

HOLOCENE EVOLUTION OF THE IJSSELDELTA

André Torres
MSc Thesis

Holocene Evolution of the IJsseldelta

Research for the fulfilment of the MSc

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1. INTRODUCTION

1.1 Background

The study of delta evolution and understanding of its dynamics are crucial for environmental planning, water control, coastal management and a broad spectre of other environmental studies. In the Netherlands, the scientific results of delta studies assume a particular importance due to the pressure on planning exerted by a high population density in a country where part of the delta plains stand below sea-level. Knowledge of subsurface geology is rather important in fluvial environments because it controls the transport of underground water through subfossil sandy channel belts with great relevance for dyke construction and management. The depth of the subsurface sand bodies also strongly affects the construction of infra-structures because in an area dominated by clays and peats, which occur widely in the Netherlands, foundations on poles must reach Pleistocene sands to assure stability of the buildings, and of course the costs will decrease if these are located in shallow sand bodies. Another very important aspect is the study and reconstruction over time of avulsions, the full or partial abandonment of a river course for a new one which can hypothetically be caused by differential sedimentation and posterior superelevation of the channel belt over the floodplain, if this process occurs in a short period of time it can have major consequences to fluvial, deltaic and even coastal dynamics because it controls the redistribution of sediments and discharge, thus influencing channel recurrence interval, density and interconnectivity (Berendsen & Stouthamer, 2001). Studies on delta morphology and dynamics can also be used for flood control and management as well as for the prospection of natural resources present in sandy fluvial deposits, like oil and gas.

Research on delta evolution has long been part of a long-term program of the Department of Physical Geography of Utrecht University. The main goals of this program are centred in the reconstruction of the evolution of the fluvial dominated deltas in the Netherlands since the Tertiary to the present together with a further understanding of these systems dynamics. The major reference is the extensive research developed in the Rhine-Meuse delta where an enormous dataset was obtained along the years and the evolution of this delta system was reconstructed in great detail, giving an important knowledge set that was summarized in the work of

Berendsen & Stouthamer (2001). Drilling has been performed in the IJssel river and delta areas during the BSc student fieldworks of 2005, 2006, 2007 and 2008. The borehole data is available in archive but also has been digitalized since 1989 when the LLG and BP-Plotter computer programs were developed and introduced so that the data can be fast and easily inserted in digital form, accessed and plotted by researchers.

1.2 Research objectives

The aim of this research is to reconstruct the palaeographic development of the IJsseldelta (Fig. 1) and to determine which tidal and fluvial processes acted at a given moment. Several questions arrive from this broad spectre relating the evolution of this area to that of the Rhine-Meuse delta, including the study of sea-level, groundwater levels and sediment load deposition over time.

The following research questions are defined:

- Which marine and fluvial processes occurred and how can they be reconstructed and/or be recognized in the deposits?
- How did the delta of the River IJssel develop during the Holocene?
- Are there indicators for variations in the River IJssel discharge?

For the achievement of the research objectives an exhaustive study and analysis was carried out of: (a) the subsurface geology of the IJsseldelta, with a particular importance to the sedimentary characteristics and age of the deposits in the distal and proximal areas of the delta; (b) the reconstruction of the alluvial architecture of the delta and changes over time, determining the marine influence by the presence of marine clays and groundwater levels over time through the study of the depth of decalcified deposits and radiocarbon dating of peat samples; (c) the avulsion history, locations and parameters.

The objective of this report is to describe the reconstruction of the IJsseldelta during the Holocene. In the second chapter of this report a general description of river delta morphology and dynamics, as well as an overview to the different morphological units present in the IJsseldelta is given. Chapter 3 describes the methods used to collect and analyze the data needed to perform this research. A ^{14}C dating of key locations in the IJsseldelta was not possible due to the limited time scope of this

project but reference is made to the future possibility of developing this kind of analysis in the area. In Chapter 4 the data set was analyzed and four transversal cross-sections were described in detail. Chapter 5 shows an overview of the development of the study area in the Late Pleistocene and Holocene, focusing upon the building and evolution of the IJsseldelta from the 12th Century to the present.

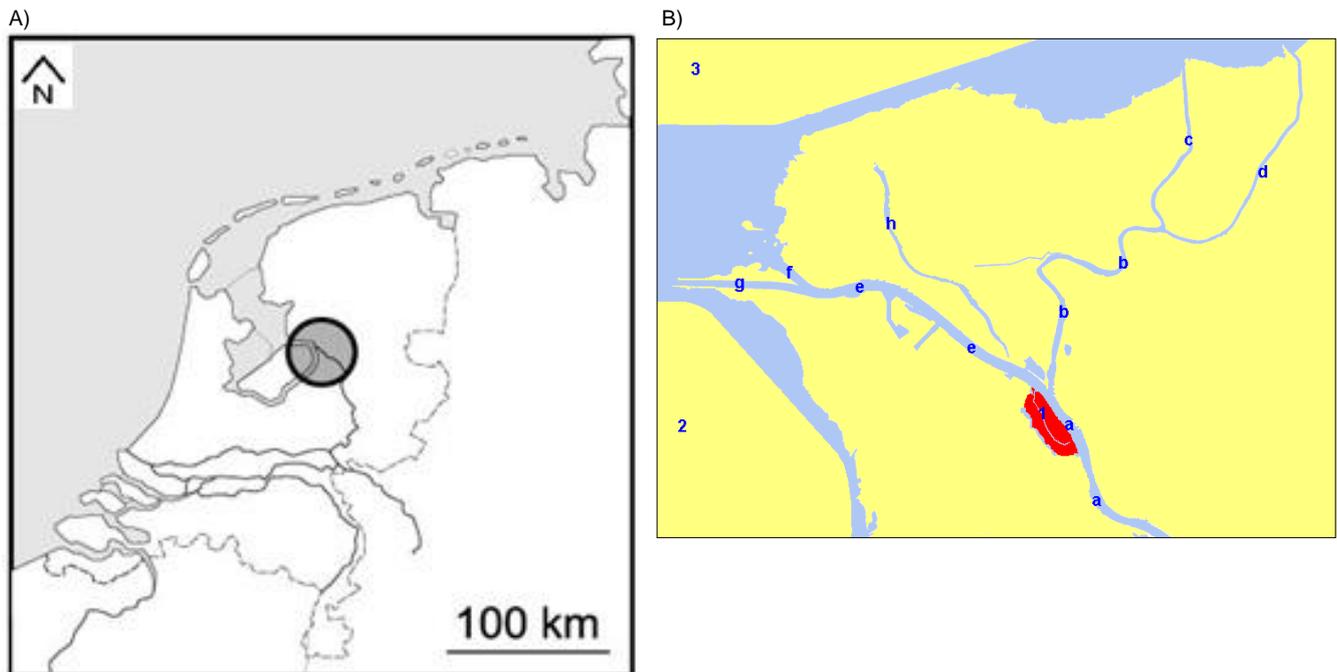


Figure 1 - A) The IJsseldelta in the Netherlands; and B) Branches of the IJsseldelta in Kampereiland:

- a. Primary channel: River IJssel
 - b. Secondary channel: IJssel (19th century); Ganzendiep (modern)
 - c. Secondary channel: Ganzendiep
 - d. Secondary channel: Goot (modern)
 - e. Primary channel: Regtediep (19th century); River IJssel (modern)
 - f. Secondary channel: Regtediep (19th century); Kattendiep (modern)
 - g. Primary channel: Keteldiep
 - h. Secondary channel: Noorderdiep (19th century); Noorddiep (modern)
- 1, Historical city center of Kampen
 - 2. Oost-Flevoland , polder Flevoland province
 - 3. Nooroostpolder, Flevolandz

The IJssel is a major tributary river from the Rhine, with a total length of about 125 kilometres, the IJssel controls 1/9 of its total discharge, forming a delta system in the last 6 kilometres of the river run that extends from around the location of Kampen to the IJsselmeer (Fig. 1). The river IJssel branches off the Nether Rhine just south of

the city of Arnhem, in its course passes along the towns and cities of Rheden, Doesburg, Dieren, Zutphen, Deventer, Zwolle (historically not on the IJssel but the Zwarte Water river), and Kampen, and discharges through a small delta into the IJsselmeer (Lake IJssel, former Zuiderzee).

