Bolt, 1978). If such uplifts were to occur in backbarrier settings, they would result in a reduction in tidal prism and hence in morphological changes comparable to those reported here.

Tectonically related upwarping of barrier sequences has been documented in the Silurian of S-W Ireland (Sloan & Williams, 1991), in the Pleistocene of the Naracoorte region, SE South Australia (Hails, 1983), in Holocene deposits of Hainan Island (15-20 m) (Wang, 1986) (all related to volcanic activity), in Holocene marine deposits of the Spencer Gulf, South Australia (2.5 m, seismotectonics), in the Penholoway shoreline deposits (6 m) and in the Wicomico barrier deposits (20 m), the last two in northern Florida (Hails, 1983). Glacio-eustatic uplift of coastal lagoon deposits has been reported, from, amongst other places, Finland (Eronen, 1983), Norway (Hafsten, 1983) and Scotland (Sissons, 1983). The exact morphodynamic responses of depositional environments affected by such changes strongly depend on the orientation of the tectonic lines with reference to the barrier orientation and the exact nature of the tectonic regime. The above examples demonstrate that preservation of coastal deposits, that have experienced rapid uplift, has occurred in the past. However, their long-term preservation potential will probably be low.

In the case of an extensive tectonic regime one might expect that the tidal volume will increase by subsidence of a backbarrier region or estuary (cf. Nelson, 1988; Sloan & Williams, 1991). This would result in an increase of the tidal prism and hence in deepening of channels and enlargement of the ebb-tidal delta.

Tectonically inactive areas

About 78% of all barrier island chains are located along passive margins (Pilkey et al., 1988), mostly in an inactive tectonic setting. In such areas a sudden decrease of the tidal volume passing through a channel can be produced by several mechanisms:

1) A washover channel changes into a tidal inlet. This can occur when a washover channel is excavated during a hurricane or severe storm (e.g. Hayes, 1967; Mehta & Brooks, 1973; Penland & Suter, 1986). Another possibility is that after a shift of the watershed (e.g. by lateral migration of the barrier island), a washover that was originally situated adjacent to a

watershed, begins to drain a part of the backbarrier area, thereby developing into an active tidal inlet.

2) Sudden infilling of part of a backbarrier drainage basin and/or blocking of the channel occurs. This can happen when massive sedimentation occurs by ice rafted deposits (cf. Dionne, 1987), aeolian transport (Schoorl, 1973), shore-parallel transport of sediment (for instance spit-growth) (cf. Moslow & Tye, 1985; Jenings & Smyth, 1988; Massa & Sanders, 1989), storm-related sedimentation (cf. Penland & Suter, 1986), mass transport deposition (cf. McKnight, 1969) or volcanic deposition (Sloan & Williams, 1991). Total infilling, however, must be considered a rare event because very large amounts of sediment have to be deposited in a relatively short time.

3) An inlet system takes over part of the drainage of an adjacent inlet system (Bruun, 1978; Moslow & Tye, 1985). This depends on the relative phase difference of the tidal wave between two competitive inlet systems. Historical data suggest that between c. 1350-1550 such an inlet take-over occurred for the drainage area of the Lauwerszee embayment (Fig. 1; Reitsma, 1991; Oost & Dijkema, 1993). Before about 1350 the precursor of the Zoutkamperlaag Inlet (west of Schiermonnikoog) was probably small and had no connection with the embayment, the latter being drained by the, at that time, large, Lauwers Inlet east of Schiermonnikoog (cf. Bosch & Vos, 1992). The watershed of Schiermonnikoog was probably breached in the period 1350-1450 (Sha, 1992, Oost & Dijkema, 1993). Around 1500 the Lauwerszee was drained by both the Lauwers Inlet and the Zoutkamperlaag Inlet. Thereafter the drainage was taken over rather quickly by the Zoutkamperlaag Inlet. As a result of the increase in tidal prism the cross-sectional area of the channels increased. Judicial information shows that in 1556 the Lauwers Inlet, had lost its connection with the Lauwerszee (Formsma 1954; 1958). The Lauwers Inlet, was largely abandoned, its channels and inlet decreased in cross-sectional area and the whole system shifted eastwards.

4) On a smaller scale, channels of the same inlet system may take over each others drainage areas, locally leading to partial or total abandonment of channels. This can happen both in the ebb-tidal delta (Moslow & Tye, 1985; Sha, 1989a; Sha & De Boer, 1991) and in the backbarrier area (Moslow & Tye, 1985). The filling up can be rather gradual, depending on the rate of take-over (Moslow & Tye, 1985). Several examples of channel abandonment can be observed in the Pinkegat Inlet, which develops cyclically from an initially single-inlet system into a multiple-inlet system, reverting back to the former within a period of 20-54 years (Figs C4 to C9, and 5) (Oost & De Haas, 1993). Owing to the considerable changes in the channel patterns of this system, many of the backbarrier channels and outer channels become abandoned in the course of each cycle (Chapter 3; see for instance Figs C6 to C9).

5) Sudden changes in tidal amplitude generated by changes elsewhere in the system (mostly changes in resonance and interference of tidal waves). The closure of the IJsselmeer (Fig. 1), for example, is known to have influenced the tidal patterns up to the inlet W of Ameland, i.e., over a distance of 45 km (Sha, 1990a; De Boer et al., 1991; Van Parreeren, pers. comm.). Before the closure, the size of the IJsselmeer basin was such that the reflected outgoing tidal wave interfered with the incoming tidal wave, resulting in a standing wave of low tidal amplitude (Klok & Schalkers, 1980). After closure, the tidal amplitude increased considerably (approximately 20% near Texel Inlet) and the catchment area of Texel Inlet within the remaining part of the Wadden Sea became larger at the cost of other inlets (Sha, 1990a).

Fossil examples

From the rock record several sequences have been described that are comparable to the sedimentary sequences formed in the Zoutkamperlaag Inlet.

Hardeberga Fm., Lower Cambrian of southern Sweden

Above a basal fluvio-deltaic deposit, the Lower Cambrian Hardeberga Formation in Scania (Hamberg, 1991) consists of three 30-50 m thick, vertically stacked barrier-related sequences. Within the backbarrier deposits of these sequences, abandoned channel fills, 1-4 m thick, are present on top of sediments deposited in an active channel setting. The abandoned channel fills comprise between 2 and 10 stacked annual microsequences of 0.2-0.8 m thick, large-scale, cross-bedded sandstone which pass upward into bioturbated and/or tidally-bedded sandstones. The channels were thought to have been abandoned by continued migration of the watersheds downwind and downdrift (cf. Fitzgerald & Penland, 1987; Hamberg, 1991). It is difficult to understand how such processes can lead to thick abandoned channel fill deposits, indicative of sedimentation over periods of up to 10 years, because abandoned channels near the watershed are mostly shallow and fill up within one year. Also, the observed bimodality and the occurrence of the large-scale cross-bedding is at variance with sedimentation near watersheds. A more likely explanation is that larger channels of the same or another inlet system took over each others drainage areas, which locally resulted in (partial) abandonment of channels (cf. Moslow & Tye, 1985).

Rocky Mountains molasse, Mesozoic of North America

Partly or completely mud-filled channels are quite common in Mesozoic sequences of the Rocky Mountains molasse (Rahmani, 1986). Brownridge and Moslow (1991) describe Lower Cretaceous estuarine channel fills of the Glauconitic Member. These tidal channel fills consist of a single fining-upward succession or multiple successions separated by erosive contacts. Part of these successions comprise massive mudstones (with thin sandstone beds), typically 1-5 m thick and occasionally up to 15 m thick, which abruptly or gradationally overlie either heterolithic point-bar deposits or crossbedded sandstones. This mudstone facies is interpreted to represent abandoned channel and channel margin deposits (Brownridge & Moslow, 1991). Especially the mudstones displaying sharp basal contacts and the thicker sequences of massive mudstone may have formed by abrupt channel and main channel abandonment, respectively.

Tegelen Formation, Pleistocene of Holland/Belgium

A part of the Tegelen Formation has been interpreted as micro- to mesotidal. Within the sediments (Turnhout Member) several fine-grained channel deposits up to 2 m thick have been observed. These have been interpreted as abandoned channel fills (Kasse, 1988). In addition, thicker channel fills reaching 10 m in thickness and consisting of muds alternating with sands were observed which probably also represent abandoned channel fills. In all cases the change from sands to clays is abrupt, a feature associated with a sudden drop in tidal amplitude (Kasse, 1988), which resulted in reduction of the tidal prism and allowed (partial) abandonment of the channels, as observed after the artificial reduction of the Zoutkamperlaag tidal prism.

CONCLUSIONS

The artificial closure of a part of the backbarrier drainage basin of the Zoutkamperlaag tidal inlet caused a sudden decrease in the tidal prism. As a result, the morphology of the inlet system was no longer in equilibrium with the new hydraulic conditions. After the closure the system adapted towards a new equilibrium. This was mainly achieved by erosion of sediment in the ebb-tidal delta and deposition in the inlet gorge and in the main channel of the backbarrier drainage basin. Also, a redistribution of sediments within the ebb-tidal delta occurred, thereby enhancing the formation of a large recurved bar. In addition, the outer channels and main channel shifted downdrift. Moreover, the observed retreat and fill of the main channel of the inlet system E of Schiermonnikoog and the subsequent shift of the watershed to the E were brought about by the reduction in the tidal prism. Of all the sedimentary features that changed, only the vertical inlet and channel fills, consisting of sands, clays or an alternation of both (abandoned channel fill), have a significant preservation potential.

Rapid changes in tidal prisms can also occur in natural settings, both in tectonically active and tectonically passive regions. In the former the size and depth of the backbarrier area can change by sudden vertical movements, the exact nature of which depends on the tectonic regime and the orientation of the coastline with reference to the tectonic lines. In tectonically passive regions a sudden decrease in tidal prism can be generated by the formation of new channels which take over the drainage of others, by sudden changes in tidal amplitude or by sudden massive sedimentation in the backbarrier area. Abandoned channel deposits in the rock record suggest that sudden decreases of tidal prisms flowing through tidal channels are not unusual. The fine-grained deposits of channel fills can thus be used as indicators of sudden changes in the hydrodynamic regime.

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REFERENCES

Aigner, T., 1985: Storm depositional systems. Lect. Notes in Earth Sci. 3. Springer., 174 pp.

Aigner, T. & Reineck, H.E., 1982: Proximality trends in modern storm sands from the Helgoländer Bight (North Sea) and their implications for basin analysis. Senckenbergiana marit., 14, 183-215.

Atwater, B.F., Obermeier, S.F., Tabaczynski, D.A., Pond, E.C. & J.R. Martin., 1991: Holocene shaking in southern coastal washington. Eos, Transactions, Am. Geophys. Un., 72, 44, 313.

Aubrey, D.G. & Weishar, L. (Eds), 1988: Hydrodynamics and Sediment Dynamics of Tidal Inlets, Preface. Lecture Notes on Coastal and Estuarine Studies, 29, Springer-Verlag, N.Y., 456 pp.

Beets, D.J., Van der Valk, L. & Stive, M.J.F., 1979: Holocene evolution of the coast of Holland. Mar. Geol., 103, 423-443.

Biegel, E.J., 1991a: De ontwikkelingen van de ebgetijde delta en het kombergingsgebied van het Friese Zeegat in relatie tot de sluiting van de Lauwerszee. Fac. of Geogr. Sci., Univ. of Utrecht, GEOPRO, 1991.07, 57 pp.

Biegel, E.J., 1991b: Equilibrium relations in the ebb tidal delta, inlet and backbarrier area of the Frisian inlet system. Fac. of Geogr. Sci., Univ. of Utrecht, GEOPRO 1991.028, 79 pp.

Biegel, E. & Hoekstra, P., 1995: Morphological response characteristics of the Zoutkamperlaag Inlet, Frisian Inlet, The Netherlands to a sudden basin area reduction. Spec. Publ. IAS, 24, 85-99.

Bishop, D.G., 1991: High-level marine terraces in western and southern New Zealand: indicators of the tectonic tempo of an active continental margin. In: Macdonald, D.I.M. (Ed.): Sedimentation, Tectonics and Eustasy, Sea-level Changes at Active Margins, I.A.S. Spec. Publ., 12, 69-78.

Bolt, B.A., 1978: Earthquake Hazards, EOS, 59, 11, 946-962.

Bosch, A. & Vos, P.C., 1992: Paleogeografische reconstructie van het lauwersmeergebied, concept. State Geol. Survey-Project 40009: 22 pp.

Brownridge, S. & Moslow, T.F., 1991: Tidal estuary and marine facies of the Glauconitic Member, Drayton Valley, central Alberta. In: Smith, D.G., Reinson, G.E., Zaitlin, B.A. & Rahmani, R.A. (Eds): Clastic Tidal Sedimentology, Can. Soc. Petr. Geol. Memoir, 16, Friesen, Altona, Manitoba, Canada, 107-122.

Bruun, P., 1978: Stability of tidal inlets, theory and engineering. Developments in geotechnical engineering 23, Elsevier, Amsterdam, 510 pp.

Carter, R.M., Abbott, S.T., Fulthorpe, C.S., Haywick, D.W. & Henderson, R.A., 1991: Application of global sea-level and sequence-stratigraphic models in Southern Hemisphere Neogene strata from New Zealand. In: Macdonald, D.I.M. (Ed.): Sedimentation, Tectonics and Eustasy, Sea-level Changes at Active Margins, Int. Assoc. Sedim. Spec. Publ., 12, 41-65.

Carter, R.W.G., Johnston, T.W., Mckenna, J. & Orford, J.D., 1987: Sea-level, sediment supply and coastal changes: Examples from the coast of Ireland. In: Carter, R.W.G. & Devoy, R.J.N. (Eds): The hydrodynamic and sedimentary Consequences of Sea-level Change, Progr. Oceanogr. 18, 79-101.

De Boer, M., 1979: Morfologisch onderzoek Ameland. Verslag van het onderzoek op het Amelander wantij in 1973. Rijkswaterstaat, Directie waterhuishouding en waterbeweging, district kust en zee, studiedienst Hoorn, Nota WWKZ-79.H005, 67 pp.

De Boer, M., Kool, G. & Lieshout, M.F., 1991: Erosie en sedimentatie in de binnendelta van het Zeegat van Ameland 1926-1984, deelonderzoek no.4. Directie Noord-Holland, Rapport ANVX-91.H202, 42 pp.

De Haas, H. & Eisma, D., 1993: Suspended sediment transport in the Dollard Estuary. Neth. J. Sea Res., 31, 37-42.

Dean, R.G., 1988: Sediment interaction at modified coastal inlets: processes and policies. In: Aubrey, D.G. & Weishar, L. (Eds): Hydrodynamics and sediment dynamics of tidal inlets, Lect. Notes on Coastal and Estuarine Studies, 29, Springer-Verlag, N.Y., 412-439.

Dean, R.G. & Walton, T.L., 1975: Sediment transport processes in the vicinity of inlets with special reference to sand trapping. In: Cronin, L.E. (Ed): Estuarine Research, Vol. II, Geology and Engineering, Acad. Press, N.Y., 129-150.

Dieckmann, R., Osterthun, M. & Partenscky, H.W., 1988: A comparison between German and North American tidal inlets. Proc. 21st Coast. Eng. Conf., ASCE, 3, 199, 2681-2691.

Dionne, J.C., 1987: Characteristic features of modern tidal flats in cold regions, In: De Boer, P.L., Van Gelder, A. & Nio, S.D. (Eds): Tide-Influenced Sedimentary Environments and Facies, Reidel Publ. Comp, Dordrecht, Holland, 301-332.

Eisma, D., Berger, G.W., Chen, W.Y. & Shen, J.,, 1987: Pb-210 as a tracer for sediment transport and deposition in the Dutch-German Wadden Sea. In: Proceedings KNGMG Symp. Coastal Lowlands, Geology and Geotechnology, Van der Linden, W.J.M. & Cloetingh, S.A.P.L., Kaasschieter, J.P.K., Van de Graaff, W.J.E., Vandenberghe, J. & Van der Gun, J.A.M. (Eds): Kluwer Publ. Dordrecht, The Netherlands, 237-253.

Eronen, M., 1983: Late Weichselian and Holocene shore displacement in Finland. In: Smith, D.E. & Dawson, A.G. (Eds): Shorelines and isostasy, Spec. Pub. Inst. Brit. Geogr., 16., 183-207.

Eysink, W.D. & Biegel, E.J., 1992: Impact of sea-level rise on the morphology of the Wadden Sea in the scope of its ecological function. Investigations on empirical morphological relations. Rijkswaterstaat, Dienst Getijdewateren, ISOS*2 Project, Phase 2, 73 pp.

FitzGerald, D.M., 1988: Shoreline erosional-depositional processes associated with tidal inlets. In: Aubrey, D.G. & Weishar, L. (Eds): Hydrodynamics and sediment dynamics of tidal inlets, Lect. Notes on Coastal and Estuarine Studies, 29, Springer-Verlag, N.Y., 186-225.

FitzGerald, D.M. & Nummedal d., 1983: Response characteristics of an ebb-dominated tidal inlet channel. J. Sedim. Petrol., 53, 3, 833-845.

FitzGerald, D.M. & Penland, S., 1987: Backbarrier dynamics of the east Friesian Islands. J. Sedim. Petrol., 57, 4, 746-754.

FitzGerald, D.M., Penland, S. & Nummedal, D., 1984a: Control of barrier island shape by inlet sediment bypassing: east frisian islands, West Germany, Mar. Geol., 355-376.

FitzGerald, D.M., Penland, S. & Nummedal d., 1984b: Changes in tidal inlet geometry due to the backbarrier filling: east frisian islands, West Germany. Shore & Beach, 52, 3-8.

Flemming, B.W., 1991: Holozäne Entwicklung, Morphologie und fazielle Gliederung der Ostfrieschischen Insel Spiekeroog (südliche Nordsee). Senckenberg-am-Meer, Bericht 91/3, 51 pp.

Flemming, B.W. & Davis, R.A., 1994: Holocene evolution, morphodynamics and sedimentology of the Spiekeroog barrier island system (southern North Sea). Senckenbergiana marit..

Flemming, B.W. & Delafontaine, 1994: seasonal rates and trends of biodeposition in a juvenile mussel bed of the east frisian wadden sea (southern North Sea). Neth. J. Aquat. Ecol., 28.

Flemming, B.W. & Nyandwi, N., 1994: Evidence for man-induced export of fine-grained sediments from backbarrier tidal flats along the southern North Sea coast. Neth. J. Aquat. Res., 28.

Flint, S., Turner, P. & Jolly, E.J., 1991: Depositional architecture of Quaternary fan-delta deposits of the Andean fore-arc: relative sea-level changes as a response to aseismic ridge subduction. In: Macdonald, D.I.M. (Eds): Sedimentation, Tectonics and Eustasy, Sea-level Changes at Active Margins, Int. Assoc. Sedim. Spec. Publ., 12, 91-104.

Formsma, W.J., 1954: Een reis naar Schiermonnikoog in 1556. Gron. Volksalmanak 1954, 65-71.

Formsma, W.J., 1958: De grens tussen de provincies Groningen en Friesland in de Wadden. Groningse Volksalmanak 1958, 27-42.

Fortuin, A.R. & De Smet, M.E.M., 1991: Rates and magnitudes of late Cenozoic vertical movements in the Indonesian Banda Arc and the distinction of eustatic effects. In: Macdonald, D.I.M. (Ed.): Sedimentation, Tectonics and Eustasy, Sea-level Changes at Active Margins, Int. Assoc. Sedim. Spec. Publ., 12, 79-90.

Gerritsen, F., 1990: Morphological stability of inlets and channels of the western Wadden Sea. Rijkswaterstaat, Dienst Getijdewateren, Nota GWAO-90.019, 86 pp.

Gerritsen, F. & De Jong, H., 1985: Stabiliteit van doorstroomprofielen in het Waddengebied. Rijkswaterstaat, Adviesdienst Vlissingen, Nota WWKZ-84.V016, 53 pp.

Hails, J.R., 1983: Coastal processes, relict shorelines and changes in sea level on selected mid- and low-latitude coasts. In: Smith, D.E. & Dawson, A.G. (Eds): Shorelines and isostasy, Spec. Pub. Inst. Brit. Geogr. 16., 29-51.

Hafsten, U., 1983: Shore-level changes in South Norway during the last 13000 years, traced by biostratigraphical methods and radiometric datings. Norsk Geogr. Tidsskrift, 37, 2, 63-79.

Hamberg, L., 1991: Tidal and seasonal cycles in a Lower Cambrian shallow marine sandstone (Hardeberga Fm.), Scania, Southern Sweden. In: Smith, D.G., Reinson, G.E., Zaitlin, B.A. and Rahmani, R.A. (Eds): Clastic Tidal Sedimentology, Can. Soc. Petr. Geol. Memoir, 16, Friesen, Altona, Manitoba, Canada, 255-273.

Hayes, M.O., 1967: Hurricanes as geological agents, South Texas coast. Am. Assoc. Pet. Geol. Bull., 51, 937-942.

Hayes, M.O., 1975: Morphology of sand accumulations in estuaries. In: Cronin, L.E. (Ed.): Estuarine Research vol. 2, Geology and Engineering, Acad. Press., N.Y., 3-22.

Hayes, M.O., 1979: Barrier Island morphology as a function of tidal and wave regime. In: Leatherman, S (Ed.): Proceedings of the coastal symposium on Barrier islands, Academic Press, N.Y., 1-28.

Hayes, M.O., 1980: General morphology and sediment patterns in tidal inlets. Sed. Geol. 26, 139-156.

Homeier, H. & Luck, G., 1969: Das Historische Kartenwerk 1:50,000 der Niedersächsischen Wasserwirtschaftsverwaltung als Ergebnis historisch-topographischer Untersuchungen und Grundlage zur kausalen Deutung der Hydrovorgänge im Küstengebiet. Göttingen.

Hubbard, D.K., Barwis, J.H. & Nummedal, D.,, 1977: Sediment transport in four South Carolina inlets. ASCE, Proc. Coastal Sediments '77, 582-601.

Hume, T.M. & Herdendorf, C.E., 1990: Morphologic and hydraulic characteristics of tidal inlets on a headland dominated low littoral drift coast, Northeastern New Zealand. J. of Coast. Research, spec. issue, Proc. Skagen Symp., 527-563.

Jarrett, J.T., 1976: Tidal prism-inlet area relationships. GITI, Rep. 3, US Army Eng. Waterways Exp. Stat. Vicksburg.

Jenings, S. & Smyth, C., 1988: The application of coastal and sea-level research to problems of coastal management in southern England. Abstracts of Int. Symp. on "Theoretical and applied aspects of coastal and shelf evolution, past and future", jointly with the inaugural meeting of IGCP Project 274 "Quaternary coastal evolution; case studies, models and regional patterns", Amsterdam, the Netherlands, 19- 24 sept. 1988, 51-52.

Johnson, J.W., 1919: Shore processes and Shoreline Development. John Wiley and Sons, 584 p.

Joustra, D.Sj., 1971: Geulbeweging in de buitendeltas van de Waddenzee. Rijkswaterstaat, studiedienst Hoorn, Studierapport WWK 71-14, 27 pp.

Kasse, C., 1988: Early-Pleistocene tidal and fluviatile Environments in the southern Netherlands and northern Belgium. Thesis, V.U. Amsterdam, Free Univ. Press, 190 pp.

Klok, B. & Schalkers, K.M., 1980: De veranderingen in de Waddenzee ten gevolge van de afsluiting van de Zuiderzee. Rijkswaterstaat, Report 78.H238, 13 pp.

Ligtendag, W.A., 1990: Van IJzer tot Jade, Een reconstructie van de zuidelijke Noordzeekust in de jaren 1600 en 1750. Report Rijkswaterstaat, 55 pp.

Massa, A.A. & Sanders, J.E., 1989: Quantitative determinations of unusually rapid rates of spit growth and longshore currents on a moderate-wave-energy microtidal coast (Fire Island NY). Abstr. Geol. Soc. of America, 21, 6, 174.

Mcknight, D.G., 1969: A recent, possibly catastrophic burial in a marine molluscan community. New Zealand J. Mar. Freshwater Res. 3, 177-179.

Mehta, A.J. & Brooks, H.K., 1973: 'Mosquito Lagoon Barrier Beach Study'. Shore and Beach, 41, 2, 27-34.

Moslow, T.F. & Tye, R.S., 1985: Recognition and characterization of Holocene tidal inlet sequences. Mar. Geol., 63, 129-151.

Nelson, A.R., 1988: Prospects for identification of small coseismic? rises in relative sea level during the Holocene along tectonically active coasts: preliminary studies in South-Central Oregon, U.S.A. Abstracts of Int. Symp. on "Theoretical and applied aspects of coastal and shelf evolution, past and future", jointly with the inaugural meeting of IGCP Project 274 "Quaternary coastal evolution", Amsterdam, the Netherlands, 19- 24 sept. 1988, 84-87.

Niemeyer, H.D., 1990: CM5 Morphodynamics of tidal inlets. In: Coastal Morphology Stichting Postacad. Ond. Civ. Techn. en Bouwtechn., Delft.

Noordstra, P., 1989: Zandhaak Schiermonnikoog, volumebepaling. Rijkswaterstaat, directie Friesland, notitie ANW 89.51., 4 pp.

Nummedal, D. & Fischer, I.A., 1978: Process-response models for depositional shorelines: The German and Georgia bights. ASCE, Proc. 16th Coastal Engineering Conf., 1215-1231.

Nummedal, D., Oertel, G.F., Hubbard, D.K. & Hine, A.C., 1977: Tidal inlet variability Cape Hatteras to Cape Canaveral. ASCE, Proc. Coastal Sediment'77, 543-562.

O'Brien, M.P., 1969: Equilibrium flow areas of inlets on sandy coasts. Journal of the Waterways and Harbours Div., ASCE, Nr. WWI, 43-52.

Oertel, G.F., 1975: Ebb-tidal deltas of Georgia estuaries. In: Cronin, L.E. (Ed.): Estuarine Research vol. 2, Geology and Engineering, Acad. Press, N.Y., 267-276.

Oost, A.P. & De Haas, H., 1992: Het Friesche Zeegat, morfologisch-sedimentologische veranderingen in de periode 1970-1987, een getijde inlet systeem uit evenwicht, deel 1 & 2. Kustgenese werkgroep 1, 68 pp.

Oost, A.P. & De Haas, H., 1993: Het Friesche Zeegat, morfologisch-sedimentologische veranderingen in de periode 1927-1970, deel 1 & 2. Kustgenese werkgroep 1, 94 pp.

Oost, A.P. & Dijkema, K.S., 1993: Effecten van bodemdaling door gaswinning in de Waddenzee. IBN-Report 025, 133 pp.

Oost, A.P., De Haas, H., IJnsen, F., Van den Boogert, J.M. & De Boer, P.L., 1993: The 18.6 year nodal cycle and its impact on tidal sedimentation. Sedim. Geol. 87, 1-11.

Oost, A.P. & De Boer, P.L., 1994: Sedimentology and development of barrier islands, ebb-tidal deltas, inlets and backbarrier areas of the Dutch Wadden Sea. Senckenb. marit. 24, 1/6, 65-115.

Penland, S. & Suter, J.R., 1986: Mapping the morphodynamic signature of hurricane impacts in the Northern Gulf of Mexico. Sediments down-under; 12th International Sedimentological Congress, Int. Assoc. Sedim., Canberra, Aug. 24-30 1986, 239.

Pilkey, O.R., Henderson, V., Keysworth, A., 1988: Controls of barrier island chain morphology and distribution: a global view. Abstracts of Int. symp. on "Theoretical and applied aspects of coastal and shelf evolution, past and future", jointly with the inaugural meeting of IGCP Project 274 "Quaternary coastal evolution; case studies, models and regional patterns", Amsterdam, the Netherl. 19-24 sept. 1988, 103-106.

Postma, H., 1961: Transport & accumulation of suspended matter in the dutch wadden sea. Neth. J. Sea res., 1, 148-190.

Postma, H., 1967: Sediment transport & sedimentation in the estuarine environment. In: Lauff, G.A. (Ed.): Estuaries, Washington D.C., 158-179.

Postma, H. (Ed.), 1982: Hydrography of the Wadden Sea: Movements and properties of water and particulate matter, Final report on Hydrography of the Wadden Sea Working Group, 2, 75 pp.

Postma, J.T., 1983: De Groningerbalg als verbinding tussen het Oort en de Eilanderbalg respectievelijk het Hornhuizerwad. Rijkswaterstaat, directie Groningen, Meet- en Adviesdienst Delfzijl, nota 83-28, 17 pp.

Postma, J.T. & Van Parreeren, D. V., 1982: Onderzoek naar de bevaarbaarheid van de wantijen Engelsmanplaat, Lutjewad en Hornhuizerwad. Rijkswaterstaat, dir. Groningen, nota 82-7, 26 pp.

Prince, R.A., Resig, J.M., Kulm, L.D. & Moore, T.C. jr., 1974: Uplifted turbidite basins on the seaward wall of the Peru Trench. Geology, 2, 607-611.

Rahmani, R.A., 1986: Mesozoic mud-filled channels; abandoned fluvial channels or active river estuaries? Reservoir, 13, 1-2.

Reineck, H.E. & Singh, I.B., 1972: Genesis of laminated sand and graded rhythmites in storm-sand layers of shelf mud. Sedimentology, 18, 123-128.

Reitsma, D.T., 1991: Historie oostelijk Waddengebied. Waddenbulletin, 26, 176-178.

Schoorl, H., 1973: Zeshonderd jaar water en land, Bijdrage tot de historische Geo- en Hydrografie van de Kop van Noord-Holland in de periode ± 1150-1750. Verhandelingen van het K.N.A.G., 2, Wolters-Noordhoff NV, Groningen, the Netherlands, 534 pp.

Sha, L.P., 1989a: Cyclic morphological changes of the ebb-tidal, Texel Inlet, The Netherlands. Geol. Mijnb., 68, 35-49.

Sha, L.P., 1989b: Sand transport patterns in the ebb-tidal delta off Texel Inlet, Wadden Sea, The Netherlands. Marine Geol., 86, 137-154.

Sha, L.P., 1989c: Variation in ebb-delta morphologies along the West and East Frisian Islands, The Netherlands and Germany. Marine Geol., 89: 11-28.

Sha, L.P., 1989d: Holocene-Pleistocene interface and three-dimensional geometry of the ebb-delta complex, Texel Inlet, The Netherlands. Marine Geol., 89, 207-228.

Sha, L.P., 1990a: Sedimentological studies of the ebb-tidal deltas along the West Frisian Islands, the Netherlands. Ph.D. Thesis University of Utrecht, Geologica Ultraiectina no. 64, Utrecht, 160 pp.

Sha, L.P., 1990b: Surface sediment characteristics in the ebb-tidal delta of Texel Inlet, The Netherlands. Sediment. Geol. 68: 125-141.

Sha, L.P., 1992: Geological Research in the Ebb-Tidal Delta of 'Het Friesche Zeegat, The Netherlands, R.G.D.-Project 40010, 20 pp.

Sha, L.P. & De Boer, P.L., 1991: Ebb-tidal delta deposits along the West Frisian Islands, The Netherlands: processes, facies architecture and preservation. In: Smith, D.G., Reinson, G.E., Zaitlin, B.A. & Rahmani, R.A. (Eds): Clastic tidal sedimentology, Can. Soc. Petrol. Geol. Mem., 16, 199-218.

Sissons, J.B., 1983: Shorelines and isostasy in Scotland, In: Smith, D.E. & Dawson, A.G. (Eds): Shorelines and isostasy, Spec. Pub. Inst. Brit. Geogr., 16., 209-225.

Sloan, R.J. & Williams, B.P.J., 1991: Volcano-tectonic control of offshore to tidal-flat regressive cycles from the Dunquin Group, Silurian: of southwest Ireland. In: Macdonald, D.I.M. (Ed.): Sedimentation, Tectonics and Eustasy, Sea-level Changes at Active Margins, Int. Assoc. Sedim. Spec. Publ., 12, 105-119.

Steijn, R.C., 1991: Some Considerations on Tidal inlets. A literature survey on hydrodynamic and morphodynamic characteristics of tidal inlets with special attention to "Het Friesche Zeegat". Delft Hydraulics, Coastal Genesis, Report H840.45., 109 pp.

Van den Berg, J.H., 1981: Rhythmic seasonal layering in a mesotidal channel fill sequence, Oosterschelde Mouth, the Netherlands. In: Nio, S.D., Shüttenhelm, R.T.E. & van Weering, Tj.C.E. (Eds): Holocene marine sedimentation in the North Sea Basin, Int. Assoc. Sedim. Spec. Publ. 5, 147-159.

Van den Boogert, J.M., 1991: Beschrijving van de Conlod programmatuur. Rijkswaterstaat, directie Friesland, rapport ANW.91.05, 107 pp.

Van den Boogert, J.M. & Noordstra, P., 1988: Beschrijving van de Conlod programmatuur. Rijkswaterstaat, directie Friesland, rapport ANW 88.30, 65 pp.

Van der Spek, A.J.F. & Beets, D.J., 1992: Mid-Holocene evolution of a tidal basin in the western Netherlands: a model for future changes in the northern Netherlands under conditions of accelerated sea-level rise? Sedim. Geol., 80, 185-197.

Van Parreeren, D. V., 1980: Waterloopkundige aspecten van de doorbaggering wantij Schiermonnikoog. Rijkswaterstaat, directie Groningen, nota 80-29, 40 pp.

Van Sijp, D., 1989: Correcties op gemeten eb- en vloedvolume bij de omrekening naar gemiddeld getij, in het Friese Zeegat. Rijkswaterstaat, directie Friesland, notitie ANW 89-02, 4 pp.

Van Straaten, L.M.J.U., 1954: Composition and structure of recent marine sediments in the Netherlands. Leidse Geol. Meded., XIX, 1-110.

Van Straaten, L.M.J.U. & Kuenen, Ph.H., 1957: Accumulation of fine grained sediments in the Dutch Wadden Sea. Geol. Mijnb., 19, 329-354.

Van Straaten, L.M.J.U. & Kuenen, Ph.H., 1958: Tidal action as cause of clay accumulation. J. Sedim. Petrol., 28, 406-413.

Veenstra, H.J. & Winkelmolen, A.M., 1976: Size, shape and density sorting around two barrier islands along the north coast of Holland. Geol. Mijnb., 55, 87-104.

Walther, F., 1972: Zusammenhänge zwischen der Größe der Ostfriesischen Seegaten mit ihrem Wattgebieten sowie den Gezeiten und Strömungen. Jber. Forschungstelle Küste, 23, 7-32.

Walton, T.L. & Adams, W.D., 1976: Capacity of inlet bars to store sand. Proc. 15th Conf. Coastal Eng., Am. Soc. Civ. Eng., Honolulu, 1919-1938.

Wang, Y., 1986: Sedimentary characteristics of the coast of Hainan Island: Effects of a tropical climate and active tectonism. Sediments down-under; 12th International Sedimentological Congress, Int. Assoc. Sedim., Canberra, Aug. 24-30, 318.

Winkelmolen, A.M., 1969: Experimental rollability and natural shape sorting of sand. 2nd ed., Holland, 141 pp.

Winkelmolen, A.M., 1982: Critical remarks on grain parameters with special emphasis on shape. Sedimentology 29, 255-265.

Winkelmolen, A.M. & Veenstra, H.J., 1974: Size and shape sorting in a Dutch tidal inlet. Sedimentology 21, 107-126.

Winkelmolen, A.M. & Veenstra, H.J., 1980: The effect of a storm surge on near-shore sediments in the Ameland-Schiermonnikoog area N. Netherlands. Geol. Mijnb., 59, 97-111.

APPENDIX 1: FIGURE CAPTIONS TO THE COLOUR PLATES (ATLAS)

Figure C4. Bathymetry of the study area in 1967. Coastlines represent the situation in 1978. The map shows the area two years before the closure of the Lauwerszee embayment (in the S). The Zoutkamperlaag Inlet system (W of the barrier island Schiermonnikoog) has a large ebb-tidal delta (N) and a deep main channel. Note the shallow (above -7.5 m DOL) remnants of an originally flood-dominated channel cutting through the triangular swash platform directly NW of Schiermonnikoog and the small shoals (-1 to -2.5 m) in the northern part of the platform (after Oost & De Haas, 1993).

Figure C5. Bathymetry of the study area in 1970, one year after closure of the Lauwerszee. Note the formation of the E-W oriented part of the recurved bar on the ebb-tidal delta (swash bar NW of Schiermonnikoog) from the small shoals of 1967. In the ebb-tidal delta an E-W oriented flood-dominated outer channel and a NW-SE ebb-dominated outer channel are visible. Also note the shallowing of the main channels in the backbarrier area of the Zoutkamperlaag Inlet and of the inlet system E of the watershed of Schiermonnikoog (after Oost & De Haas, 1992).

Figure C6. Bathymetry of the study area in 1975. The triangular swash platform retreated coastwards in the period 1970-1975. The shoal in the NW and the E-W oriented swash bar have become higher. Note the rotation of the outer channels. The southern part of the inlet gorge has filled up. The shallowing of the main channels in the backbarrier area of the Zoutkamperlaag Inlet system and of the system E of Schiermonnikoog has continued. The shift of the interjacent watershed has commenced. The Pinkegat system, W of Ameland, has developed into a double inlet system (after Oost & De Haas, 1992).

Figure C7. Bathymetry of the study area in 1979. The erosion and retreat of the ebb-tidal delta of the Zoutkamperlaag Inlet has continued and the formation of the N-S oriented part of the recurved bar has begun. Rotation of the outer channels and shallowing of the main channel in the backbarrier drainage basin of the Zoutkamperlaag Inlet have continued. Pronounced retreat and shallowing of the system E of Schiermonnikoog is evident. Note the abandonment of smaller channels S of Ameland due to the marked changes in the configuration of the Pinkegat system (after Oost & De Haas, 1992).

Figure C8. Bathymetry of the study area in 1982. Erosion and retreat of the ebb-tidal delta of the Zoutkamperlaag Inlet and bar formation off Schiermonnikoog have continued. The shallowing of the main channel in the backbarrier drainage basin of the Zoutkamperlaag system has decelerated slightly. A pronounced shift of the watershed S of Schiermonnikoog has occurred in combination with the erosion of channels at the western side (after Oost & De Haas, 1992).

Figure C9. Bathymetry of the study area in 1987. Erosion and retreat of the ebb-tidal delta of the Zoutkamperlaag Inlet and formation of the recurved bar, which reached full maturity and became supratidal in 1987, have continued. The eastern branch of the outer ebb-dominated channel has been abandoned. The shallowing of the main channel in the backbarrier area of the

Zoutkamperlaag Inlet and erosion and channel formation at the western side of the watershed S of Schiermonnikoog have continued. Note that the Pinkegat Inlet has once more become a single inlet system. In this cyclic process several larger outer channels (compare 1982 and 1987) and smaller backbarrier channels have been abandoned (after Oost & De Haas, 1992).