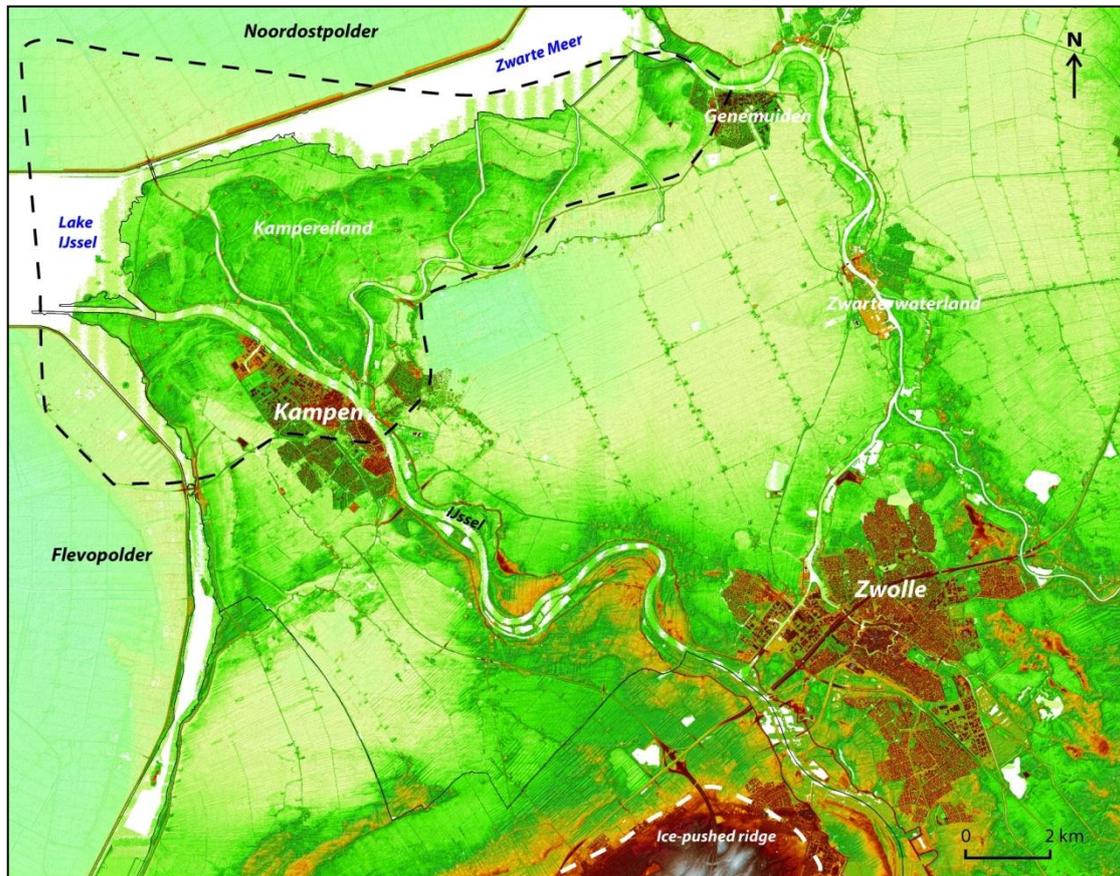


Figure 2 - Digital elevation model of the IJsseldelta (Rijkswaterstaat-AGI, 2005). Colours in order of increasing height: white-light green-green-yellow-red. The black dashed-line represents the approximate distribution of IJsseldelta sediments.

Although the IJssel is a relatively recent river dating to Roman times, it drains through a much older valley between the Veluwe and Salland. The IJsseldelta is situated in the Dutch province of Overijssel, immediately north of Kampen, and its main overhead area is known by the name of Kampereiland (Fig. 2). It is a relatively minor delta in the Netherlands and represents the distributary mouth of the river IJssel, with the area of Kampereiland retaining the sedimentary evidences of several former branches of the delta. The distributaries of the IJssel in the delta area were the “Zuiderdiep”, “Rechterdiep”, “Noorderdiep”, “Garste”, “Ganzendeiep” and the “Goot”.

2. RIVER DELTA MORPHOLOGY

In situations when the sediment supply of a contemporary river to the coastline is faster and more effective than its redistribution by basal energy, a discrete lobe shaped shoreline protuberance that narrows in the direction of the feeding river, may develop forming what is called a *delta* (Reading, 1996).

In river deltas the main sediment supply consists of mud, silt and sand, gradients are low to moderate and the depositional patterns change accordingly to long-term changes in input characteristics and base level. The variability of the discharge and sediment supply from a river, together with the grain size and accommodation space that affect the gradient of the delta plain and channel morphology, are major controls for delta development and architecture (Fig. 3).

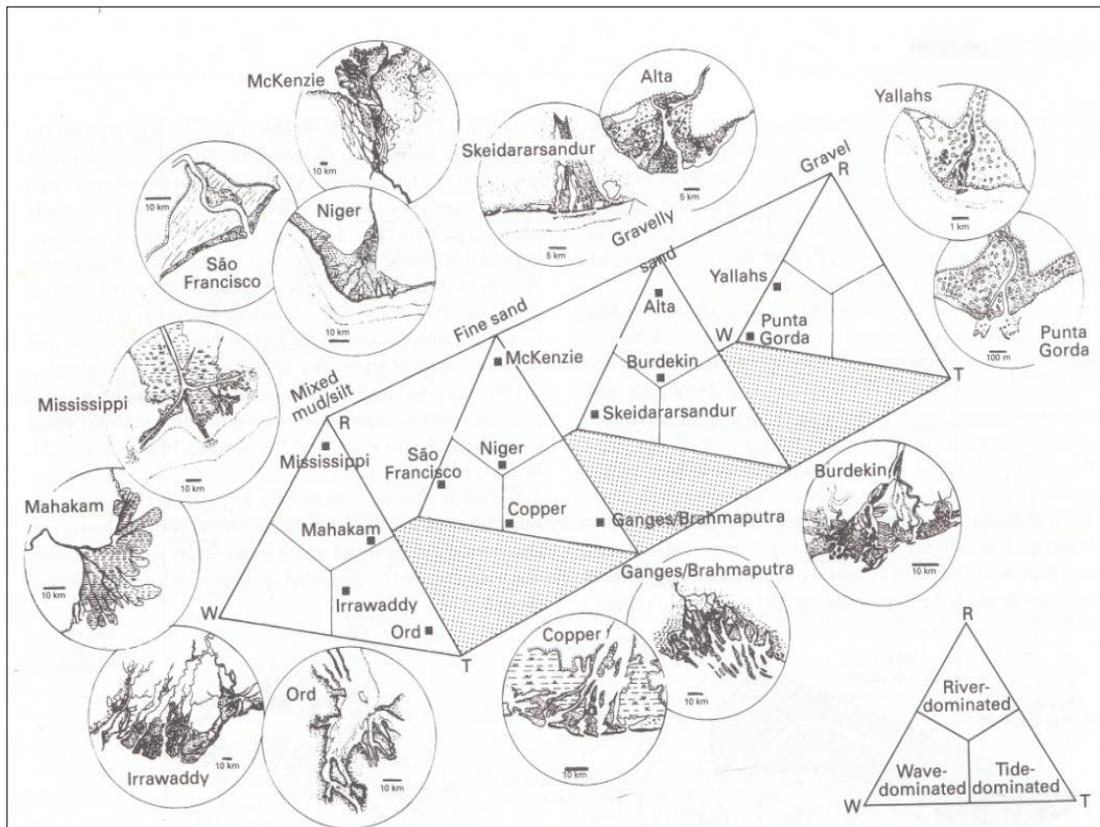


Figure 3 Classification of modern deltas based on process (fluvial, wave or tide) of sediment dispersal at the delta front (after Galloway, 1975) and on prevailing grain size. Extracted from Reading (1996).

2.1 Deltaic architecture

Gilbert (1885) made the first description of the deltaic depositional architecture facies as a threefold structure that included a subaerial topset, subaqueous foreset and bottomset beds (Fig. 4).

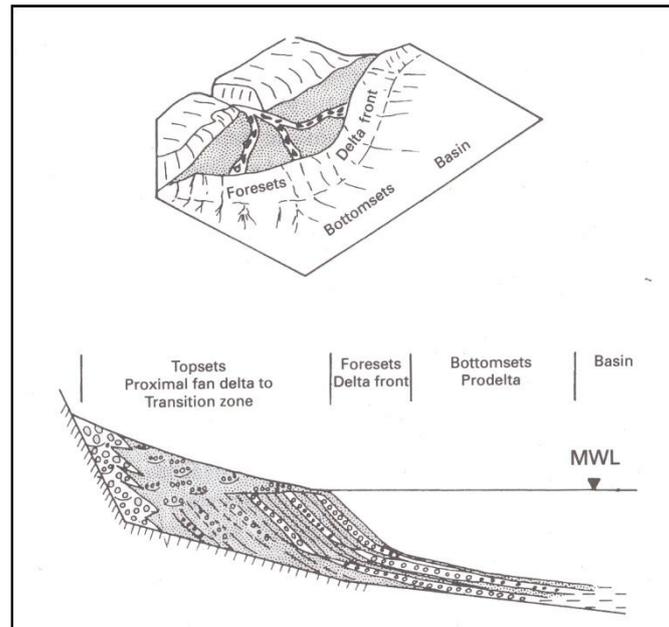


Figure 4 Gilbert-type fan delta; plan view and cross-section (Reading, 1996).

The Gilbert classification applies to deltas with high delta front slopes, thus developing in basin margins with high gradients and coarse-grained systems. The topset beds are deposited by shifting subaerial channels and include the proximal delta fan and a transition zone. Foreset beds develop in the delta front zone, where the sediments deposited in the river mouth are mobilized further into the distal zone direction through grain flows and/or debris flows. Homopycnal conditions (equal density between river and basin waters), that generate intense mixing in the river mouth and produce considerable sedimentation of bedload, favour the formation of foreset beds. In the distal, low-gradient, prodelta zone the bottomset beds are deposited from suspended load and gravity flows (Reading, 1996).

This idea was later extended by Barrell (1912), who divided the delta structure in topset, foreset and bottomset based on characteristics like bedding, texture, colour or fauna of the elements.

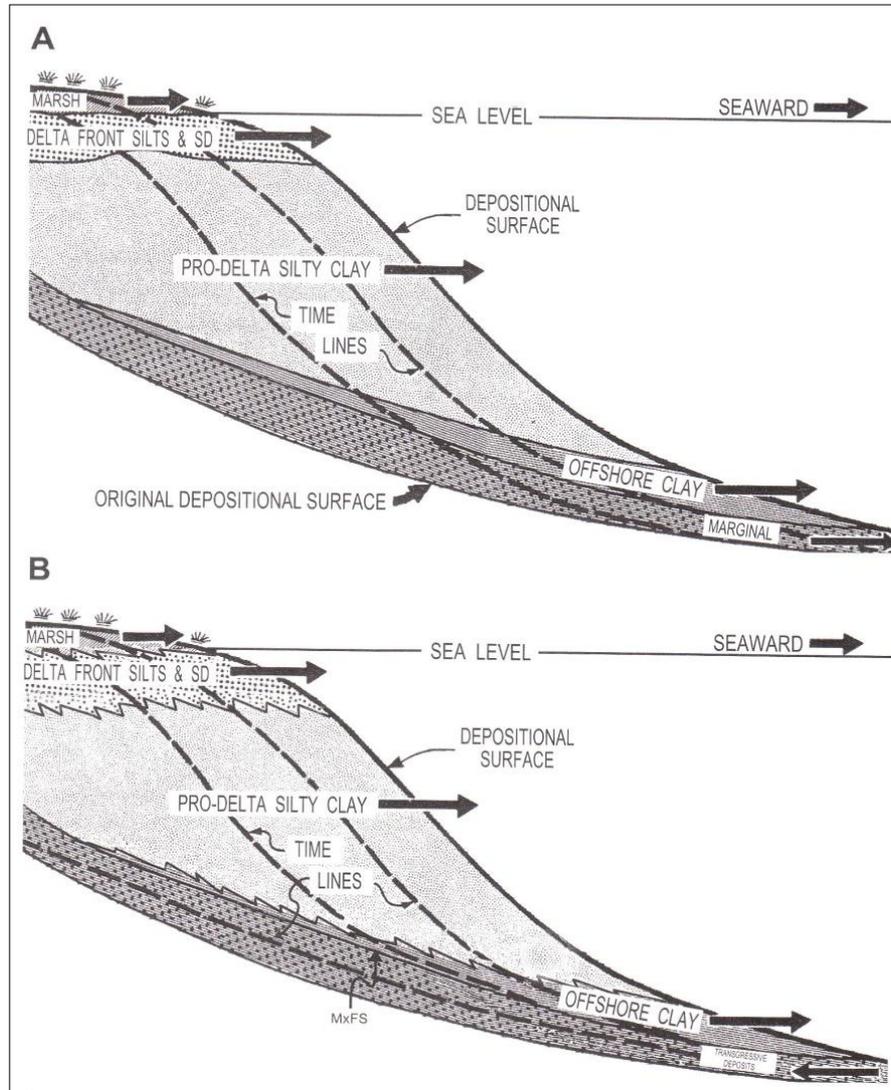


Figure 5 - A) Seaward progradation of delta as illustrated by Scruton (1960) and B) modified version of Scruton diagram (Gani and Bhattacharya, 2005).

Scruton (1960) constructed a dip-oriented cross-section of the delta structure explaining the processes of delta progradation and development of different facies by time lines (Fig. 5). Two distinct phases were observed: the first relates to the delta seaward progradation, representing a constructive phase; and a destructive phase when the delta top is reworked by a marine transgression. After the delta starts to prograde, these transgressive deposits should be underlain by a transgressive

surface and overlain by a maximum flooding surface (Gani and Bhattacharya, 2005). Thus, instead of migrating seaward, the transgressive deposits should migrate landward and the boundary between delta front and prodelta should be somewhat lower in relation to the sloping surface of the depositional system, known as the clinoform (Fig. 6).