Figure C10. Bathymetry of the study area in 1991. Only sounding data of the ebb-tidal deltas were available. The ebb-tidal delta of the Zoutkamperlaag Inlet has retreated further, but at a reduced rate. The recurved bar encloses a small embayment and is in decay, having become intertidal and breached by several channels (up to 5 m deep) (after Oost & De Haas, 1993).

Figure D1. Erosion and sedimentation patterns in the period 1970-1987, reflecting the effects of the closure of the Lauwerszee embayment. Visible are the erosion of the ebb-tidal delta of the Zoutkamperlaag Inlet, the formation of the recurved bar, sedimentation and erosion due to outer channel shift, the vertical fill of the inlet and main back-barrier channel of the Zoutkamperlaag Inlet, the formation of new channels at the western side of the watershed and the fill of the main channel SE of Schiermonnikoog. The erosion and sedimentation pattern in the Pinkegat system is a poor reflection of the rapid changes that have occurred in that system (compare Figs. C4 to C10; after Oost & De Haas, 1992).

CHAPTER 6

THE INFLUENCE OF BIODEPOSITS OF THE BLUE MUSSEL *MYTILUS EDULIS* ON FINE-GRAINED SEDIMENTATION IN THE TEMPERATE-CLIMATE DUTCH WADDEN SEA

ABSTRACT

The filter feeder *Mytilus edulis* significantly influences the fine-grained sedimentation in the Dutch Wadden Sea. Depending on the size of the population, variable, but large amounts of sediment are filtered from the water column and compressed into faeces and pseudo-faeces (in total 2.6-15.1 * 10^9 kg.y-^{1} minerals dry weight). Experiments show that these biodeposits have a larger grainsize than the original sediment and behave hydrodynamically as grains of silt/sand size. Also, they are fairly resistant to decay during phases of transport and rest. In this way biodeposition, in combination with deposition of sand, leads to the formation of mussel mudmounds and muddy sandflats surrounding the mussel beds. The concentration of fine-grained material in the sediment decreases with increasing distance from the mussel beds. This decrease is due to the prevailing relatively high-energy conditions. A smaller part of the fine-grained sediment is deposited on more distant tidal flats and in neighbouring channels. After some time the larger part of the biodeposits is resuspended. Experiments show that also the resuspended particles have a significantly larger grainsize than the original suspended matter.

From late autumn to winter, sedimentation rates on mussel beds are relatively low. Erosion of the mud occurs during storms and ice-coverage. The average suspended sediment concentration of the water increases during the winter season. Resuspended or ice-rafted sediments are exported to the North Sea or deposited on tidal flats and tidal marshes. The latter in particular occurs in late autumn.

Apart from mussels many other animals are also capable of biodepositing substantial amounts of sediment. Where this process is important relative to physical and chemical sedimentary processes, it can substantially influence sedimentation patterns. Biodeposition is often quite influential in lagoons and estuaries. In such environments the characteristic low variability of species and the high biomass frequently result in high amounts of a few kinds of filter feeders. Although direct evidence in the fossil record is often lacking, biodeposition must have played an important role in depositing the fine-grained siliciclastics of many shallow-marine sedimentary successions.

INTRODUCTION

The Dutch Wadden Sea is a partly intertidal backbarrier area. The tidal regime is micro- to mesotidal. The sediment consists for 80-90% of sands (Md: 170-190 μm), the rest being finer, muddy sediments (Van Straaten, 1964). Within the Wadden Sea fine-grained sedimentation is influenced by *Mytilus edulis* L., which is one of the major filter feeders in the area.

It has long been recognized that plants and animals can influence sedimentation. Filter feeders, which produce biodeposits from the indigestible part of suspensions, are a well known example (Verwey, 1952; Arakawa, 1963, 1965, 1968, 1971, 1972; Haven & Morales-Alamo, 1972). From intertidal flats to the deep sea, biodeposits[1] of invertebrates can form a substantial part of the sediment (Schäfer, 1953; Youngbluth et al., 1989; Flemming & Delafontaine, 1994), and influence the physical, chemical and ecological conditions (Arakawa et al., 1971; Pryor, 1975; Dankers et al., 1989).

Estuaries and lagoons have a relatively great biomass as compared to the open sea (Nybakken, 1982). Due to the abundance of fauna, the amount of biodeposition is often large enough to significantly influence sedimentation patterns, notwithstanding the high input of other sediments (see Rhoads & Stanley, 1965; Reineck & Singh, 1971). Filtration and biodeposition by mussels (*Mytilus* and *Modiolus* spp.) in estuaries and lagoons is mostly studied from a biological point of view (e.g., Verwey, 1952; Tammes & Dral, 1956; Widdows et al., 1979; Dankers & Koelemaij, 1989). Less frequently attention is paid to the sedimentological implications (cf. Van Straaten, 1954; Flemming & Delafontaine, 1994).

M. edulis filters sea water to obtain food. All suspended particles greater than 2 to 5 μm in diameter are completely filtered out by the gills and removed from the ventilation current (Tammes & Dral, 1956; Vahl, 1972, Bayne et al., 1976, 1977; Widdows, 1984). In the case of *Mytilus edulis*, absorption efficiency and the total amount of food consumed are inversely related. At low particle concentrations, all suspended matter is filtered from the water and ingested, and passes through the stomach to the digestive gland. After absorption of the food the remainder is excreted as 'glandular faeces'. When more material enters the stomach, part of it bypasses the digestive gland and is excreted as 'intestinal faeces' (Widdows et al., 1979).

At a particle concentration of $\pm$ 5 mg.l^{-1} a threshold is reached above which further material filtered by the gills cannot be fully ingested and is excreted as pseudofaeces (Widdows et al., 1979). *M. edulis* is, to some extent, capable of regulating the material intake by adaptation of the gill and palp size (Essink et al., 1986), by changing the filtration rate (Dral, 1967; Bayne et al., 1976), and by changing the porosity of the gills (Bayne et al., 1977). Absolute filtration rates decrease when concentrations of suspended matter exceed

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[1] Biodeposits are defined as faecal pellets and pseudo-faeces, whereas the entire filtration process and the deposition is termed biodeposition (cf. Haven & Morales-Alamo, 1972).

120 mg.l^{-1} for mussels of 3 cm shell length and above 190 mg.l^{-1} for mussels of 7 cm length (Widdows et al., 1979).

This paper discusses the influence of the filter feeder *M. edulis* L. (the common blue mussel) on fine-grained sedimentation in the Dutch Wadden Sea. The production of biodeposits by *M. edulis* was examined experimentally. Also, the impact of filtration and biodeposition by *M. edulis* on sedimentation and its role in the seasonal variation of fine-grained sedimentation in the Dutch Wadden Sea is discussed.

EXPERIMENTS

Little experimental work has been carried out to study the influence of the mussel on the sedimentation of fine-grained sediments. Three experiments were done in order to study (1) in which way the mussels influence the suspended sediment in the water, (2) hydrodynamic differences between biodeposits and the original suspension, and (3) the resistance of biodeposits against disintegration during periods of transport and rest.

Experiment 1: Influence of *Mytilus edulis* on suspended sediment in the water

Methods

One kilogram of mussels was rinsed superficially and barnacles were removed. The animals were put in artificial sea water (without sediment) for 48 hours to clear the intestines (De Vooys, 1987). Fresh aerobe fine-grained sediment from the Wadden Sea, disaggregated by prolonged ultrasonic treatment, was used to prepare, with an Atterberg cylinder, a suspension with a grainsize of 30 μm and less.

The mussels were put into an aquarium with 7.5 litres of sediment-free artificial sea water, and an airlift system was installed to produce current velocities of 0.2-0.4 m.s^{-1}. Subsequently the suspension was added. At regular time intervals the turbidity of the water and the grainsize distribution of the particles in suspension were measured with a Malvern Laser particle sizer, until the particle concentration was too low to be measured. The experiment was repeated without mussels.

The same mussels were then left in the aquarium with a high concentration suspension during one night. The following morning the water was probed again. The experiment was also repeated without mussels.

Results

In water with mussels to which an addition of suspended fine-grained sediment had been added, the initial turbidity strongly decreased and the mean grainsize increased within one hour (Fig. 1a,b; Table I). In the experiments without mussels the water remained turbid and the grainsize did not increase (Fig. 1c,d).

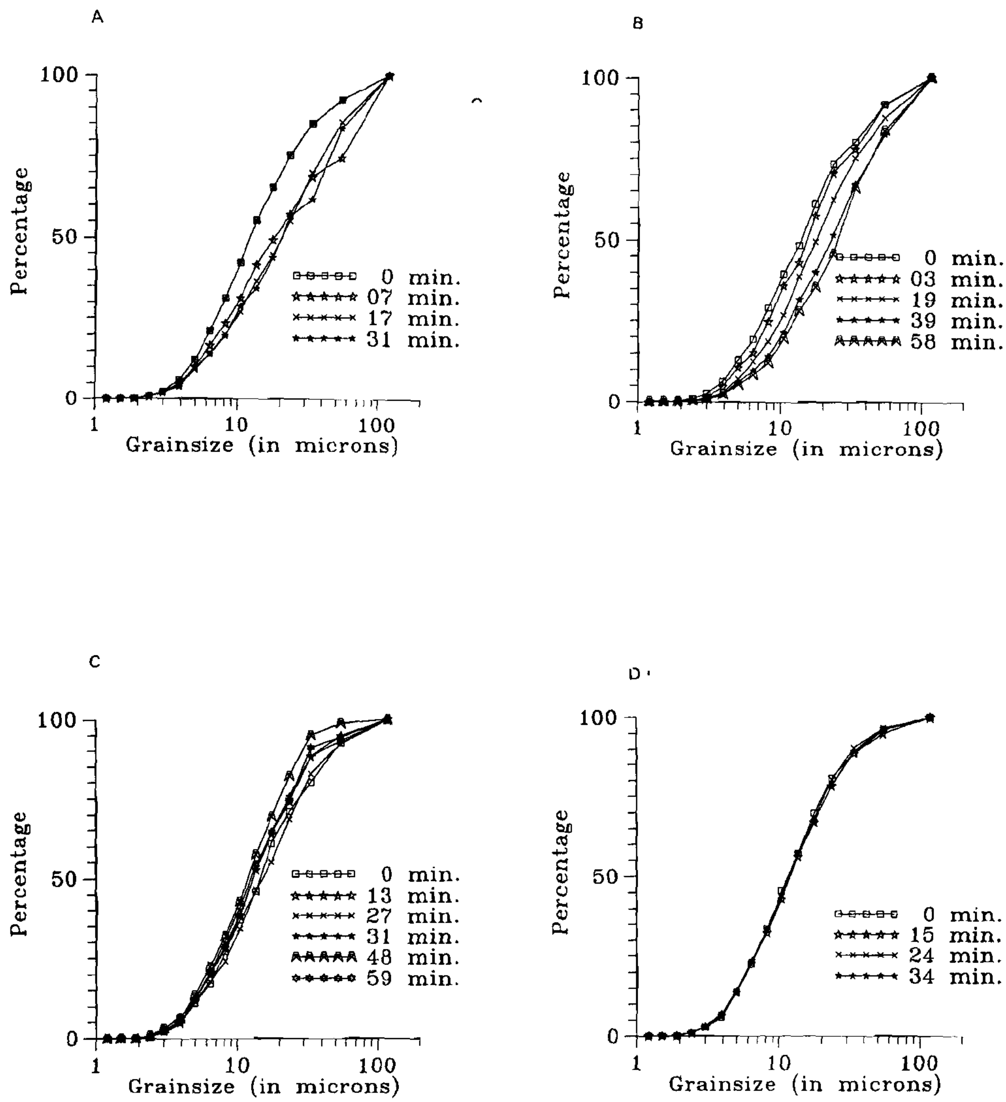
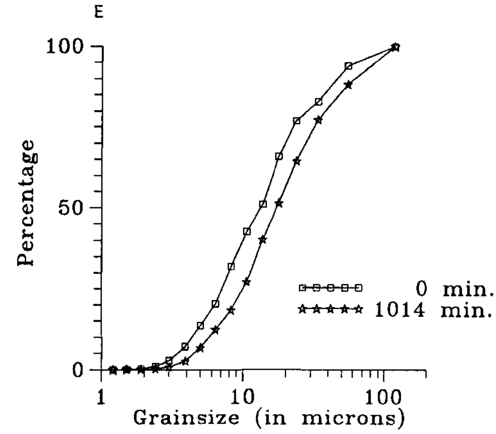
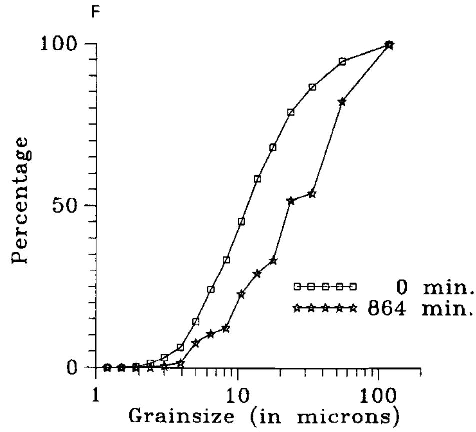
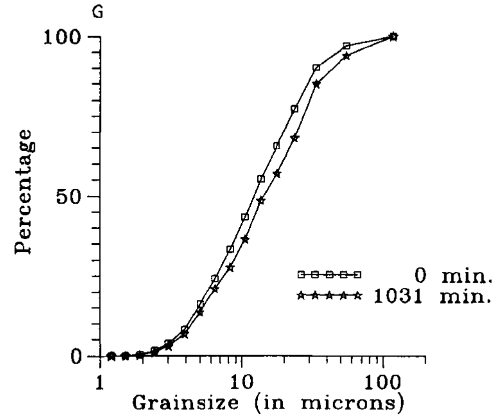


Figure 1a-d: The effects of filtration by *M. edulis*. For explanation see Table I.







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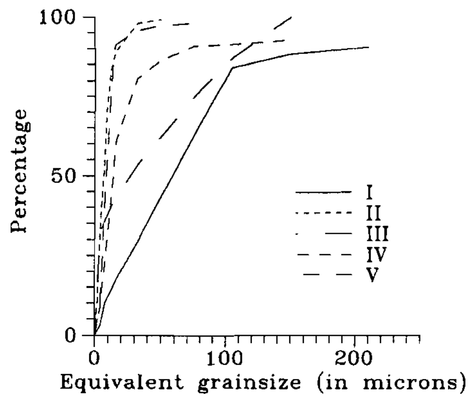


Figure 1e-g: The effects of filtration by *M. edulis*. For explanation see Table I.

Figure 2 (lower right): The equivalent quartz grainsize of; I: faecal pellets; II: original suspension; III: faecal pellets after 36 h of rest in water of 19°C; IV: faecal pellets after 72 h of rotation; V: the original suspension after 36 h of rest and 72 h of rotation.

Table I: The experiments of Fig. 1

Fig. No.	Exp. No.	Mussels present?	Initial concentration	Measurement intervals
1a	1	Yes	44 $mg.l^{-1}$	0, 7, 17, 31 min.
1b	2	Yes	63 $mg.l^{-1}$	0, 3, 19, 39, 58 min.
1c	3	No	16 $mg.l^{-1}$	0, 13, 27, 31, 48, 59 min.
1d	4	No	72 $mg.l^{-1}$	0, 15, 24, 34 min.
1e	5	Yes	63 $mg.l^{-1}$	0, 1014 min.
1f	6	Yes	87 $mg.l^{-1}$	0, 864 min.
1g	7	No	59 $mg.l^{-1}$	0, 1031 min.

In the experiments with a high suspension concentration the water had become clear after one night. Concurrently the grainsize of the suspended matter in the agitated water had increased (Fig. 1e,f) and biodeposits were found on the bottom between the shells and in the corners of the aquarium, where the current velocity was low. In the experiments without mussels only a slight increase in grainsize was observed (Fig. 1g) and the water had remained turbid.

Experiment 2: Comparison of the biodeposits with the original suspension

Methods

At the end of experiment 1 the pellets produced by the mussels were collected. Observations by microscope showed that the clay pellets were surrounded by an organic film. The pellets had a diameter of up to 0.5 mm. The settling velocity of the original suspension and that of the biodeposits were measured in a sedimentation balance (Fig. 2).

Results

Measurements with the sedimentation balance showed that the equivalent quartz grainsize of the biodeposits was about ten times greater (M=60 μm) than that of the original suspension material (Fig. 2).

Experiment 3: Resistance of pellets to decay during rest and transport

Abrasion of pellets is mainly caused by water movements (Van Straaten, 1964). Moreover, microbial degradation of the organic matter may alter the grainsize of pellets (cf. Newell, 1965; Frankenberg & Smith, 1967; Gowing & Silver, 1983; Jacobsen & Azam, 1984; Boede, 1985; Peduzzi & Herndl, 1986; Dittmann, 1987). Other alteration mechanisms are changes in the water content (Smith & Frey, 1985), coverage by sediment (Van Straaten, 1954), and destruction of pellets by larger organisms (Watling, 1989; Dittmann, 1988). The effect of transport and of decay during rest on the grainsize of the pellets were studied in the laboratory.

Methods

After storing the pellets of experiment 1 in artificial sea water at 19°C for 36 hours, the equivalent grainsize was measured in a sedimentation balance (Fig. 2). Subsequently the pellets were transported during 72 hours, at 15.7 $cm.s^{-1}$, after which the settling velocity was measured again (Fig. 2). The original suspension was treated in the same way (Fig. 2).

Results

Originally, the pellets produced during experiment 1 had an equivalent quartz grainsize of M=60 μm, ten times larger than that of the original suspension material (Fig. 2, exp. 2). After the resting phase of 36 hours the equivalent grainsize of the pellets had decreased to M=26 μm. After subsequent transport at 15.7 $cm.s^{-1}$ during 6 hours the bulk of these pellets was still intact. After 72 hours of transport the average equivalent grainsize had been reduced to from M=26 μm to M=13.5 μm. This is still about twice the equivalent size of the original particles in suspension (M=6.5 μm). Even equivalent grainsizes of more than 100 μm were observed (Fig. 2).

The original suspension which had not been subjected to filtering by *Mytilus edulis* had become slightly coarser after the period of rest followed by the subsequent transport of 72 hours than it was before treatment (from M=6.5 to M=8.5 μm). The coarser part of this suspension (above D^{90}) also increased in grainsize (Fig. 2).

DISCUSSION OF THE EXPERIMENTS

Experiment 1: Influence of *Mytilus edulis* on suspended sediment in the water

The rapid decrease in turbidity during the experiments with mussels confirms that *M. edulis* effectively traps sediment. This is not surprising, since the average pumping rate of 1 kg of mussels is of the order of 100 $l.h^{-1}$ (cf. Dankers et al., 1988). At that rate the total volume of the aquarium is filtered several times within one hour.

The observed increase of the grainsize of the remaining suspension suggests that fine material of up to ± 23 μm was trapped more effectively than coarser material (cf. Fig. 1). This is in contradiction with the normally observed filtering activity of *M. edulis*, which filters out particles with a grainsize >2-5 μm up to at least 100 μm (Bayne et al., 1976, 1977) at a high and constant rate. The obvious explanation for the observed increase in grainsize is that parts of the biodeposits are resuspended in the agitated water. Such a biologically induced increase in grainsize is also observed in experiment 3.

A spontaneous formation of microflocs with a typical diameter up to 125 μm (cf. Eisma, 1986) is not likely, because a strong increase in grainsize was not observed in the experiments without mussels, in particular not over short time intervals (cf. Fig. 1). The relatively small increase in grainsize of the suspension over one night indicates some (bacteria-driven?) formation of microflocs.

Experiment 2: Comparison of the biodeposits with the original suspension

The experiments show that the coarser biodeposits settle at flow velocities of 0.2-0.4 $m.s^{-1}$. Due to their large water content these grains of up to 500 μm are hydrodynamically equivalent to quartz grains of on the average 60 μm. This is nine times larger than that in the original suspension (M= 6.5 μm). The biodeposits thus settle much faster than the original suspension.

Experiment 3: Resistance of pellets to decay during rest and transport

The third experiment shows that pellets are fairly resistant to disintegration during resting phases. The observed decrease in grainsize and in settling velocity must be the result of decay of the organic mucus layer and perhaps of an increase in water content. Part of the pellets was still present after the resting phase. This implies that biodeposits can survive a period of several hours of rest, after which further transport is possible. This is also supported by direct observations in the Dutch Wadden Sea, where large amounts of deposited pellets can be observed to be transported again during the next flood on the intertidal flats (Oost, unpubl. data).

During one flood or ebb period, water in the Dutch Wadden Sea moves over a distance of about 3.6 km on average (cf. Zimmermann, 1976). This is comparable to the distance travelled by the pellets during 6 hours in the third experiment. Thus, biodeposits rolling over the sediment surface can be transported over several km from the area where they were produced before full disintegration. In the Wadden Sea intact pellets may be transported over even larger distances, because the low specific density of the pellets may result in saltation and suspension transport (cf. Winkelmolen & Veenstra, 1974). Haven & Morales-Alamo (1972) arrived at a similar conclusion for biodeposits of oysters (*Crassostrea virginica*).

Even after 72 hours the grainsize of the pellet sample (mainly in suspension) was still larger than that of the original suspension. This indicates that the resuspended biodeposits have disintegrated into particles which have a larger size than those in the original suspension. Important causes are likely particle attachment by polysaccharides in the biodeposits (Eisma, 1986) and enhanced binding of particles by Van der Waals and intermolecular forces upon compression. Thus, resuspension of biodeposits is of influence on the size of floccules, a possibility mentioned by Eisma (1986) and Eisma et al. (1991b).

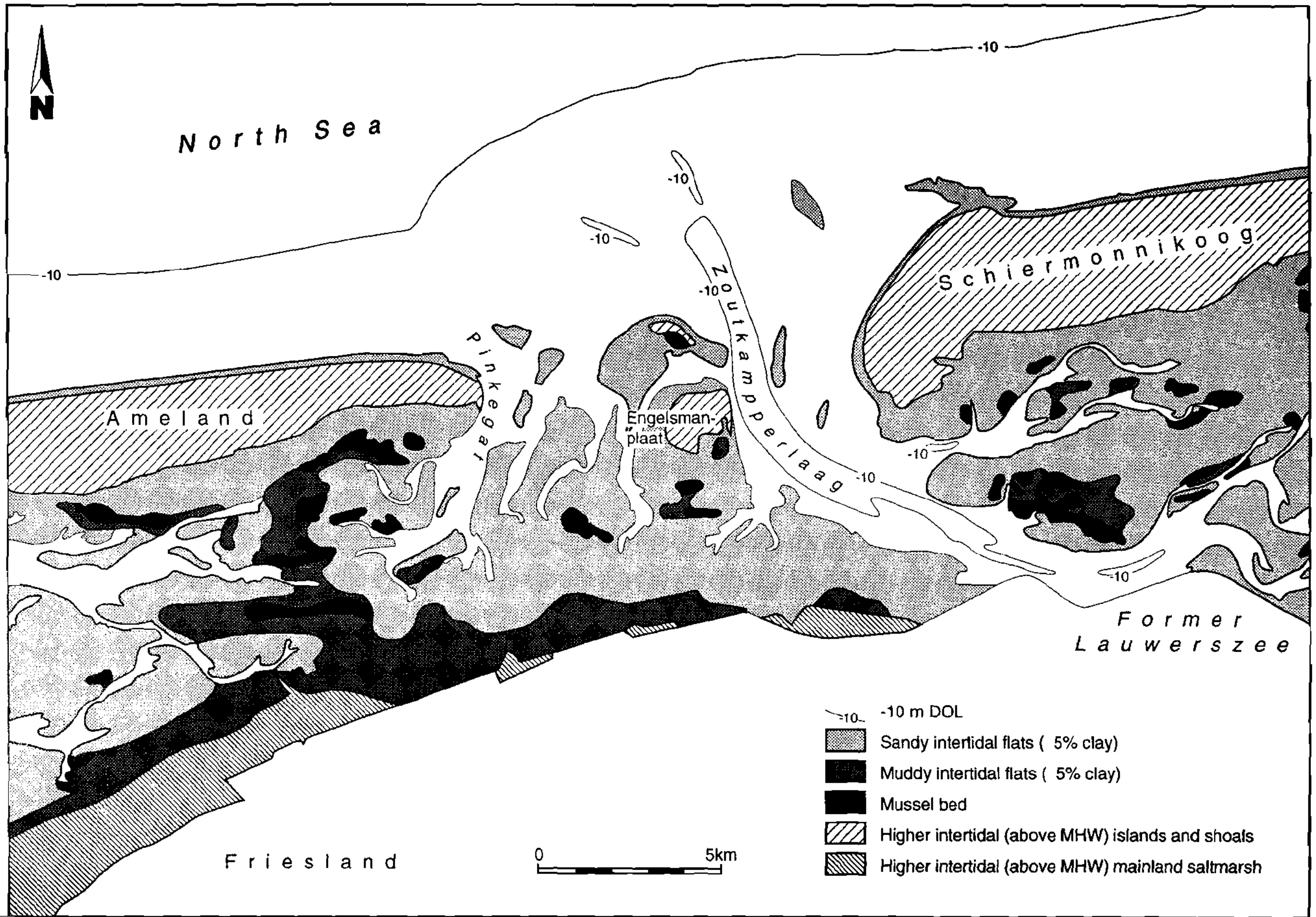
QUANTITATIVE IMPORTANCE OF FILTER FEEDING BY MUSSELS

The macrozoobenthos of the Dutch Wadden Sea is characterized by high biomass values and low species diversity, with only 5 or 6 species accounting for 80-90% of the macrozoobentic biomass (Beukema, 1982a). In the seventies of this century about 23% of the intertidal macrozoobenthos bottom fauna and 67% of the subtidal bottom fauna of the Western Dutch Wadden Sea consisted of mussels (Beukema, 1976; De Wilde & Beukema, 1984; Dankers, 1986; Dekker, 1987). The high percentage in the subtidal environment is partly the result of mussel cultures (Dankers, 1986; Dankers et al., 1989). Intertidal mussel beds were an important biogenic structure in the eastern Wadden Sea, before shell fishing was intensified and several severe winter storms occurred (Verwey, 1952; Beukema, 1993; Van Straaten, pers. comm.; Dijkema, pers. comm.). In 1978 the intertidal flats of the eastern Wadden Sea were covered for 4.5% with natural mussel colonies (Dankers, 1986) (Fig. 3). Natural intertidal mussel beds in the Dutch Wadden Sea have been almost completely eliminated in the summer of 1990 by mussel fishery and storms. Intertidal mussel beds were still largely absent in 1993 (Beukema, 1992, 1993; Dankers, 1993), and were still rare in the beginning of 1994. In summer 1994 a reasonably good spatfall occurred in some areas. Below the newly formed mussel beds a thick layer of fine-grained sediment was present (Dankers, pers. comm.). Thus mussels were (in the eastern Wadden Sea), and still are (in culture plots in the western Wadden Sea) a major component of the macrobenthos.

Table II: Estimates of the production of faeces and pseudofaeces by filter feeders in the Dutch Wadden Sea.

Reference	Filter feeder(s)	Location	Amount in 10^6 kg
Verwey, 1952	Mytilus edulis	West. Wad. Sea 1,560 km^2	105-255 (1949)
Kamps, 1962	Mytilus edulis	Groninger Wad 500 km^2	1,312.5
De Glopper, 1964	Mytilus, Cerastoderma & Mya	Groninger Wad 500 km^2	2,500-3,000
Verwey, 1981	Mytilus edulis	West. Wad. Sea 1,560 km^2	1,000
Dankers et al., 1989	Mytilus edulis	West. Wad. Sea 1,560 km^2	15,168

Based on microscope studies, Van Straaten (1954) concluded that the fine-grained muddy deposits in the Wadden Sea were formed by sedimentation of grains, floccules and faecal pellets. Verwey (1952; Table II) estimated the annual biodeposition by *M. edulis* to be of the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North Sea
-10
-10
-10
-10
-10
-10
-10
-10
Ameland
Pinkegat
Zoutkampperlaag
Engelsman-
plaat
Schiermonnikoog
Former
Lauwerszee
Friesland
0
5km
-10 m DOL
Sandy intertidal flats (5% clay)
Muddy intertidal flats (5% clay)
Mussel bed
Higher intertidal (above MHW) islands and shoals
Higher intertidal (above MHW) mainland saltmarsh

to be of the order of 10^7 kg. Kamps (1956), De Glopper (1964) and Verwey (1981) even estimated the annual biodeposition of the three important species of filter feeders (*Cerastoderma edule*, *Mya arenaria* and *Mytilus edulis*) to be of the order of 10^9 kg (Table II). Recently Dankers et al. (1989) even calculated the total annual production of faeces and pseudo-faeces to be of the order of 10^{10} kg (Table II).

Annual amount of biodeposits in the Wadden Sea

The amount of suspended matter which is, on the average, annually filtered from the water column by *M. edulis* in the Dutch Wadden Sea can be calculated from the size of the mussel population, the size classes, their daily filtration volume, the amount of suspended matter per litre, the percentage of suspended matter that cannot be filtered by the mussels, the percentage of organic matter and the number of days during which mussels are active (Dankers et al., 1989). These factors are discussed below:

Population size

The size of the mussel population in the Dutch Wadden Sea fluctuates strongly from year to year, due to climate-influenced population dynamics and fishery (Table III; Beukema, 1982a; Dankers & Koelemaij, 1989; Dankers et al., 1989; Dijkema, 1989; Beukema, 1993; Dankers, 1993; cf. Nehls & Thiel, 1993). The Wadden Sea is a temperate tidal flat area, characterized by strong seasonal differences. The water temperature is above 15°C (up to 21°C) from the middle of May to the middle of September and predominantly between -1°C to +4°C during the winter (Anonymous, 1967; Postma & Dijkema, 1982). In the Dutch Wadden Sea the average number of days with ice is about 10 (Eisma, 1980). In about 40% of the winters, temperatures are low enough to allow substantial ice coverage of the tidal flats (IJnsen, 1988). Although mussels can survive prolonged periods of frost (Bayne et al., 1976), the abrasion of beds by moving ice can be catastrophic as was illustrated by the almost complete destruction of the mussel population in December 1938 (Kuenen, 1942) and in the winter of 1962/63 (Dankers & Koelemaij, 1989; Dijkema, pers. comm.), the local destruction in 1971/72 (Essink, 1978) and the strong reduction of the population in the three cold winters of 1984-1987 (Dankers et al., 1989).

Figure 3 (opposite page): Mussel beds on the watershed south of Ameland in 1975-1978. Mud flats near the mussel colonies consisted of biodeposits of the mussels (after Dijkema 1989). Mussel beds and adjacent mudflats were completely absent in 1993.

Table III: Variation in mussel biomass in the Dutch Wadden Sea, based on estimates for various years in 10^6 kg.

Area	Lowest		Highest	
Eastern Wadden Sea Intertidal	6	1987[1]	180	1961[2]
Western Wadden Sea Intertidal	3-4	1987[3]	24	1977[4]
Western Wadden Sea Subtidal	40	*	165	1981[5]
Western Wadden Sea Culture	33	1977-1987[6]	136	1977-1987[6]

[1]: Wensink & Reitsma in Dankers et al., 1989
[2]: Kamps (1962) in Dankers & Koelemaij, 1989
[3]: Annual average based on data of Wensink & Reitsma and Beukema (both in Dankers et al., 1989) and data of Dankers et al., 1989.
[4]: Beukema et al., 1978
[5]: Dekker (1987) in Dankers & Koelemaij, 1989, extremely high biomass (Dankers et al., 1989)
[6]: Dankers & Koelemaij, 1989
*: Based on the factor 33/136 for the subtidal culture plots.

Wind speeds are, on the average, 7 $m.s^{-1}$ in summer and 15 $m.s^{-1}$ in winter. On the average, about 9 major storms, causing a strong rise of the water level, occur each year (cf. IJnsen, 1977). Waves during storms have heights of up to 2 m (e.g., De Haas & Eisma, 1993). Storms in winter and early spring can have a devastating effect on mussel colonies and can destroy them (almost) completely (Kamps, 1956; Van Straaten, 1964; Seed, 1976; Dankers et al., 1989; Dankers & Koelemaij, 1989; Flemming et al., 1993) or kill them by sediment coverage (Kranz, 1974). Occasionally, intensive predation by starfish in summer causes a strong reduction of the subtidal mussel population (Dankers et al., 1989).

Spatfall depends on the water temperature (between 5°C-22°C) and salinity (between 1.5%-4.0%) (Verhagen, 1982), and occurs from March to June (De Blok & Geelen, 1958), sometimes followed by a second peak in late summer (Anonymous, 1967). Settling of the young shells happens around July (Verhagen, 1982). Recruitment is negatively related to the size of the total stock (Seed, 1976; Beukema, 1982a; Verhagen, 1982), so that any destruction of the mussel population tends to be restored. The calculation below was based on average annual populations (Dankers et al., 1989).

Volume of water filtered

The amount of water filtered per gram mussel depends on the size of the mussels, the pumping velocity of the mussels, the concentration of suspended sediment, and the duration of the water coverage (Dankers, 1986).

The smaller the size of a mussel, the larger is the relative pumping rate (Bayne & Newell, 1983). In general, mussels pump at high rates. The average amount of water

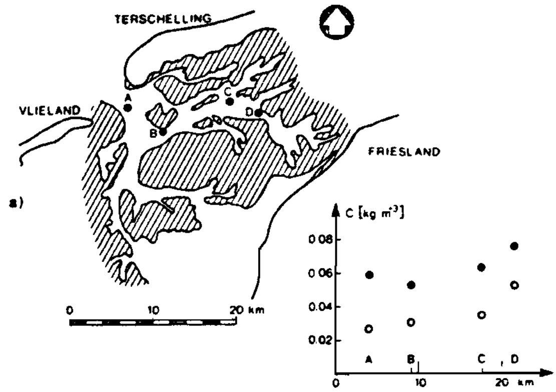

Figure 4: Suspended-sediment concentrations in the Vlie Inlet system. Dark circles represent the autumn-winter situation and open circles the spring-summer situation for the period 1977-1981 (Dronkers, 1984).

pumped by 1 kg of mussels of average length is 130 l/h in the Wadden Sea (Dankers & Koelemaij, 1989). In the present calculations the pumping rates applied by Dankers et al. (1989), who made an estimate for the filtration in different environments have been used. For the intertidal mussels it was assumed that they were covered with water during 18 h.

Although the blue mussel is able to adapt its feeding apparatus to a certain extent (Essink et al., 1986), filtering activity decreases when suspended sediment concentrations become too high (Widdows et al., 1979). For 7 cm-long mussels, the production of biodeposits increases with increasing suspended sediment concentration up to about 150 mg/l; at higher concentrations a decrease of production occurs; the total production starts to decrease at values above circa 190 mg/l (Widdows et al., 1979). In the Wadden Sea such high concentrations normally (De Haas & Eisma, 1993) are found in areas where mussels do not form colonies. It is not known if there is a causal relationship. During storms a reduction in pumping due to too high suspension concentrations occurs. Furthermore, algal blooms of *Phaeocystis* sp., in particular in spring, halve the pumping rate (Prins et al., 1994; Smaal, pers. comm.). In the period from 1974 until 1989 the duration of *Phaeocystis* blooms (periods with >1000 cells.ml^{-1}) varied strongly from year to year, but was on average 73 days (Cadée & Hegeman, 1986, 1991).

Also the temperature is of influence on the pumping velocity. Although mussels are poikilotherm, laboratory experiments indicate that they can maintain a constant pumping rate in water with temperatures ranging from 5°C to 20/22°C (Bayne et al., 1976; Widdows et al., 1979; Jϕrgensen; 1990). Recent research even shows that the pumping rate is not significantly influenced by temperatures down to at least -1°C (Loo, 1992; Smaal, pers. comm.). Only after strong and abrupt changes in temperature the animals need about 14 days to acclimatize (Bayne et al., 1976; Jϕrgensen, 1990). Such changes, however, do normally not occur under natural conditions, and it is assumed that pumping rates are only reduced, when sea ice is formed (on average 10 days per year).

Amount of suspended matter per litre

Eisma and Kalf (1987) give values of around 10 $mg.l^{-1}$ for suspended matter concentration in the North Sea area in front of the Wadden Sea. In the Wadden Sea the concentration increases in the direction of the mainland (Fig. 4) (Van Straaten, 1954; Postma, 1954; Eysink, 1979; Oost & De Boer, 1994). Close to the mainland concentrations can be as high as 130 $mg.l^{-1}$ (Eysink, 1979). Within the Wadden Sea the average suspended matter concentration is considered to be of the order of 42.5 $mg.l^{-1}$ (De Wit et al., 1982).

Percentage which cannot be filtered out by the mussels

As stated above, particles >2-5 μm in diameter are filtered out completely by the blue mussel. Smaller particles are partly retained (Tammes & Dral, 1956; Bayne et al., 1976, 1977). Normally the grainsize of suspended matter has been measured with a coulter counter. Eisma et al. (1991a) demonstrated that such measurements are sensitive to the fragility of flocs in the water, and that the actual in situ floc size is often much greater. Measurements over half a tidal cycle in the Ems-Dollard Estuary indicate that the amount of grains with an *in situ* size smaller than 5 μm is mostly less than 1% by weight of the total suspended matter (Eisma & Li, 1993). In the calculation this small percentage has been neglected. It is assumed that all of the suspended matter, which is pumped in by the mussel, is filtered.

Amount of organic matter retained

In spring and summer about 20% of the suspended matter consists of particulate organic matter (POM; Cadée, 1982; Eisma & Kalf, 1987). In autumn and winter it is about 5-10% (Cadée, 1982). An average percentage of 15% POM in suspended matter was assumed by Dankers et al. (1989; see also Verwey, 1952) in their calculations of the filtration by mussels. Dankers et al. (1989) demonstrate that only a small part of the organic matter is consumed. The larger part is returned with the pseudo-faeces, which consist for ± 15% of organic matter. Faeces consist for ± 9% of organic matter (Dankers et al., 1989).

Annual amounts of biodeposits in the Dutch Wadden Sea

Based on the above data, the amount of faeces and pseudofaeces formed annually by the common blue mussel in the Wadden Sea has been calculated. It is assumed that they can maintain their full pumping activity during 9 months of the year. Three months are lost, due to high suspended matter concentrations during algal blooms, to storms or to sea-ice.

Table IV: Estimated annual production of biodeposits (in 10^6 kg.y^{-1} dry weight) in the Dutch Wadden Sea (after Dankers et al., 1989)

Area	Faeces	Pseudofaeces
Eastern Wadden Sea Intertidal*	19-567	161-4,844
Western Wadden Sea Intertidal	11-76	94-646
Total Intertidal	30-643	255-5,490
Western Wadden Sea Subtidal, natural	169-697	1,434-5,914
Western Wadden Sea Subtidal, culture	133-547	1,064-4,386
Dutch Wadden Sea Total	332-1,887	2,753-15,790
Mineral component	302-1,716	2,340-13,422

*: The eastern Wadden Sea only consists for 20-30% of subtidal zones, which are mainly channels through which high amounts of sediment are transported. This strongly restricts the amount of subtidal mussels (cf. Dankers et al., 1989), which is therefore neglected.

Following the detailed calculations for the western Wadden Sea by Dankers et al. (1989) for the various environments, the minimum and maximum amount of biodeposits formed by the mussel population (based on Table III) in the whole Dutch Wadden Sea is calculated to be of the order of 3.1 * 10^9 kg.y^{-1} to 17.7 * 10^9 kg.y^{-1} (Table IV). The amount of non-organic mineral components is 2.7-15.1 * 10^9 kg.y^{-1}. These figures are for years with a low and a high biomass, respectively.

SEDIMENTS FORMED

The amount of biodeposits annually produced by *M. edulis* is so high that it significantly influences sedimentation patterns in the Wadden Sea (Fig. 3). Part of the biodeposits is not transported, but accumulates with sands in and below the subtidal (Dankers, 1986) and intertidal mussel colonies. Another part is transported and deposited in the mussel-induced muddy sandflats, which commonly surround the intertidal mussel colonies, or is incorporated in other sediments of the backbarrier area. The larger part of these biodeposits, however, is resuspended immediately after formation or after a short time.

Mussel beds

Subtidal mussel beds occur mainly in the western Wadden Sea on those places where the animals can survive. Natural intertidal mussel beds mainly occur along the larger tidal channels, around the ends of smaller gullies (Verwey, 1952; Dijkema, 1989) or as series of colonies oriented parallel to the mainland on the watersheds. Maps and aerial photographs suggest that in particular larger, older colonies influence the position of tidal drainage channels (Van Straaten, 1954; De Vries, pers. comm.; Dankers, pers. comm.; pers. obs.). This most likely results from the fact that they increase the stability of the sediment and locally reduce the tidal volume. In reverse, a position in zones where sufficient water is supplied during each tide, provides the mussels with large amounts of food. Several studies indicate that within a backbarrier area mussel beds tend to newly form at more or less the same place after destruction (Thiesen, 1968; Seed, 1976; Dankers & Koelemaij, 1989; Nehls & Thiel, 1993), which indicates that certain areas are more favourable for settlement and growth of mussels than others. Moreover, parts of the sediment accumulations often survive erosion and serve as a nucleus for the establishment of a new colony (Dankers, pers. comm.). Colonies in particular occur in the more wind-sheltered areas such as behind the islands (Nehls & Thiel, 1993), near the watershed and along gullies (Fig. 3; cf. Dankers & Koelemaij, 1989). Thus the long-term distribution pattern is relatively stable.

The surface area of individual beds can be up to several km^2 (e.g., Dijkema, 1989; De Vries, pers. comm.). The beds mostly consist of series of separate mudmounds, which are up to several hundreds of metres wide and several tens of metres broad, separated by small erosive channels and tidal pools (Van Straaten, 1949; Dankers, 1993). Erosion by waves, that come mainly from the west, often causes the intertidal mudmounds to be asymmetric with a steeper side dipping to the east, and to migrate eastward (Van Straaten, 1964; Fig. 5). They thus can serve as an indicator of the wave direction.

The biodeposits of the mussels partly settle between the shells of the colony. Resuspension of the (disintegrated) biodeposits is restricted by the protection of byssus threads (Van Straaten, 1954; McMaster, 1958; Ehlers, 1988) and by the dead and living shells. In addition, other organisms, such as algae, have a stabilizing effect on the sediment (Kuenen, 1942; Wanless et al, 1981; Dame & Dankers, 1988; Vos et al., 1988). The current velocity needed to erode the clay-rich sediments, which remain after the decay of the biodeposits, is higher than the velocity to entrain whole pellets (e.g., Nowell et al., 1981), in particular after some time has passed (Creutzberg & Postma, 1979). These mechanisms result in a strong sedimentation of fine-grained sediment in and around mussel colonies (e.g., Van Straaten, 1951; Beukema, 1976; Dittmann, 1987; Flemming & Delafontaine, 1994). In older mussel beds the clay-rich layers may be quite consolidated.

The mussels move by using their foot and by releasing byssus threads (Kuenen, 1942; Maas Geesteranus, 1942; De Blok & Geelen, 1958; McMaster, 1958; Seed, 1976). In this way the living animals remain at the sediment-water interface. The continuous movement of

Table V: Estimate of the total amount of suspended matter in intertidal mussel beds and muddy sandflats at end of the summer in the period 1975-1978 (Based on Dijkema, 1989)[1]

Area	Mussel beds[2]	Muddy Sandflats[3]
Western Wadden Sea	$0.9 * 10^9$ kg	$0.04 * 10^9$ kg
Eastern Wadden Sea	$2.2 * 10^9$ kg	$0.12 * 10^9$ kg
TOTAL	$3.1 * 10^9$ kg	$0.16 * 10^9$ kg

[1]: Map compiled from data collected in the period 1975-1978, during which the general mussel-bed and sediment distribution pattern were relatively stable (Dijkema, pers. comm.).

[2]: For the calculations an average mud (clay and silt) content of 40% (Favejee, 1951; Wiggers, 1960), a mean thickness of 20 cm of the mussel beds and an amount of 0.93 g per cm^{-3} wet sediment is assumed (Van Straaten, 1964; Ehlers, 1988; Flemming & Delafontaine, 1994, De Vries, pers. comm.; Dijkema, pers. comm.; Flemming, pers. comm.). This is a low estimate: compression in old mussel beds leads to consolidation of the mud layers and hence to higher amounts of mud. The sand stored in the intertidal mussel beds is 1.5 times the amount of suspended matter.

[3]: For the calculations a low average mud (clay and silt) content of 5% (Wiggers, 1960; De Glopper, 1967; Winkelmolen & Veenstra, 1974; Veenstra, 1978; Dijkema, 1989), a mean thickness of 15 cm is assumed and an amount of 0.373 g per cm^{-3} wet sediment for the mud (grain support, matrix in pores) is assumed (Flemming & Delafontaine, 1994; De Vries, pers. comm.; Dijkema, pers. comm.; pers obs.).

the shells creates holes, which are filled with clay and/or sand (Kamps, 1956; Oenema, 1988). Also coarser material such as shells and pebbles may be trapped (Dankers, pers. comm.). Locally this results in a high sand content within the mussel beds, up to 90% >50 µm (e.g., Favejee, 1951). Thus, the sediment of the mussel beds is accumulated by a combination of biodeposition and physical accretion. This process results in a strong vertical accretion of up to several $dm.y^{-1}$.

Although the blue mussel can live up to levels just below the neap high-water level (Verwey, 1952), the mussel beds on the intertidal flats of the Wadden Sea normally grow only up to about mean sea level (Van Straaten, 1964; pers. comm.; Dankers, 1986), so that the mussels are covered with water about half of the time (Dijkema, 1989). The height is restricted, because feeding duration decreases during vertical growth (Van Straaten, 1964; Seed, 1976; Dijkema, 1989). Also, at higher levels the mussels have to spend more energy to withstand the increasingly higher wave energy and the longer duration of aerial exposure (McMaster, 1958; Van Straaten, 1964; Seed, 1976; Widdows, 1984). The vertical growth also slows down because the increase in wave energy results in an increased removal of sediments from the mussel bed. Thus, the vertical accretion decreases asymptotically with increasing height (cf. Oenema, 1988; Flemming & Delafontaine, 1994). Due to these restrictions, mature intertidal mussel beds decrease in height from lower intertidal flats to flats at mean sea level (Dittmann, 1987). In the Dutch Wadden Sea they reach a thickness of up to 50-65 cm (Fig. 5; Van Straaten, 1964; De Vries, pers. comm.; Dijkema, pers. comm.). From the German Wadden Sea heights as much as 180 cm have been reported (Linke, 1954).

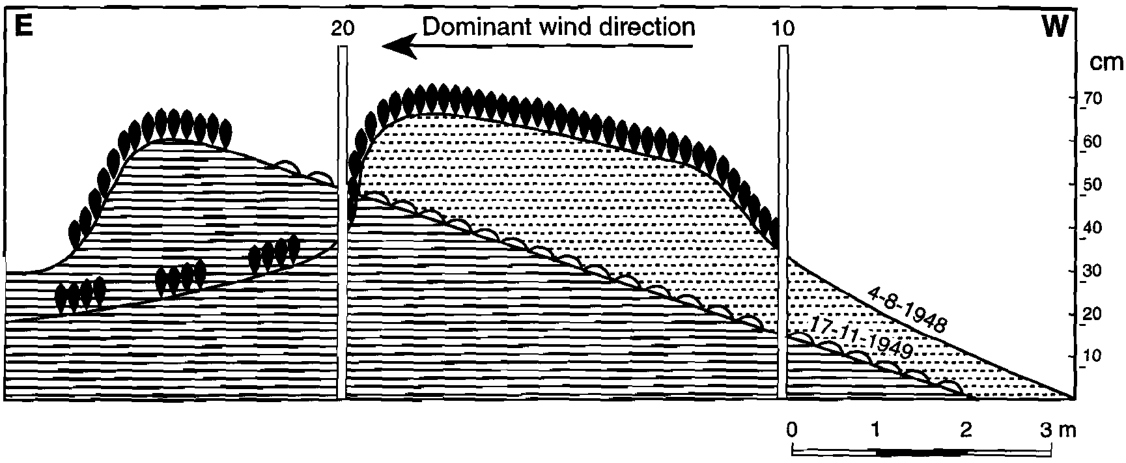

Figure 5: Transect through a mudmound and its eastward displacement in 470 days (Van Straaten, 1951)

The total area covered by mussel beds and the amount of sediment stored beneath them can be substantial. In the years 1975-1978, when the mussel population was larger than average, but not at its maximum, approximately 7.7 * 10^9 kg sediment was present below the intertidal mussel mounds (41.5 km^2) in the Dutch Wadden Sea (Table V) due to the activity of the mussels. By comparing the amount of fine-grained sediment below the intertidal mussel beds (3.1 * 10^9 kg) with the annual amount filtered from the water column (1.9 * 10^9 $kg.y^{-1}$, based on data of Beukema et al. (1978), Dankers et al. (1989) and Dankers & Koelemaij (1989)), and taking also resuspension into account, it is clear that these accumulations must have been built up over a period of at least several years. Indeed, natural mussel colonies have been observed to exist for several years in succession. The sediment shows annual growth patterns of dead shells (winter/storm deposits) alternating with muddy sand (Flemming & Delafontaine, 1994). In the same period approximately 35 km^2 culture beds and 35 km^2 wild subtidal mussel beds were present in the western Wadden Sea, with an average sediment thickness of 0.15 m (Dankers et al., 1989; Dankers pers. comm.). Assuming identical fine-grained sediment concentrations as in the mussel beds on the intertidal flats, an additional 9.8 * 10^9 kg of sediments was present in these subtidal beds.

The elevation of the mudmounds above the intertidal flats indicates that the shoal height (related to sealevel) would normally, i.e., without mudmounds, be lower (cf. Eysink, 1979; cf. Eysink & Biegel, 1992). The same goes for the subtidal shoals. Upon the disappearance of the intertidal mussel population by fishery and storms, the high mudmounds were removed. Because of the equilibrium height of the tidal flats it seems likely that in particular in the Eastern Dutch Wadden Sea the sand which formed part of the mussel beds was net exported from the backbarrier area, whereas the fine-grained part was resuspended. The destruction of the natural mussel population in the last years thus led to a substantial loss of sediment (in the period 1975-1978 some $8.3 * 10^6$ m^3 of sediment was present in the intertidal mudmounds alone).

Biological relevance

The formation of dense colonies is advantageous for mussels. Beds with a dense population of mussels (up to 3.500 $sp.m^{-2}$) are more resistant to wave action than solitary settlements on a tidal flat (Seed, 1976). Also, the shells and the mudmound topography increase bottom roughness and stimulate mixing of the water column, thus favouring the availability of food (cf. Dankers et al., 1989). Moreover, the elevation of the beds and the size of the population may serve to outcompete other filter feeders, such as *Cerastoderma edule*. At increasing heights *Cerastoderma edule* is handicapped relative to the mussel colony, because of a decreasing rate of growth and spat settlement (Jensen, 1991), increasing negative influence of low temperatures (Beukema, 1982a, b; Dörjes, 1982; BOEDE, 1985; Beukema & Cadée, 1986; Jensen, 1991; Hertweck, 1992) and because of a reduced food supply due to the filtration by the mussel population (Kamermans, 1993).

Mussel-induced muddy sandflats

In experiment 3 it was demonstrated that biodeposits are quite resistant to decay and can be transported over several kilometres. Pseudo-faeces are less resistant against transport than faeces (Oenema, 1988; Dankers, pers. comm.). The transported biodeposits accumulate for a large part on the intertidal muddy sandflats (5-8% clay, 42-95% sand), which commonly surround intertidal mussel colonies (Fig. 3; Wiggers, 1960; Dijkema, 1978; Lambeek, 1991). Also in tidal channels adjacent to subtidal colonies an increase in fine-grained sediment has been observed (Dankers, 1986). Once the biodeposits have settled, they disintegrate by bioturbation, bacterial activity and sediment coverage, and normally they loose their form (Van Straaten 1954; Oenema, 1988). As in the mussel beds, relatively high current velocities are needed to resuspend the fine-grained deposits, which result from this disintegration (cf. Creutzberg & Postma, 1979; Vos et al., 1988). The fine-grained material is mixed with sand by biogenic (Van Straaten, 1954; Haven & Morales-Alamo, 1972) and physical processes.

Mussel-induced muddy sandflats can form on locations where, in the absence of biodeposits, no strong accumulation of fine-grained sediments takes place (Fig. 3) (Van Straaten, 1964; Flemming, pers. comm.; cf. Haven & Morales-Alamo, 1972). Indeed, with the decrease of the number of mussel beds (due to intense fishing (Beukema, 1993; Dankers, 1993)), muddy sandflats also disappeared. In 1993 mussel beds and the surrounding muddy sandflats thus were completely absent in large parts of the Dutch Wadden Sea (pers. obs.; Dankers, pers. comm. Dijkema, pers. comm.; De Vries, pers. comm.). Upon re-establishment of new young mussel concentrations in 1994 the surrounding sandflats became muddier again (De Vries, pers. comm.).

From spring to early autumn intertidal deposits of muddy sand with a thickness of 15 cm on the average (De Vries, pers. comm.; Dijkema, pers. comm.) form in the vicinity of mussel colonies over areas of up to several km^2 (cf. Dijkema, 1989), in particular during relatively quiet weather conditions (Abrahamse & Buwalda, 1964). These mussel-induced muddy sandflats are best developed in relatively sheltered parts of the backbarrier area, in particular near the islands and on the tidal watersheds (cf. Dijkema, 1989). Mussel-induced muddy sandflats are mostly absent where currents are strong, e.g., in the vicinity of larger channels. It is estimated that in the period 1975-1978 at the end of the summer, $0.4 * 10^9$ kg of fine-grained sediments was present in intertidal mussel-induced muddy sandflats due to the influence of mussels (Table V). This is a low estimate; the silt and clay content of the muddy sandflats varies between 5-48% (Dijkema, 1989). The estimate is based on comparison of the sediments of the muddy sandflats with nearby normal sands, which almost totally lacked clay and silt (cf. Veenstra, 1978; Lambeek, 1991). This learns that at least 5% of the silt and clay deposited can at least be attributed to the influence of mussels.

Biological relevance

Biodeposits have a lower feeding value for the mussel than the original suspension. Therefore, the formation of transportable biodeposits enhances the feeding effectivity of the mussels. Reprocessing is avoided because the biodeposits are not easily resuspended, and can be transported over a relatively long distance away from the mussels who produced them. Once the biodeposits settle, they form a substratum for bacteria and several other organisms (Newell, 1965; Frankenberg & Smith, 1967; Van Es, 1982; BOEDE, 1985; Watling, 1989; Dittmann, 1987). Upon resuspension the material will have an increased feeding value for the, amongst others, bacteriophage mussels.

Moreover, pelletisation of suspended matter makes organic matter from the water column available to other benthic organisms. Field experiments suggest that mussel biodeposition significantly alters the fauna in favour of deposit feeders and omnivores (Dittmann, 1987).

Deposition in other parts of the backbarrier area

Pellets in the Wadden Sea are transported through the gullies towards areas where they can settle (Verwey, 1952). Transported material thus also accumulates on the lower and higher intertidal flats in physical structures and burrows (Fig. 6; Van Straaten, 1954).

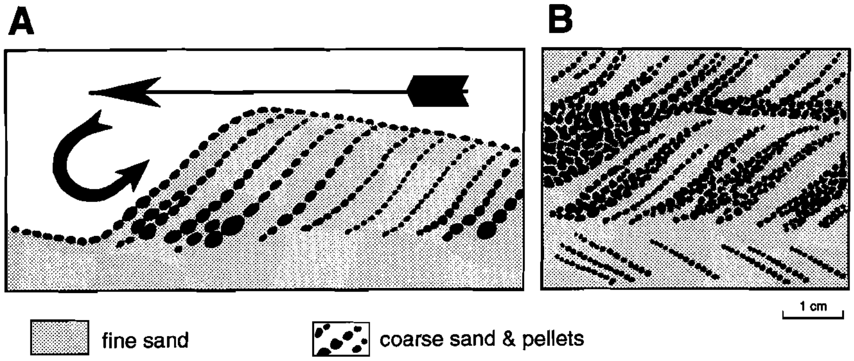

Figure 6: A: Biodeposits as part of foresets of current ripples; B shows detail (Van Straaten, 1951).

The deposits are partly reworked by deposit feeders, which also produce pellets (e.g., *Corophium volutator, Macoma balthica*) or loosely bound biodeposits (e.g., *Arenicola marina*). On the lower intertidal flats pellets are in particular deposited in ripple troughs (Van Straaten, 1954) to produce flasers or even well developed continuous layers, resulting in wavy or lenticular bedding (Reineck & Wunderlich 1968; Oost & Baas, 1994).

The amount of (disintegrated) mussel biodeposits admixed with sediments of the intertidal sandflats, mudflats and channels is not known. At the end of the summer about 3% by volume (i.e. ± 2% by weight) of the intertidal sand-flat sediments consists of fine-grained (<63 μm) material.

Although the importance of biodeposits formed by *M. edulis* decreases with increasing distance from the mussel beds, the processes which concentrate fine-grained material in the more quiet and/or higher parts of the backbarrier area (Oost & De Boer, 1994) also influence the distribution of the relatively light pellets. Part of the (partly disintegrated) pellets is therefore concentrated on the mudflats bordering the tidal marshes of the islands and of the mainland. From spring onwards, but in particular in late summer and early autumn (Stratingh & Venema, 1855; Van Es et al., 1980) fine-grained material accumulates on these mudflats until the autumn storms remove it partly or fully (Fig. 7; Kamps, 1956).
The high mud content in channel sediments bordering muddy tidal flats (Van Straaten, 1954) suggests that the pellets are partially deposited in the channels (cf. De Glopper, 1964). In particular in abandoned channels (Van den Berg, 1981; Oost et al., 1993) and abandoned gullies (pers. obs.) deposition during summer is dominated by fine-grained sediments. Experiment 2 demonstrated that the biodeposits behave hydrodynamically as quartz grains of silt/sand size. This explains the often quite thick layers of fine-grained sediment which are

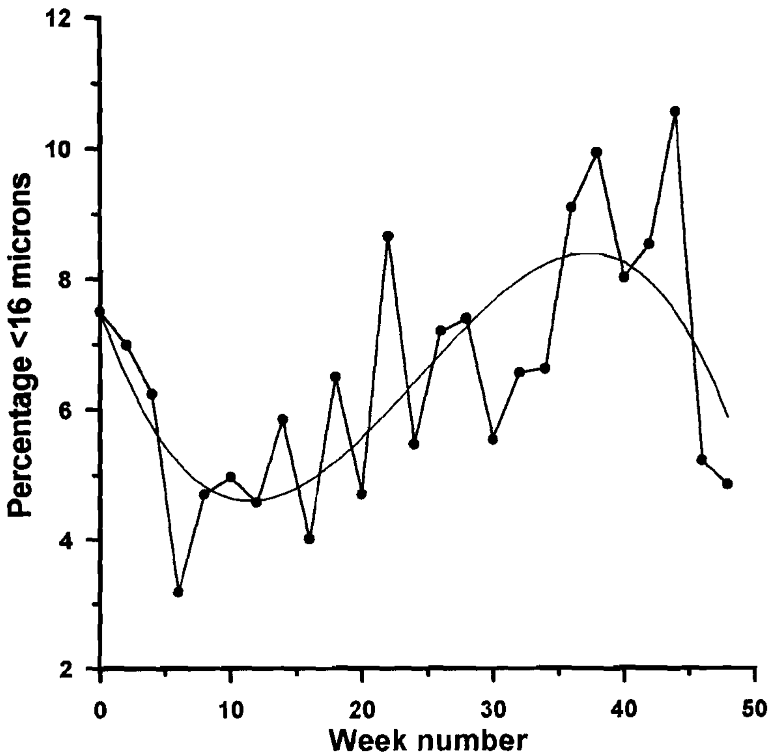

Figure 7: Annual variation of the percentage of sediment <16 μm in the top 0.5 cm of the intertidal mudflats near the mainland (average of 15 separate observations). Broken line is third order polynomial fit (data from Kamps, 1956).

frequently observed in various tidal deposits (cf. Allen, 1984). Terwindt & Breusers (1972) observed mud layers of 10 to 25 mm thick in subtidal estuarine deposits in the southern Netherlands. They argued that these can only form by settling from suspension if subsequent mud layers of a few mm accumulated over several slack water periods (cf. Postma, 1981). However, clay layers of up to 2 cm thick, which have been formed during the turn of the tide, were indeed observed in double mud drapes in subtidal deposits of the Eastern Scheldt (Visser, 1980). A logical explanation is that this fine-grained sediment originally consisted of pellets. Similar clay layers of several cm thick have also been observed in channel deposits in the Wadden Sea (e.g., Van Straaten, 1954; pers. obs.).

ANNUAL CYCLE OF FINE-GRAINED SEDIMENTATION

As discussed above, biodeposits of the mussel contribute to the deposition of fine-grained sediments in the Wadden Sea. Fine-grained sedimentation varies over the year by variations in the biological and physical conditions.

Biological conditions

The amount of biodeposits produced fluctuates over the seasons, as a result of seasonal variations in growth rate, population size and pumping rate of the mussels.

The growth rate per kg mussels varies over the year because of seasonal differences in food supply and metabolism. Primary food is almost exclusively produced by small-sized algae (Beukema, 1981). Food supply is much higher in spring (import from the North Sea) and summer (in situ production) than in autumn and winter, due to the available light (Beukema, 1981; Postma, 1981, 1984; Peinert et al., 1982; Verhagen, 1982; Dankers et al., 1989). Consequently the growth season of bivalves is concurrent with blooms of algae in spring and early summer (Kuenen, 1942; De Wilde, 1975; Beukema, 1982a; Verhagen, 1982; cf. Zwarts, 1991). Dankers et al. (1989) observed an increase in weight and length of mussels in the western Wadden Sea from April to August. Afterwards the mussels grow more slowly. They also decrease in weight (up to 50%) until December, after which they gain weight again. The exact nature of these changes is not yet fully understood (Dankers et al., 1989), but seems to be related to food limitations and reproduction (Anonymous, 1967).

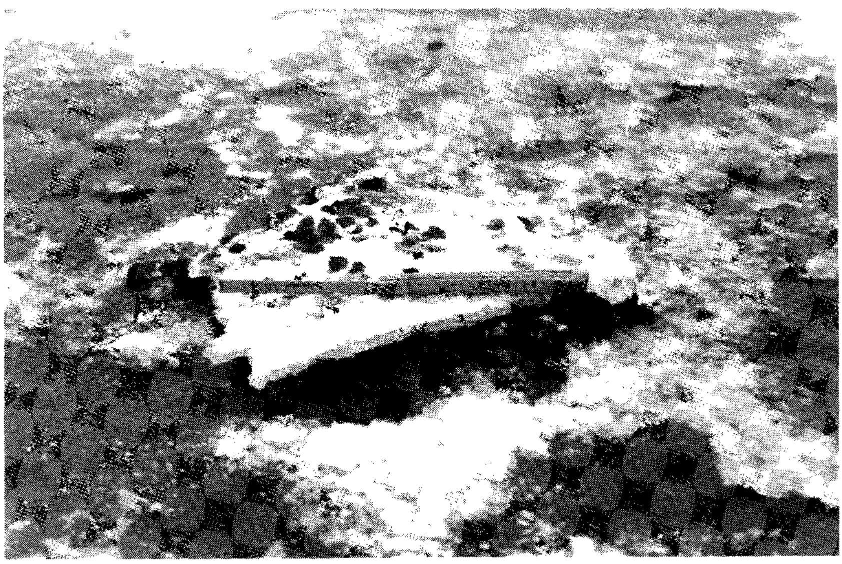

Figure 8: Mussels transported by ice, Jade Bay Germany.

Due to seasonal changes in environmental conditions mussel populations also vary seasonally (Verhagen, 1982; Dankers et al., 1989). The population tends to decrease during the winter season (August/September-February), and to increase afterwards (Bayne & Worall, 1980; Verhagen, 1982; Dankers et al., 1989). Important reductions of the population have been observed due to the destruction of mussel beds and the underlying fine-grained deposits by moving ice (Fig. 8) and by erosion during storms (Kuenen, 1942; Kamps, 1956; Van Straaten, 1964; Dankers et al., 1989; Dankers & Koelemaij, 1989; Flemming et al., 1993; Dijkema, pers. comm.). When ice forms or storms occur in the winter season, the filtration rate of the mussel is also somewhat reduced (cf. Smith & Frey, 1985).

As a result of these seasonal variations, the total amount of faeces and pseudo-faeces produced also fluctuates, being lowest in winter and highest in spring to early autumn (cf. Dankers, 1986). Furthermore, due the shorter length of the day in winter, the number of diatoms decreases (De Wilde & Beukema, 1984), and with that also their sediment stabilizing effect (cf. De Boer, 1981; Vos et al., 1988). The effect is in particular important during low-energy (fair weather) conditions (cf. De Jonge, 1992; Eisma, 1993).

Physical conditions

Over the seasons, physical conditions, such as wind force, temperature and tides, fluctuate markedly in the temperate Wadden Sea area. These physical mechanisms cause the suspended sediment concentrations in the Wadden Sea to be higher in autumn and winter than during summer (Fig. 4, Dronkers, 1984; cf. De Wilde & Beukema, 1984; cf. Hickel, 1984).

Wind waves are among the most important factors causing (re)suspension of sediment (Dronkers, 1984; De Jonge, 1992). Wind waves cause an increased bottom stress and a reduction of the slack water periods during which sediment can settle (Dronkers, 1984). The annual suspended sediment flux between the shoals and the channels under the influence of wind waves in the Dutch Wadden Sea area is estimated to be 1.5-2 * 10^{11} kg (Table VI).

Winds are predominantly from the SW-NW (Vroom et al., 1989), and they increase the height of the water level (IJnsen, 1987). Therefore, suspended sediment transport through the Wadden Sea, under the influence of wind waves, is mainly to the east (De Groot, 1962; e.g., De Boer, 1979; e.g., Reineck, 1980; e.g., De Boer et al., 1991). In autumn and winter winds are strongest and many storms occur. As a result, the annual maximum high-water levels occur mainly between October and February (cf. IJnsen, 1974). The waves generated by the stronger winds in this period result in a strong resuspension of fine-grained sediment. During storms, suspended sediment concentrations temporarily reach values of 0.8 $g.l^{-1}$ (De Haas & Eisma, 1993) to even several $g.l^{-1}$ (Van Sijp, 1988). A value of ± 50 $g.l^{-1}$ was reported south of Ameland by De Boer (1979). At such high concentrations (>5-10 $g.l^{-1}$) viscosity of the water increases. This decreases the influence of waves on the bottom (Eisma, 1993), probably resulting in an equilibrium suspension concentration and layer thickness (Genuchten, 1984).

Table VI: Estimated orders of magnitude of annual sediment transport and deposition in the Dutch Wadden Sea

Flux or deposition	Amount
Filtration of minerals by mussels	2.6-15.1 * 10^9 kg.y^{-1}
Fine-grained sediments below intertidal mussel beds, 1975-1978	3.1 * 10^9 kg
Fine-grained sediments in mussel-induced muddy sandflats, 1975-1978	0.16 * 10^9 kg
Sand below intertidal mussel beds in the period 1975-1978	4.65 * 10^9 kg
Long-term average net sedimentation of suspended matter, including the Dollard area	1.4$^{(1)}$-3.0$^{(2)}$ * 10^9 kg.y^{-1}
Long-term average net sedimentation of sand, including Dollard area(3)	14.0-36.9 * 10^9 kg.y^{-1}
Annual sum of wind-induced flux of suspended sediment between the shoals and the channels and *vice versa* (up to watershed of Borkum)(4)	1.5-2.0 * 10^{11} kg.y^{-1}
Annual sum of flux of suspended sediment through the inlets (up to the Dollard)	0.3$^{(5)}$-1.3$^{(6)}$ * 10^{11} kg.y^{-1}

(1) Eysink (1979),
(2) Essink et al. (1983)
(3) Rijkswaterstaat (1981)
(4) Based on De Jonge (1992) and Eysink (1979)
(5) Based on a mean concentration of 10 mg.l^{-1} in the North Sea ((Postma, 1981; Eisma & Kalf, 1987), a duration of the tide of 12,42 h, and a tidal prism of 4,378 * 10^9 l (Vroom et al., 1989)
(6) Based on a mean concentration of 42,5 mg.l^{-1} in the backbarrier area (Dankers et al., 1989), a duration of the tide of 12,42 h, and a tidal prism of 4,378 * 10^9 l (Vroom et al., 1989).

Thick layers of mud settle quickly at the end of storms, because resuspended sediments have a large median grainsize (cf. Anderson, 1976; Peinert et al., 1982). Quick re-settling is also suggested by the similarity of the suspended sediment distribution patterns in winter and summer (Fig. 4). Moreover, the similarity indicates that the influence of storms on the distribution pattern is subordinate to other effects (Dronkers, 1984).

The lower temperatures in winter result in a decrease in settling velocity of suspended sediment (settling is ± 17% slower in sea water of 6°C than in sea water of 20°C), mainly due to the increase in viscosity of the water (Anderson, 1973; Piekhaar, 1980; Eisma, 1993). Furthermore, sea ice may form, and freeze onto the tidal flats (Abrahamse & Buwalda, 1964; Reineck, 1980; Keuper, 1985). Even in the tidal channels it may freeze onto the sediment surface (IJnsen, 1988). This happens in particular at low tide, and parts of the sediment are incorporated in the ice (Stratingh & Venema, 1855; e.g., Abrahamse & Buwalda, 1964; De Vries, pers. comm.; pers. obs.). During the flood the ice is partly lifted and transported by currents and wind, and deposited on the intertidal flats and supratidal marshes or carried during ebb through the channels to the North Sea. Upon thaw large amounts of floating ice are transported from the Wadden Sea through the channels and over the ebb-tidal deltas (e.g., Keuper, 1985). Due to the higher temperatures of the North Sea water (Keuper, 1985) the ice melts quickly and looses its sediment, resulting in fine-grained suspended sediments.

An additional factor is that in the winter season tides are stronger on the Northern Hemisphere, thus generating stronger tidal currents with a greater erosive power.

Annual change in the erosion and deposition pattern

The above discussed seasonality in biological and physical conditions results in a marked annual cycle of erosion and deposition of biodeposits and of fine-grained sediments in general (Fig. 9):

Total filtration of suspended matter is relatively strong from spring to autumn. Part of the (pseudo-)faeces accumulates beneath the mussel beds, which also grow due to physical accretion. A smaller part accumulates on the mussel-induced muddy sandflats from spring to autumn. Moreover, a part accumulates on the sandflats, in the channels and on the mudflats.

During winter biological and physical conditions result in a slower growth or even erosion of the mudmounds (e.g., Oenema, 1988; Flemming & Delafontaine, 1994), and in a decrease of the concentration of fines in the mussel-induced muddy sandflats (e.g., Abrahamse & Buwalda, 1964; Dijkema, pers. comm.), in the intertidal mudflats (Fig. 7; Kamps, 1956; Eysink, 1979), and probably also on the sandflats (Van Es et al., 1980). Comparable annual changes in mud content of intertidal flats have been observed in the eastern Scheldt (Oenema, 1988). Since long-term net sedimentation rates are low ($mm.y^{-1}$), while high deposition rates occur in summer ($cm.y^{-1}$), erosion of fines from the sediment indeed must occur in the winter season (e.g., Stratingh & Venema, 1855; Kamps, 1956; De Boer, 1979; De Haas & Eisma, 1993).

In abandoned channels and gullies fine-grained sedimentation is more important in the summer half year, while sands dominate sedimentation during the winter half year (Van den Berg, 1981; Oost et al., 1993). This is partly due to the stronger tides and storms and related currents in winter.

Depending on the wind direction and the phase of the tide resuspended matter is transported partly towards the open sea or re-deposited in the Wadden Sea during or shortly after storms (Stratingh & Venema, 1855; Kamps, 1956; Van Straaten & Kuenen, 1957; De Haas & Eisma, 1993). On the supratidal marshes adjacent to the intertidal area, deposition of fluid mud layers up to several dm thick occurs (Stratingh & Venema, 1855; Kamps, 1956; Dijkema et al., 1988; De Vries, pers. comm.; Dijkema, pers. comm.; pers. obs.). This in particular happens in the second half of the autumn, a period formerly called the 'mud months' by the owners of coastal salt marshes (e.g., Stratingh & Venema, 1855; Kamps, 1956). Furthermore, ice on the supratidal marshes often contains fine-grained sediments

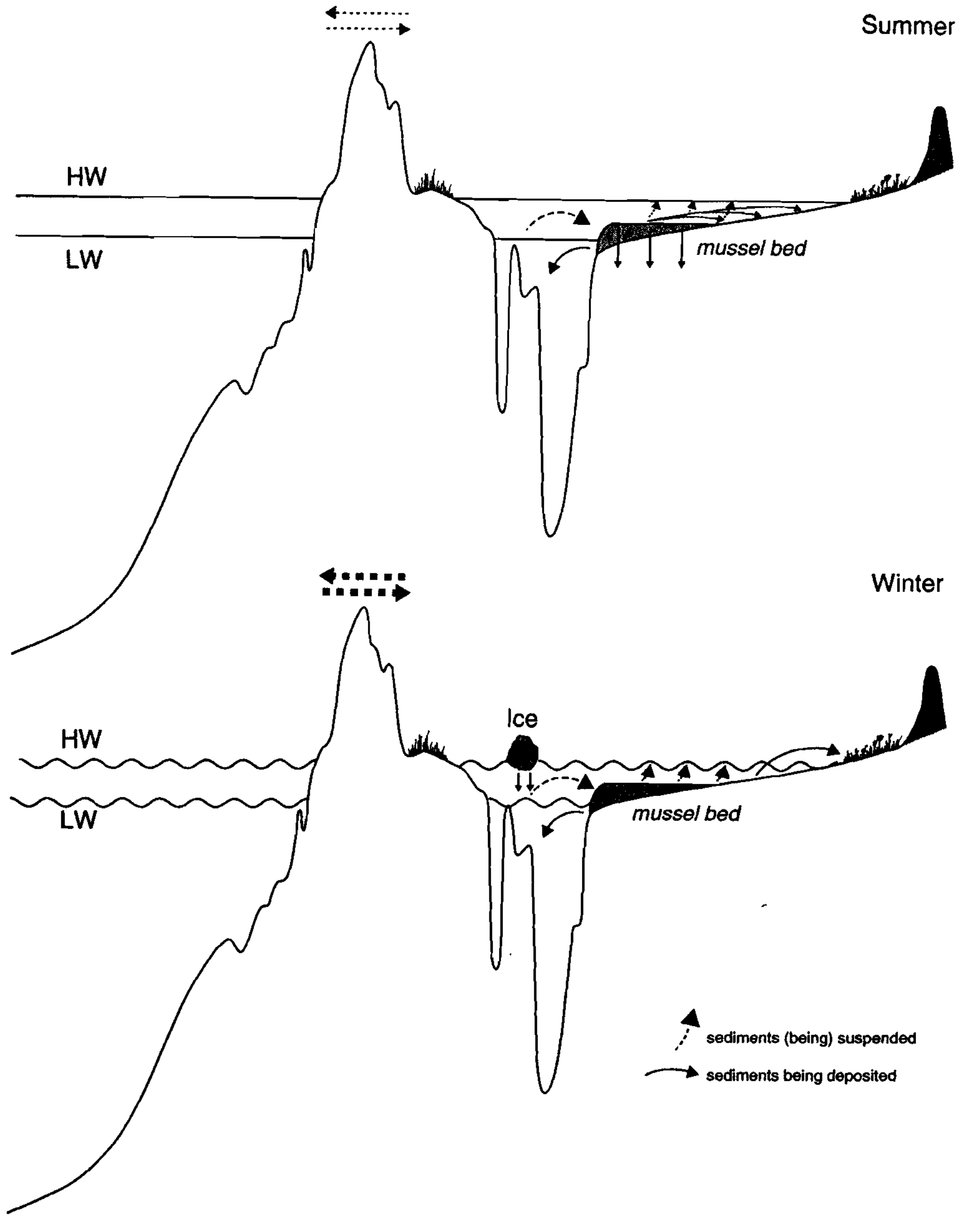

Figure 9: Annual cycle in the sedimentation pattern for fine-grained sediments. Arrows give directions and relative magnitudes of transport.

from the mudflats in front of the marshes. This is another way of mud transport to the marsh areas (Stratingh & Venema, 1855; Dijkema et al., 1980).

The resuspended matter, which is exported out of the Wadden Sea by ice and by storm-induced transport during winter is concentrated along the North Sea coast of the Dutch Wadden Sea and transported eastward (cf. Otto et al., 1990). As a consequence suspended matter concentrations in the North Sea, near the Wadden Sea, are higher during the winter than during the summer (cf. Eisma & Kalf, 1987). Part of the sediment re-enters the back-barrier area with the flood. An unknown part of the sediment is transported beyond the estuarine circulation zone, which extends into the North Sea (Postma, 1981, 1984), and is 'lost' for the Wadden Sea. This in particular happens when strong outflow during ebb occurs, after storm flooding of the backbarrier basin. Ice rafting is probably of minor importance, because ice is concentrated on the ebb-tidal deltas and on the coast of the barrier islands (cf. Keuper, 1985). After the winter the cycle starts again; the exact timing and strength of sedimentation and erosion depending on both biotic and physical conditions.