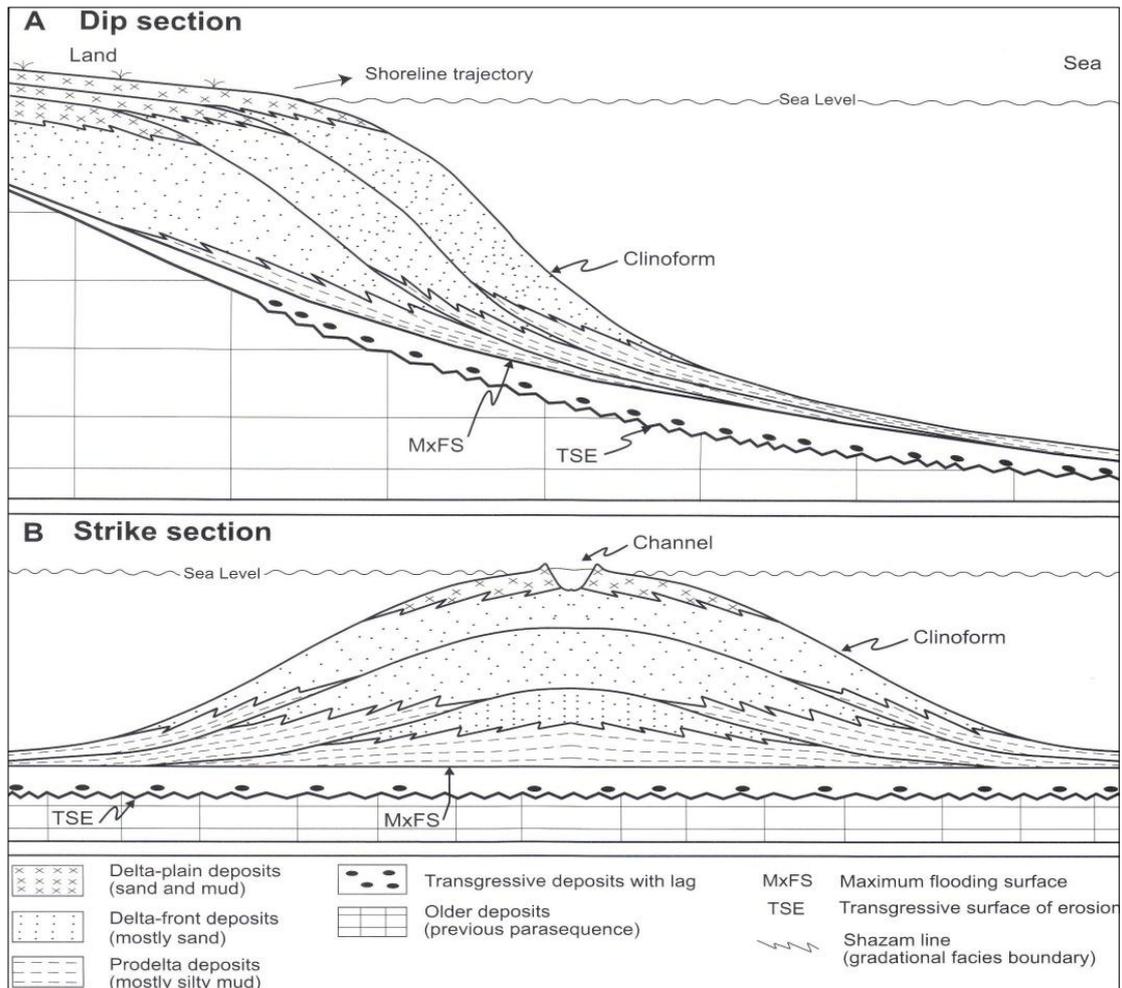


Figure 6 - Generalized intra-parasequence bedding geometry and facies architecture of a prograding delta during a slow sea-level rise. A) Dip section shows the seaward-dipping bedset boundaries (clinoforms), which follow time lines and are bidirectional in B) strike section (Gani and Bhattacharya, 2005). Note facies dislocations across bedset boundaries. Figures not to scale.

In a strike and dip intra-parasequence of a prograding delta during a slow sea-level rise (Gani and Bhattacharya, 2005), a sharp erosional surface is formed between the prodelta and the delta front deposits as a result of the opposite actions of the prograding delta trajectory and the regressive coastline (Fig. 6). During the first

phase of the marine transgression, the older deposits are reworked by the marine action and pushed inland, forming a erosional surface above the transgressive lag deposits. When sea level reaches the maximum hold, which represents the maximum flooding surface, the new delta starts to build, prograding seaward and producing seaward dipping bedsets. In the strike view, there are bi-directional clinoforms, where the coarser delta front sediments pass laterally to prodelta muddy sediments.

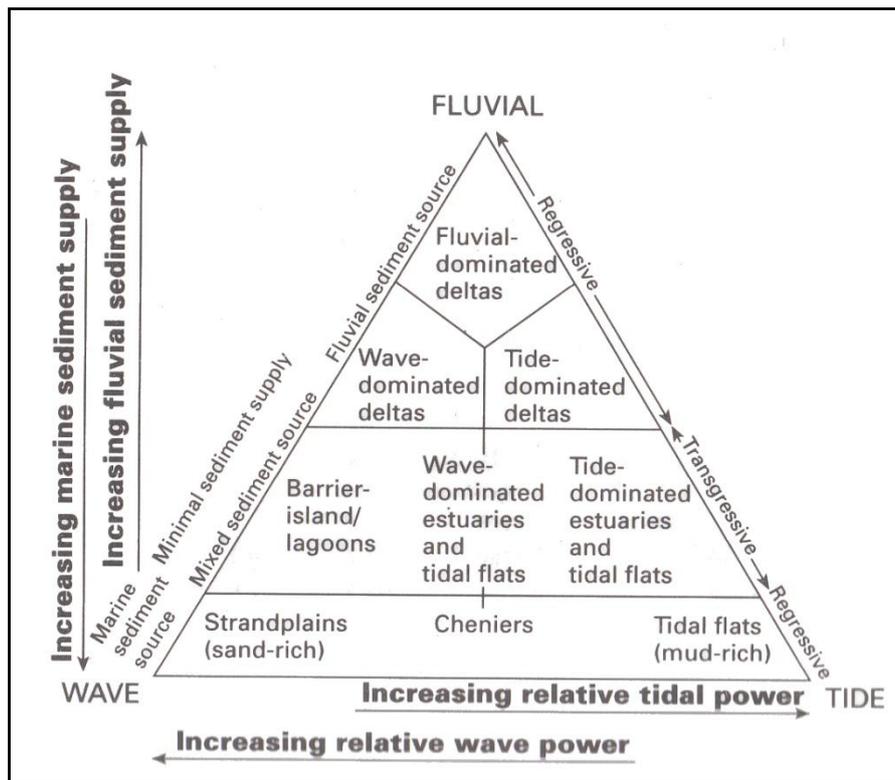


Figure 7 Ternary process-based classification for all coastal systems other than coarse-grained systems (modified from Dalrymple et al., 1992). It is extended downwards from Galloway (1975) to include additional coastal systems (Reading, 1996).

A process-based classification for deltas was developed by Galloway (1975), establishing the classic ternary diagram for deltas (Fig. 7). In this diagram the fluvial input is plotted against tidal and wave reworking processes but grain size is ignored, and coarse-grained systems are not taken into account. Even so, this diagram represents a good model for discrimination of fine-grained deltaic systems.

2.2 Prograding delta morphology

The components of a prograding delta can be differentiated based on the gradient of the profile but also the dominant process involved, which will influence the type of sediments found in each zone. The process-based terminology divides the delta in *i) delta plain*, which is the mainly subaerial zone dominated by rivers; *ii) delta front*, where the interaction between fluvial and tidal or wave induced basinal processes take place; and *iii) pro-delta*, a zone of quiet sedimentation with slow gravity flows and mass flow deposition (Reading, 1996).

2.2.1 Delta Plain

In the delta plain, several active and residual or abandoned distributary channels cross the lowland with the presence of associated levees, and shallow water or emergent areas in between. In the upper delta the basinal processes are absent and the morphogenesis is similar to that found in alluvial areas. Lower delta channels are still affected by fluvial processes but here there can be some incipient tidal influence. The distributary channels in the delta plain differ from the fluvial ones in several aspects: there can be some action of basinal processes as well as salt water wedge penetration in the lower reaches; avulsions are more frequent because active distributary channel gradients become lower as a result of delta progradation and also the width-depth ratio is lower in the delta plain channels because of their shorter active duration and limited lateral migration potential. Interdistributary areas of the delta plain are composed by swamps that change to marshes in a seaward direction; features like levees, crevasse channels, crevasse splays and minor deltas, can also be formed. The tidally influenced channels have low sinuosity, higher width-depth ratios and can be funnel-shaped. Tidally influenced interdistributary areas include tidal flats and channels, sand and muds are deposited by migrating point bars in a small fining upward sheet sequences.

2.2.2 Delta Front

In this zone, the distributary channels of the delta enter the basin interacting with tidal and/or wave basinal processes (Fig. 8). The sediments are coarser, mainly sandy

and show a delta progradation characteristic coarsening upward sequence. Depending on the dominant process, a different morphology will be formed: if the fluvial processes are strong, mouth bars with small crests and subaqueous levees made of coarser material will develop; in the case of a major tidal influence, tidal sandbars and ebb and flood tidal deltas will be formed. Where wave processes prevail, swash bars, beach ridges and beach spits develop.

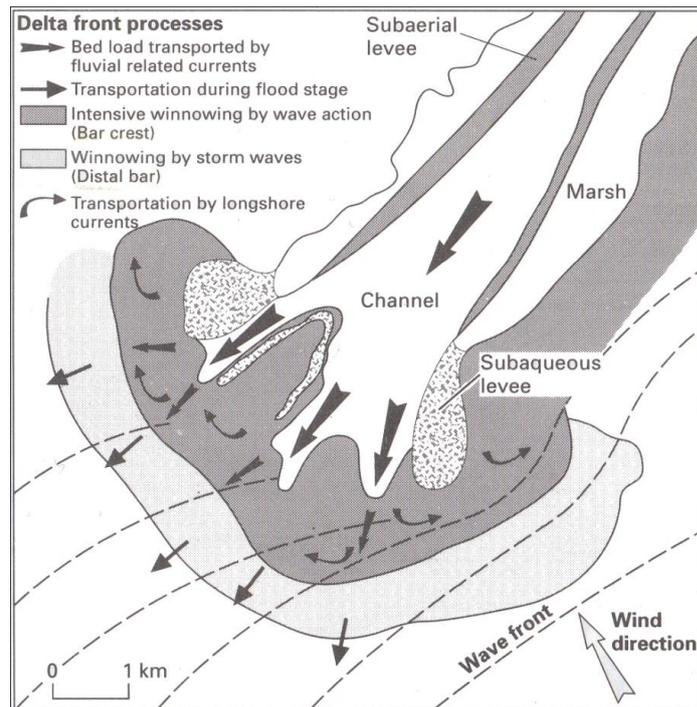


Figure.8 Delta front processes (Reading, 1996).

2.2.3 Pro-delta

The pro-delta zone is not influenced by tidal or wave processes. The deposits reflect variations in fluvial discharge and sediments carried by the river, and consist of well laminated often bioturbated silts and clays with some intermixing of sand. The finer particles are carried seaward in hypopycnal plumes (when the receiving basin water is denser than the incoming flow) that flow buoyantly until the sediments will flocculate and settle. In the more rare case of an existing hyperpycnal plume (when the incoming flow is denser than the receiving basin) the laminated and bioturbated sediments are entrained by sharp-based beds.

2.2.4 The IJsseldelta

The delta features described previously are present in the IJsseldelta (Fig.9). A delta plain area extends to the west and northwards from the city of Kampen until a coastal ridge that represents a more elevated area in the landscape, in the outer reaches of the coastal rampart, the delta plain continues until the IJsselmeer. Adjacent to the former most important branches of the IJsseldelta in Kampereiland there are natural levee deposits that appear as slightly salient ground in the topography of Kampereiland. The underwater pro-delta deposits extend further and reach the Noordoostpolder and the Flevoland polder. The IJsseldelta features will be further explained in Chapter 4 of this report.

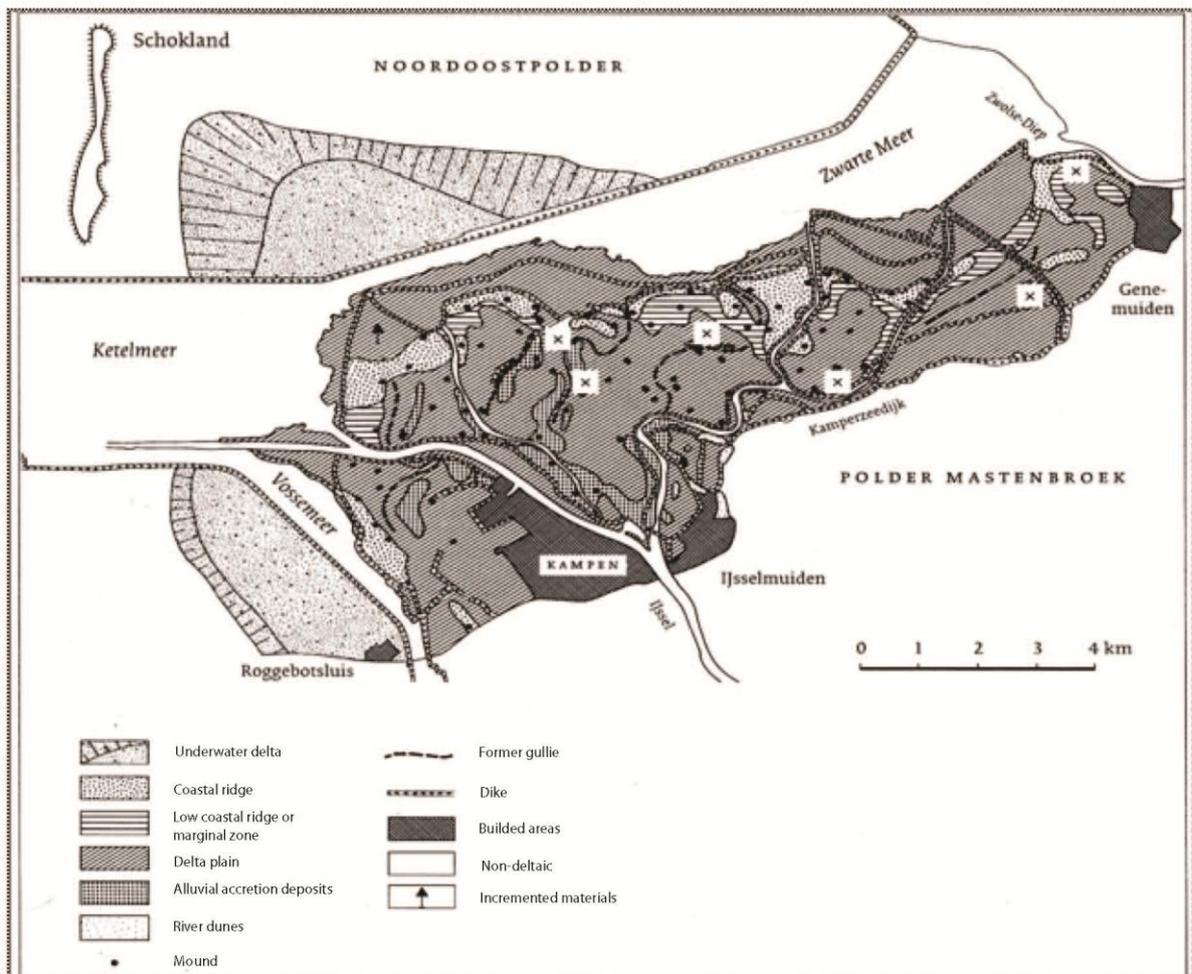


Figure 9 – Geomorphological map of the IJsseldelta (Dirkx, 1996). The main geomorphological features of the IJsseldelta are represented, it is important to point that the underwater delta sediments extend past Kampereiland and are found in the Noordoostpolder and Flevopolder.

3. METHODS AND APPROACH

To achieve the proposed goals a diverse set of research methods has been applied. First of all, it was needed to know the deposits present in the delta with the highest resolution possible, for that the sampling obtained by hand coring provided a valuable data set. Sediment core descriptions are the main basis for the research, with the analysis and comparison of the boreholes in the database making possible to reconstruct the sedimentary architecture in the area and the evolution of former channel belts as well as the extent of marine and fluvial transported clays, for example. By plotting selected boreholes in four transversal transects it was possible to more easily trace and compare the sedimentary distribution and architecture of the delta from the proximal to consecutive more distal zones.

Historical data from old maps, shipping journals, major storms and floods or dike infrastructure construction can be very useful, when crossed with the coring data, to help reconstruct the development of the delta. Even so, these data must be carefully used because there can be precision errors or equivocal leads.

3.1 Digital elevation model

The digital elevation model of the study area was plotted using the actual elevation database of the Netherlands (AHN, Rijkswaterstaat-AGI, 2005) and GIS-software (ArcMap). The elevation map of the IJsseldelta (Fig. 10) was used for an initial *a priori* analysis of the study area, morphological features like gullies and levees, or even the presence of former channel belts and crevasse splays could be inferred and then confirmed by borehole drilling. In this way the elevation map was the first tool used for the planning of the drilling campaign, and borehole locations were selected for areas that pose interesting questions due to the morphology of the surface. Furthermore, the elevation map values (scale 1:10,000) were used to establish the z-coordinate (elevation relative to Ordnance Datum = NAP \approx mean sea level) of individual boreholes.

Through a cross-relating the information provided by elevation map with that contained in the borehole logs it was possible to construct a detailed geomorphogenetic map of the study area (Fig. 22, Chapter 4).



Figure 10 Digital elevation model of the IJsseldelta. Colours in order of increasing height: white-light green-green-yellow-red. Dashed-lines represent the approximate location of former channels.