THE INFLUENCE OF *MYTILUS EDULIS* AND FILTER FEEDERS IN GENERAL ON SEDIMENTATION PATTERNS

Summarizing: the high amounts of faeces and pseudo-faeces which are produced by mussels influence the suspended sediment concentration in the water column and the sedimentation (patterns) of fine-grained sediments in the Wadden Sea.

Mussel filtration activity forms a significant element in the suspended matter cycle in the backbarrier system, in particular because the filtering process increases the effective grain-size, as demonstrated by the above experiments. Filtration of non-organic suspended particles by *M. edulis* in the Dutch Wadden Sea is of the order of 2.6-15.1 * 10^9 $kg.y^{-1}$ (cf. Table VI). Thus, the maximum amount that can be filtered by the mussels is 5 times greater than the long-term average net deposition of fine-grained matter in the Dutch Wadden Sea, including the Dollard estuary (1.4-3.0 * 10^9 $kg.y^{-1}$; Eysink, 1979; Essink et al., 1983). In years with a low mussel biomass, all material which is filtered, can theoretically be deposited (if no substantial other fine-grained deposition occurs). In years with a high mussel biomass more than 80% of the annually formed biodeposits has to be resuspended. Even compared to the total flux of suspended sediment, which annually moves through the backbarrier area between the shoals and channels due to wind and waves (Table VI), the filtration is significant (0.01-0.1 times the flux).

Mussels also affect the sedimentation (patterns) of mud and sand. As demonstrated by the experiments, the biodeposits have a considerably larger quartz-equivalent grainsize than the original suspension. This allows settling of fine-grained sediments in areas where this would normally be difficult or impossible. Substantial amounts (of the order of 10^9 kg) are stored beneath the mussel beds, which, in the intertidal zone, mainly occur along the larger tidal channels, the ends of smaller gullies and on the watersheds. Besides fine-grained sediment these mudmounds mainly contain sand. These accumulations could not occur in the absence of mussel colonies. Another part of the fine-grained biodeposits settles in the vicinity of the mussel colonies thus forming muddy sand flats. Observations indicate that these muddy sand flats cannot exist in the absence of the filter-feeding mussel colonies. Besides these two areas, the biodeposits settle in other parts of the Wadden Sea. After some time most of the biodeposits is resuspended. Resuspension is in particular strong in the winter season. A part of the mud is then transported out of the backbarrier area, and another part is deposited on the tidal marshes.

Siliciclastic biodeposits in other marine deposits

The influence of filtration and biodeposition on fine-grained sedimentation patterns has been reported from a number of areas (Table VII). The relative importance depends on the ratio between the amount of biodeposition and the physical/chemical sedimentation. For instance, in the Gulf of Maine physical sedimentation rates are low (Youngbluth et al., 1989). Thus, the, in an absolute sense, low rates of biodeposition by *Meganyctiphanes norvegica* (Table VII) locally dominate sedimentation (Youngbluth et al., 1989). Even if physical sedimentation rates are high, biodeposition can still be an important factor in the overall sedimentation pattern, when the number of animals is substantial, as was illustrated by the above description of the impact of mussels in the Dutch Wadden Sea. This and other examples (*Calianassa major* (Pryor, 1975), *Macoma balthica* (Risk & Moffat, 1975) and *M. edulis* (Flemming & Delafontaine, 1994)) demonstrate that the biodeposition can be locally several $kg.m^{-2}.day^{-1}$ (Table VII).

In particular estuaries and (intertidal) lagoons are characterized by a low faunal diversity in combination with a high biomass (Nybakken, 1982). The results often in high densities of certain biodepositing filter or deposit feeding species. Therefore, such shallow marine areas can be strongly influenced (at least locally) by filtration and biodeposition.

Siliciclastic biodeposits in ancient marine deposits

Judging from the abundance of fossils of filter feeders in ancient lagoonal and estuarine deposits, biodeposition must also have been important in these environments in the geological past. At present, important biodepositing filter feeders are some species of polychaetes, ascidia, crustacea and bivalves. Moreover, brachiopods are likely to also have formed important deposits in the geological past, in particular in the Paleozoic (cf. Rudwick, 1970). Where fossils of these groups of filter feeders are present, their influence on sedimentation has to be considered.

An example are the bioherms of the Triassic bivalve *Placunopsis ostracina*, which lived in an oyster-like fashion (Bachmann, 1979). Around and between the bioherms fine-grained marls and pelletised marls settled. The settling on a living ammonite indicates that

Table VII: Overview of various filterfeeders and their filtration of sediment from literature.
a.a. = annual average, d.a. = daily average, bur = burrow, max = maximum

Animal	Feeding mode	Environment	Density (animals.m^{-2})	Biodeposits produced (dry weight)		References
				(mg.sp^{-1}.h^{-1})	(g.m^{-2}day^{-1})	
Onuphis microceph. (Polychaeta)	filter	Sand-dominated tidal creek and beachface-shoreface, wave dominated zone Atlantic coast of United States	max: 120; average: 26	1.7	max (120): 4.8	Pryor, 1975
Styela plicata (Ascidia)	filter	Subtidal, Hiroshima bay, Japan	---	Feb: 1.4 Apr: 1.1 Oct: 7.6 Dec: 1.2	---	Arakawa et al., 1971
Ciona intestinalis (Ascidia)	filter	Subtidal, Hiroshima bay, Japan	---	Feb: 0.5 Apr: 0.7 Dec: 1.3	---	Arakawa et al., 1971
Meganyctiphanes norvegica (Crustacea)	filter	Offshore, Gulf of Maine & off Georges Bank	10-2857m^{-3}		0.06-0.2 min: 0.05-0.2	Youngbluth et al., 1989
Calianassa major (Crustacea)	mainly filter	Atlantic coast of United States	lagoon: 450 burrows beach-foreshore: 2-8 10 m water depth: 20 1 burrow can contain >10 animals	144.7 mg/burrow	max (450 bur.): 1562.4 min: 1406.2-1531.2	Pryor, 1975
Barnacle community	filter	Boat basin, Fort Pierce, Florida	---		flux: 20-50 min: 15-38	Hoskin, 1980

Table VII (cont'd): Overview of various filterfeeders and their filtration of sediment from literature.
a.a. = annual average, d.a. = daily average, bur = burrow, max = maximum

Animal	Feeding mode	Environment	Density (animals.m^{-2})	Biodeposition rate (dry weight)		References
				($mg.sp^{-1}.h^{-1}$)	($g.m^{-2}.day^{-1}$)	
Geukenisia demissa	filter	Transitional marsh, Durant Marsh, Sapelo Island, Georgia; flooded: 2.8 $h.day^{-1}$	46.7	Spring: 8 Summer: 29 Autumn: 25 Winter: 5	1.0 3.8 3.3 0.7	Smith & Frey, 1985
Geukenisia demissa	filter	Ponded water low marsh, Durant Marsh, Sapelo Island, Georgia; flooded: 5.6 $h.day^{-1}$	18.3	Spring: 29 Summer:114 Autumn: 66 Winter: 6	3.1 11.7 6.8 0.6	Smith & Frey, 1985
Choromytil. chorus	filter	Queule River Estuary, Chile			a.a.: 271	Jaramillo et al., 1992 In: Smaal & Prins, 1993
Mytilus chilensis	filter	Queule River Estuary, Chile			a.a.: 234	Jaramillo et al., 1992
Mytilus galloprovincialis	filter	Subtidal, Hiroshima bay, Japan	---	Feb: 3.1; Apr: 2.0 Aug: 0.6; Oct: 6.2 Dec: 2.2		Arakawa et al., 1971
Mytilus galloprovincialis	filter	Subtidal, rope culture			702	P. Camacho et al., 1991 In: Smaal & Prins, 1993
Mytilus edulis	filter	Subtidal rocks, Balthic Sea		a.a.: 2.9863 min: 2.4384		Kautsky & Evans, 1987 In: Dittmann, 1987
Mytilus edulis	filter	Laboratory, intertidal rocks, Japan			a.a.: 32.6	Tsuchiya, 1980 In: Dittmann, 1987
Mytilus edulis 1 yr	filter	Intertidal flats, macrotidal German Wadden Sea			max: 530 (partly accretion of sand)	Flemming & Delafon-taine, 1994

Table VII (cont'd): Overview of various filterfeeders and their filtration of sediment from literature.
a.a. = annual average, d.a. = daily average, bur = burrow, max = maximum

Animal	Feeding mode	Environment	Density (animals.m^{-2})	Biodeposits produced (dry weight)		References
				($mg.sp^{-1}.h^{-1}$)	($g.m^{-2}.day^{-1}$)	
Mytilus edulis	filter	Intertidal flats, mesotidal German Wadden Sea	average: 1,300	d.a.: 3.1 a.a.: 2.2	a.a.: 68.5; min: 57.5, max: 159.7	Dittmann, 1987
Mytilus edulis	filter	Intertidal flats, mesotidal Dutch Wadden Sea, flooding 18 $h.day^{-1}$, total production	tidal flat average: 5; mus. beds: 333-1017	a.a. faeces: 5.4 a.a pseudof: 45.9	flats: 6; mus. beds: 410-1252	Beukema, 1982; Dankers et al. & Dankers & Koelemaij, 1989
Mytilus edulis	filter	Subtidal flats, mesotidal Dutch Wadden Sea (wild), total production	786	a.a. faeces: 7.2 a.a. pseudof: 61.1	1,288	Dankers & Koelemaij, 1989; Dankers et al., 1989
Crassostrea virginica	filter	Laboratory, York River, Virginia		May-Oct.: 9.6 min: 7.4-8.7 max: 23.3		Haven & Morales-Alamo, 1972
Crassostrea gigas	filter	Subtidal, Hiroshima Bay, Japan	420,000/raft (198 m^2)	a.a.: 5.6; Feb: 3.7 Apr: 1.7; Aug:4.2; Oct: 12; Dec: 5.7		Arakawa et al., 1971
Crassostrea gigas	filter	Departure Bay, subtidal			3.9	In: Smaal & Prins, 1993
Cerastoderma edule	filter	Intertidal, Dutch Wadden Sea; flooded 12 $h.day^{-1}$	late winter: 48 summer: 155	a.a. faeces: 3.9 a.a. biodep.: 9.8	a.a. faeces: 4.7 a.a. biodep: 19.9	Verwey, 1952; Beukema, 1982; Dankers et al., 1989
Macoma balthica	filter/ depos.	Intertidal flats, macrotidal Minas Basin, Bay of Fundy, Canada, flood. 12 $h.day^{-1}$, active 200 $days.yr^{-1}$	maximum: 3,500; average: 670	normal: 20.8 a.a. 5.7	max (3,500, normal rate): 875	Risk & Moffat, 1977
Macoma balthica	filter/ depos.	Subtidal, 15°C	---	d.a. 21.7		Tunniclife In: Risk & Moffat, 1977
Mya arenaria	filter	---	---	d.a. 1.2		Haven & Morales-Al., 1966

Placunopsis must have been a filter feeder (Bachmann, 1979). The fine-grained marls and pellets thus may well be the biodeposits of *Placunopsis*.

Also the oyster-related *Gryphea* sp. were filter feeders. These fossils often occur in huge numbers in living position in the fine-grained sediment, as is the case with mussels. Most likely, *Gryphea* influenced its environment in the same way as mussels do, to its own benefit. This is also suggested by the observation of pelletal grains within the doublets of *Gryphea* shells (Oost & Molenaar, unpubl. data). The large numbers of *Gryphea* in fine-grained Liassic deposits suggest that the animal may have had a significant influence on local sedimentation patterns.

More than 70% of the sedimentary record consists of fine-grained siliciclastic sediments (Pryor, 1975; Potter et al., 1980). In carbonate settings well recognizable faecal pellets and peloids form an important part of the sedimentary record (cf. Lees, 1975), because pelletisation enhances early lithification and preservation (Bathurst, 1971; Wanless et al., 1981). It depends on the firmness of the pellets whether they can still be recognized in sedimentary deposits. For instance, the faecal pellets of *Cerastoderma edule* (an important filterfeeder in the Wadden Sea which filters volumes of water comparable to *Mytilus edulis*) are much firmer than those of *M. edulis*, and have been recognized at depths of over 0.6 m beneath the sediment surface (Van Straaten, 1955; pers. comm.). Siliciclastic biodeposits commonly disintegrate or are strongly deformed during (early) diagenesis. For this reason there are only a few examples of well recognizable fossil siliciclastic pellet deposits (e.g., the Upper Devonian Catskill Formation; Cuomo & Rhoads, 1987), although at present faecal pellets are abundant in many places (Reineck & Singh, 1980). That (deformed) pellets and pel-aggregates can be the principal component of (in particular) marine shales and muds is observed in thin sections. They form up to mm-large, flattened, semi-concentric masses defined by fabric, organic content, and colour (Pryor, 1975; Potter et al., 1980). Furthermore the common presence of (pelletal) glauconite, which often forms by alteration of faecal pellets (Pryor, 1975; Blatt et al., 1980), or a local change in clay mineralogy, brought about by the digestive action (cf. Pryor, 1975), may indicate that a major part of the original sediment consisted of biodeposits.

Also, sedimentary structures can indicate the presence of important sources of biodeposits. When part of the lee-side of ripples or dunes consists of mm-sized clay particles next to sand grains (cf. Fig. 6) it can be expected that these were originally deposited as pellets. This can even result in sedimentary structures such as wave or current ripples, which almost exclusively consist of mud (Fig. 10). Furthermore, the occurrence of several-cm-thick clay layers in tidal areas may be indicative of biodeposits.

Figure 10: Ripples totally consisting of biodeposits (original grainsize mainly clay to silt) of *Corophium volutator*, Dollard Estuary, The Netherlands.

CONCLUSIONS

The common blue mussel *M. edulis* strongly influences the sedimentation of fine-grained sediments in the Dutch Wadden Sea by producing faecal pellets and pseudofaeces. Experiments demonstrate that the pellets behave hydrodynamically as grains of silt/sand size. The biodeposits are fairly resistant to phases of transport and rest. Upon disintegration they often fall apart in particles (flocs) which are larger than the particles in the original suspension.

The amount of annually filtrated suspended sediment is equal to, or larger than the annual average net deposition of fines in the Wadden Sea. The larger part is resuspended. It is clear that the biodeposits have a great influence on sedimentation patterns, considering their different hydrodynamic properties as compared to those of suspended fine-grained sediment. Under natural conditions part of the biodeposits is stored beneath the mussel beds, which accrete by the accumulation of clays and sands. The growth of the beds is defined by the interaction between physical and biological parameters, such as wave action and food supply. In the course of several years large amounts of sediment can thus be stored. Around the mussel beds, muddy sandflats occur, due to sedimentation of the biodeposits of mussels.

The biologically and physically induced seasonality of biodeposition is part of the annual cycle in the sedimentation pattern of fine-grained sediments. In general, fine-grained sediments are stored beneath the mussel beds and on the intertidal flats in summer, and they are (partly) resuspended in winter. The resuspended sediments are partly deposited on the tidal marshes and/or exported towards the North Sea.

The amount of biodeposits produced in other areas demonstrates that biodeposition often plays an important role in the deposition of fine-grained sediments. It is in particular important in estuaries and (intertidal) lagoons, which are characterized by a high biomass and a low faunal diversity. Similar conditions prevailed in such environments in the geological past, and it can be expected that biodeposits form a substantial part of shallow marine fine-grained siliciclastic deposits in the sedimentary column.

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REFERENCES

Abrahamse, J. & Buwalda, J.D., 1964: Wadlopen, 'land'-loperij bij laagwater. In: Anderson, W.F., Abrahamse, J., Buwalda, J.D. & Van Straaten, L.M.J.U. (Eds): Het Waddenboek. Thieme, Zutphen, 9-36.

Allen, J.R.L., 1984: Sedimentary structures their character and physical basis. Developments in Sedimentology 30 A&B, Elsevier, Amsterdam, 593 & 663 pp.

Anderson, F.E., 1973: Observations of some sedimentary processes acting on a tidal flat. Marine Geology, 14, 101-116.

Anderson, F.E., 1976: Rapid settling rates, observed in sediments, resuspended by boat waves over a tidal flat. Neth. J. Sea Res., 10, 44-58.

Anonymous, 1967: Mussel farming in the Netherlands, Chapter 2, 49-103.

Arakawa, K.Y., 1963: Studies on the Molluscan faeces (I). Publ. Seto Marine Biological Laboratory, 11, 2, 185-208.

Arakawa, K.Y., 1965: Studies on the Molluscan faeces (II). Publ. Seto Marine Biological Laboratory, 13, 1, 1-21.

Arakawa, K.Y., 1968: Studies on the Molluscan faeces (III). Publ. Seto marine Biological Laboratory, 16, 2, 127-139.

Arakawa, K.Y., 1971: Studies on the faecal pellets of marine invertebrates (excluding molluscs) I. Publ. Seto Marine Biological Laboratory, 19, 4, 231-241.

Arakawa, K.Y., 1972: Studies on the Molluscan faeces (IV). Publ. Seto Marine Biological Laboratory, 19, 6, 347-357.

Arakawa, K.Y., Kusuki, Y. & Kamigaki, M., 1971: Studies on biodeposition in oyster beds (I) - economic density for oyster culture. Venus, 30, 113-128.

Bachmann, G.H., 1979: Bioherme der Muschel *Plancunopsis ostracina* v. Schlotheim und ihre Diagenese, N. Jb. Geol. Paläont. Abh., 158, 3, 381-407

Bathurst, R.G.C., 1971: Carbonate sediments and their diagenesis. Elsevier, NY, 620 pp.
Bayne, B.L., 1984: Ecological consequences of stress. In: Bayne, B.L., Brown, D.A. Burns, K., Dixon, D.R., Ivanovici, A, Livingstone, D.R., Lowe, D.M., Moore, M.N., Stebbing, A.R.D. & Widdows, J. (Eds): The effects of stress and pollution on marine animals. Praeger Sci. N.Y., 141-157.
Bayne, B.L. & Newell, R.C., 1983: Physiological energetics of marine molluscs. In: K.M. Wilbur (Ed.): The Mollusca, Vol. III, Acad. Press, NY, 407 pp.
Bayne, B.L. & Worall, C.M., 1980: Growth and production of mussels *Mytilus edulis* from two populations. Mar. Ecol. Progr. Ser. 3, 317-328.
Bayne, B.L., Thompson, R.J. & Widdows, J., 1976: Feeding and digestion. In: Bayne, B.L. (Ed.): Marine mussels: their ecology and physiology, Int. Biol. Progr. 10. Cambridge Univ. Press, Cambridge, 121-158.
Bayne, B.L., Thompson, R.J. & Widdows, J., 1977: Physiological measurements on estuarine bivalve molluscs in the field. In: Keegan, B.F., O'Ceidigh, P. & Boaden, P.J.S. (Eds): Biology of benthic organisms, 11th Eur. Symp. on Mar. Biol. Galway, 57-68.
Bernard, F.R., 1974: Particle sorting and labial palp function in the pacific oyster *Crassostrea gigas* (Thunberg, 1975). Biol. Bull., 146, 1-10.
Beukema, J.J., 1976: Dierlijk leven op en in de bodem. In: Abrahamse, J., Joenje, W. & Van Leeuwen-Seelt, N., Waddenzee, Natuurgebied van Nederland, Duitsland en Denemarken, De Lange/Van Leer, Deventer, 125-131.
Beukema, J.J., 1981: The role of the larger invertebrates in the Wadden Sea ecosystem. Marine zoology, 4, 211-221.
Beukema, J.J., 1982a: Annual variation in reproductive success and biomass of the major macrozoobenthic species living in a tidal flat area of the Wadden Sea. Neth. J. Sea Res., 16, 37-45.
Beukema, J.J., 1982b: Calcimass and carbonate production by molluscs on the tidal flats in the Dutch Wadden Sea: II the edible cockle, *Cerastoderma edule*. Neth. J. Sea Res., 15, 391-405.
Beukema, J.J., 1992: Expected changes in the Wadden Sea benthos in a warmer world: Lessons form periods with mild winters. Neth. J. Sea Res., 30, 73-79.
Beukema, J.J., 1993: Increased mortality in alternative bivalve prey during a period when the tidal flats of the Dutch Wadden Sea were devoid of mussels. Neth. J. Sea Res., 31, 395-406.
Beukema, J.J., De Bruin, W. & Jansen, J.J.M., 1978: Biomass and species richness of the macrobenthic animals living on the tidal flats of the Dutch Wadden Sea: long-term changes during a period with mild winters. Neth. J. Sea Res., 12, 58-77.
Beukema, J.J., Cadée, G.C. 1986: Zoobenthos responses to eutrophication of the Dutch Wadden Sea. Ophelia, 26, 55-64.
Blatt, H., Middleton, G.V. & Murray, R.C., 1980: Origin of sedimentary rocks, 2nd Ed. Prentice Hall, New Jersey, 782 pp.
BOEDE, 1985: Biological research Ems-Dollard Estuary by a survey of the ecosystem research during 1973 and 1982/by biological sturdy of the Ems-Dollard Estuary (BOEDE), Rijkswaterstaat comm. no. 40, 182 pp.
Cadée, G.C., 1982: Tidal and seasonal variation in particulate and dissolved organic carbon in the western Wadden Sea. Neth. J. Sea Res., 16, 228-249.
Cadée, G.C. & Hegeman, J., 1986: Seasonal and annual variation in *Phaeocystis pouchetii* (Haptophyceae) in the westernmost inlet of the Wadden Sea during the 1973 to 1985 period. Neth. J. Sea Res., 20, 29-36.
Cadée, G.C. & Hegeman, J., 1991: Historical phytoplankton data of the Marsdiep. Hydrobiol. Bull., 24, 111-118.
Creutzberg, F. & Postma, H., 1979: An experimental approach to the distribution of mud in the Southern North Sea. Neth. J. Sea Res., 13, 99-116.
Cuomo, M.C. & Rhoads, D.C., 1987: Biogenic sedimentary fabrics associated with pioneering polychaete assemblages: modern and ancient. J. Sed. Petr., 57, 537-543.

Dame, R.F. & Dankers, N., 1988: Uptake and release of materials by a Wadden Sea mussel bed. J. Esp. Mar. Biol. Ecol., 118, 207-216.

Dankers, N., 1986: Onderzoek naar de rol van de mossel en de mosselcultuur in de Waddenzee. Rijksinstituut voor Natuurbeheer, rapport 86/14, 36 pp.

Dankers, N., 1993: Integrated estuarine management - obtaining a sustainable yield of bivalve resources while maintaining environmental quality. In: R.F. Dame (Ed.): Bivalve filter feeders in estuarine and coastal ecosystem processes. Springer, Berlin, 479-511.

Dankers, N. & Koelemaij, K., 1989: Variations in the mussel population of the Dutch Wadden Sea in relation to monitoring. Helogolander Meeresunters., 43: 529-535.

Dankers, N., Koelemij, K. & Zegers, J., 1989: De rol van de mossel en de mosselcultuur in het ecosysteem van de Waddenzee. RIN-Report 89/9: 66 pp.

De Blok, J.W. & Geelen, H.J.F. 1958: The substratum required for the settling of mussels (Mytilus edulis L.). Archives Neerlandaises de zoologie, XIII, Supplement, Vol. Jub, 446-460.

De Boer, M., 1979: Morfologisch onderzoek Ameland. Verslag van het onderzoek op het Amelander Wantij in 1973. Rijkswaterstaat, Studiekring Hoorn, Nota WWKZ-79.H005, 67 pp.

De Boer, M., G. Kool & M.F. Lieshout, 1991: Erosie en sedimentatie in de binnendelta van het Zeegat van Ameland, 1926-1984, Deelonderzoek no.4. - Rijkswaterstaat, Directie Noord-Holland, Report ANVX-91.H202, 42 pp.

De Boer, P.L., 1981: Mechanical effects of micro-organisms on intertidal bedform migration. Sedimentology, 28: 129-132.

De Groot, A.J., 1962: Mangaantoestand van nederlandse en duitse Holocene sedimenten in verband met slibtransport en bodemgenese. PhD-Thesis, utrecht, 205 pp.

De Haas, H. & Eisma, D., 1993: Suspended-sediment transport in the Dollard estuary. Neth. J. Sea Res. 31, 37-42.

De Glopper, R.J., 1964: Landaanwinning in het Waddengebied. In: Anderson, W.F., Abrahamse, J., Buwalda, J.D. & Van Straaten, L.M.J.U. (Eds): Het Waddenboek. Thieme, Zutphen, 152-177.

De Glopper, R.J., 1967: Over de bodemgesteldheid van het Waddengebied. Van zee tot land, Rijksdienst voor de IJselmeerpolders, rapporten en mededelingen inzake de droogmaking, ontginning en sociaal-economische opbouw der IJsselmeerpolders, 43, 67 pp.

De Jonge, V.N., 1992: Physical processes and dynamics of microphytobenthos in the Ems Estuary (The Netherlands). Min. of Transport, Publ. Works and Water Management, Tidal Water Division, 176 pp.

De Vooys, C.G.N., 1987: elimination of sand in the blue mussel, Mytilus edulis. Neth. J. Sea Res., 21, 75-78.

De Wilde, P.A.W.J., 1975: De rol van bodemdiatomeeen en detritus als voedselbron voor bentische evertebraten in de Waddenzee. In: Swennen, C., De Wilde, P.A.W.J. & Haeck, J. (Eds): Werkgroep Waddengebied, mededeling nr. 1: Symposium Waddenonderzoek, 39-49.

De Wilde, P.A.W.J. & Beukema, J.J., 1984: The role of the zoobentos in the consumption of organic matter in the Dutch Wadden Sea. Neth. Inst. for Sea research, Publ. Series, 10, 145-148.

De Wit, J.A.W., Schotel, F.M. & Bekkers, L.E.J., 1982: De waterkwalitieit van de Waddenzee 1971-1981. Rijkswaterstaat, Rijksinstituut voor zuivering van afvalwater, Hoofdafdeling Oppervlaktewater Lelystad, Nota 82.065, 67 pp.

Dekker, R., 1987: The importance of the subtidal macrobenthos as a food source for the Wadden Sea ecosystem. Proc. fifth intern. Wadden Sea Symposium, Biol. Med. Del. Ser., 37, 27-36

Dijkema, K.S., 1989: Habitats of The Netherlands, German and Danish Wadden Sea, 1:100.000. RIN, Veth Foundation, Texel, Leiden, 30 pp.

Dijkema, K.S., Reineck, H.E. & Wolff, W.J., 1980 (Eds): Geomorphology of the Wadden Sea Area. Rep. 1 of the Wadden Sea Working Group. Balkema, Rotterdam, 135 pp.

Dijkema, K.S., Van den Bergs, J., Bossinade, J.H., Bouwsema, P., De Glopper, R.J. & Van Meegen, J.W.Th.M. 1988: Effecten van rijzendammen op opslibbing en omvang van de vegetatiezones in de Friese en Groninger landaanwinningswerken. Rijkswaterstaat, Dir. Groningen Nota GRAN 1988-2010, RIN-Report 88/66, Texel & RIJP-Report 1988-33 Cbw, Lelystad: 119 pp.

Dittmann, S., 1987: Die Bedeutung der Biodeposite für die Benthosgemeinschaft der Wattsedimente. Unter besonderer Berücksichtigung der Miesmuschel *Mytilus edulis L.*. PhD-thesis, Georg-August-Univrsität, Göttingen, 182 pp.

Dörjes, J. 1982: Das Watt als Lebensraum. In: Reineck, H.E. (Ed.): Das Watt, Ablagerungs- und Lebensraum, 3rd ed., Waldemar Kramer, Frankfurt am Main, 107-143.

Dral, A.D.G., 1967: The movements of the laterofrontal cilia and the mechanisms of particle retention in the mussel *Mytilus edulis* L.. Neth. J. Sea Res., 3, 391-422.

Dronkers, J., 1984: Import of fine marine sediment in tidal basins. Neth. Inst. for Sea research, Publ. Series, 10, 83-105.

Eisma, D., 1980: Natural forces. In: Dijkema, K.S., Reineck, H.E. & Wolff, W.J., (Eds): Geomorphology of the Wadden Sea Area. Rep. 1 of the Wadden Sea Working Group. Balkema, Rotterdam, 20-31.

Eisma, D., 1986: Flocculation and de-flocculation of suspended matter in estuaries. Neth. J. Sea Research, 20, 183-199.

Eisma, D., 1993: Suspended matter in the aquatic environment. Springer-Verlag, NY, 315 pp.

Eisma, D. & Kalf, J., 1987: Distribution, organic content and particle size of suspended matter in the North Sea. Neth. J. Sea Res., 21, 265-285

Eisma, D., Bernard, P., Cadée, G.C., Ittekot, V., Kalf, J., Laane, R., Martin, J.M., Mook, W.G., Van Put, A. & Schumacher, T., 1991a: Suspended-matter particle size in some West-European estuaries; Part I: particle size distribution. Neth. J. Sea Res., 28, 193-214.

Eisma, D., Bernard, P., Cadée, G.C., Ittekot, V., Kalf, J., Laane, R., Martin, J.M., Mook, W.G., Van Put, A. & Schumacher, T., 1991b: Suspended-matter particle size in some West-European estuaries; Part II: a review on floc formation and break-up. Neth. J. Sea Res., 28, 215-220.

Eisma, D. & Li, A., 1993: Changes in suspended-matter floc size during the tidal cycle in the Dollard Estuary. Neth. J. Sea Res., 31, 107-117.

Ehlers, J., 1988: Morphodynamics of the Wadden Sea. Balkema, Rotterdam, 397 pp.

Essink, K., 1978: The effects of pollution by organic waste on macrofauna in the eastern Dutch Wadden Sea. Neth. Inst. Sea Res., Publ. Ser., 1, 135 pp.

Essink, K., Reenders, R., De Reus, J.H. & De Wit, J.A.W., 1983: Problematiek van het slibstorten in de Waddenzee en het Eems-Dollard estuarium. Rijkswaterstaat WWKZ-82.HO22/MAD-82.35/RIZA-83.072.

Essink, K., Tydeman, P., De Koning, F. & Kleef, H.L., 1986: On the adaption of the mussel Mytilus edulis L. to different enironmental suspended matter concentrations. Dienst getijdewateren, Groningen, report no. GWAO-86.156, 15 pp.

Eysink, W.D., 1979: Morfologie van de Waddenzee, gevolgen van zand- en schelpenwinning, verslag literatuuronderzoek. Waterl. Lab., Report R-1336, 92 pp.

Eysink, W.D. & Biegel, E.J., 1992: Impact of sea-level rise on the morphology of the Wadden Sea in the scope of its ecological function. Investigations on empirical morphological relations. Rijkswaterstaat Dienst Getijdewateren, ISOS*2 Project, Phase 2, 73 pp.

Favejee, J.C.L., 1951: The origin of the "Wadden" mud. Mededelingen Landbouwhogeschool te Wageningen Nederland, 51, 5, 113-141.

Flemming, B.W. & Delafontaine, M.T., 1994, submitted: Seasonal rates and trends of biodeposition in a juvenile mussel bed of the East Frisian Wadden Sea (Southern North Sea).

Flemming, B.W., Delafontaine, M.T. & Bartholomä, A., 1993: Morphodynamics and mass balance of biodepostion in a backbarrier mussel bank (Spiekeroog, Southern North Sea). In: Amler, M.R.W. & Tietze, K.W. (Eds): Geologica et Palaeontologica, Sediment 93, Marburg, Kurzfassungen, 30-31.

Frankenberg, D. & Smith, K.L.jr, 1967: Coprophagy in marine animals. Limnol. Oceanogr., 12, 443-450.

Genuchten, P.M.B., 1984: Slib in de Waddenzee, literatuuronderzoek. Rijkswaterstaat, nota WWKZ-84.H006, MAD 84.26 33 pp.

Gowing, M.M. & Silver, M.W., 1983: Origins and microenvironments of bacteria mediating faecal pellet decomposition in the sea. Mar. Biol. 73, 7-16.

Haven, D.S. & Morales-Alamo, R., 1972: Biodeposition as a factor in sedimentation of fine suspended solids in estuaries. Geol. Soc. Am. Mem. 133, 121-130.

Hertweck, G., 1982: Die Bewohner des Wattenmeeres in ihren Auswirkungen auf das Sediment. In: Reineck, H.E. (Ed.): Das Watt, Ablagerungs- und Lebensraum, 3rd ed., Waldemar Kramer, Frankfurt am Main, 145-172.

Hickel, W., 1984: Seston in the Wadden Sea of Sylt (German Bight, North Sea). Neth. Inst. for Sea research, Publ. Series, 10, 113-131.

Hoskin, C.M., 1980: Flux of barnacle plate fragments and faecal pellets measured by sediment traps. Journ. of Sed. Pet., 50, 1213-1218.

IJnsen, F., 1974: Onderzoek naar de basispeilen in het gebied van de Westelijke Waddenzee, RWS, Studiedienst Hoorn, Nota 74.10, 25 pp.

IJnsen, F., 1977: Over de invloed van troposferische circulatiepatronen op het ontstaan van stormvloeden in de Noordzee. Rijkswaterstaat, Dir. Friesland, Afd. ANP, rapport 77.01, 27 pp.

IJnsen, F., 1987: Getijverlaging door afwaaiing, Harlingen, Terschelling, Griend, steekproef 1940-1985. Rijkswaterstaat, Direction Friesland, Reportno. ANW 87-20, 15 pp.

IJnsen, F., 1988: IJsgang in de Waddenzee. Rijkswaterstaat, Direction Friesland, Reportno. ANW 88-02, 34 pp.

Jacobsen, T.R. & Azam, F. 1984: Role of bacteria in copepod faecal pellet decomposition: colonization, growth rates and mineralization. Bull. Mar. Sci. 35, 3, 495-502.

Jaramillo, E., Bertrán, C. & Bravo, A, 1992: Mussel biodeposition in an estuary in southern Chile. Mar. Biol., 94, 127-142.

Jensen, K.T., 1991: Dynamics and growth of the cockle, *Cerastoderma edule*, on an intertidal mudflat in the Danish Wadden Sea: Effects of submersion time and density. Neth. J. Sea Res., 28, 335-345.

Jϕrgensen, C.B., 1990: Bivalve filter feeding: Hydrodynamics, bioenergetics, physiology and ecology. Olsen & Olsen, Fredensborg, 140 pp.

Kamermans, P., 1993: Food limitation in cockles (*Cerastoderma edule* (L.)): Influences of location on tidal flat and of nearby presence of mussel beds. Neth. J. Sea Res., 31, 71-81.

Kamps, L.F., 1956: Slibhuishouding en landaanwinning in het oostelijk Waddengebied. Int. Rep. RWS, Direction Landaanwinning, Baflo, 93 pp.

Kamps, L.F., 1962: Mud distribution and land reclamation in the Eastern Wadden shallows. Rijkswaterstaat Comm., 4, The Hague, 73 pp.

Kautsky, N. & Evans, S., 1987: Role of biodeposition by *Mytilus edulis* in the circulation of matter and nutrients in a Baltic coastal ecosystem. Mar. Ecol. Prog. Ser., 38, 201-212.

Keuper, F.J., 1985: IJsverplaatsing Waddenzee, januari-februari 1985. Rijkswaterstaat, Dir. waterhuishouding en waterbeweging, Adviesdienst Hoorn, Notitie WWKZ-85.H228, 6 pp.

Kranz, P.M., 1974: The anistrophic burial of bivalves and its paleoecologcial significance. Journal of Geology, 82: 237-265.

Kuenen, D.J., 1942: On the distribution of mussels on the intertidal sand flats near Den Helder. Archives Neerlandaises de zoologie, VI, 117-160.

Lambeek, J.J.P., 1991: Available sediment data of the Dutch Wadden Sea (part 1) Rijkswaterstaat notitie GWAO-91.10090. 62 pp. + 16 appendices.

Lees, A., 1975: Possible influence of salinity and temperature on modern shelf carbonate sedimentation. Marine Geology, 19, 159-198.

Linke, O., 1954: Die Bedeutung der Miesmuscheln für die Landgewinnung im Wattenmeer. Natur & Volk, 84, 253-261.

Loo, L.O., 1992: Filtration, assimilation, respiration and growth of *Mytilus edulis* L. at low temperatures. Ophelia, 35, 123-131.

Maas Geesteranus, R.A., 1942: On the formation of banks by Mytilus edulis L. Archives Neerlandaises de zoologie, VI, 283-326.

McMaster, R.L., 1958: Modification of underwater surface sediment layers by sea mussels (*Mytilus edulis*). Journ. of Sed. Pet., 28, 515-516.

Nehls, G. & Thiel, M., 1993: Large-scale distribution patterns of the mussel *Mytilus edulis* in the Wadden Sea of Schleswig-Holstein: do storms structure the ecosystem? Neth. J. Sea Res., 31, 181-187.

Newell, R.C., 1965: The role of detritus in the nutrition of two marine deposit feeders, the prosobranch *Hydrobia ulvae* and the bivalve *Macoma balthica*. Proc. Zool. Soc. London, 114, 25-45.

Nowell, A.R.M., Jumars, P.A. & Eckman, J.E., 1981: Effects of biological activity on the entrainment of marine sediments. Marine Geol., 42, 133-153.

Nybakken, J.W., 1982: Marine Biology, an ecological approach. Harper & Row, N.Y., 446 pp.

Oenema, O., 1988: Early diagenesis in recent fine-grained sediments in the Eastern Scheldt. Thesis Utrecht University, 222 pp.

Oost, A.P. & De Boer, P.L., 1994: Sedimentology and development of barrier islands, ebb-tidal deltas, inlets and backbarrier areas of the Dutch Wadden Sea. Senckenbergiana Maritima, 24, 65-115.

Oost A.P. & Baas J.H., 1994: The development of small-scale bedforms in tidal environments: an empirical model for unsteady flow and its applications. Sedimentol., 41, 883-903.

Oost, A.P., De Haas, H., IJnsen, F., Van den Boogert, J.M. & De Boer, P.L., 1993: The 18.6 year nodal cycle and its impact on tidal sedimentation. Sedimentary Geology, 87: 1-11.

Otto, L., Zimmermann, J.T.F., Furnes, G.K., Mork, M., Saetre, R. & Becker, G., 1990: Review of the physical oceanography of the North Sea. Neth. J. Sea Res., 26, 161-238.

Peduzzi, P. & Herndl, G.J. 1986: Role of bacteria in decomposition of faecal pellets egested by the epiphyte-grazing gastropod *Gibbula umbilicaris*. Mar. Biol. 92: 417-424.

Peinert, R., Saure, A., Stegmann, P., Stienen, C., Haardt, H., & Smetacek, V., 1982: Dynamics of primary production and sedimentation in a coastal ecosystem. Neth. J. Sea Res., 16, 276-289.

Pérez Camacho, A., González, R. Funetes, J., 1991: Mussel culture in Galicia (N.W. Spain). Aquaculture, 92, 263-278.