3.2 Data sampling

Most lithological information about the IJsseldelta is archived as an extensive database available in digital form (stored in the digital database program LLG) that results from the boreholes made by the BS students of Physical Geography, Utrecht University, in the year 2007. Cores were drilled in order to form cross-sections with a borehole spacing of ~100 meters and the deepest possible depth, with the ultimate goal of reaching the underlying older Pleistocene deposits. The core descriptions in the database present information about the characteristics, distribution and location of the deposits present in the delta, including information about texture, depth, grain-size values, color, concentration of calcium (Ca) and iron (Fe) in the deposits, gravel and organic admixtures, as well as the depth of groundwater.

Additional boreholes were made during the fieldwork needed for this research and the borehole descriptions are also available in the LLG digital database with the boreholes being labeled as 200810 (standing for the year and group) followed by a sequential borehole number. Although the applied methods were the same, the

sampling strategy differed substantial from the one of the BSc students in 2007. In this case the boreholes were made in selected key locations where further information or confirmation of the available data was needed (e.g. insufficient depth of the available cores or unreliable borehole descriptions) to complete the cross-sections or in areas not included in the cross-sections that posed interpretation issues and where lithologic data was lacking.

When assembled in a wider frame, the descriptions of individual boreholes provide an overview to the vertical/lateral distribution and geometry of the sediments in the IJsseldelta, thus being crucial for the reconstruction of the delta.

3.2.1 Coring techniques

The standard drilling equipment that was used according to the method described in Berendsen & Stouthamer (2001) is:

- *Edelman hand auger* (Fig. 11), for the upper 1 meter or until the groundwater table;
- *Gauge* (Fig. 11), with a length of 0.8 to 1.0 m and a diameter of 3 cm, extra extension rods of 1 m length can be assembled for further depths. This device is used for clay and peat, to a depth of 15 m, Holocene channel belt sand bodies can be sampled with the gauge although in this way it is impossible to obtain sand samples;
- *Van der Staay suction corer*, for sampling sand and fine gravel under the groundwater table up to a depth of 15 m. The suction corer cannot be used in floodbasin deposits but, although not being advisable, it can penetrate thin layers of clay and peat if they underlie channel belt sand bodies.

For each borehole the coordinates x- and y- were obtained with a GPS device and located in the topographic map of the Netherlands, scale 1:10,000 with an accuracy of ± 5 m, the z- coordinate (elevation relative to Ordinance Datum = NAP \approx mean sea level) was taken from the digital elevation model of the study area, scale 1:10,000.



Figure 11 Edelman coring equipment, including, rods, corer and gauge (Berendsen & Stouthamer, 2001).

3.2.2 Borehole description

The boreholes were described using the method used by Berendsen & Stouthamer (2001) which was based upon the system defined by De Bakker & Schelling (1966) for weight percentage of organic matter and Verbraeck (1984) for sand and gravel grain size as well as gravel content classification.

Boreholes in the database are labeled with a unique number that includes the year, working group and sequential or the borehole. During sampling in field, the boreholes were described in intervals of 10 cm (or less when necessary), taking into account texture (Fig. 12), organic material content, gravel content, median grain size, colour, iron and calcium carbonate content (with and 5% HCl solution), groundwater levels, shells and other attributes. The percentage of organic matter and gravel content were estimated in the field using the mentioned classifications, grain size of sand and gravel was estimated using a sand ruler.

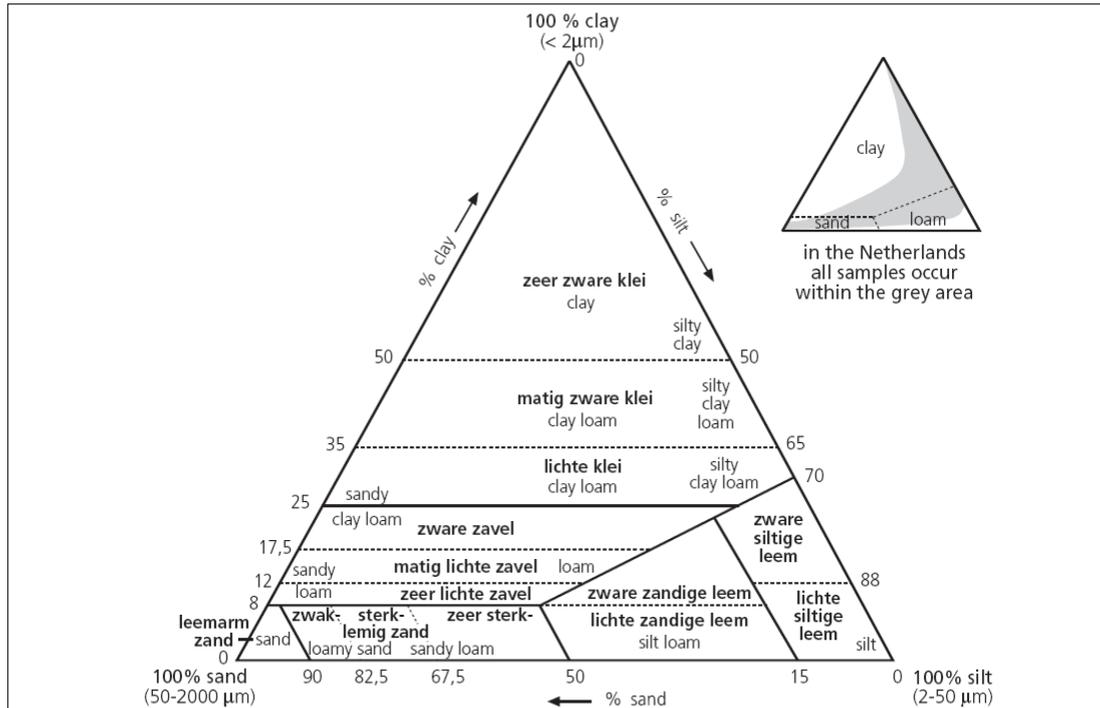


Figure 12 - Texture classification of clastic sediments, consisting of a mixture of sand, silt and clay. In bold are represented the Dutch classifications for textural units (after Verbraeck, 1984, in Berendsen & Stouthamer, 2001).

Because the Dutch classification of textural units differs from the standard U.S.D.A nomenclature, the first were converted to English according to Berendsen & Stouthamer (2001) as is shown in Figure 12.

3.2.3 Lithostratigraphic units

The deposits in the boreholes descriptions were grouped into lithostratigraphic units according to their lithologic composition and architectural elements (facies defined by grain size, bedform composition, internal sequence and external geometry). The subdivision of the deposits into lithostratigraphic units was based in the lithogenetic units defined by Berendsen (1982) which was adapted to the particular characteristics of the deposits found in the IJsseldelta (Table I).

Lithostratigraphic unit *Bc*, that previously corresponded to “Echteld Formation, crevasse splay deposits”, was changed to group all delta deposits of the Echteld Formation, reflecting the complex nature of delta progradation (Table I). Furthermore, this unit was subdivided so that it could discriminate specific deposits in the delta and if the sedimentation occurred under the influence of the river IJssel (Unit Bc1), delta

bulk progradation (Unit Bc), quieter lagoon conditions (Unit Bc3) or in underwater channels (Unit Bc4).

ADAPTED LITHOSTRATIGRAPHIC UNITS (STRAT)	
P	Middle and Lower Pleistocene formations
Df	Drente Formation, glaciofluvial deposits
Td	Boxtel Formation, coversands
Tp	Boxtel Formation, fluvioperiglacial deposits
To	Boxtel Formation, organic deposits
TD	Boxtel Formation, Delwijnen Member, aeolian river dune deposits
Kb	Kreftenheye Formation; channel deposits
KWk	Kreftenheye Formation; Wijchen Member; floodbasin or levee deposits
KWr	Kreftenheye Formation; Wijchen Member; residual channel deposits or peat
Bb	Echteld Formation; channel deposits
Bo	Echteld Formation; natural levee deposits
Br	Echteld Formation; residual channel deposits
Bk	Echteld Formation; floodbasin deposits
Bd	Echteld Formation; dyke breach deposits
Bc	Echteld Formation; delta deposits
Bc1	Echteld Formation; deposits within the Ramspolzand under the influence of the IJssel but in a quiet environment (more lagoon influence)
Bc2	Echteld Formation; Ramspolzand; delta deposits
Bc3	Echteld Formation; deposit within the Ramspolzand under the influence of the delta but in a quiet environment (more lagoon influence)
Bc4	Echteld Formation, deposits of underwater channels inside the Ramspolzand
Vg	Nieuwkoop Formation, gyttja
Vr	Nieuwkoop Formation; Phragmites peat
Vz	Nieuwkoop Formation; Carex peat
Vb	Nieuwkoop Formation; Wood peat
Vo	Nieuwkoop Formation; undifferentiated
WDk	Naaldwijk Formation; Walcheren deposits, floodbasin and natural levee deposits

Table 1 – Adapted lithostratigraphic units (STRAT).

The lithostratigraphic units were used to complete the descriptions of individual boreholes and also in the cross-sections to trace architectural elements and build geological profiles based on the lithology and genesis of the deposits.

3.3 Inter-core correlations

After defining the lithostratigraphic units at core level the following step was to correlate these units with the adjacent boreholes. The inter-core correlations were established by tracing the boundary surfaces for each lithostratigraphic unit across the lithologic profiles, taking into account the sedimentary characteristics available in the borehole descriptions, the depositional environment and the relative time interval of deposition. Again, the borehole descriptions stored in the LLG digital database were the main data used to perform this task.

More complete and higher-resolution borehole descriptions were used as reference to more precisely define sedimentary facies and bounded surfaces, that were then traced along the other cores of the transect.

3.4 Transversal cross-sections

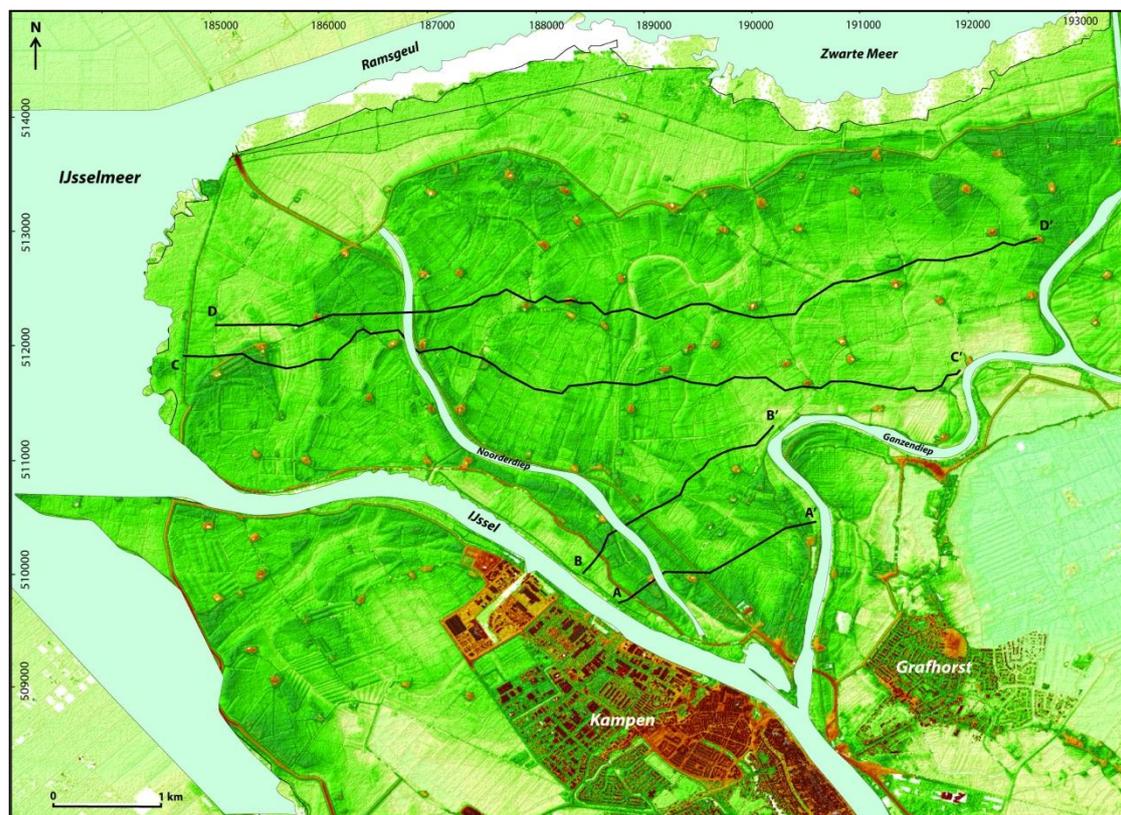


Figure 13 – Digital elevation model of the IJsseldelta and location of the cross-sections. Colours in order of increasing height: white-light green-green-yellow-red. In this figure it is used the Dutch Coordinate System (m).

Selected cores were plotted in cross-sections so that relations between them were better seen and used for geomorphogenetic mapping. The cross-sections were constructed in a south to north sequence so that the transversal profiles of the delta were represented, interpreted and the different sections of the delta longitudinal morphology compared, giving a better insight into the delta evolution. A geomorphogenetic map was built using borehole data from the delta area and taking into account the lithology and morphogenetic factors.

To achieve the purposed goals in this research, the chosen approach to the research was to use the selected borehole database in order to elaborate several cross-sections that cover the study area from the fluvial dominated delta apex near Kampen to the center of Kampereiland, where marine and fluvial influences interfinger. Four cross-sections were planned and constructed with SW-NE (cross-sections A-A' and B-B') and W-E directions (cross sections C-C' and D-D') as is shown in Figure 13. The position of these four cross-sections gives a good insight to the 2D geometry and facies characterization/distribution of the delta deposits in Kampereiland. In the cross-sections it is represented the lithology as well as the adapted lithostratigraphic units (Table I).

3.5 Dating and correlating channel belt fragments

By the elaboration of detailed geological and geomorphological maps that show the patterns of Holocene channel belts sandy deposits and their contemporary natural levees in a matrix of clayey and peaty materials, where the older channel belts have been eroded by the more recent ones, the relative age of channels can be obtained by cross-cutting relationships. The relative depth of overbank deposits of channel belts can also be an indicator of relative ages of residual channels.

After a palaeogeographic reconstruction is made, it is possible to reconstruct the periods of activity of individual channels and of the related natural levee deposits in a river system, bringing further light to discover the location and relative dates of avulsions.

3.6 Geomorphogenetic mapping

The geomorphogenetic map is essentially based on borehole descriptions and maps made by the undergraduate students of Physical Geography, Utrecht University, during the fieldwork campaign of 2007 (Fig. 22). As the name implies, this map represents, at a scale of 1:25,000, the morphology and genesis of Kampereiland taking into account factors like the lithological profile, age of the deposits and processes involved in landscape formation and evolution. In this map, the main focus is set into the genesis of the landscape and the processes involved on its development, giving a valuable overview of the extent of marine (tidal) and fluvial influence in the area for the period of the delta formation.

GEOMORPHOGENETIC UNITS	
AEOLIAN LANDFORMS	
Eo	Aeolian coversand riverdune; are mapped on the basis of the relief. The detailed subdivision in map units is based on differences in profile build-up.
FLUVIAL LANDFORMS	
Fs1	Levees and pointbars; the profile starts fining downward between 0 and 50 cm below the surface and grades from sand to gravel at least within 120 cm below surface level.
Fs3	Levees and pointbars; the profile starts fining downward between 100 and 150 cm below surface level, passing into loam at approximately 150 cm below surface level and into sand at approximately 180 cm below surface level .
Fs31	Lateral accretion ridges.
Fs32	Mouth bar.
Fs33	Mouth bar.
Fs5	Levee.
Fs7	Residual channel; mapping of this unit is initially based on the relief, the profile is characterized by one thicker clay layer compared to the surrounding "levees" and "levees and pointbars". Peat, clayey peat or peaty clay does not appear in residual channels.
Fs8	Residual channel; the residual channels contains at least a 10 cm thick layer of peat, clayey peat or peaty clay.
Fs9	Residual channel; the residual channel contains water.
Fc3	Crevasse splay; the profile starts fining downward between 100 and 150 cm below surface level and contains loam and sand at least within 200 cm below surface level.
Fk	Floodbasin