Piekhaar, R.S., 1980: Oriënterend onderzoek naar de sedimentatie-eigenschappen van zee- en rivierslib. Rijkswaterstaat, Directie waterhuishouding en waterbeweging, district kust en zee, nota WWKZ 80.S004, rapport 181, 1-36.

Postma, H., 1954: Hydrography of the Dutch Wadden Sea. Arch. Neerl. Zool., 10, 405-511.

Postma, H., 1981: Exchange of materials between the North Sea and the Wadden Sea. Marine Geology, 40, 199-213.

Postma, H., 1984: Introduction to the symposium on organic matter in the Wadden Sea. Neth. Inst. for Sea research, Publ. Series, 10, 15-22.

Postma, H., & Dijkema, K.S., 1982: Hydrography of the Dutch Wadden Sea: Movements and properties of water and particulate matter. Final report on 'Hydrography' of the Wadden Sea Working Group, Report 2. Balkema, Rotterdam, 75 pp.

Potter, P.E., Maynard, J.B. & Pryor, W.A., 1980: Sedimentolgy of shale, study guide and reference source. Springer-Verlag, NY, 303 pp.

Prins, T.C., Dankers, N. & Smaal, A.C., 1994, in press: Seasonal variation in the filtration rates of a semi-natural mussel bed in relation to seston composition. J. Exp. Mar. Biol. Ecol. 18 pp.

Pryor, W.A., 1975: Biogenic sedimentation and alteration of argillaceous sediments in shallow marine environments. Geological society of America Bulletin, 86, 1244-1254.

Reineck, H.E., Wunderlich, F., 1968: Classification and origin of flaser and lenticular bedding. Sedimentology, 11, 99-104.

Reineck, H.E., 1980: Sediments and dynamical processes. In K.S. Dijkema, Reineck, H.E. & Wolff, W.J., (Eds): Geomorphology of the Wadden Sea area, final report of the section 'Geomorphology' of the Wadden Sea Working Group. Balkema, Rotterdam p. 32-49.

Reineck, H.E. & Singh, I.B., 1971: Der Golf von Gaeta/Tyrrhenisches Meer, 3, Die Gefüge von Vorstrand- und Schelfsedimenten. Senckenbergiana maritima 3, 185-201.

Reineck, H.E. & Singh, I.B., 1980: Depositional Sedimentary Environments, Scnd. Rev. Ed.. Springer-Verlag, NY, p.

Rhoads, D.C. & Stanley, D.J., 1965: Biogenic graded bedding. J. Sed. Petr., 35, 956-963.

Risk, M.J. & Moffat, J.S., 1977: Sedimentological significance of faecal pellets of *Macoma Balthica* in the Minas Basin Bay of Fundy. J. Sed. Petr., 47, 1425-1436.

Rudwick, M.J.S. 1970: Living and fossil brachiopods. Biological Series, Hutchinson & Co., London, 199 pp.

Schäfer, W., 1953: Zur Unterscheidung gleichförmiger Kotpillen meerischer Evertebraten. Senckenbergiana 34, 81-83.

Seed, R. 1976: Ecology. In: Bayne, B.L. (Ed.): Marine mussels: their ecology and physiology. International biological prog. 10, Cambr. Univ. Press, Cambridge, 13-65.

Smaal, A.C. & Prins, T.C., 1993: The uptake of organic matter and the release of inorganic nutrients by bivalve suspension feeder beds. In: Dame, R.F. (Ed.): Bivalve filter feeders, NATO ASI Ser., G33, 271-298.

Smith, J.M. & Frey, R.W., 1985: Biodeposition by the ribbed mussel *Geukensia demissa* in a salt marsh, Sapelo Island, Georgia. Journ. of Sed. Pet., 55, 817-828.

Stratingh, G.A. & Venema, C.A., 1855: De Dollard of geschied,-aardrijks- en natuurkundige beschrijving van dezen boezem der Eems. Reprinted in 1979 by 'de Landelijke Vereniging tot Behoud van de Waddenzee' & 'Stichting Het Groninger Landschap', 333 pp.

Tammes, P.M.L. & Dral, A.D.G., 1956: Observations on the straining of suspensions by mussels. Archives Neerlandaises de zoologie, tome XI, 87-112.

Terwindt, J.H.J. & Breusers, H.N.C., 1972: Experiments on the origin of flaser, lenticular and sand-clay alternating bedding. Sedimentology, 19, 85-98.

Thiesen, B.F. 1968: Growth and mortality of culture mussels in the Danish Wadden Sea. Meddelelser fra Danmarks Fiskeri-og Havundersøgelser. N.S., 6, 47-78.

Tsuchiya, M., 1980: Biodeposit production by mussel *Mytilus edulis* L. on rocky shores. J. Exp. Mar. Biol. Ecol., 96, 199-212.

Vahl, O. 1972: Efficiency of particle retention in *Mytilus edulis*. Ophelia, 12, 42-52.

Van den Berg, J.H., 1981: Rhythmic seasonal layering in a mesotidal channel fill sequence, Oosterschelde Mouth, the Netherlands. In: Nio, S.D., Shüttenhelm, R.T.E. & Van Weering, T.C.E. (Eds): Holocene marine sedimentation in the North Sea basin. Int. Ass. Sed., Spec. Publ. 5, 147-159.

Van Es, F.B., 1982: Some aspects of the flow of oxygen and organic carbon in the Ems-Dollard Estuary. PhD-thesis, Groningen, 121 pp.

Van Es, F.B., Van Arkel, M.A., Bouwman, L.A. & Schröder, H.G.J., 1980: Influence of organic pollution on bacterial, macrobenthic and meiobenthic populations in intertidal flats of the Dollard. Neth. J. Sea Res., 14, 288-304.

Van Sijp, D., 1988: Troebelheidsmeting Nieuwlandsrijd Ameland. Rijkswaterstaat, Directie Friesland, notitie ANW 88-28, 4 pp.

Van Straaten, L.M.J.U., 1949: Quelques particularités du relief sous-marin de la Mer des Wadden (Hollande). Comptes-Rendus du Congrès Sédimentation et Quarternaire France, 139-145.

Van Straaten, L.M.J.U., 1951: Texture and genesis of Dutch Wadden Sea sediments. Proceedings of the third international congress of sedimentology Groningen - Wageningen, Netherlands, 225-244.

Van Straaten, L.M.J.U., 1954: Composition and structure of Recent marine sediments in the Netherlands. Leidse Geol. Mededel. 19, 1-110.

Van Straaten, L.M.J.U., 1964: De bodem der Waddenzee. In: Anderson, W.F., Abrahamse, J., Buwalda, J.D. & Van Straaten, L.M.J.U. (Eds): Het Waddenboek. Thieme, Zutphen, 75-151.

Van Straaten, L.M.J.U. & Kuenen, Ph.H., 1957: Accumulation of fine grained sediments in the Dutch Wadden Sea. Geologie en Mijnbouw, 19, 329-354.

Veenstra, H.J., 1978: Rapport van het Geologisch Instituut van de Rijksunviersiteit Groningen inzake een onderzoek naar de herkomst van sediment nabij de Blauwe Slenk in de Waddenzee, 14 pp. In: Rijkswaterstaat: Hydrografisch-sedimentologisch onderzoek met betrekking tot de winning van zand in de Waddenzee c.a., Interimrapport Werkgroep I, 13 pp.

Verhagen, J.H.G., 1982: A distribution and population model of the mussel Mytilus Edulis in Lake Grevelingen. Waterloopkundig laboratorium delft hydraulics laboratory. Publication 279, 1-11.

Verwey, J., 1952: On the ecology of distribution of cockle and mussel in the Dutch Waddensea, their role in sedimentation and the source of their food supply, with a short review of the feeding behaviour of bivalve mollusks. Archives Néerlandaises de Zoologie, tome X, 2e Livr, 171-239.

Verwey, J., 1981: The blue mussel Mytilus edulis. In: Dankers, N., Kühl, H. & Wolff, W.J. (Eds): Invertebrates of the Wadden Sea, Balkema, Rotterdam, 115-116.

Visser, M.J., 1980: Neap-spring cycles reflected in Holocene subtidal large-scale bedform deposits: a preliminary note. Geology, 8, 543-546.

Vos, P.C., De Boer, P.L. & Misdorp, R., 1988: Sediment stabilisation by benthic diatoms. In: De Boer, P.L., Van Gelder, A. & Nio, S.D. (Eds): Tide-influenced sedimentary environments and facies: Reidel, Dordrecht, 511-526.

Vroom, M.G., Steyaert, F.H.I.M., Misdorp, R., Venema, H., Elgershuizen, J.H.B.W., Rakhorst, H.D. & Damoiseaux, M.A. (1989): Wadatlas. Min. Verkeer & Waterstaat, Stadsdrukkerij, Amsterdam, 95 pp..

Wanless, H.G., Burton, E.A. & Davis, J., 1981: Hydrodynamics of carbonate faecal pellets. Journ. of Sed. Pet., 51, 27-36.

Watling, L., 1989: Small-scale features of marine sediments and their importance to the study of deposit feeding. In: Lopez, G., Taghon, G. & Levinton, J. (Eds): Ecology of marine deposit feeders. Lecture notes on coastal and estuarine studies, 31, 269-290.

Widdows, J., 1984: Physiological measurements. In: Bayne, B.L., Brown, D.A. Burns, K., Dixon, D.R., Ivanovici, A, Livingstone, D.R., Lowe, D.M., Moore, M.N., Stebbing, A.R.D. & Widdows, J. (Eds): The effects of stress and pollution on marine animals. Praeger Sci. N.Y., 3-45.

Widdows, J., Fieth, P. & Worrall, C.M., 1979: Relationships between seston, available food and feeding activity in the common mussel *Mytilus edulis*. Marine Biol., 56, 195-267.

Wiggers, A.J., 1960: Die Korngrössenverteilung der holozänen Sedimente im Dollart-Ems-estuary. In: Van Voorthuysen, J.H. & Kuenen, Ph.H. (Eds): Das Ems-Estuarium (Nordsee), ein sedimentologisches Symposium, Verh. Kon. Ned. Geol. Mijnbouwk. Gen., XIX: 75-81.

Winkelmolen, A.M. & Veenstra, H.J., 1974: Size and shape sorting in a Dutch tidal inlet. Sedimentology, 21: 107-126.

Youngbluth, M.J., Bailey, T.G., Davoll, P.J., Jacoby, C.A., Blades-Eckelbarger, P.I. & Griswold, C.A., 1989: Faecal pellet production and daily migratory behavior by the euphausiid *Meganyctiphanes norvegica* effect benthic-pelagic coupling. Deep-Sea Research, 36, 1491-1501.

Zimmerman, J.T.F., 1976: Mixing and flushing of tidal embayments in the Western Dutch Wadden Sea I. Distribution of salinity and calculation of mixing time scales. Neth. J. Sea Res., 10, 149-191.

Zwarts, L., 1991: Seasonal variation in body weight of the bivalves *Macoma Balthica, Scrobicularia plana, Mya arenaria* and *Cerastoderma edule* in the Dutch Wadden Sea. Neth. J. Sea Res., 28, 231-245.

CHAPTER 7

TIME AS AN INDEPENDENT VARIABLE FOR CURRENT RIPPLES DEVELOPING TOWARDS LINGUOID EQUILIBRIUM MORPHOLOGY

Terra Nova, 5, 29-35

ABSTRACT

Flume experiments show that current ripples on very fine sand surfaces always develop towards a linguoid shape with constant height and wavelength provided that sufficient time is allowed for their formation. Straight and sinuous current ripples only reflect intermediate stages in ripple development and may be regarded as non-equilibrium bedforms. The time period which current ripples require to reach linguoid equilibrium morphology is related to an inverse power of flow velocity. In the transitional stage from current ripples to upper stage plane bed (i.e. washed-out ripple stage) only the equilibrium wavelength remains constant, whereas equilibrium height rapidly decreases to zero. Our observations imply that bed-roughness parameters in sediment transport calculations can be simplified when equilibrium conditions are attained, and that inferences about flow energy from the dimensions of current ripples in very fine sand need to be regarded with caution.

INTRODUCTION

It is commonly believed that the equilibrium morphology of current ripples depends on bed shear stress and grain size (Menard, 1950; Allen, 1968, 1984; Yalin, 1977, 1985; Reineck & Singh, 1980), although some studies indicate a very weak relationship or none at all (Harms, 1969; Yalin, 1975; Jopling & Forbes, 1979; Middleton & Southard, 1984). In many of these studies (e.g. Allen, 1968, 1984; Reineck & Singh, 1980) small scale, straight and sinuous current ripples are regarded as equilibrium bedforms at low flow velocities. However, Richards (1980) showed theoretically that current ripple dimensions only scale on grain size, and not on flow strength. The importance of the time factor for the morphological development (wavelength and height) of current ripples has been largely ignored. Most studies mention only the approximate time periods that current ripples require to reach equilibrium dimensions, although Yalin (1975) quantified the development in time of current ripple length from initial flat bed conditions.

Under sufficiently high flow velocities, the time for ripples to reach their equilibrium wavelength is relatively short (tens of minutes to several hours). At low flow velocities, near the threshold value for sediment movement, much longer periods (tens of hours to several days) are required to reach equilibrium wavelength (cf. Yalin, 1975). The aim of this study was to explore the relationships between equilibrium ripple morphology, flow velocity and time.

EXPERIMENTAL METHODOLOGY

A series of flume experiments was performed to study the development of bedforms under steady, uniform flow, starting from initial flat bed conditions. The flume tank used is a rectangular channel in which the water flows in a continuous loop. The flow is driven by a chain with paddles (Fig. 1). A well-sorted (σ = 0.49), siliciclastic sediment with a median grain size of 95 µm was used to avoid the formation of dunes (Allen, 1968, 1984; Southard & Boguchwal, 1990a; Van den Berg & van Gelder, 1992). We completed 9 runs at different flow velocities (Figs 2 and 3), each run lasting between 1 hour and 4 days. Every run started with a flat bed and 32-34 cm of standing water before a steady, uniform flow was imposed over the bed. No sediment was added nor removed from the flume during experiments, so primary bedforms were erosive.

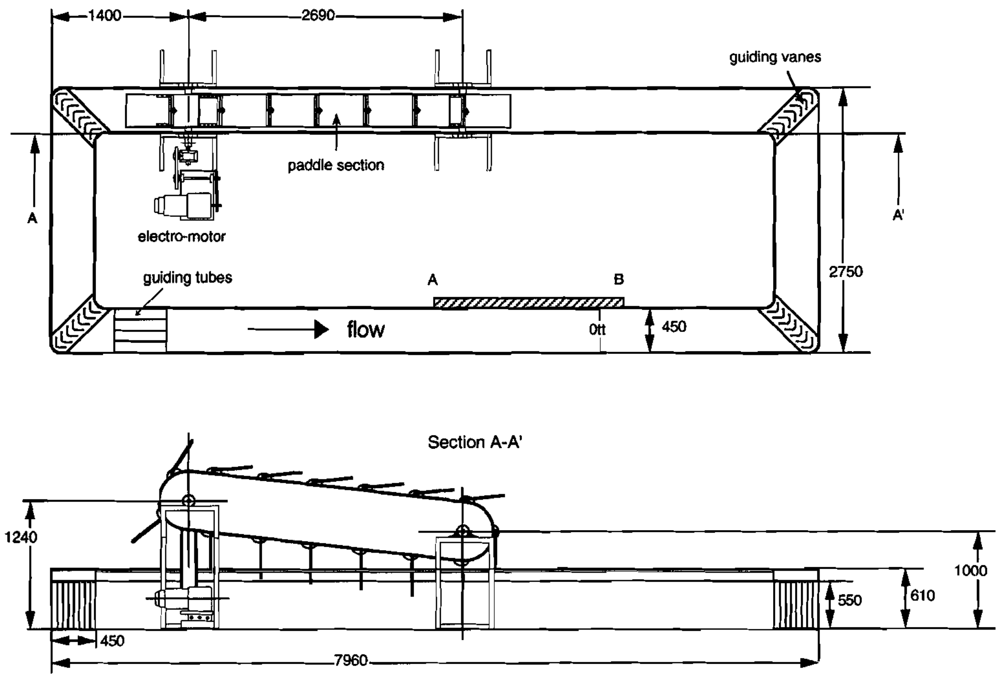

Figure 1. Flume, plan view and cross-section. All dimensions are in mm. See Winkelmolen (1976) for a full technical description. Ott = location of Ott-type current meters. Section A-B is the transect along which bedform development was studied.

The development of bedforms was monitored along the side-wall of the flume (A-B in Fig. 1) by following a standard procedure, in which flow velocity, water depth, water temperature, and ripple heights and wavelengths were recorded at regular time intervals varying from 30 s up to one hour depending on flow velocity. Flow velocity was measured in the centre of the flume using Ott-type current meters located at 4 different heights above the bed. Depth-averaged flow velocity was calculated from a best-fit logarithmic curve through these measurements. The maximum variation in local water depth during the runs was 3 cm depending on the maximum bedform height. Water temperature ranged between 13.8 and 17.8°C for successive runs, but was constant within 1.4°C for each run. The influence of these temperature variations on bedform development is considered negligible (cf. Southard & Boguchwal, 1990b). Pilot studies indicated that the depth-averaged flow velocity near the side-wall of the flume was 10.5% smaller than along the central transect. The measured flow velocities have been multiplied by 0.895 to correct for this wall effect. The influence of the side-wall effect on bedform morphology was verified at the end of several experiments by comparing bedform heights and wavelengths along several transects parallel to the flume wall. Apart from minor upstream and downstream deflections of ripple crests near the wall, which had no measurable influence on average bedform size, no additional side-wall effects on bed morphology were observed. The data derived from the experiments is summarized in Table I.

DEVELOPMENT OF THE BEDFORMS

The development of bedform height and wavelength at a depth-averaged flow velocity of 34 $cm.s^{-1}$ is shown in Fig. 2. Average current ripple height and wavelength increased rapidly during the first hours of the run towards constant values of 13.3 mm and 121 mm, respectively, after approximately 7 hours. Once at these values, the bed configuration is apparently in dynamic equilibrium with the physical conditions (Fig. 2). During their development towards equilibrium conditions, the current ripples evolved from straight, via sinuous, to an equilibrium linguoid shape. The latter shape was maintained until the end of the experiment.

The results of subsequent runs taken at higher and lower flow velocities were similar to those shown in Fig. 2. The development of average bedform height and wavelength can be approximated empirically by the best-fit equation:

$$d_t=d_e-(d_e-d_o)e^{-ct} \tag{1}$$

applicable for all runs, where d is the bedform height or wavelength in mm, c is a constant proportional to the time taken to reach equilibrium dimensions, and t is the time lapse in hours. The subscripts for d indicate actual (t), equilibrium (e), and initial (o) average height and wavelength values.

Table I. Experimental results of basic flow parameters, equilibrium times and bedform dimensions.

Measured and computed basic flow data

Run no.	t_{tot} (h)	h_o (m)	h (m)	U ($m.s^{-1}$)	T (°C)	ν (10^{-6} m^2s^{-1})	bed st.	q (m^2s^{-1})	Fr (-)	Re (10^{-5})
47	95.58	0.32	0.331	0.277	16.4-17.8	1.064	CR	0.092	0.154	0.862
42	72.55	0.34	0.344	0.341	17-17.8	1.068	CR	0.117	0.186	1.098
46	7	0.324	0.331	0.399	15.5-16.1	1.112	CR	0.132	0.221	1.187
45	4.85	0.332	0.345	0.471	14.2-15	1.149	CR	0.162	0.256	1.413
44	3.9	0.34	0.361	0.536	13.8-14.2	1.170	CR	0.193	0.285	1.652
43	4.02	0.34	0.359	0.591	16-16.6	1.099	CR	0.212	0.315	1.931
49	2.5	0.33	0.346	0.652	17.3-17.4	1.071	CR	0.225	0.354	2.105
11a	nr	nr	0.315	0.808	14	1.172	WOR	0.254	0.459	2.170
10a	nr	nr	0.315	0.816	14	1.172	WOR	0.257	0.464	2.194
48	1	0.33	0.33	0.839	14.4-14.6	1.157	WOR	0.277	0.466	2.392
25a	nr	nr	0.325	0.844	14.5	1.156	WOR	0.274	0.473	2.374
7a	nr	nr	0.32	0.850	13.5	1.188	WOR	0.272	0.480	2.290
9a	nr	nr	0.32	0.881	13.7	1.181	WOR	0.282	0.497	2.387
8a	nr	nr	0.325	0.925	13.5	1.188	UPB	0.301	0.518	2.530
8	nr	nr	0.338	0.950	16	1.110	UPB	0.321	0.522	2.894

Dimensions of bedforms and equilibrium times

Run no.	H_e (mm)	H_{max} (mm)	$T_e(H)$ (h)	L_e (mm)	L_{max} (mm)	L_o (mm)	$T_e(L)$ (h)	L_e/H_e (-)	L_{max}/H_{max} (-)
47	12.3	24.1	20.24	117.5	206	49.5	25.84	9.6	8.5
42	13.3	25	6.55	120.7	196.5	36.8	8.26	9.1	7.9
46	12.8	24.4	1.94	115.8	186.9	49.1	3.44	9.0	7.7
45	12.9	26.9	0.86	114	203.1	42.9	1.35	8.8	7.6
44	14.3	29.5	0.46	118	197.2	40.2	0.50	8.3	6.7
43	12.8	23.9	0.42	113.5	175.4	49	0.50	8.9	7.3
49	13.4	29.6	0.31	114.6	188.5	68.7	0.40	8.6	6.4
11a	10.7	13	nr	nr	nr	nr	nr	-	-
10a	6.3	8	nr	nr	nr	nr	nr	-	-
48	4.8	6.8	0.12	114.1	141	79	0.24	23.8	20.7
25a	3.7	5	nr	nr	nr	nr	nr	-	-
7a	3.2	4	nr	nr	nr	nr	nr	-	-
9a	1.4	2	nr	nr	nr	nr	nr	-	-

Column headings and symbols:
t_{tot} = total run time, h_o = initial flow depth, h = average flow depth during equilibrium conditions, U = depth-averaged flow velocity corrected for side-wall effects, T = water temperature, ν = kinematic viscosity, q = $U.h$ = flow discharge, Fr = $U/\sqrt{(gh)}$ = Froude number, Re= Uh/ν = Reynolds number, H_e, L_e = equilibrium bedform height resp. wavelength, H_{max}, L_{max} = maximum bedform height resp. wavelength, $T_e(H), T_e(L)$ = equilibrium time for bedform height resp. wavelength, L_o = initial bedform wavelength, CR = current ripple, WOR = washed-out ripple, UPB = upper stage plane bed, nr = not recorded. Runs with the extension 'a' are results from unpublished work by Baas.

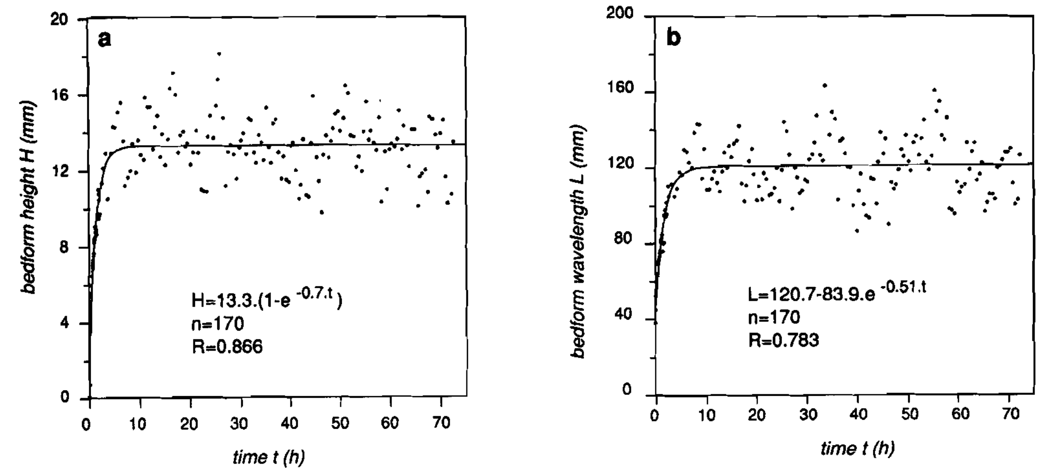

Figure 2. Development of current ripple height (a) and wavelength (b) with time at a depth-averaged flow velocity U of 34 cm.s^{-1}. Initial water depth was 34 cm. During the run, water depth ranged from 33 to 36 cm, depending on bedform height. Temperature ranged from 17 to 17.8°C. Dots indicate data points calculated from average dimensions of 10 to 50 current ripples at a given time. Lines give best-fit curves through the data points using Equation (1). Increase of data point scatter with time is caused by an increase of the three-dimensionality of the bed surface. Equilibrium current ripple height and wavelength were reached after 6.5 and 8.3 hours, respectively. It is unknown as yet whether the difference in these time values is significant. H = current ripple height (mm); L = current ripple wavelength (mm); t = time (h); n = number of data points; R = correlation coefficient.

The time span T_e for the current ripples to reach equilibrium dimensions is defined as the time required to reach 99% of the equilibrium current ripple height or wavelength value. The value of T_e was calculated from Equation (1) for each run, using the relations:

$$d_t=0.99d_e \quad (2)$$

$$t=T_e \quad (3)$$

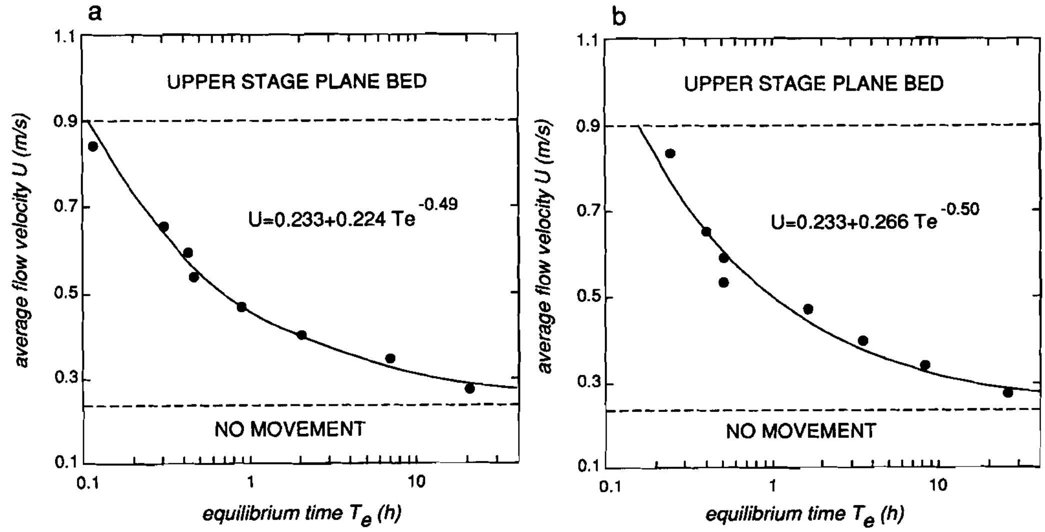

Figure 3. Depth-averaged flow velocity U as a function of the time period T_e required to reach equilibrium ripple height (a) and wavelength (b) starting from flat bed conditions. Dashed lines separate the current ripple field from no movement conditions at 23.3 $cm.s^{-1}$ (Miller et al., 1977) and from the upper stage plane bed field at about 90 $cm.s^{-1}$. Equations represent best-fit curves through the data points.

The time taken to reach equilibrium bedform morphology for different depth-averaged flow velocities is shown in Fig. 3. The relationship between equilibrium time and flow velocity is described by a power function. Equilibrium conditions are established after a few minutes at flow velocities approaching the upper stage plane bed phase and only after tens of hours near the critical flow velocity of initial sediment movement.

DISCUSSION

Our results contrast with the common idea that the equilibrium dimensions of current ripples depend on bed shear stress, which is proportional to the square of the flow velocity and the shear velocity (Menard, 1950; Allen, 1968, 1984; Yalin, 1977, 1985; Reineck & Singh, 1980), although the results of other studies demonstrate a very weak relationship or none at all (Harms, 1969; Yalin, 1975; Jopling & Forbes, 1979; Richards, 1980; Middleton & Southard, 1984). Our results support the latter studies. From our experiments we conclude that flow velocity merely defines the time span needed for current ripples to obtain equilibrium height and wavelength values, and a linguoid form.

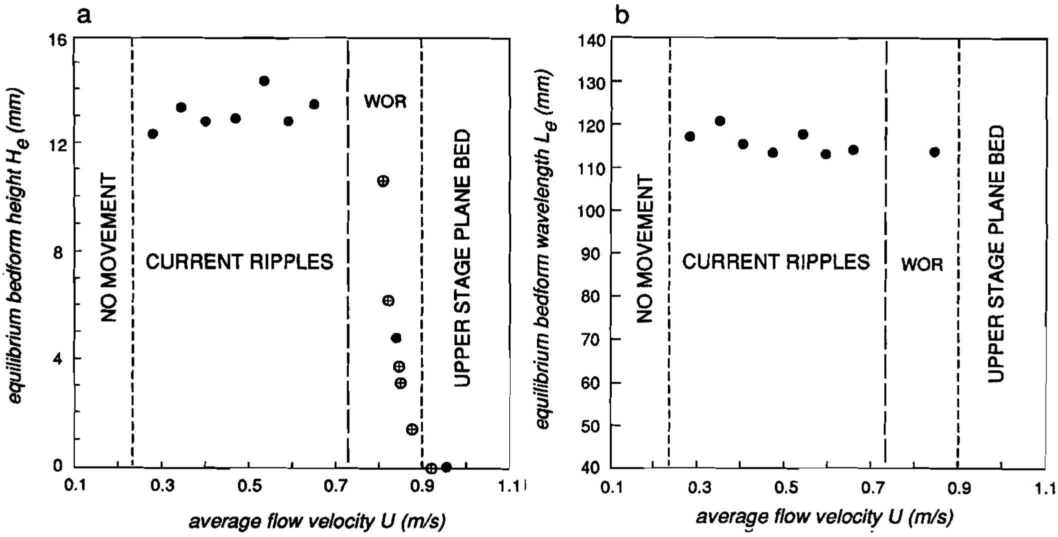

Figure 4. Equilibrium bedform height (a) and wavelength (b) as a function of depth-averaged flow velocity. Closely spaced dashed lines separate the current ripple stability field from no movement at 23.3 cm.s^{-1} and the upper stage plane bed field at about 90 cm.s^{-1}. Within the current ripple field the widely spaced dashed line separates stable current ripples from washed-out ripples (= WOR). Dots indicate measurements from this study. Circled plus signs (a) are results from earlier, unpublished work by the first author, in which only the bedform height was studied in detail. Average equilibrium current ripple height is 13.1 mm. Average equilibrium ripple wavelength is 116 mm.

Figure 4 shows that, for the given grain size, equilibrium current ripple height and wavelength are constant within the current ripple stability field with average values measuring 13.1 mm and 116 mm, respectively. Even at low flow velocities just above the onset of sediment movement (Miller et al., 1977), current ripples reach identical equilibrium dimensions, although this takes tens of hours to achieve (Fig. 3). Pilot studies indicate a period of several days to weeks in order to form equilibrium current ripples when the flow velocity is extremely close to initial sediment movement. Only in the narrow stability field of washed-out ripples (Fig. 4) does the bedform height decrease rapidly towards zero at the boundary with upper stage plane bed conditions (see also Jopling & Forbes, 1979). Washed-out ripples are characterized by a more symmetrical geometry than ordinary current ripples (Figs 5 and 6) and a smaller equilibrium height. The different geometry is most likely caused by the development of a high-concentration bedload layer, which hinders the development of equilibrium current ripple height. With increasing flow velocity, and thus increasing bedload concentration, this effect becomes increasingly important. Current ripples are suppressed

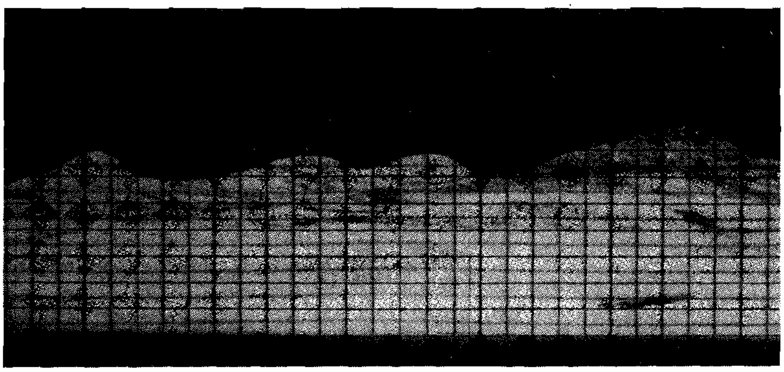

Figure 5. Equilibrium current ripples at a velocity of 47 cm.s^{-1}, showing the irregular bedform dimensions due to the three-dimensional plane form of the current ripples. Grid size is 1 cm^2. Flow is from left to right.

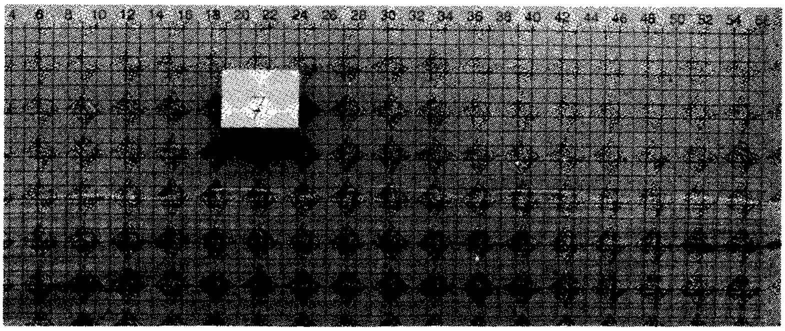

Figure 6. Washed-out ripples at a velocity of 84 cm.s^{-1}, 7 minutes after the start of the experiment, showing the small height and symmetrical form of the bedforms. Grid size is 1 cm^2. Flow is from left to right.

entirely at a depth-averaged flow velocity of about 90 cm.s^{-1} (Fig. 4a). However, the equilibrium wavelength appears to be constant and independent of flow velocity within the entire current ripple stability field (Fig. 4b). Apparently, the wavelength of the washed-out ripples is not influenced by the dense bedload layer and reaches practically the same equilibrium value (114 mm) as current ripples formed at lower velocities. The same trend was observed for the transition from dunes to upper stage plane bed conditions in medium sand (Bridge &

Best, 1988). They found that both ordinary dunes and washed-out dunes had a mean wavelength of 70-80 cm.

The development of current ripples from straight, via sinuous, to an equilibrium linguoid shape shows that a differentiation of these crestline patterns (e.g. Reineck & Singh, 1980) does not depend on flow velocity, but rather reflects different stages during development.

Experimental results similar to ours have not been reported previously. We ascribe this to the fact that the development towards equilibrium conditions is extremely slow at low current velocities. From our earlier experiments with the same grain size it proved to be impossible to define equilibrium conditions visually.

An important implication of our results is that incompletely developed current ripples in the fossil record, formed under hydrodynamic conditions comparable to those in our flume experiments, could only have been generated under conditions of sufficiently low flow velocities and sufficiently short time periods. Our observations also imply that bed-roughness parameters in very fine sand can be largely simplified in sediment transport calculations when current ripples show constant equilibrium dimensions.

CONCLUSIONS

Our results show that equilibrium current ripples have a linguoid morphology and that the period of time needed to reach equilibrium dimensions depends on the flow velocity. If flow velocity is near the threshold value for sediment movement, equilibrium dimensions are reached only after periods of tens of hours to several days. As flow velocities become higher, equilibrium dimensions are reached within shorter time periods. Research papers should, therefore, refer more accurately to the time period in which equilibrium was reached. In our experiments with very fine sand current ripples always obtained a constant linguoid shape at equilibrium conditions, and developed via straight and sinuous to linguoid shapes at all velocities. Hence, straight and sinuous ripples do not represent an equilibrium phase.

From this new evidence, current ripples in the fossil record, formed under hydrodynamic conditions comparable to those in our flume experiments, must be reinterpreted in terms of equilibrium versus non-equilibrium conditions first, rather than considering them to be diagnostic of palaeo-current velocity. Straight and sinuous current ripples are representative of non-equilibrium conditions, whereas linguoid ripples represent bed surfaces that are in equilibrium with the flow conditions.

Bed-roughness parameters in sediment transport calculations can be assumed constant if equilibrium conditions can be ensured.

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REFERENCES

Allen, J.R.L., 1968: Current Ripples: their relation to patterns of water and sediment motion, North Holland Publishing Company, Amsterdam, 433 pp.

Allen, J.R.L., 1984: Sedimentary structures: their character and physical basis, Developments in Sedimentology, 30A, Elsevier, Amsterdam, 593 pp.

Bridge, J.S. and Best, J.L., 1988: Flow, sediment transport and bedform dynamics over the transition from dunes to upper-stage plane beds: implications for the formation of planar laminae, Sedimentology, 35, 753-763.

Harms. J.C., 1969: Hydraulic significance of some sand ripples, Bulletin of the Geological Society of America, 80, 363-396.

Jopling, A. and Forbes, D.L., 1979: Flume study of silt transportation and deposition, Geografiska Annaler, 61A, 67-85.

Menard, H.W., 1950: Current-ripple profiles and their development, Journal of Geology, 58, 152-153.

Middleton, G.V. and Southard, J.B., 1984: Mechanics of sediment movement, Society of Econ. Paleont. Mineral., Short Course, 3, 401 pp.

Miller, M.C., McCave, I.N. and Komar, P.D., 1977: Threshold of sediment motion under unidirectional currents, Sedimentology, 24, 507-527.

Reineck, H.E. and Singh, I.B., 1980: Depositional sedimentary environments; second, revised and updated edition, Springer-Verlag, Berlin, 549 pp.

Richards, K.J., 1980: The formation of ripples and dunes on an erodible bed, Journal of Fluid Mechanics, 99, 597-618.

Southard, J.B. and Boguchwal, L.A., 1990a: Bed configurations in steady unidirectional water flows. Part 2. Synthesis of flume data, Journal of Sedimentary Petrology, 60, 658-679.

Southard, J.B. and Boguchwal, L.A., 1990b: Bed configurations in steady unidirectional water flows. Part 3. Effects of temperature and gravity, Journal of Sedimentary Petrology, 60, 680-686.

Van den Berg, J.H. and van Gelder. A., 1992: The transition of ripples to upper plane bed in flows over very fine sand and silt: an analysis of flume experiments and deposits of the Yellow River, China, In Puigdefabregas, C. and Marzo, M., eds., Proceedings of the 4th International Congress on Fluvial Sedimentology, Barcelona, International Association of Sedimentologists, Special Publication, 13.

Winkelmolen, A.M., 1976: An inexpensive infinite flume, Geologie en Mijnbouw, 55, 51-60.

Yalin, M.S., 1975: On the development of sand waves in time, Proceedings of the 16th Congress of the International Association for Hydraulic Research, Sao Paulo, Brazil, 2, 212-219.

Yalin, M.S., 1977: On the determination of ripple length, Journal of the Hydraulic Division, ASCE, 103, 439-442.

Yalin, M.S., 1985: On the determination of ripple geometry, Journal of Hydraulic Engineering, 111, 1148-1155.

CHAPTER 8

THE DEVELOPMENT OF SMALL-SCALE BEDFORMS IN TIDAL ENVIRONMENTS: AN EMPIRICAL MODEL FOR UNSTEADY FLOW AND ITS APPLICATIONS

Sedimentology, 41, 883-903

ABSTRACT

Dimensions and plan morphology of current ripples are generally considered to vary with flow velocity and grain size. Recently, however, it has been shown that for sand of D_{50}=0.095 and 0.238 mm the equilibrium dimensions are identical at all velocities within the stability field of ripples and that the plan form of equilibrium ripples is linguoid. On this basis, an empirical unsteady flow model has been developed and tested with flume experiments in order to predict ripple development in natural depositional environments. The model includes the development of washed-out ripples and upper stage plane bed. The unsteady flow model explains the development and preservation of small-scale bedforms in various tidal environments more accurately than previous models. Such bedforms can serve, therefore, as indicators of prevailing hydrodynamic conditions.

INTRODUCTION

Current theories consider the dimensions and plan form of ripples to depend on flow velocity (e.g. Allen, 1968, 1969, 1984; Harms, 1969; Banks & Collinson, 1975; Reineck & Singh, 1980). The general opinion is that the plan morphology of ripples changes with increasing flow velocity from straight, via sinuous to linguoid (Allen, 1968; Harms, 1969). This hypothesis, however, complicates the interpretation of several processes and structures observed in tidal environments. For example, why are linguoid ripples abundant on tidal flats during slack water, when flow velocity has gradually decreased and passed through the stage in which straight ripples should have been formed?

The idea that the morphology of ripples is defined by flow velocity was questioned by Baas et al. (1993), who demonstrated that the time factor is important in ripple dynamics. Steady, uniform flow experiments in a flume showed that the equilibrium dimensions of current ripples in very fine sand (D_{50}=0.095 mm) are constant and independent of flow velocity. Only the time required to achieve an equilibrium geometry depends on flow strength (Baas et al., 1993; Baas, 1994). Starting from flat bed conditions, ripples develop in sequence from an incipient form via straight and sinuous forms towards linguoid forms (Fig. 1). These linguoid ripples increase in average height and wavelength until average dynamic equilibrium values are eventually reached for the given grain size.

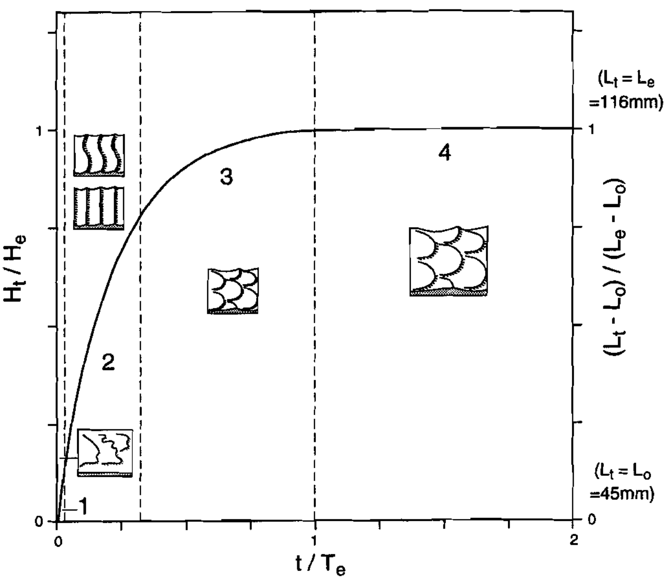


Figure 1: Standardized development plot of ripple height and wavelength in very fine sand (D_{50}=0.095 mm), sorting coefficient (Folk, 1968) = 0.49. The stages of development are separated by vertical, dashed lines. Initial and equilibrium values are indicated along the ordinates.

Similar results were obtained from flume experiments with fine sand, D_{50}=0.238 mm (Baas, 1993). Equilibrium heights and wavelengths in fine sand were observed to be larger than in very fine sand (17.0 mm versus 13.1 mm, and, 141.1 mm versus 115.7 mm). This is in agreement with earlier observations of a proportional relationship between grain size and ripple wavelength (Yalin, 1985), and it is also suggested by the data of Khandriche et al. (1986). In a 2.4 m wide flume, Simons et al. (1965) observed that ripples formed in 0.19 mm, 0.27 mm and 0.46 mm sand, initially have a uniform shape and wavelength (straight to sinuous ripples), and with continued flow change into linguoid ripples. Similar observations were made by Jain & Kennedy (1971), Mantz (1978) in 0.015 mm and 0.066 mm sand and Costello & Southard (1981) in 0.510 mm, 0.600 mm and 0.660 mm sand. It is therefore concluded that the above development stages of ripples observed in 0.095 mm and 0.238 mm sand also apply to other grain sizes. Hence, the plan form of ripples in natural environments reflects their development stage.

In order to understand the processes governing the development of current-related bedforms in tidal environments the objective of this paper is to develop an empirical unsteady flow model. This model is based on the above mentioned flume studies of Baas et al. (1993)

and Baas (1994), and is tested with flume experiments. The aim of the model is to explain bedform sequences in various tidal subenvironments, which so far have been poorly understood with existing theories of small scale bedform dynamics.

EMPIRICAL MODEL FOR BEDFORM DEVELOPMENT IN UNSTEADY FLOW

The development of small-scale bedforms under unsteady flow conditions can be computed by extending the equations of the empirical steady flow model (Baas et al., 1993; Baas, 1994) to an unsteady flow model for bedform development. The results of theoretical calculations with this model of bedform development during an unsteady tidal flow are verified empirically by simulating these flow conditions in a flume.

Theoretical model

The empirical bedform stability model for steady flow (Baas et al., 1993; Baas, 1993; Baas, 1994) is based on two sets of best-fit equations, quantifying the relationship between the dimensions of the bedforms in 0.095 mm sand, the 10°C-equivalent, depth-average flow velocity, the time necessary to reach equilibrium bedform morphology, and the development time for ripple height and wavelength. The equations are:

$$\frac{H_t}{H_e} = 1 - (0.01)^{\frac{t}{T_{e,H}}} \tag{1a}$$

$$\frac{L_t - L_o}{L_e - L_o} = 1 - (0.01)^{\frac{t}{T_{e,L}}} \tag{1b}$$

$$U_{10} = 0.233 + 0.255 T_{e,H}^{-0.442} \tag{2a}$$

$$U_{10} = 0.233 + 0.297 T_{e,L}^{-0.47} \tag{2b}$$

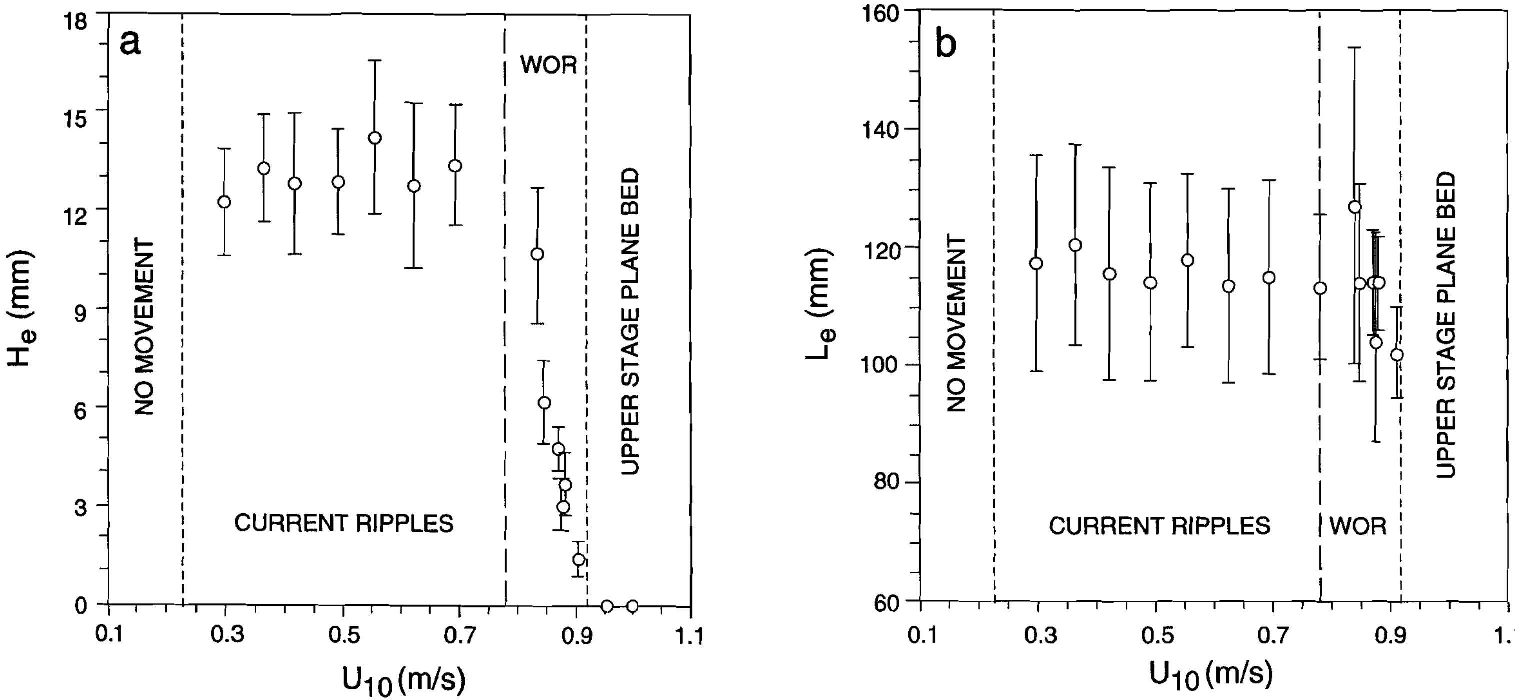


Figure 2: (a) Equilibrium bedform height (H_e) against 10°C-equivalent flow velocity (U_{10}) for D_{50}=0.095 mm. Vertical bars indicate standard deviations. WOR=washed-out ripples. (b) Equilibrium bedform wavelength (L_e) against 10°C-equivalent flow velocity (U_{10}) for D_{50}=0.095 mm.

where H is the bedform height (mm), L is the bedform wavelength (mm), t is the development time (h), T_e is the equilibrium time (h), and U_{10} is the depth-average flow velocity equivalent to a water temperature of 10°C. Subscripts indicate the actual (t), equilibrium (e) and initial (o) heights and wavelengths. The equilibrium time T_e is defined as the time needed to reach 99% of the equilibrium bedform dimensions from initial flat bed conditions.

Equation (1) describes the development of ripples (Fig. 1) and washed-out ripples (Southard & Harms, 1972; Jopling & Forbes, 1979) from initial flat bed conditions. In 0.095 mm sand, ripples are stable between the threshold flow velocity of general motion of sediment of 0.233 $m.s^{-1}$ and a flow velocity of 0.78 $m.s^{-1}$ (both 10°C-equivalent value calculated from Miller et al., 1977). Washed-out ripples occur as the transitional phase between ripples and upper stage plane bed (Southard & Harms, 1972; Jopling & Forbes, 1979) in a narrow velocity range between 0.78 $m.s^{-1}$ and 0.92 $m.s^{-1}$. Four different stages of ripple development were distinguished: (1) incipient ripples a few grain diameters high at the very first stage of ripple formation with an average initial wavelength of 45 mm; (2) straight to sinuous ripples (up to t/T_e-values of 0.3); (3) non-equilibrium linguoid ripples (between t/T_e-values of 0.3 and 1); (4) equilibrium linguoid ripples (at $t/T_e \geq 1$) with an average height and wavelength of 13 mm and 116 mm (Fig. 2). As stated above, these equilibrium dimensions are independent of flow velocity (Baas et al., 1993; Baas, 1994). The equilibrium wavelength of washed-out ripples is independent of flow velocity as well. Their equilibrium height, however, decreases rapidly with increasing flow velocity (Fig. 2a).

Equation (2) shows an inverse power relation between equilibrium time and flow velocity for both linguoid ripple and washed-out ripple height and wavelength (Fig. 3). At flow velocities near the transition towards upper stage plane bed it takes only a few minutes to reach equilibrium conditions, while near the threshold of sediment movement hundreds of hours are required.

Steady, uniform flow conditions hardly exist in natural environments. Therefore, the empirical steady flow model was used as the basis for an empirical unsteady flow model. The new model calculates the development of bedforms in unsteady flows by dividing flow velocity curves into small velocity increments, $\Delta U_{10}/\Delta t$, and integrating the contributions of these increments to bedform development according to the equation:

$$S=\sum \frac{\Delta t_i}{T_{e,i}} \tag{3}$$

where S is the cumulative development stage, Δt_i is the time period of the i-th increment, and $T_{e,i}$ is the equilibrium time associated with ΔU at the i-th time increment.

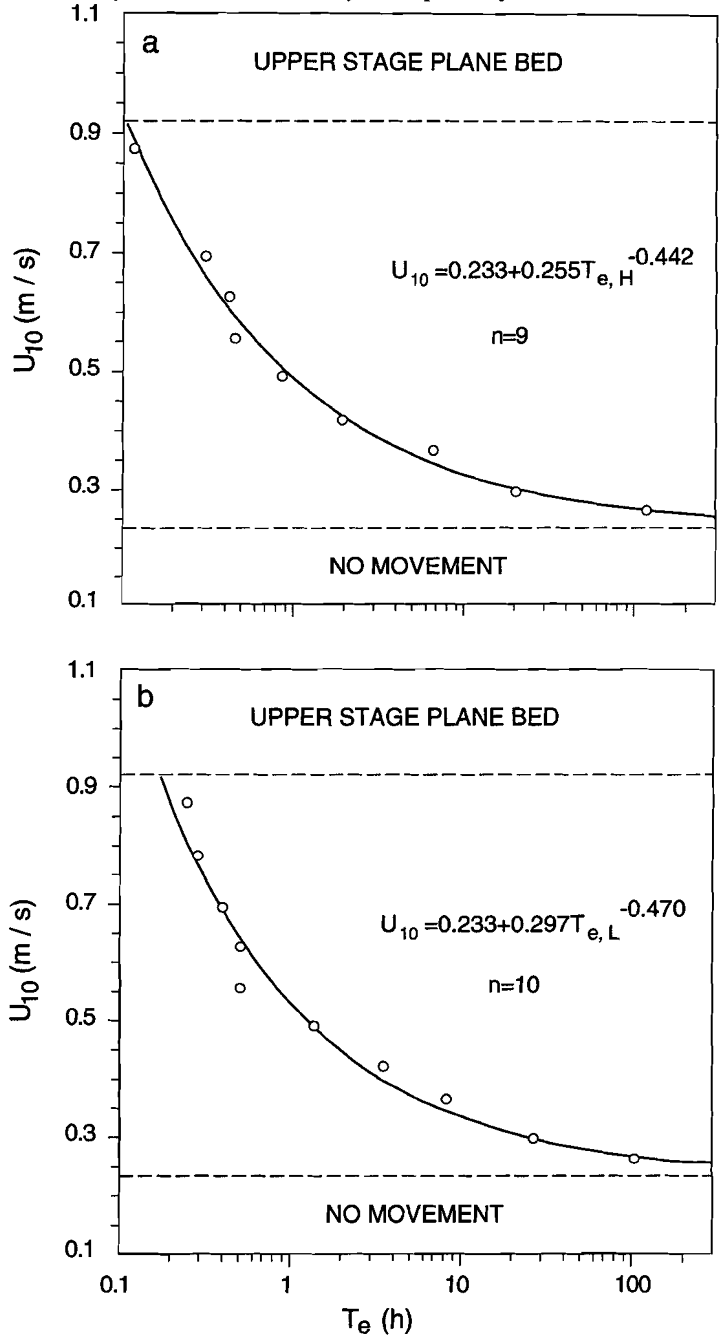


Figure 3: (a) 10°C-equivalent flow velocity (U_{10}) as a function of equilibrium time (T_e) for bedform height in very fine sand (D_{50}=0.095 mm), including a best-fit power function through the data points. (b) The same type of diagram for bedform wavelength. n=number of measurements.

Combining Eqs (2) and (3) gives a cumulative development stage in terms of flow velocity:

$$S_H = \sum \Delta t_i \left(\frac{U_{10,i} - 0.233}{0.255}\right)^{\frac{1}{0.442}} \tag{4a}$$

$$S_L = \sum \Delta t_i \left(\frac{U_{10,i} - 0.233}{0.297}\right)^{\frac{1}{0.47}} \tag{4b}$$

where $U_{10,i}$ is the 10°C-equivalent flow velocity at the *i*-th time increment.

Equation (4), rewritten for bedform height and wavelength development in very fine sand, gives:

$$\frac{H_t}{H_e} = 1 - (0.01)^{S_H} \tag{5a}$$

$$\frac{L_t - L_o}{L_e - L_o} = 1 - (0.01)^{S_L} \tag{5b}$$

where H_e=13.1 mm, L_o=44.6 mm, and L_e=115.7 mm. Equilibrium conditions are reached when S_H and S_L are equal to or greater than 1. Non-equilibrium conditions have values ranging from 0 to 1.

This model includes the time periods involved in the adaptation from upper stage plane bed to washed-out ripples, from high velocity washed-out ripples to low-velocity washed-out ripples, and from washed-out ripples to 'normal' ripples. Adaptation times for reverse changes in bedform type cannot be obtained from the original steady flow model, and are therefore approximated in the unsteady flow model by a vertical step function suggesting zero adaptation time. This seems a legitimate approximation, because at the high flow velocities involved these changes in bedform type take place rapidly (in the order of minutes).

The empirical unsteady flow model requires two corrections for the relation between flow velocity and equilibrium time. Firstly, Eq. (2) is based on equilibrium flow conditions, while at non-equilibrium conditions, i.e., during bedform development, the flow velocity is slightly higher due to a smaller bed roughness. The actual effect per time increment depends on both the bed roughness and the flow strength. It is largest in rapid flow over a flat bed, and smallest in slow flow over bedforms approaching equilibrium dimensions. In the experiments described in this paper an average correction of 2%, based on results of Baas (1994), was applied to flow velocities at non-equilibrium conditions.

Secondly, in the velocity ranges between 0.30 $m.s^{-1}$ and 0.42 $m.s^{-1}$ for ripple height, and between 0.30 $m.s^{-1}$ and 0.49 $m.s^{-1}$ for ripple wavelength, the best-fit curves of Eq. (2) predict slightly smaller equilibrium times than those given by the data points in Fig. 3. In these flow velocity ranges, the unsteady flow model has been corrected by using the linearly interpolated equilibrium times of the experimental data points instead of the equilibrium times given by the best-fit curves of Eq. (2).

Input parameters

Tidal flood or ebb phases in semi-diurnal environments can be approximated by a sinusoidal velocity curve, with the maximum flow velocity after 3 h. In natural semi-diurnal environments, the period of an ebb-flood cycle is 12.3 h. The time for one flood or ebb phase was chosen at 6 h, leaving out about 10 min for slack water conditions. Based on observations at the turning of tidal currents, this is a reasonable approximation, especially for subtidal environments.

Due to experimental restrictions, the sine-curve was approximated by raising the flow velocity in small steps from zero velocity to a maximum value at 3 h and subsequently lowering it in similar steps. In this way, four runs were conducted along different flow velocity curves shown in Fig. 4 and Table I. The velocity curves were chosen in such a way that the maximum flow velocity corresponds to a specific bedform type in the bedform stability diagram at a median grain size of 0.095 mm (Fig. 4). These curves are comparable to those at which natural bedforms are generated (see below). The experimental velocity curves were used as input for the theoretical calculations of bedform development (Eq. 5). All calculations started from flat bed conditions. The results of the calculations are summarized in eight bedform development curves (solid lines in Fig. 5) and in Tables II-V.

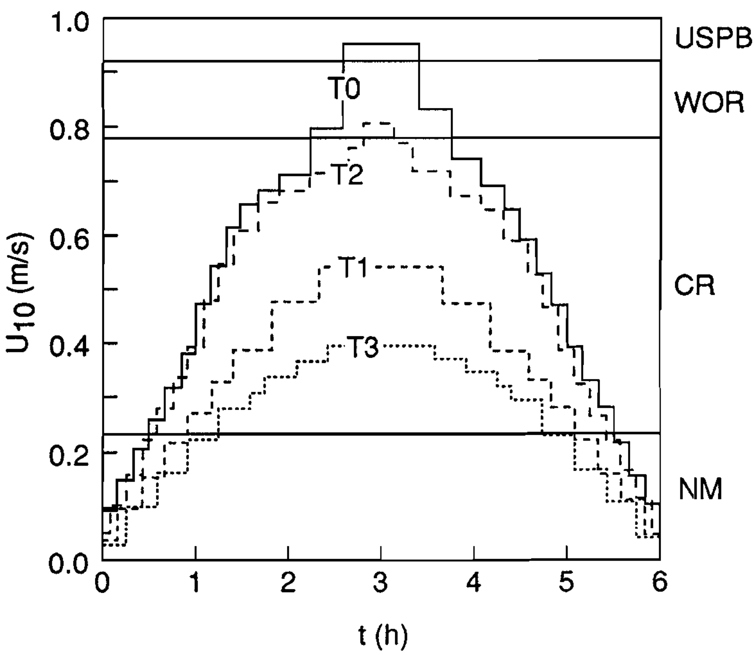


Figure 4: Flow velocity curves for the conducted experiments.

Table I: Hydraulic results and bedform types.

	T0	T1	T2	T3
h_o (m)	0.32	0.32	0.326	0.332
T (°C)	17-18.5	16.2-16.7	16.5-16.9	16.5-16.8
$U_{10,max}$ (m/s)	0.947	0.539	0.802	0.390
Bedform types	NM-CR_e-WOR-UPB-WOR-CR_e-NM	NM-CR_e-NM	NM-CR_e-WOR-CR_e-NM	NM-CR_n-NM

NM = no movement; CR_n = non-equilibrium ripples; CR_e = equilibrium ripples; WOR = washed-out ripples; UPB = upper stage plane bed.

Model calculations

Run T3

The maximum flow velocity of run T3 is 0.39 $m.s^{-1}$, which is the lowest velocity of the conducted runs (Fig. 4 and Table I). The tidal velocity curve passes through the lower part of the stability field of ripples between 1.25 h and 4.75 h. Before and after, no sediment movement occurs.

The empirical model produces S-shaped development curves for ripple height and wavelength (solid lines in Fig. 5a-b). The plan form of the ripples changes from incipient into straight and sinuous at 2 h and finally into linguoid at 3.6 h. Linguoid ripples continue to grow until the flow velocity becomes too low for sediment transport, reaching an ultimate height of 10.9 mm and an ultimate wavelength of 98 mm. Thus, non-equilibrium ripples are predicted to form after one half of a semi-diurnal tidal cycle.

Run T1

The maximum flow velocity in run T1 is 0.54 $m.s^{-1}$ (Fig. 4 and Table I), denoting a position in the middle of the ripple field. The calculated development curves are given in Fig. 5c-d. As soon as grains begin to move after 0.92 h incipient ripples are formed. They change into straight and sinuous ripples at 1.52 h, which subsequently evolve into linguoid ripples at 2.12 h. Equilibrium height and wavelength are reached at 2.71 h and 3.03 h, respectively. Thereafter, equilibrium linguoid ripples are stable until the end of the run.

Run T2

In run T2 the maximum flow velocity is 0.80 $m.s^{-1}$. The flow velocity is high enough to form washed-out ripples around 3 h (Fig. 4 and Table I). The unsteady flow model starts

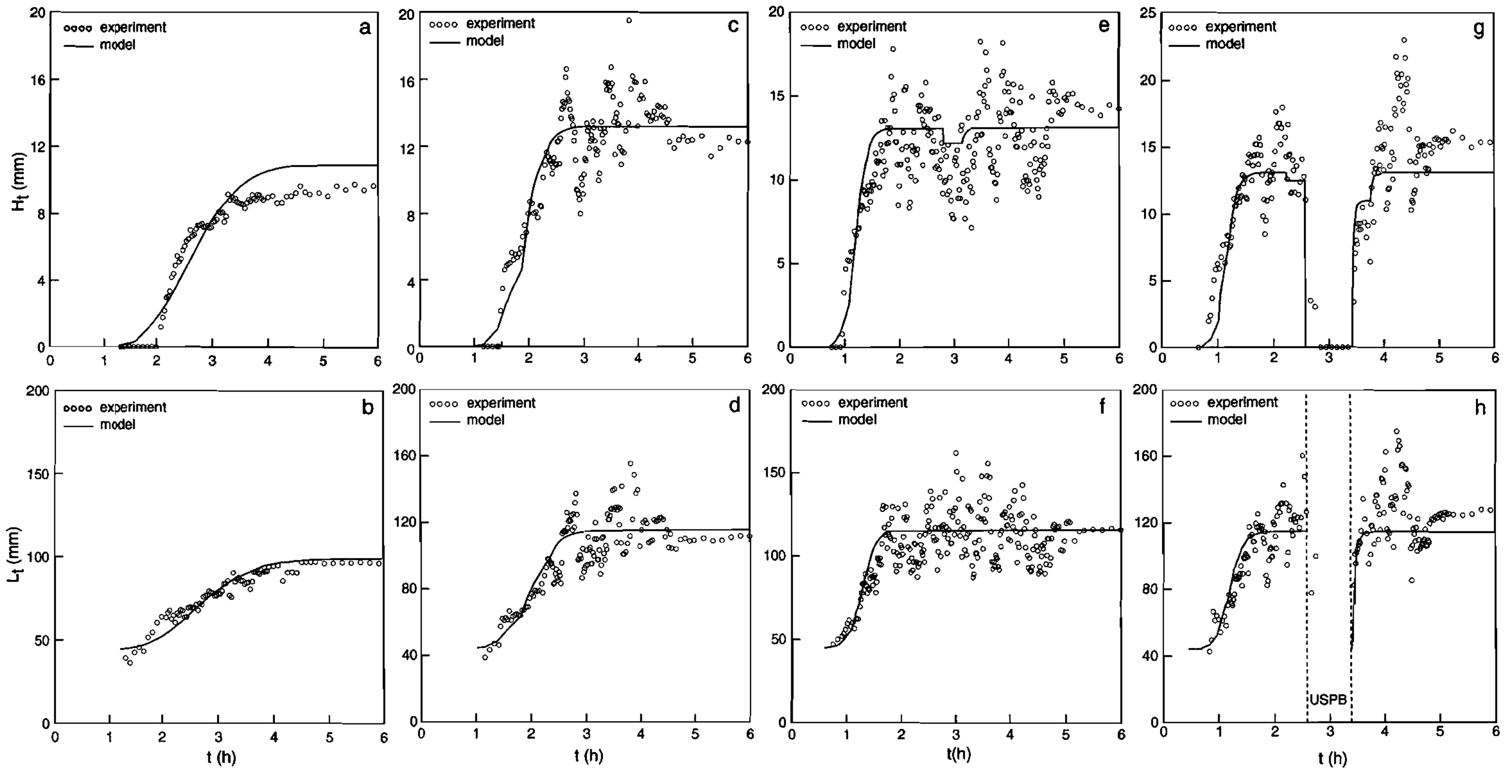


Figure 5: Calculated and measured development of bedform (a) height in run T3, (b) wavelength in run T3, (c) height in run T1, (d) wavelength in run T1, (e) height in run T2, (f) wavelength in run T2, (g) height in run T0, (h) wavelength in run T0.

calculating bedform heights and wavelengths at 0.58 h and predicts an equilibrium time of 1.68 h for height and of 1.82 h for wavelength (Fig. 5e-f). Equilibrium conditions are maintained until 2.83 h. Thereafter, the flow velocity curve enters the washed-out ripple field and a new equilibrium height of 12.2 mm is established (cf. Fig. 2a). As mentioned above, the unsteady flow model approximates this change in bedform type by a jump in height from 13.1 mm to 12.2 mm (Fig. 5e). At 3.17 h flow velocity re-enters the ripple field. The system needs about 0.1 h to re-establish equilibrium ripple height. The empirical model computes a constant equilibrium wavelength between 1.82 h and 6 h (Fig. 5f).

Run T0

Run T0 has a maximum flow velocity of 0.95 $m.s^{-1}$. The flow velocity curve covers all stability fields from no movement to upper stage plane bed (Fig. 4 and Table I). According to the calculated development curves, ripples develop first and reach their equilibrium height and wavelength at 1.57 h and 1.72 h, respectively (Fig. 5g-h). Bedform height decreases from 13.1 mm to 12.5 mm at 2.25 h, when flow velocity enters the stability field of washed-out ripples. Washed-out ripples change into an upper stage plane bed at 2.58 h. During flow deceleration, the upper stage plane bed is replaced by washed-out ripples (H_e=11 mm) at 3.42 h, and finally equilibrium ripple height is re-established at 3.88 h.

The empirical unsteady flow model calculates two separate development curves for bedform wavelength (Fig. 5h). The first curve starts at the threshold velocity of sediment movement. It follows an S-shaped path towards an equilibrium ripple wavelength of 115.7 mm at 1.72 h. Wavelength is not defined between 2.58 h and 3.42 h, when an upper stage plane bed is stable. The second development curve starts at 3.42 h, when the flow velocity re-enters the washed-out ripple field. Equilibrium wavelength is re-established rapidly at 3.65 h and remains constant thereafter.

Experimental control of the model calculations

The experiments were undertaken with the equipment described in Baas et al. (1993) and Baas (1994). The rectangular, infinite flume tank has a length of 7.96 m and a width of 0.45 m (Fig. 6; Winkelmolen, 1976). The sediment used in the experiments is a well-sorted, natural sand with a median grain diameter of 0.095 mm. Flow velocities, flow depths and water temperatures were measured concurrently at time intervals of 5 minutes during bedform development. Bedform dimensions were measured at time intervals ranging from 1 to 10 minutes, depending on the rate of morphological change. Bedform heights and wavelengths were averaged for each point in time. Each average is based on 10-40 ripples. The initial flow depth, water temperature, maximum flow velocity and bedform types of all experimental runs are given in Table I. Average bedform heights and wavelengths, measured in the experiments, are plotted as dots in Fig. 5. Tables II-V summarize the experimental results.

Table II: Development stages and bedform dimensions of run T3.

	Model	Experiment
Boundary NM-CR	1.25 h	1.25 h
Incipient CR	1.25-2 h	1.25-2.1 h
Str./sin. CR	2-3.6 h	2.1-±4.5 h
Linguoid CR	3.6-6 h	±4.5-6 h
Boundary CR-NM	4.75 h	4.75 h
H (at 6 h)	10.9 mm	9.7 mm
L (at 6 h)	98.0 mm	96.1 mm
L/H (at 6 h)	9.0	9.9
H_{max} (at 6 h)	-	14.2 mm
L_{max} (at 6 h)	-	134.4 mm

NM = no movement; CR = ripples; str = straight; sin = sinuous.

Table III: Development stages and bedform dimensions of run T1.

	Model	Experiment
Boundary NM-CR	0.92 h	0.92 h
Incipient CR	0.92-1.52 h	0.92-1.48 h
Str./sin. CR	1.52-2.12 h	1.48-2.24 h
Linguoid CR	2.12-6 h	2.24-6 h
$T_{e,H}$	2.71 h	2.6 h
$T_{e,L}$	3.03 h	2.65 h
H_e	13.1 mm	13.2 mm
L_e	115.7 mm	113.6 mm
L_e/H_e	8.9	8.6
H_{max}	-	33.1 mm
L_{max}	-	196.6 mm
Boundary CR-NM	5.08 h	5.08 h

NM = no movement; CR = ripples; str = straight; sin = sinuous.

Table IV: Development stages and bedform dimensions of run T2.

	Model	Experiment
Boundary NM-CR	0.58 h	0.58 h
Incipient CR	0.58-1.03 h	0.58-0.97 h
Str./sin. CR	1.03-1.34 h	0.97-1.54 h
Linguoid CR	1.34-2.83 h	1.54-2.83 h
$T_{e,H}$	1.68 h	1.8 h
$T_{e,L}$	1.82 h	1.67 h
H_e	13.1 mm	12.5 mm
L_e	115.7 mm	113.1 mm
L_e/H_e	8.9	9.0
H_{max}	-	28.0 mm
L_{max}	-	185.9 mm
Boundary CR-WOR	2.83 h	2.83 h
$T_{e,H}$	2.83 h	2.92 h
H_e	12.2 mm	10.2 mm
L_e/H_e	9.5	11.1
H_{max}	-	25.1 mm
Boundary WOR-CR	3.17 h	3.17 h
Linguoid CR	±3.27-6 h	±3.43-6 h
$T_{e,H}$	3.27 h	3.43 h
H_e	13.1 mm	13.1 mm
L_e/H_e	8.9	8.6
H_{max}	-	30.3 mm
Boundary CR-NM	5.42 h	5.42 h

NM = no movement; CR = ripples; WOR = washed-out ripples; str = straight; sin = sinuous

In general, the measured paths of bedform development correspond well with calculated paths. Most differences can be attributed to the data scatter, due to the dynamics of current ripple. The standard deviations of equilibrium ripple height and wavelength average 2 mm and 16.3 mm, respectively (Baas, 1994). Data scatter also explains why the ripple heights and wavelengths in the last hour of the experimental runs are often consistently different from the calculated values (e.g. Fig. 5d-e). This is caused by a "freezing" of deviating bed-form dimensions due to the low flow velocities in this time-period.

Table V: Development stages and bedform dimensions of run T0.

	Model	Experiment
Boundary NM-CR	0.5 h	0.5 h
Incipient CR	0.5-0.99 h	0.5-0.84 h
Str./sin. CR	0.99-1.26 h	0.84-1.29 h
Linguoid CR	1.26-2.25 h	1.29-2.25 h
$T_{e,H}$	1.57 h	1.65 h
$T_{e,L}$	1.72 h	1.69 h
H_e	13.1 mm	13.9 mm
L_e	115.7 mm	116.9 mm
L_e/H_e	8.9	8.4
H_{max}	-	24.6 mm
L_{max}	-	169.1 mm
Boundary CR-WOR	2.25 h	2.25 h
$T_{e,H}$	2.25 h	2.34 h
H_e	12.5 mm	12.9 mm
L_e/H_e	9.3	9.1
H_{max}	-	18.5 mm
Boundary WOR-UPB	2.58 h	2.58 h
$T_{e,H}$	2.58 h	2.78 h
$T_{e,L}$	2.58 h	2.78 h
Boundary UPB-WOR	3.42 h	3.42 h
$T_{e,H}$	3.57 h	3.53 h
$T_{e,L}$	3.65 h	3.85 h
H_e	11 mm	8.9 mm
L_e	115.7 mm	126.1 mm
L_e/H_e	10.5	14.2

H_{max}	-	12.5 mm
L_{max}	-	183.8 mm
Boundary WOR-CR	3.75 h	3.75 h
Linguoid CR	±3.88-6 h	±4.2-6 h
$T_{e,H}$	3.88 h	4.2 h
H_e	13.1 mm	15.9 mm
L_e/H_e	8.9	7.9
H_{max}	-	27.6 mm
Boundary CR-NM	5.5 h	5.5 h

Table V (cont'd): Development stages and bedform dimensions of run T0. NM = no movement; CR = ripples; WOR = washed-out ripples; UPB = upper stage plane bed; str = straight; sin = sinuous

The time required for linguoid ripples to change into washed-out ripples was about 5 minutes in runs T2 and T0, and it took about 10 minutes to change from washed-out ripples to upper stage plane bed in run T0. These values warrant the assumption of the model that these changes occur rapidly.

Computed washed-out ripple heights are often slightly larger than measured heights (Tables IV and V, and Fig. 5e-g). A detailed study on equilibrium heights in the stability field of washed-out ripples is needed to improve the empirical model on this point.

Equilibrium heights and equilibrium times of ripples in the second half of run T0 are larger than the heights calculated by the model (Table V and Fig. 5g). This deviation is caused by the large ripple height occurring around 4.3 h, and the fixation of the heights at 15.5 mm at the end of the run. In the latter case, time was too short for the ripples to return to smaller equilibrium heights. The deviation is within the error range of two times the standard deviation of equilibrium height.

Apart from these deviations, Fig. 5a-h and Tables II-V demonstrate that the empirical unsteady flow model provides a reliable tool to predict bedform development in tidal currents.

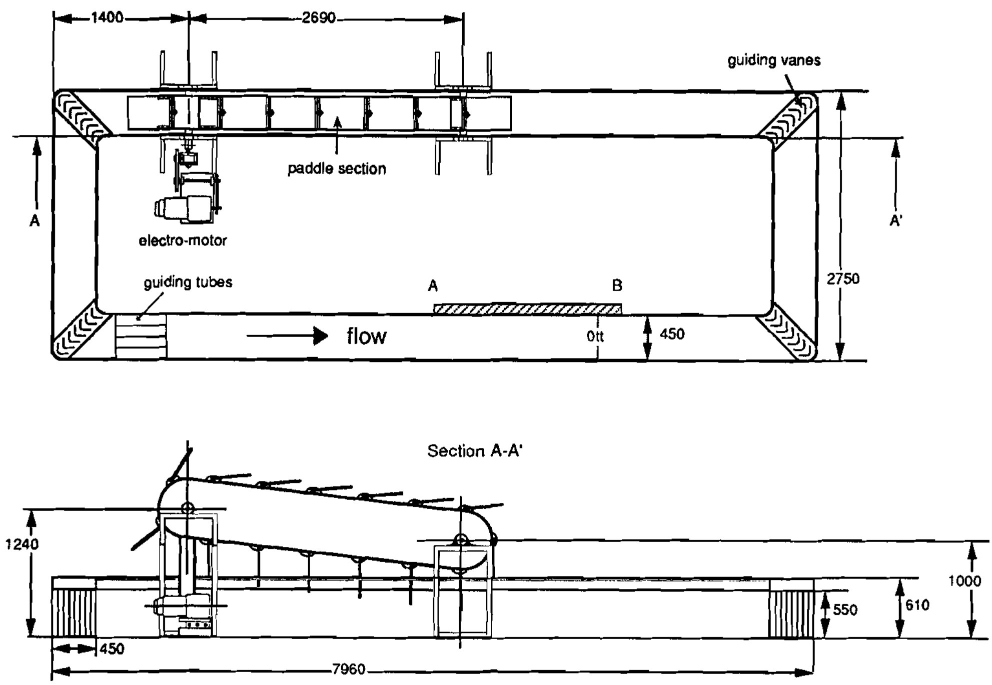


Figure 6: Flume, plan view and cross-section. All dimensions are in millimetres. See Winkelmolen (1976) for a full technical description. Ott=location of current meters. Section A-B is the transect along which bedform development was studied.

APPLICATION OF THE EMPIRICAL UNSTEADY FLOW MODEL TO CURRENT-RELATED STRUCTURES IN TIDAL ENVIRONMENTS

Introduction

The empirical unsteady flow model can be used as a first order approximation to understand bedform development in 0.095 mm sand in tidal environments. Although tidal flow velocity curves in nature can be more complex (e.g. Fig. 7), the simplest tidal velocity curve, the sine-curve, was used. This allows comparison of the model with the field observations in a semi-quantitative way and demonstrates the general applicability of the unsteady flow model in nature. The maximum tidal flow velocity governs the development path of the bedforms.

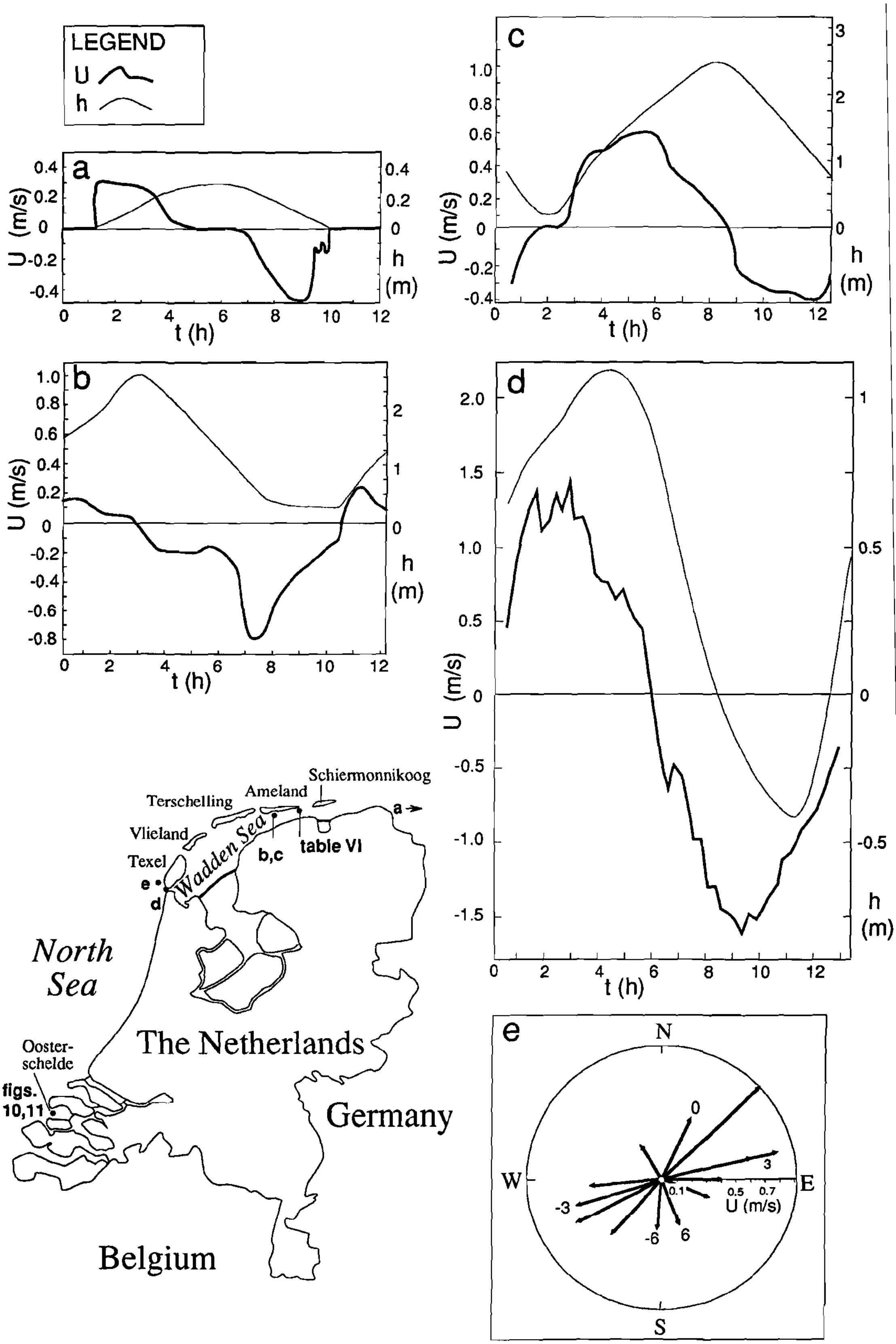
Obviously, when the maximum flow velocity is below the threshold velocity of sediment motion a flat bed will persist for 6 h. When the flow velocity reaches a maximum value within the stability field of ripples, it depends on the exact position within that field whether the ripples reach an equilibrium geometry. Fig. 8 shows that, according to half a sine-shaped flow velocity curve, a maximum flow velocity of about 0.45 $m.s^{-1}$ (0.436 $m.s^{-1}$ for height and 0.474 $m.s^{-1}$ for wavelength) is required to form equilibrium ripples after exactly 6 h. The theoretical value of 0.45 $m.s^{-1}$ is in agreement with the experimental results, which indicate a maximum flow velocity between 0.39 $m.s^{-1}$ and 0.54 $m.s^{-1}$.

At maximum tidal flow velocities between 0.78 $m.s^{-1}$ and 0.92 $m.s^{-1}$, washed-out ripples are formed around 3 h and ripples are present in the remaining time periods except in the first period of no movement. For maximum flow velocities higher than 0.92 $m.s^{-1}$ the sequence of bedform types is initial flat bed - linguoid ripples - washed-out ripples - upper stage plane bed - washed-out ripples - linguoid ripples. For all tidal flow velocity curves, ripples are frozen when the flow velocity becomes lower than 0.233 $m.s^{-1}$. In conclusion, in very fine sandy tidal environments ripples should be dominant at slack water conditions.

Based on the observation that ripple dimensions for other grain sizes are also independent of flow velocity (Baas, 1993a), it is concluded that a similar unsteady flow model can be made for sands with other grain sizes. Therefore, a new interpretation of bedforms and bedform sequences in various tidal sub-environments is given.

Depending on the tidal amplitude each sub-environment is characterized by a specific hydrodynamic character, illustrated by flow velocity curves from the mesotidal Wadden Sea (Fig. 7). The character of the flow velocity curve depends on the topographic position within the tidal system and on the tidal range. The topographic position determines the duration of the flow, especially on the emerging intertidal areas, and it influences the strength of the flow. Furthermore, the strength of the flow seems to increase with the tidal range (cf. De Boer et al., 1989) and the tidal volume (Fig. 7).

Figure 7 (next page): Typical flow velocity curves along the mesotidal Dutch/German coast. Numbers on map refer to figures. Note that all velocity curves approach a sine-curve. (a) Flow velocity curve on the intertidal flats of Germany (modified after Reineck, 1978). (b) Flow velocity curve in an intertidal gully behind Ameland (modified after Van Straaten, 1954). (c) Flow velocity curve in a shallow subtidal channel (modified after Van Straaten, 1954). (d) Flow velocity curve in an inlet channel (modified after Postma, 1954). (e) Ellipsoidal flow velocity curve in the Dutch offshore region (modified after Sha, 1990).



LEGEND
U
h
a
U (m/s)
h (m)
t (h)
b
c
d
e
N
S
W
E
U (m/s)
0
3
-3
-6
6
0.1
0.5
0.7
Schiermonnikoog
Ameland
Terschelling
Vlieland
Texel
Wadden Sea
b,c
table VI
e
d
North Sea
The Netherlands
Oosterschelde
figs. 10,11
Germany
Belgium

Intertidal environments

Intertidal flats

On intertidal flats flow velocities are relatively low. For instance, in the mesotidal Dutch Wadden Sea the maximum velocity usually does not exceed 0.4 $m.s^{-1}$, contrary to more than 1 $m.s^{-1}$ in the subtidal inlets (Fig. 7; Eisma, 1980). The tidal flow velocity curve on tidal flats normally approaches a sine-curve (e.g. Fig. 7a). The period during which the flood and ebb currents are active is largely restricted by the height of the tidal flat with reference to the low water level. The higher parts of an intertidal flat are submerged only during 5-6 h around high water.

The relatively low flow velocities on mesotidal intertidal flats and the limited period during which the tidal current is active, strongly limit the development of ripples. The higher the intertidal flat, the less time is available to form equilibrium ripples within one ebb or flood period (see Eqs 4a and 4b). The model predicts that, going from the higher to the lower intertidal flat, current ripples will in general evolve towards equilibrium dimensions, i.e., from straight, via sinuous to linguoid patterns (cf. Wunderlich, 1967).

Intertidal flats, Bay of Fundy, Canada

A part of the macrotidal Bay of Fundy consists of a flat, fine sandy intertidal area. The sedimentary sequence deposited here shows an alternation of parallel-laminated sands and (minor) current ripples. Single ripple sets have a thickness of 10-30 mm (Dalrymple et al., 1990). The current ripples are mostly straight-crested on the higher parts of the tidal flats and linguoid on the lower parts (Dalrymple, pers. comm.).

The tidal flow velocity curve is asymmetrical. Directly after submergence of the flat, upper stage plane bed conditions are established (Dalrymple, pers. comm.). Flow velocities reach maximum values of 1.5-3 $m.s^{-1}$ (Dalrymple et al., 1990) and decrease thereafter. During flow deceleration, rapid settling from suspension results in thick parallel-laminated sand layers, deposited at upper stage plane bed conditions (Dalrymple et al., 1990). During the final stage of the tidal flow, when the flow velocity enters the stability field of ripples, ripples start to form on the flat bed. Several tens of minutes are needed to reach equilibrium dimensions (cf. run T0). The straight-crested ripples observed on the upper intertidal flats can therefore be explained to result from a very rapid decrease in flow velocity, whereas the linguoid ripples on the lower intertidal flats obviously had sufficient time to reach equilibrium dimensions, because of the longer duration of the flow.

Equilibrium ripple height restricts the depth to which ripples erode the underlying deposits. In the Bay of Fundy the layers of parallel-laminated sand are too thick to be completely reworked into ripples and thus they are partially preserved. Hence, the preservation of parallel lamination in tide-influenced sediments, as in the Bay of Fundy, can be interpreted as the result of high rates of deposition of sediment settling from suspension at upper stage plane bed conditions, especially during flow deceleration.

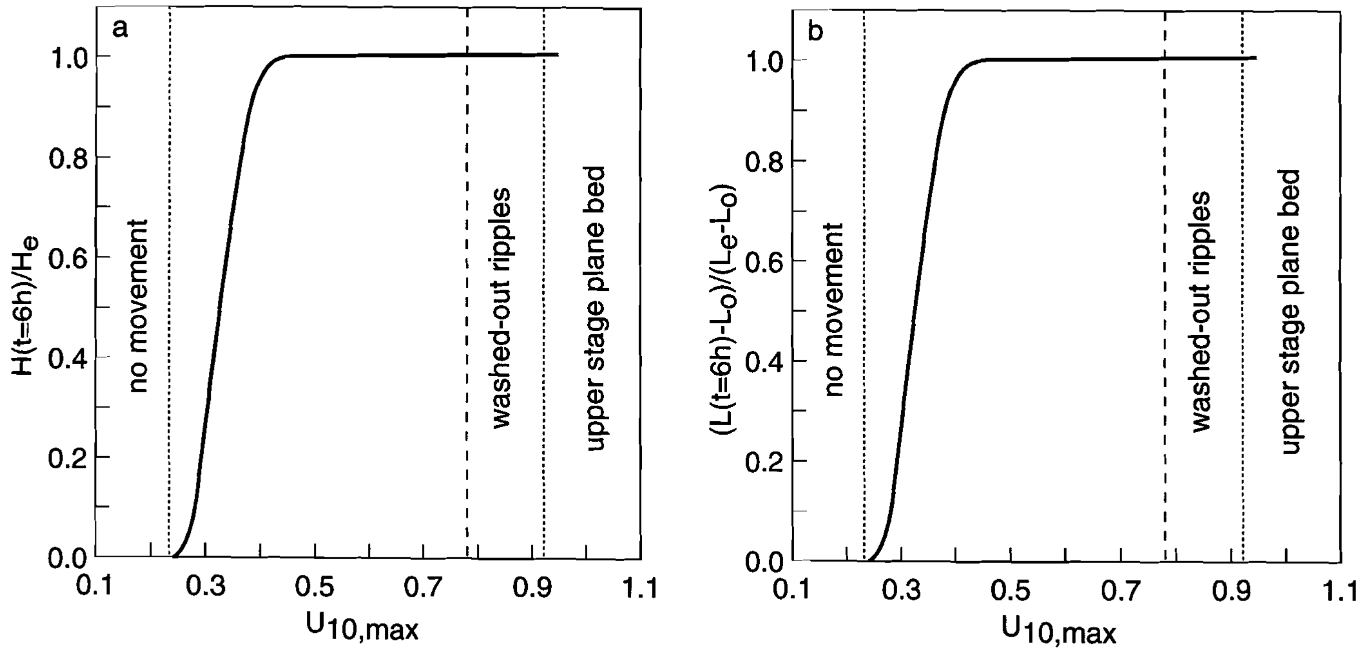


Figure 8: (a) Standardized bedform height after 6 h as a function of maximum tidal flow velocity under a sine-shaped velocity curve. (b) Standardized bedform wavelength after 6 h as a function of maximum tidal flow velocity.

Intertidal flats, Bay of Mont-Saint-Michel, NW France

Tessier (1992) described and discussed climbing ripples in the upper intertidal zone (h=0.5-1 m below spring tide level; Tessier, pers. comm.) of the Bay of Mont-Saint-Michel. The climbing ripples are linguoid in shape. Ripple-sets consist of series of laminae, each of which is normally graded (Fig. 9). In each set of climbing ripples the foresets thicken towards the middle and become sigmoidal. Towards the top of the set the laminae become overall finer grained. Furthermore, the slope and thickness of these laminae decrease (Fig. 9; Tessier, 1992).

According to Tessier (1992) each lamina represents one flood period. One set of climbing ripples is thought to represent a full neap-spring-neap cycle (Fig. 9). Towards spring tide the foresets of the linguoid ripples thicken and become sigmoidal. Towards neap tide the laminae become thinner again and grain size decreases.

The observation that linguoid ripples are stable during most of the neap-spring-neap cycle at the prevailing low flow velocities can be explained with the empirical unsteady flow model. Individual lamina formed in one flood (Fig. 9) show that the amount of sediment deposited is small, indicating low flow velocities and thus a large equilibrium time (see Eqs 1 and 2). During each tide the linguoid ripples continue to migrate as long as the flow velocity is above the threshold value of sediment motion for the sediment size in question. The ripples continue to develop until their equilibrium linguoid dimensions are reached (cf. run T3). Major changes in ripple shape do not occur within one tide, because the development stages and equilibrium dimensions are identical at all flow velocities.

Around spring tide, very fine sand dominates deposition, allowing the formation of relatively large ripples. Around neap tide, finer sediments (mainly silts) are deposited, enhancing the formation of smaller ripples (Yalin, 1985; Khandriche et al., 1986). These smaller neap tide ripples reshape the larger spring tide ripples into smaller linguoid forms (Fig. 9).

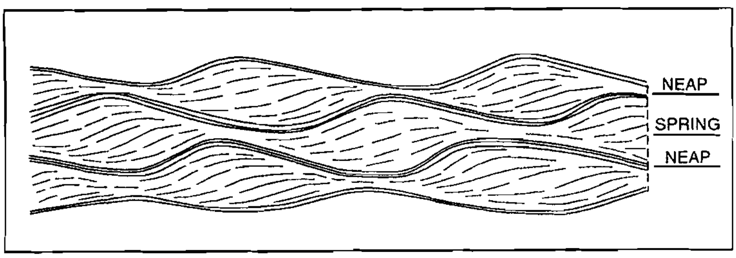


Figure 9: Sets of climbing ripples (courtesy Tessier, 1992).

Intertidal gullies, Pinkegat, the Netherlands

In intertidal gullies flow directions are in general more or less bi-directional. In the mesotidal Dutch Wadden Sea the flow velocity in gullies is typically less than 0.2 $m.s^{-1}$ during the longer part of the tide, but values as high as 0.8 $m.s^{-1}$ at peak ebb discharge have been observed (Fig. 7b; Van Straaten, 1954; Oost & De Boer, 1994). This implies that during a relatively long period no or limited sediment movement occurs. Flow strength increases over a period of 3 to 4 h to, at maximum, ebb velocities that allow washed-out ripples to develop in very fine sand.

Observations and measurements in deep, narrow intertidal gullies in the tidal flats of the Pinkegat inlet system (Fig. 7) show that ripples are predominantly linguoid. In shallower, narrow intertidal gullies ripples are smaller and have a more sinuous to straight plan form (Table VI). In deep, broad intertidal gullies ebb-ripples are generally straight to sinuous at the edges of the gully and are more linguoid towards the centre. The dimensions of the ripples measured in the gullies of the mesotidal Pinkegat-system are largely similar to experimentally formed ripples (Baas et al., 1993; Baas, 1993a). Only during the latest part of the ebb flow, they were slightly flattened due to shallow flow depths.

Table VI: Current ripple data from intertidal gullies and intertidal flats of the Pinkegat inlet system (see Fig. 7 for location). All observations were done in wave-sheltered parts of gullies.

Sample	D50 (mm)	Sorting	Ripple plan form	n	H (mm)	H_{max} (mm)	L (mm)	Remarks
91wad1	0.146	0.32	ling (sin)	36	11.5 ± 6.1	25	121.1 ± 34.9	i.g.
91wad2	0.151	0.32	str	38	10.5 ± 3.7	17	94.6 ± 16.0	i.f.
91wad3	0.169	0.31	str	22	5.1 ± 2.0	9	55.8 ± 17.2	i.g.
91wad4	0.166	0.30	str (sin)	20	9.8 ± 4.2	19	93.9 ± 16.8	i.g.
91wad5	0.135	0.32	ling	95	8.6 ± 5.0	27	100.5 ± 33.9	f.r.

Explanation: str = straight ripples; sin = sinuous ripples; ling = linguoid ripples; i.g. = intertidal gully; i.f. = intertidal flat; f.r. = flattened in last run-off phase.

As long as the flow velocity is within the ripple field and development time is sufficient, ripples in an intertidal gully continue to develop until equilibrium dimensions are met. This especially applies to the deeper parts of gullies, where the current lasts longer and the maximum flow velocities (Fig. 7b) are higher than in the shallower parts (cf. Eqs 4a and 4b). Therefore, linguoid ripples are more dominant in the deeper parts of intertidal gullies and non-equilibrium ripples at the margins.

Subtidal environments

In subtidal environments flow velocity curves are essentially sinusoidal. This allows predictions of the development of current-related structures using the empirical unsteady flow model. Because flow velocities are often high, equilibrium ripples develop quickly. The main factors restricting current ripple generation are therefore bioturbation and wave-reworking (see discussion below).

Subtidal channels, Pinkegat, the Netherlands

The flow pattern in subtidal channels is generally bidirectional with only a few minutes for slack water (Fig. 7c-d). In lagoons or on tidal flats, such channels are usually slightly dominated by the ebb current. In some cases, for instance in estuaries, either the flood or the ebb current dominates so strongly over the reverse current that locally the flow can be considered uni-directional. Within the channels the velocity curve approximates a sine-function (Fig. 7c-d). The flow velocities in subtidal channels are usually higher than on intertidal flats. For the mesotidal Dutch Wadden Sea, typical maximum flow velocities in the main inlets are between 1 $m.s^{-1}$ and 2 $m.s^{-1}$ (Fig. 7d), whereas maximum velocities in the back-barrier channels are usually less than 1 $m.s^{-1}$ (Fig. 7c; Eisma, 1980). At these maximum velocities upper stage plane beds or, depending on grain size, dunes are generated, which will normally merge into ripples during flow deceleration. Within the channels of the mesotidal Wadden Sea all bedform development curves corresponding to those in runs T0 to T3 are to be expected, because maximum flow velocities vary strongly with the tidal volume.

The development of subtidal ripples was observed by the authors at fair weather conditions during an ebb period in a 3 m deep subtidal channel of the mesotidal Pinkegat inlet system (Fig. 7), using under-water video recordings. The median grain size of the well-sorted channel sand was 0.140 mm. Around the turn of the tide flat bed conditions were generated by bioturbating shrimp, crab and flatfish. At 1.23 h after slack tide, sediment started to move at flow velocities of 0.19-0.20 $m.s^{-1}$ at 1 m above the bed. Incipient current ripples began to form after 1.67 h at a flow velocity of 0.31 $m.s^{-1}$. Straight ripples were formed after 1.88 h at a velocity of 0.36 $m.s^{-1}$. Bioturbation regularly destroyed these bedforms. Subsequently, at 2.87 h and 0.51 $m.s^{-1}$, bioturbation decreased and the ripples grew and developed into linguoid ripples. At 3.35 h and 0.55 $m.s^{-1}$, the concentration of suspended sediment load strongly increased and washed-out ripples were formed. At 3.47 h and 0.56 $m.s^{-1}$, ripples started to disappear due to erosion and a shell lag was exposed. At 4.00 h and 0.64 $m.s^{-1}$ ripples had completely disappeared. A strong increase in the concentration of clay-floccules, mainly due to re-suspension processes (de Haas & Eisma, 1992; Eisma & Li, 1993), reduced visibility and prevented further observations during the waning phase of the ebb.

The development stages observed agree with the empirical unsteady flow model. However, the appearance of the subsequent stages was delayed by bioturbation, which strongly

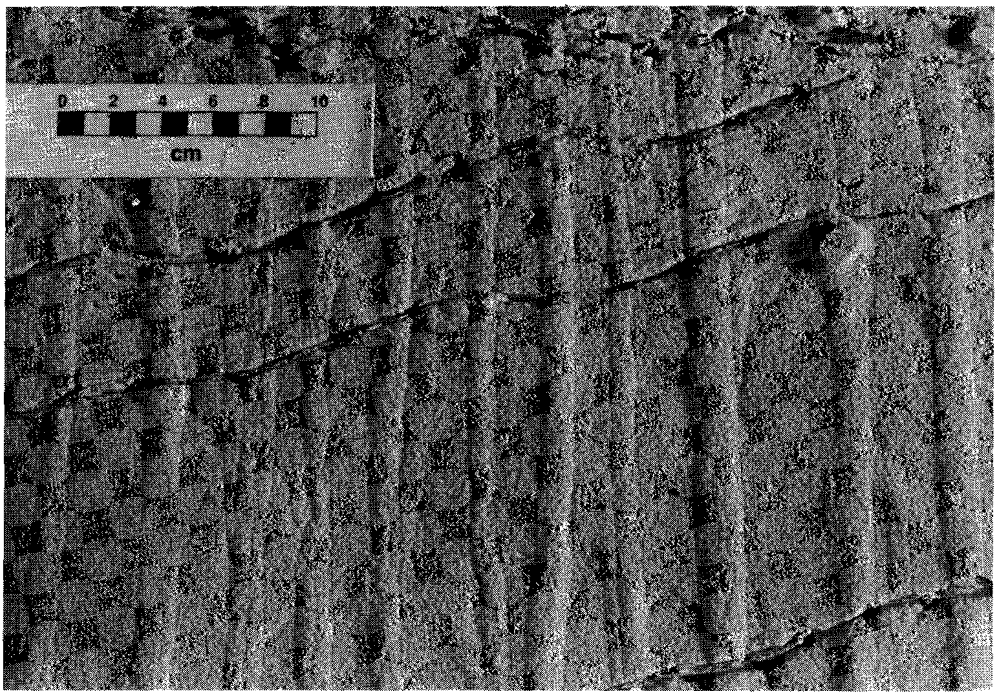


Figure 10: Fully preserved coflow ripples on the lower foresets of a large subtidal dune from the Oosterschelde (see Fig. 7 for location).

hindered bedform buildup at low flow velocities. As soon as bioturbation was prevented by the velocity of the flow, ripples rapidly developed to equilibrium dimensions. Washed-out ripples were formed at lower flow velocities than expected, probably because of high concentrations of suspended load and net sediment supply. These ripples were destroyed when the flow velocity entered the upper stage plane bed field (cf. run T0).

Linsen bedding, wavy bedding and flaser bedding

Linsen, wavy bedding and flasers (Terwindt, 1967; Reineck & Wunderlich, 1968; Terwindt & Breusers, 1971; Allen, 1984) are common in offshore deposits and in lagoonal subtidal to intertidal deposits (Reineck & Singh, 1980). They may include wave ripples, current ripples or both. In all these bedding types, fully preserved current ripples without internal erosion surfaces have been observed (cf. Van Straaten, 1954; De Raaf & Boersma, 1971). They are covered with clay which settled around the turn of the tide.

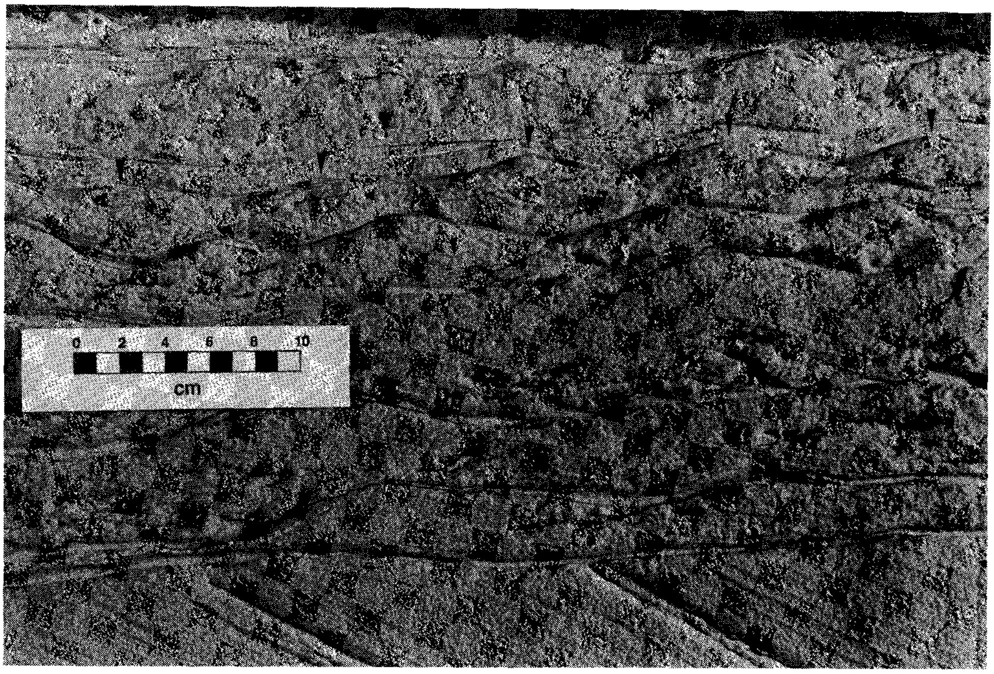


Figure 11: Backflow and coflow ripples on the bottomsets and lower foresets of a large subtidal dune from the Oosterschelde (see Fig. 7 for location). The arrows point to the tops of ripples where evidence of flow reversals is visible.

A well preserved example of wavy bedding was observed in the bottomsets and foreset toes of a large subtidal dune from the mesotidal Oosterschelde (Figs 7 and 10; see also Visser, 1980). The small scale bedforms represent coflow flood ripples on the foresets of the sandwave and have been preserved completely. The covering clay drape prevented erosion in a later stage (Terwindt & Breusers, 1982). Full preservation is incompatible with the conventional bedform theory, because the ripple dimensions should decrease when flow velocity decreases and the older ripples should thus be eroded.

These current ripples are well preserved, because they are stable up to the moment that they become inactive. The equilibrium forms are independent of flow velocity and thus do not decrease in size in a decelerating flow, whereas the non-equilibrium forms continue to develop at all flow velocities as long as they exceed the threshold velocity of sediment movement (cf. runs T1 and T3). Hence, the field observations in the Oosterschelde (Fig. 10) agree with the model prediction that no degeneration of ripples should occur upon decreasing flow velocity.

In the bottomsets and lower foresets of another large tidal dune in the Oosterschelde (De Boer et al., 1989), the effect of flow reversals is visible (Fig. 11). Backflow ripples were reworked into coflow ripples when tidal flow velocity decreased. The coflow ripples used the backflow ripples as a nucleus, while maintaining a constant wavelength and height. Shortly thereafter tidal flow velocity became low enough to allow suspended clay to settle, covering the then inactive ripples. Similar observations have been made on intertidal flats by Reineck (1961). Hawley (1981) showed experimentally that after a flow reversal newly formed ripples use older ripples as a nucleus. With previous models it is difficult to understand why the coflow and backflow ripples have identical wavelengths although they are formed at different flow velocities.

According to the empirical unsteady flow model, the height and wavelength of the backflow ripples already represent an appropriate development stage to start the coflow ripple development, when the local current is reversed 180°. Only the orientation is incorrect. Obviously, the coflow changes the orientation of the ripple crests without significantly changing the ripple size, and ripple development is continued from thereon.

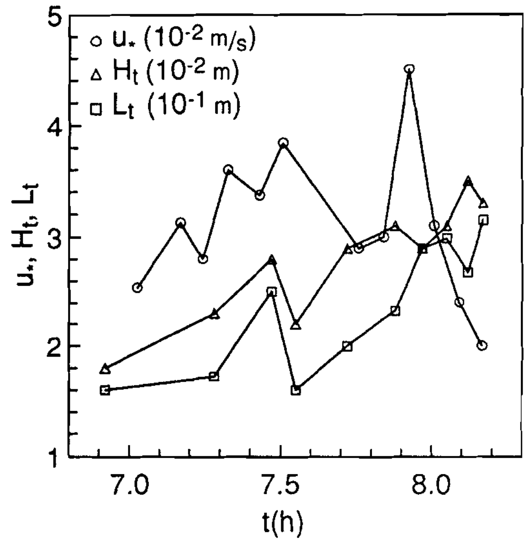


Figure 12: Development of shear velocity (u_*), current ripple height (H_t), and current ripple wavelength (L_t) in Puget Sound (h=31 m), Washington (modified after Kachel & Sternberg, 1971).

Deeper subtidal environments

Tidal flow patterns are often elliptical in deeper subtidal environments on the continental shelf (Fig. 7e; e.g. Dietrich et al., 1975) and the tidal flow velocity curve approximates a sine-function. Flow velocities are generally high compared to those on the intertidal flats. For the mesotidal Dutch offshore area, for instance, typical maximum flow velocities near the coast are up to 1 m.s^{-1} (Fig. 7e; Sha, 1990). Further offshore, throughout the southern North Sea, they amount up to 0.5 m.s^{-1} (Walker, 1984), as opposed to 0.4 m.s^{-1} for the tidal flats. At such flow velocities the generation of equilibrium linguoid ripples is to be expected.

Deeper subtidal environment, Puget Sound, Washington

The development of ripples during an ebb tidal flow at a depth of 31 m in Puget Sound, Washington (Fig. 12) was studied by Kachel & Sternberg (1971). Once sand (D_{50}=0.370 mm) began to move, the ripples increased in size throughout the 1.2 h observation period. The flow velocity and bed shear stress attained maximum values after 0.67-0.75 h, and decreased thereafter. This observation is in sharp contrast with the assumption that the ripples should decrease in size when flow velocity decreases.

The continued growth in dimensions of the ripples is, however, in close agreement with the empirical unsteady flow model (cf. run T3). The most likely explanation is that the ripples continued to grow to reach equilibrium dimensions during the deceleration of the flow.

Deeper subtidal environment, Sable Island Bank, Scotia Shelf

Amos et al. (1988) studied the generation of ripples on the tide-influenced Canadian continental shelf at water depths of 22 m by photographing the surface at regular time intervals of 30 minutes. They observed that straight and linguoid ripples were formed at the same location under identical conditions of low wave energy and moderate flow velocity, whereas during periods of high flow velocity only linguoid ripples were present (Fig. 13).

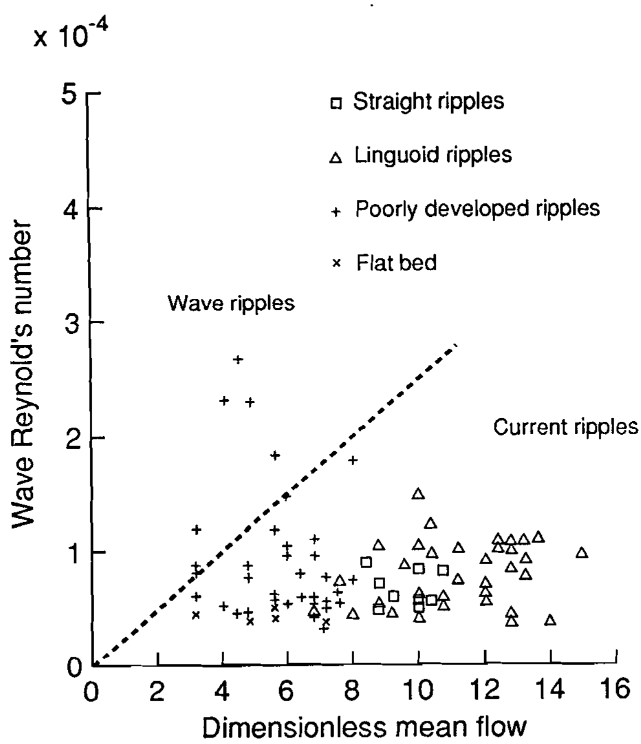


Figure 13: Scattergram of bedform types at Sable Island Bank (h=22 m), Scotia Shelf. Dimensionless mean flow (U_{100}/W_s) is plotted against wave Reynolds number ($U_{mb}d_o/v$) (modified after Amos et al., 1988). Bedforms type shown are linguoid ripples, straight ripples and poorly developed ripples (largely due to bioturbation at low flow velocities).

According to the empirical unsteady flow model, straight ripples are non-equilibrium ripples, which will develop into linguoid ripples with continued flow. The higher the flow velocity, the faster this development takes place. Thus the two types of ripples observed represent different development stages at the same flow velocity (Fig. 13). The lack of observations of other than linguoid ripples at higher flow velocities is most likely an artifact. When the flow velocity increases the period during which straight ripples exist is shortened and the chance that they will be photographed in a given time-interval decreases considerably (cf. Eqs 1 and 2). Additionally, the linguoid ripples could have been formed in an earlier stage of the tidal flow at lower flow velocities.

BIOTURBATION AND WAVE-REWORKING

Sediment transport can be influenced strongly by animals living in or on the sediment surface. The influence can be through binding the sediment, increasing the erodibility of the sediment or destructing surface relief (e.g. Jumars & Nowell, 1984). Mucous adhesion (Nowell et al., 1981) and burrows can strongly reduce the erodibility of the sediment.

Bioturbation can enhance the erodibility by (1) increasing the bottom roughness by (1a) local variations in fluid momentum impinging on the bed, for instance brought about by protruding worm tubes (Jumars & Nowell, 1984); (1b) by producing an irregular surface, for instance by *Arenicola arenaria* (Van Straaten, 1954; Nowell et al. 1981; Eckman et al., 1981; Oost & De Boer, 1994); (1c) particle exposure to the flow (Jumars & Nowell, 1984); (2) particle momentum, for instance actively bringing sediment grains in suspension (Nowell et al.., 1981; Jumars & Nowell, 1984) or (3) sorting of sediment, for instance the fining upward sequences made by *Corophium volutator*. These increases in erodibility enhance sediment transport. It is thus expected that current ripple formation is enhanced. A possible exception is when clay is brought into suspension, which may cause drag reduction (cf. Best & Leeder, 1993).

Some forms of bioturbation tend to destroy small-scale structures on the surface, and thus reduce bottom roughness. Intensive bioturbation can obliterate ripples formed during one tide. This was demonstrated by the earlier described video observations on the bottom of a subtidal channel in the backbarrier area of the Pinkegat inlet system. Similar observations were done by Van Straaten (1954; 1964), Rhoads & Young (1970) and Amos et al. (1988) (see also Fig. 13). On subtidal flats, in small channels and on intertidal flats, where flow velocities are low and slack water periods are long, ripples will often become reworked by bioturbation into more or less flat beds. Regeneration of ripples during the subsequent tidal flow will thus start from flat beds.

In our observations, at low flow velocities, the effect of destruction of incipient ripples by bioturbation is stronger than the combined effect of the increased bottom roughness and sediment suspension by bioturbation and the natural rate of formation of current ripples. At progressively higher flow velocities the bioturbation is suppressed and ripple formation becomes dominant.

Wave-formed structures also increase the bottom roughness. Amos & Collins (1978) pointed out that the relative energy of waves and tidal flow determine the exact nature of the bedforms generated. On the intertidal sandflats of the Wash (UK) they found that current ripples dominated on the lower tidal flats whereas wave ripples dominated on the higher flats. Thus, the higher the intertidal flat the greater the chance that wave reworking will take place.

DISCUSSION

The above examples demonstrate that ripple development over half a tidal flow is predictable when the flow velocity curve is known. Also, the bedforms observed can be understood well when one has a general idea about the relevant flow velocity curve. Because ripple development is independent of flow velocity, this can be extended to any flow velocity profile as long as the flow is unidirectional. Even if the current is not uni-, but bidirectional at 180°, height and wavelength of current ripples, which were already formed during the reverse flow will fit in the development curve of the ripples which are to be formed.

A major limitation to the predictive power of the unsteady flow model, however, is that momentarily ripple development curves are only known for well sorted very fine sand (Baas et al., 1993) and fine sand (Baas, 1994). Development curves for other grainsizes and the effects of factors like form, sorting, mineralogy and settling from suspension still have to be investigated. Another limitation is that rapid deposition from suspension suppresses the development of ripples.

Notwithstanding these limitations it is still possible to deduct the flow conditions which have prevailed from observations in the fossil record, assuming unidirectional flow, low settling from suspension and that general development curves at steady flow will be comparable with the profiles of Fig. 1, albeit with other equilibrium dimensions. Next to the above-mentioned examples of tidal environments the following general conclusions can be drawn:

1) Non-equilibrium linguoid, sinuous, straight and incipient current ripples are indicative of flow conditions where flow velocities were too low ànd time was too short to reach equilibrium conditions. Such ripples will mostly point to environments, where flow is regularly changing in direction and/or velocity, which also enables bed-form destroying conditions, such as waves, suspension-fallout, bioturbation etc., to be more effective at the sediment-water interface.

2) Equilibrium current ripples indicate flow conditions where flow velocities were high enough and time was long enough to reach equilibrium conditions. Such ripples can, for instance, be formed in a short time at high flow velocities or over a longer period at low flow velocities. This provides an explanation for the presence of well-developed current ripples formed under conditions of flow speeds just in excess of threshold, as are sometimes observed on the continental shelf (Jon J. Williams, pers. comm.).

3) Upper stage plane beds will not be preserved under steady state conditions of sediment supply. During flow deceleration they will always be changed into current ripples. Upper stage plane beds can only be preserved under conditions of net sediment deposition.

Deposition from suspension at upper stage plane bed conditions will result in a vertical sequence of horizontal laminae. This type of sequence will normally be topped by ripples formed during the deceleration of the flow. The erosion by ripple troughs may partly destroy the underlying horizontal laminae. Hence, the preservation of upper stage plane bed laminae depends on the depth of erosion of the ripples. Furthermore, the ripples and upper stage plane bed laminae may be separated by an unstructured interval, if the flow deceleration occurred very rapidly and consequently the rate of deposition of suspended sediment was too high to develop any structures (cf. Lowe, 1988).

Theoretically, an upper stage plane bed which is not topped by current ripples can only be preserved during extremely rapid, practically unrealistic, flow deceleration of the order of a few seconds. Nevertheless, horizontal lamination has been observed in various depositional environments. These deposits may form partly as upper stage plane bed laminae which are covered during the deceleration of the flow by sediment raining down from an oversaturated suspended load. At such conditions ripple development will be suppressed due to the high rates of deposition from suspension (Lowe, 1988).

CONCLUSIONS

In contrast to current theories (e.g. Allen, 1968, 1969, 1984; Harms, 1969; Banks & Collinson, 1975; Reineck & Singh, 1980) it is concluded that equilibrium dimensions of ripples are constant and independent of flow strength. This allows the construction of an empirical unsteady flow model, which calculates the development of bedforms in unsteady flows by dividing flow velocity curves into small velocity increments, $\Delta U_{10}/\Delta t$, and integrating the contributions of these increments to bedform development. For any flow velocity curve and over any period, theoretical predictions and calculations of bedform development can be made with this model, as long as the current is unidirectional. This was illustrated by calculations of bedform development in semi-diurnal tidal flows, which were compared with flume experiments and field data. Theoretically, the model can be extended in a broad range of sedimentary environments.

The empirical unsteady flow model demonstrates that bedform development in a tidal environment depends on the character of the tidal flow velocity curve, which, in turn, is defined by the topographical position (i.e., the sub-environment) and the tidal range. For instance, parallel laminae with superimposed linguoid ripples can be understood as being formed at upper stage plane bed conditions during high rates of settling from suspension, followed by reworking of the deposited sediment into ripples during flow deceleration. Due to the short duration of the flow and the low flow velocities, it is also concluded that the higher an intertidal flat the greater the chance that ripples will not reach equilibrium dimensions within one ebb or flood period. Such ripples will be straight, sinuous or non-equilibrium linguoid at the end of the period. Comparable conclusions can be made for intertidal gullies. Another example are the fully preserved current ripples, often observed in tidal deposits, particularly in flaser, wavy and linsen bedding. They are a logical consequence of the independence of flow velocity, which allows ripples to continue to develop or to maintain equilibrium dimensions during waning tidal flow.

The dimensions of current ripples preserved in the sedimentary record, can now be used to determine whether equilibrium conditions have been reached during the flow. In combination with other data, as for example, grain size fluctuations (e.g. mud drapes), valuable information on the prevailing hydrodynamic conditions can thus be extracted from outcrops and cores.

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REFERENCES

Allen, J.R.L., 1968: Current Ripples: their Relation to Patterns of Water and Sediment Motion. North Holland Publishing Company, Amsterdam, 433 pp.

Allen, J.R.L., 1969: On the geometry of current ripples in relation to stability of fluid flow. Geogr. Annlr., A51, 61-96.

Allen, J.R.L., 1984: Sedimentary Structures: their Character and Physical basis. Developments in Sedimentology, 30A/B, Elsevier, Amsterdam, 593/663 pp.

Amos, C.L., Bowen, A.J., Huntley, D.A. & Lewis, C.F.M., 1988: Ripple generation under combined influences of waves and currents on the Canadian continental shelf. Continental Shelf Research, 8, 1129-1153.

Amos, C.L. & Collins, M.B., 1978: The combined effects of wave motion and tidal currents on the morphology of intertidal ripple marks: The Wash, U.K. J. sedim. Petrol., 48, 849-856.

Baas, J.H., 1993: Dimensional analysis of current ripples in recent and ancient depositional environments. Geologica Ultraiectina, 106, 199 pp.

Baas, J.H., 1994: A flume study on the development and equilibrium morphology of current ripples in very fine sand. Sedimentology.

Baas, J.H., Oost, A.P., Sztano, O., De Boer, P.L. & Postma, G., 1993: Time as an independent variable for current ripples developing towards linguoid equilibrium morphology. Terra Nova 5, 29-35.

Banks, N.L. & Collinson, J.D., 1975: The size and shape of small-scale current ripples: an experimental study using medium sand. Sedimentology, 22, 583-599.

Best, J.L. & Leeder, M.R., 1993: Drag reduction in turbulent muddy seawater flows and some sedimentary consequences. Sedimentology, 40, 1129-1137.

Costello, W.R. & Southard, J.B., 1981: Flume experiments on lower-flow-regime bed forms in coarse sand. J. sedim. Petrol., 51, 849-864.

Dalrymple, R.W., Knight, R.J., Zaitlin, B.A. & Middleton, G.V., 1990: Dynamics and facies model of a macrotidal sand-bar complex, Cobequid Bay-Salmon River estuary (Bay of Fundy). Sedimentology, 37, 577-612.

De Boer, P.L., Oost, A.P. & Visser, M.J., 1989: The diurnal inequality of the tide as a parameter for recognizing tidal influences. J. sedim. Petrol., 59, 912-921.

De Haas, H. & Eisma D., 1992: Suspended sediment transport in the Ems-Dollard estuary. Neth. Jour. of Sea Res., 31: 37-42.

De Raaf, J.F.M. & Boersma, J.R., 1971: Tidal deposits and their sedimentary structures. Geologie en Mijnbouw, 50, 479-504.

Dietrich, G., Kalle, K., Krauss, W. & Siedler, G., 1975: Allgemeine Meereskunde: Eine Einführung in die Ozeanographie; 3rd Edition. Gebrüder Borntråger, Berlin, 593 pp.

Eckman, J.E., Nowell, A.R.M. & Jumass P.A., 1981: Sediment destabilization by animal tubes. Journal of Marine Res., 39, 361-373.

Eisma, D., 1980: Natural Forces. In: Geomorphology of the Wadden Sea Area (Ed. by K.S. Dijkema, H.E. Reineck & W.J. Wolff). Wadden Sea Working Group, Report 1, 20-31.

Eisma, D. & Li, A., 1993: Changes in suspended-matter floc size during the tidal cycle in the Dollard estuary. Neth. Jour. of Sea Res., 31: 107-117.

Folk, R.L., 1968: Petrology of Sedimentary Rocks. Hemphills, Austin, Texas, 170 pp.

Harms, J.C., 1969: Hydraulic significance of some sand ripples. Bull. geol. Soc. Am., 80, 363-396.

Hawley, N., 1981: Flume experiments on the origin of flaser bedding. Sedimentology, 28, 699-712.

Jain, S.C. & Kennedy, J.F., 1971: The growth of sand waves. Proceedings of the International Symposium on Stochastic Hydraulics, Pittsburgh, 449-471.

Jopling, A. & Forbes, D.L., 1979: Flume study of silt transportation and deposition. Geografiska Annaler, 61A, 67-85.

Jumars, P.A. & Nowell, A.R.M., 1984: Effects of benthos on sediment transport: difficulties with functional grouping. Continental Shelf Research, 3, 115-130.

Jumars, P.A., Nowell, A.R.M. & Self, R.F.L., 1981: A simple model of flow-sediment-organism interaction. Mar. Geol., 42, 155-172.

Kachel, N.B. & Sternberg, R.W., 1971: Transport of bedload as ripples during an ebb. Mar. Geol., 10, 229-244.

Khandriche, A., Werner, F. & Erlenkeuser, H., 1986: Auswirkungen der Oststürme vom Winter 19-78/79 auf die Sedimentation im Schlickbereich der Eckernförder Bucht (Westliche Ostsee). Meyniana, 38, 125-152.

Lowe, D.R., 1988: Suspended-load fallout rate as an independent variable in the analysis of current structures. Sedimentology, 35, 765-776.

Mantz, P.A., 1978: Bedforms produced by fine, cohesionless granular and flakey sediment under subcritical water flows. Sedimentology, 25, 83-103.

Miller, M.C., McCave, I.N. & Komar, P.D., 1977: Threshold of sediment motion under unidirectional currents. Sedimentology, 24, 507-527.

Nowell, A.R.M., Jumars, P.A. & Eckman, J.E., 1981: Effects of biological activity on the entrainment of marine sediments. Mar. Geol., 42, 133-153.

Oost, A.P. & De Boer, P.L., 1994: Sedimentology and development of barrier islands, ebb-tidal deltas, inlets and backbarrier areas of the Dutch Wadden Sea. Senckenb. maritima, 24, 65-115.

Postma, H., 1954: Hydrography of the Dutch Wadden Sea. Archs Néerl. Zool., 10, 405-511.

Reineck, H.E., 1961: Sedimentbewegungen an Kleinripplen im Watt. Senckenb. Lethaea, 42, 51-67.

Reineck, H.E., 1978: Das Watt, Ablagerungs- und Lebesraum; 2nd Edition. Waldemar Kramer, Frankfurt am Main, 185 pp.

Reineck, H.-E. & Singh, I.B., 1980: Depositional Sedimentary Environments; Second, Revised and Updated Edition. Springer-Verlag, Berlin, 549 pp.

Reineck, H.E. & Wunderlich, F., 1968: Classification and origin of flaser and lenticular bedding. Sedimentology, 11, 99-104.

Rhoads, D.C. & Young, D.K., 1970: The influence of deposit-feeding organisms on sediment stability and community trophic structure. J. Mar. Res., 28, 150-178.

Sha, L.P., 1990: Sedimentological studies of the ebb-tidal deltas along the West Frisian Islands, the Netherlands. Geologica Ultraiectina, 64, 159 pp.

Simons. D.B., Richardson, E.V. & Nordin Jr., C.F., 1965: Sedimentary structures generated by flow in alluvial channels. In: Primary Sedimentary Structures and their Hydrodynamic Interpretation (Ed. by G.V. Middleton). Spec. Publs. Soc. econ. Paleont. Miner., 12, 34-52.

Southard, J.B. & Harms, J.C., 1972: Sequence of bedform and stratification in silts, based on flume experiments. Bull. Am. Ass. Petrol. Geol., 56, 654-655.

Terwindt, J.H.J., 1967: Mud transport in the Dutch delta area and along the adjacent coastline. Neth. J. Sea Res., 3, 505-531.

Terwindt, J.H.J. & Breusers, H.N.C., 1971: Experiments on the origin of flaser, lenticular and sandy-clay alternating bedding. Sedimentology, 19, 5-98.

Terwindt, J.H.J. & Breusers, H.N.C., 1982: Flume experiments on the origin of flaser bedding. Sedimentology, 29, 903-907.

Tessier, B., 1992: Enregistrement des cycles tidaux en accretion verticale dans un milieu actuel (la Baie du Mont-Saint-Michel) et dans une formation ancienne (la Molasse Marine Miocene du Bassin de Digne): mesure du temps et application à la reconstitution des paléoenvironnements. U.F.R. des Scienc. de la Terre et de l'Amenagement Regional, Univ. de Caen, Thèse, 122 pp.

Van Straaten, L.M.J.U., 1954: Composition and structure of Recent marine sediments in the Netherlands. Leidse Geol. Mededel., 19, 1-110.

Van Straaten, L.M.J.U., 1964: De bodem der Waddenzee. In: Het Waddenboek, (Ed. by J. Abrahamse, J.D. Buwalda & L.M.J.U. Van Straaten), pp. 75-151. Nederlandse Geologische Vereniging, 223 pp.

Visser, M.J., 1980: Neap-spring cycles reflected in Holocene subtidal large-scale bedform deposits: A preliminary note. Geology, 8, 543-546.

Walker, R.G., 1984: Shelf and shallow marine sands. In: Facies Models, Second Edition (Ed. by R.G. Walker), pp. 141-170. Geoscience Canada, Reprint Series 1, 317 pp.

Winkelmolen, A.M., 1976: An inexpensive infinite flume. Geologie en Mijnbouw, 55, 51-60.

Wunderlich, F., 1967: Feinblätterige Wechselschichtung und Gezeitenschichtung, Senckenberg. Lethaea, 48, 337-343.

Yalin, M.S., 1985: On the determination of ripple geometry. Journal of Hydraulic Engineering, 111, 1148-1155.

APPENDIX 1: LIST OF SYMBOLS

Δt_i = time period of *i*-th increment of change in flow strength (h)
d_o = orbital diameter of wave oscillations (m)
D_{50} = median grain size (m)
h = flow depth (m)
h_o = initial flow depth (m)
H_e = equilibrium bedform height (mm)
H_{max} = maximum bedform height (mm)
H_t = bedform height at time t (mm)
L_o = initial bedform wavelength (mm)
L_e = equilibrium bedform wavelength (mm)
L_{max} = maximum bedform wavelength (mm)
L_t = bedform wavelength at time t (mm)
S = cumulative development stage of bedforms
t = time (h)
T = water temperature (°C)
T_e = equilibrium time, general (h)
$T_{e,H}$ = equilibrium time for height (h)
$T_{e,i}$ = equilibrium time at the *i*-th time increment (h)
$T_{e,L}$ = equilibrium time for wavelength (h)
U_{10} = 10°C-equivalent depth-averaged flow velocity ($m.s^{-1}$)
$U_{10,i}$ = 10°C-equivalent flow velocity at *i*-th time increment ($m.s^{-1}$)
U_{100} = instantaneous near-bed flow velocity ($m.s^{-1}$)
U_{mb} = mean deviation of the near-bed oscillating flow velocity ($m.s^{-1}$)
W_s = grain settling velocity in standing water ($m.s^{-1}$)
υ = kinematic viscosity of water ($m^2.s^{-1}$)

CHAPTER 9

EPILOGUE

INTRODUCTION

In the previous chapters the development of the Wadden Sea and the sedimentary processes active have been studied on several temporal and spatial scales: from the whole eastern Dutch Wadden Sea down to current ripples and the production of pellets. Complex behaviour and interaction of processes is observed on many different scales. In this chapter it is discussed at which scale of time and space one should look to make valid interpretations and predictions for such a complex system and its sedimentary successions. This problem is discussed with the focus on the barrier-related deposits discussed in this thesis.

BARRIER SYSTEMS AS HIERARCHICAL SYSTEMS

Historical perspective

Sedimentology was largely restricted to sedimentary petrology before 1950. Modern, comparative sedimentology started with the publication of Kuenen & Migliorini (1950). In their paper on turbidity currents as a cause of graded bedding, the results of flume experiments were related to fossil turbidite deposits. Under the impact of this concept geologists started to look at sedimentary rocks as sediments with modern analogues, which to some extent could be simulated by experiments (Reading, 1981). Since then an increasing effort has been made to understand the processes of sediment deposition, both by experiments and by observations of present-day depositional environments (Reading, 1981).

Initially this served merely to interpret the conditions under which specific sedimentary rocks had been formed in the geological past. A wide range of facies models was developed. With the elaboration of these models, sedimentologists became increasingly aware of the relation between sedimentological features of fossil deposits and the parameters which had controlled their formation. The facies-model approach has been so successful, because sedimentary successions can be grouped into a limited number of facies models. This implies that, although variability is large, the variation in sedimentary sequences largely concentrates around the facies models.

With the re-emergence of stratigraphy as a primary tool for the analysis of sedimentary successions, more variables were introduced for facies interpretation, especially sea-level changes (Walker & James, 1992). At the same time larger data sets became available from sedimentary successions, because of a better resolution of geological observations due to better tools, and because the possibilities to store and process data increased dramatically due to faster and more powerful computer systems.

Table I: Order of magnitude of time spans, during which sedimentary features are active.

Morphological unit	Order of magnitude of active existence time
Barrier system	1,000 - 10,000 yr
Whole backbarrier area	1,000 - 10,000 yr
Barrier chain	1,000 - 10,000 yr
Barrier island	1,000 yr
Tidal system	1,000 yr
Estuary	100 - 10,000 yr
Aeolian dune-complex	100 - 1,000 yr
Shore	100 - 1,000 yr
Giant subtidal dune	100 - ? yr
Backbarrier drainage area	100 - 1,000 yr
Backbarrier channel system	10 - 1,000 yr
Ebb-tidal delta	10 - 1,000 yr
Tidal flat system	10 - 1,000 yr
Inlet channel	10 - 1,000 yr
(Inter)tidal flat	10 - 1,000 yr
Tidal marsh	10 - 1,000 yr
Creek system	10 - 1,000 yr
Marsh shoal system	10 - 1,000 yr
Barrier shoal	10 - 1,000 yr
Ebb-delta shoal	10 - 1,000 yr
Backbarrier channel	10 - 100 yr
Backbarrier tidal shoal	10 - 100 yr
Channel bend	10 - 100 yr
Creek	10 - 100 yr
Aeolian dune	10 - 100 yr
Saw tooth bar	10 - 100 yr
Swash bar	10 yr
Creek bend	1 - 100 yr
Outer channel	1 - 100 yr
Breaker bar	1 - 10 yr
Marginal flood channel	1 - 10 yr
Gully	1 - 10 yr
Mussel bed	1 yr
Ebb- & flood chute	1/2 yr - 10 yr
Subtidal megaripple	neap-spring cycle - 1 yr
Intertidal megaripple	neap-spring cycle - 1 yr
Internal bioturbation	h - 1 yr
Marsh shoal layer	h - neap-spring cycle
Hummock	h
Current ripple	minute - neap-spring cycle
Wave ripple	minute - ebb-flood cycle
Aeolian ripple	minutes
Surface bioturbation	seconds - ebb-flood cycle
Faeces & pseudo-faeces*	seconds - ebb-flood cycle

* Merely existence time outside musselbeds, not active, but passive.

In order to make better predictions of the genesis and lateral variability of sedimentary deposits, there is a growing need to understand the dynamics and the development of sedimentary systems in more detail. At which scale of time and space one should do research to make valid interpretations and predictions, is presently one of the major problems in sedimentology. In addition, it is one of the major problems for the prediction of future developments of active sedimentary systems.

Hierarchy

Schumm and Lichty (1965) tackled this question to some extent for alluvial systems. Their idea is that there consists a hierarchy of cause-effect relations, determining the development of landforms. Also, they stated that the distinction between cause and effect is a function of time and space, because factors that determine the character of landforms can change from dependent to independent variables as the limits of time and space change. For instance, over a long (geological) span of time changes in drainage network morphology depend on (in decreasing hierarchical order): initial relief, geology, climate, vegetation, the relief or the volume of the system above base level, and the runoff and sediment yield within the system. The drainage network morphology in its turn influences hill-slope morphology. However, over a very short time span the hill-slope morphology is constant and independent of the drainage network morphology, the runoff and sediment yield, the relief, etcetera (Schumm & Lichty, 1965).

From the previous chapters it appears that also the elements constituting a barrier system can be thought of as being hierarchically structured (cf. Allen & Starr, 1985; cf. O'Neill et al., 1986). For the sake of the argument the discussion is here restricted to the morphological elements of the barrier system, but it might be expanded to include also other elements (such as biota). Largely independent external variables outside the barrier system, such as geology, initial relief, tides and climate, control its development. To determine the hierarchical level, an estimate was made of the order of magnitude of the time spans, during which some of the most important morphological elements are active. The estimate was mainly based on the information from the Dutch coast (Table I). From Table I it appears that the spatial extent of a morphological element generally increases with the increase of time during which they are active (cf. Schumm & Lichty, 1965; cf. Allen & Starr, 1985; cf. O'Neill et al., 1986).

The various morphological elements form a nested hierarchical structure, meaning that a high order (level) morphological element encompasses lower order morphological elements, and on its turn forms part of an even higher order one. A straightforward example is that there can be no channel bends, without a channel, which, in its turn, forms part of a larger channel system. The lower order morphological elements normally show a higher frequency of characteristic behaviour (development) than do the higher order morphological elements, which 'filter' the high frequency 'signal' of the lower order morphological elements. An example is that only a restricted part of the ripples, which are generated on a shoal during one tide, adds to the accretion of the shoal.

In general, the higher order morphological elements constrain, control and bind the lower ones (cf. Wijnberg & Terwindt, in press). Each morphological element is influenced (mostly by waves, currents or supply of sediment) by one or more morphological elements of a higher order, and (influenced) external variables. For instance, the hydrodynamic/-morphologic development of a channel meander will to a large extent control the dynamics of a megaripple in the meander (chapter 1). The lower order morphological elements are, so to say, to a large extent the product of the higher order ones, and on their turn they control the development of still lower order morphological elements. Thus, it depends on the level at which one observes, whether a morphological element should be seen as a cause or as a result.

Furthermore, elements of equal rank influence each other especially if they form part of the same higher order morphological element. This is illustrated by the development of the outer channels of the Zoutkamperlaag (chapter 4): The marginal flood channel W of the barrier island Schiermonnikoog can exist only when the other outer channels are concentrated along the westernmost side of the ebb-tidal delta, and it is abandoned when other outer channels get positioned in the middle of the ebb-tidal delta. Another example is the enhancement of current ripple formation by an increase of bottom roughness due to bioturbation (chapter 8).

Reversely, there is a feed-back (see also below: self organisation) of the lower order morphological elements to the higher order ones. In general such direct feed-back is small, because it forms only a small part of the larger element and exists only for a limited period of time. Even the spatial and temporal integrated feed-back to the higher order element can only be a part of the influence exerted by the lower ones, because such processes are not 100% efficient, and there are other influences as well.

Self organisation

In the current approach it is assumed that on each level of the tidal hierarchical system, the morphological elements interact with the, mostly hydrodynamic, variables which form it, in a non-linear way. Such a feed-back may either be self-enhancing or self-arresting (Bull, 1991; Dronkers, 1994). As a result, the hierarchical system is for an important part self-organizing on each level. In the Wadden Sea this often leads to dynamic equilibria, for instance the dynamic equilibrium maintained by small-scale current ripple-fields once they have formed from initial flat bed conditions (chapter 7). Other illustrations are the cross-sectional area of a channel (chapter 3) and the asymptotical growth of mussel beds to the mean sea-level height (chapter 6). In each case the process of development is driven by and constrained by the influence of the higher order morphological elements and the external variables which again may be influenced by them.

Actually, a morphological element can only develop a characteristic dynamical morphology, because it is self-organizing (within the constraints of the external variables and the higher-order elements). This implies that, although variability is large, the variation of the morphology largely concentrates around attractors, as for instance, the equilibrium ripple form[1]. This is also one of the major keys to the success of facies models in sedimentology. Within one sedimentary sub-environment the limited variability in morphology and morphological dynamics only allows a limited variability of sedimentary deposits to be formed.

It also implies that on any level of the hierarchy, a morphology most likely generates some feed-back, which is part of its reason of existence. This is mostly a hydrodynamic counter force. For a complete barrier system it may be expected that not only tidal range, the amount of sea-level rise, and sediment supply are of importance for its formation and maintenance, but also the possibility to generate some counter force to overall landward transport of sediments. This possibility is created by the initial relief. It primarily determines the tidal prism and hence the strength of the outflowing ebb-current. Thus, under the conditions of decreasing rates of sea-level rise in the Holocene and the relatively constant initial relief gradient of the Dutch coast, the tidal prism became increasingly important and helped establishing barrier systems everywhere along the coast (chapter 1).

It may be expected that self-organisation at each level also implies that although lower order morphological elements exert influence on a process, they will normally not lead to a complete disruption at much higher levels. Even in truly chaotic systems the influence of lower order elements will lead to unpredictability of the deterministic behaviour of a higher order morphological element, but not of its overall development (cf. Dronkers, 1994). This is comparable with the weather: although we cannot predict the weather for a certain day in next winter, we can predict that on average it will be colder than in next summer. Likewise, we cannot predict the exact development of the Pinkegat system, but we can expect that it will follow the same cyclic development pattern as it did over the past 200 years.

Perturbations

On each level (sudden) perturbations may occur, disturbing the system. If such perturbations are small, they may disturb the development of a morphological element temporarily, after which it may return to its normal development. If they are strong they may disrupt the morphological element so strongly that it cannot return to its normal development; it has passed a threshold (see below). Although the effects of perturbations may totally disturb the development of the morphological elements on a certain level, these are incorporated in the

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[1] An attractor may also involve the preferent position of the morphological element. For example, drainage channels tend to remain positioned in the middle of their drainage area (chapter 3). For the Zoutkamperlaag this may imply that after some time the main backbarrier channel of the Zoutkamperlaag along the southernmost side of the drainage basin will be left in favour of a channel in the middle of this basin.

development on higher levels, as part of the system. An example is the destruction of ripples by shrimps and crabs, which, on a higher level, forms a normal part of the development of channels (chapter 8). Even storms, which can influence the coast for decennia or even can destroy (parts of) islands and shoals, are in the long run (and on a large scale) a part of the natural development of a barrier complex (Stolk, 1989; chapter 2).

Even the effects of relatively large-scale artificial perturbations in the Wadden system seem to be incorporated on a very high level (parts of the barrier system). During the formation and dyking/closure of the Zuider Zee, Middelzee and Lauwerszee embayments large parts of the inlet system changed, but the system as a whole managed to sustain (chapters 2 and 5). Natural developments may produce comparable perturbations, e.g., upon the flooding of a peat-filled palaeo-valley (chapter 2) or the rapid infill or closure of part of a drainage basin by a spit.

Thresholds and stability

Due to feed-backs the morphological elements can generate some inertia to changes, and thus stabilize themselves. For instance, the bottom roughness of ripples decelerates the flow and thus slightly extends the energy domain under which ripples are stable. Morphological elements may change, or cease to exist, when a threshold is passed. The threshold may be of a hydrodynamical, biological, morphological, or other character. An example of hydrodynamical thresholds is the appearance of current ripples at the threshold velocity of sediment movement (chapter 7). An example of a biological threshold is the erosion of dunes, once the vegetation is removed (chapter 2). An example of a morphological threshold is the repeated break-through of the spit E of Ameland by newly formed inlets of the Pinkegat (chapter 3). On a slightly higher order level is the (expected) passage of a morpho-

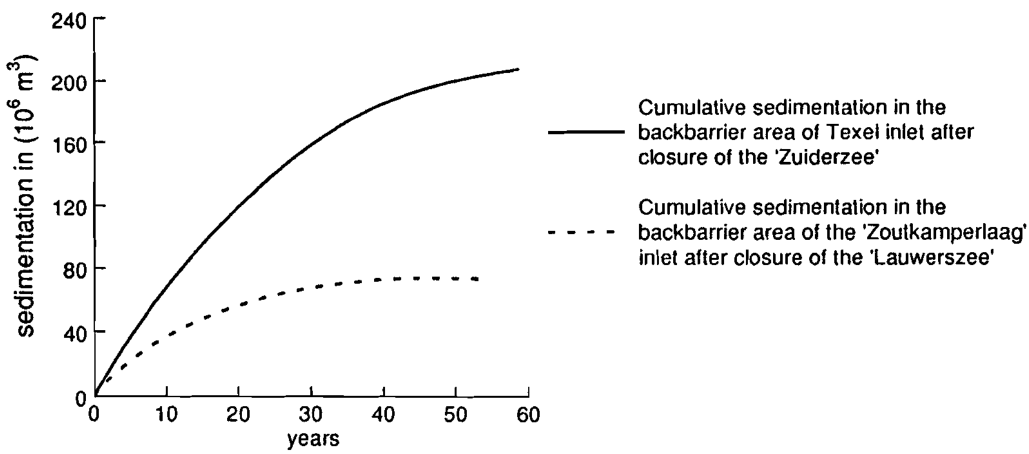

Figure 1: Asymptotical sedimentation after the closure of parts of tidal basins (Vroom et al., 1989).

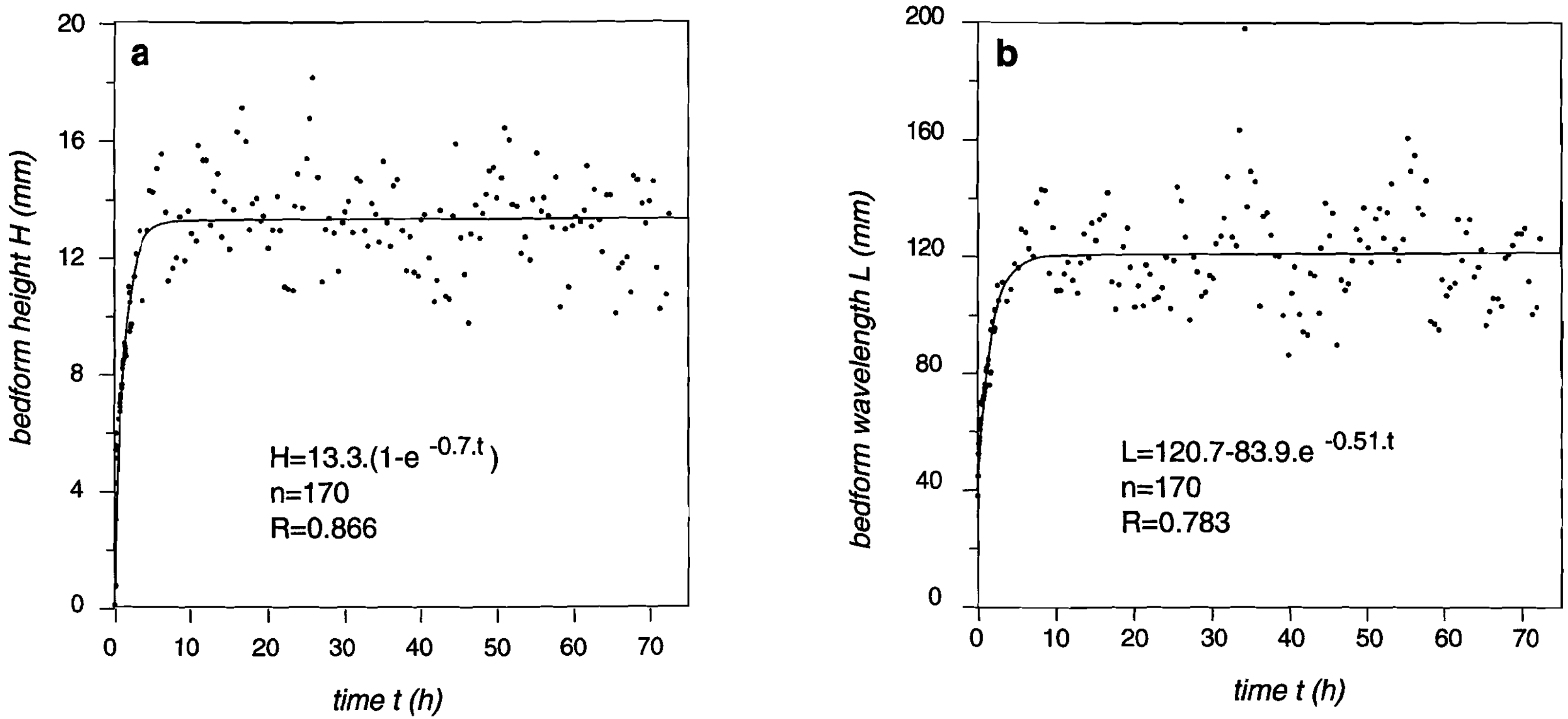

Figure 2: Development towards equilibrium bedform wave height and length from plane bed, note the small variability during the first stage of development (chapter 7).

logical threshold by the disappearance of the Engelsmanplaat, which is due to the cyclic development with superimposed eastward migration of the Pinkegat inlet. Afterwards, it is expected that the Pinkegat system and the Zoutkamperlaag system will merge (chapter 3), and develop a new equilibrium. The latter example also shows that the threshold cannot only be passed by the influence of external variables and higher order morphological elements. It can also be passed by the influence exerted by lower order element, although it may be argued that the gradual eastward shift of the Pinkegat system is ultimately generated by external variables, i.e., tides, waves, and sediment supply. The passage of a threshold can also be partly or even exclusively due to an 'external' perturbation, such as a storm generating a new outer channel or closing another one (chapter 4). Bull (1991) points out that for alluvial systems some time may pass between the perturbation and the actual passage of the threshold (reaction time). The delayed shift of the morphological watershed after the closure of the Lauwerszee embayment is an illustration of such a reaction time in the Wadden Sea system (chapter 5).

After the passage of a threshold within the barrier system of the Wadden Sea, new morphological elements are normally quickly generated relative to the existence time of the variable. Again, like with the perturbations, these changes are also incorporated in the higher level variables. This is, for example well illustrated by the relatively rapid: a) decay of the island Bosch during the decay of the Lauwers Inlet (chapter 2), b) formation of the new inlets in the Pinkegat inlet (chapter 3), and c) formation of marginal flood channels W of Schiermonnikoog (chapter 4). Observations in nature (chapters 2, 3, 4 & 5), experiments (chapters 7 & 8), and model calculations (chapter 8) indicate that the switch towards the new dynamics is often asymptotical in character (Fig. 1), the first part being very quick, with often a small variability (Fig. 2).

Summarizing: The above discussion shows that barrier systems can be approached as hierarchical systems. In such systems the nature of the problem determines on which level observations have to be done and wether the observed phenomena should be considered mainly as causes or results (cf. Allen & Starr, 1985; O'Neill et al., 1986). Such an approach may lead to unexpected insights, especially where the scales of the hierarchical level were beyond the human scale. Also, it helps to avoid scale-confusion and to understand stability and instability of systems or parts of them.

PRESERVATION POTENTIAL

From the above hierarchical approach two conclusions can be drawn about the preservation potential:

1) During 'normal' development morphological elements on each level often tend to evolve around attractors. As stated, the morphological element are each controlled and constrained

by the higher order morphological element: a ripple develops upon a megaripple, in a channel, in a channel system. From this it follows that also the preservation potential of a morphological element is for a large part controlled by the dynamics of the higher order morphological element(s) of which it forms part. In other words: the larger morphological element provides the 'accommodation space' for the smaller ones. Since the larger morphological elements are generally characterized by slower dynamics, the time over which a small sedimentary structure is preserved will, on average, increase with incorporation in increasingly larger sedimentary structures of increasingly higher order morphological elements. For example, if a ripple structure becomes part of a subtidal megaripple it can be preserved during several neap-spring cycles. Only if this megaripple itself becomes incorporated in a larger structure, such as a lateral accretion in a channel bend, it is preserved over a longer period. Thus, in order to be preserved in the fossil record the structure has to be incorporated in the sedimentary structures formed by the highest order scale. For transgressive barrier systems such as the Wadden Sea these are the deposits left beneath the ravinement surface (cf. Sha, 1990). Therefore, preservation of sediments will be favoured especially in inlets, because of their large depths (Fig. E1[2]).

2) Perturbations and/or the passage of thresholds can put a sudden end to a normal development and thus preserve ('freeze') the sedimentary structures of a morphological element. As stated above, the effects of the passage of thresholds and perturbations will normally be incorporated in the higher order morphological elements. An example are the abandoned channel deposits in ebb-tidal deltas, after the sudden abandonment of channels. Thus, the higher the scale on which the perturbation occurs, the higher also the scale of the lower order morphological elements of which the sedimentary structures are preserved. When the system is perturbated or a threshold is passed at a high level, large series of lower level sedimentary structures can be preserved (Fig. E1).

The perturbations or the passage of a threshold as such may not leave any traces in the sedimentary record, except the preservation of fairly large sedimentary sequences, perhaps capped by deposits showing a sudden change in sedimentary conditions. Thus, probably many of the fossil transgressive barrier sequences, which are often fairly complete, are most likely preserved, because of the passage of a threshold or a large-scale perturbation, which left no further traces besides the preservation of the sequence.

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[2] For figure E1 see appendix A.

REFERENCES

Allen, T.F.H., T.B. Starr, 1982: Hierarchy: perspectives for ecological complexity, The University of Chicago Press, Chicago, 310 pp.

Bull, W.B., 1991: Geomorphic responses to climate change. Oxford Univ. Press, 326 pp.

Dronkers, J., 1995: Holland of Bolland, de toekomst van de nederlandse delta. Rede uitgesproken naar aanleiding van de aanvaarding van het ambt van bijzonder hoogleraar in de Fysica van Kustsystemen aan de Universiteit Utrecht op 18 januari 1995, 28 pp.

Kuenen, Ph.H. & Migliorini, C.I., 1950: Turbidity currents as a cause of graded bedding. J.Geol., 58, 91-127.

O'Neill, R.V., D.L. DeAngelis, J.B. Waide & T.F.H. Allen, 1986: A hierarchical concept of ecosystems, Monographs in Population Biology 23, 253 pp.

Reading, H.G., 1981: Sedimentary Environments and Facies, Blackwell Sc. Publ., 4th pr., 569 pp.

Schumm, S.A. & Lichty, R.W., 1965: Time, space, and causuality in geomorphology. Am. J. of Science, 263, 110-119.

Sha, L.P., 1990: Sedimentological studies of the ebb-tidal deltas along the West Frisian Islands, the Netherlands. Phd Thesis, University of Utrecht, Geol. Ultraiectina, 64, Utrecht. 160 pp.

Stolk, A. 1989: Zandsysteem kust - een morfologische karakterisering. Technisch Rapport 1, Kustverdediging na 1990. Dept. of Physical Geography, report GEOPRO 1989-02, 97 pp.

Vroom, M.G., Steyaert, F.H.I.M., Misdorp, R., Venema, H., Elgershuizen, J.H.B.W., Rakhorst, H.D. & Damoiseaux, M.A., 1989: Wadatlas, Min. Verkeer en Waterstaat, Stadsdrukkerij, Amsterdam, 95 pp.

Walker, R.G. & James, N.P., 1992: Facies models: response to sea level change. Geol. Ass. Can., Love Printing Serv. Ltd, Stittsville, Ontario, 409 pp.

Wijnberg, K.M. & Terwindt, J.H.J., in press: Extracting decadal morphological behaviour from high-resolution, long-term bathymetric surveys along the Holland coast using eigenfunction analysis.

Curriculum Vitae

Albert Peter Oost was born in Emmen, The Netherlands on 26th August 1961. In 1979 he received his secondary school diploma (Atheneum B) in Ter Apel, and started his academic education at the University of Groningen where he obtained his 'Kandidaats' diplomas in Geology in August 1983. His 'Doctoraal'(MSc) diploma (Cum Laude) in Sedimentology with Marine Geology and Economic Geology as secondary subsidiary subjects was obtained at Utrecht University in August 1988. Between March 1989 and March 1995 he was employed as PhD student and researcher at the Institute of Earth Sciences of Utrecht University and studied the sedimentary development of the Dutch Wadden Sea. In April 1995 he continues as a post-doc at the KVI 'Kernfysisch Versneller Instituut' at Groningen University.