Fk1	Floodbasin; the profile does not present fining downward or upward sequences and is almost entirely non-calcareous. It might consist of peat, clayey peat and peaty clay or humic-rich clay up to 150 cm below the surface although the profile mainly consists of clay down to 200 cm below the surface.
Fk22	Floodbasin
Fk23	Floodbasin
Fk25	Floodbasin
Fk32	Floodbasin
Fu3	Embanked floodbasin; the profile contains channel bed load sediments (mostly calcareous sand) between 100 cm and 200 cm below surface level. Sometimes gravel layers and clay bands occur in the channel deposits.
Fu5	Embanked floodbasin; the profile consists mainly of loam until a depth of 10 cm below surface level, passing down into clay loam or clay. The thickness of the levee deposits does not play a role as it is highly dependent on the outward sedimentation. Levee deposits may even be entirely lacking.
Fo1	Dyke breach deposits; the profile is characterized by the presence of a highly calcareous sandy clay or loamy layer of 40-150 cm thick. The map indicates on which units the dike breach deposits are located. Thus, the profile build-up is given for at least 240 cm below surface, and up to 350 cm below surface level.
LAGUNAR LANDFORMS	
Lw	Coastal ramparts
Lk1	Lagunar plain
Lk2	Lagunar plan
Lk3	Lagunar plain

Table II – Geomorphogenetic units.

The geomorphogenetic classification used to build this map follows the units described by Berendsen (1982) posteriorly completed and upgraded during the following years (Table II). This classification focuses on the genesis of the landscape and the geomorphogenetic factors behind its formation, with a clear separation between fluvial and marine environments, and landscape morphology related to older terrain landforms. The age of the deposits and morphological units is also taken into account and is determined through several dating methods described by Berendsen & Stouthamer (2001). Another important feature to the unit classification is the lithological profile, especially the first 0-200 cm counting from the surface that gives an important insight into the landscape geomorphology. In this way, the depth, succession and relative position of the deposits, when compared with ideal profiles for different situations, is also an important tool to understand the genesis of the landscape

3.7 Historical data

Historical archives, maps, and other written documents, as shipping journals and data from former engineering works, can be used to know which channels were active for boat routes at each moment and help reconstruct the age of river systems. Although historical evidence can be very useful, it must be used carefully, many times descriptions of locations and events can be inaccurate, often there are errors in the documents that derive from wrong interpretations and bad translations. Furthermore, the final age of a river channel may be younger in historical evidence because the channel can still contain stagnant water after the formation of gyttja and peat starts.



Figure 14 – The IJsseldelta in the “Kaart van Overijssel” by N. ten Have (circa 1690)

Much of the historical information for the area of Kampereiland consisted mainly of old maps which, although not accurate, show which main branches of the River IJssel were active at a certain time. The map of the Dutch Province of Overijssel (“Kaart van Overijssel” in Dutch) made by N. ten Have in the end of the 17th century (*circa* 1690 AD) may not be precise for locations but shows relevant information for the main river channels in the IJsseldelta about the year 1690 (Fig. 14). It is quite visible in the referred map that at this date the main discharge of the IJssel was drained by the Rechterdiep and Zuiderdiep, the Noorderdiep is still active at this time, and the Gazendiep follows a course similar to the actual. Other historical maps from the same period, like the 1658 AD historical map of the Netherlands “Der Zuyder Zee” by Janssonius (Fig. 15), show similar information but with lower resolution and detail.



Figure 15 – Historical map of the Netherlands (1658 AD) by Janssonius.

The dates of the human settlements and farms were very useful to know where the higher grounds were located. Of course, usually the structures were built in anthropogenic mounds but, in several situations, the alluvial accretion and coastal ridges were used for settlements locations. This important data was assembled and studied by Dirx (1996) and is approached with more detail in Chapter 5.

Records of major storm surges, floods and dyke works are registered in the Explanation for the Soil Maps of “West Lelystad” (soil map sheet 20), “Oost Lelystad” (soil map sheet 20) and West Zwolle (soil map sheet 21) by Eilander et al. (1990). Storms affected the area in the 12th and 13th centuries, eroding large areas of peat and giving origin to the “Zuiderzee”. The “Julianavloed” flood in 1164 AD marks the beginning of this phase, during other floods in 1212, 1214 and 1248 AD the seawater penetrated Lake Almere and the barriers along the Dutch coast were breached, marine creeks protruded further inland washing away sediments and, as a result, an inland sea developed and the margins of the lagoon area became a more saline environment (Fig. 15).

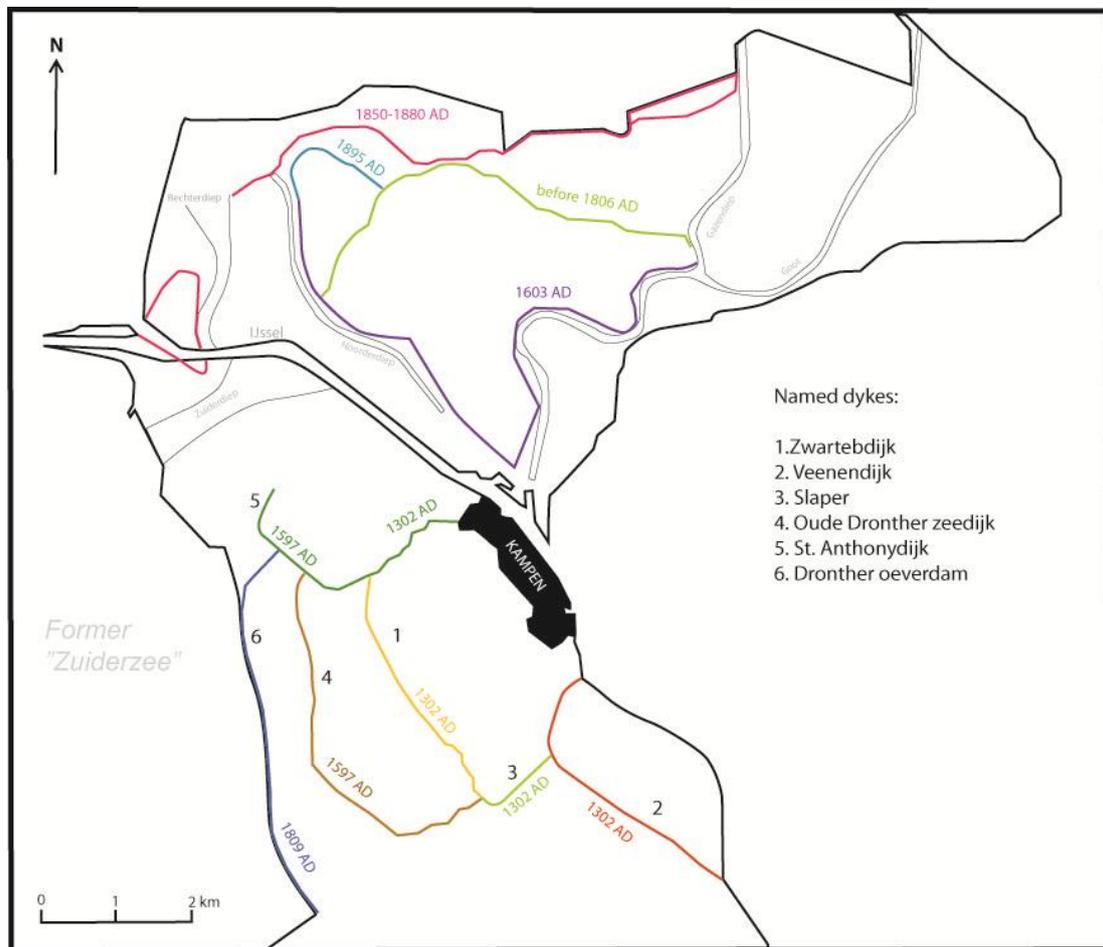


Figure 16 - Major dykes constructed in the IJsseldelta before the 20th century. Adapted from Eilander et al. (1990)

Valuable data was given by the recorded time when the dikes were constructed, revealing environmental changes in the morphology of the landscape thus being an

important asset to understand changes in the delta dynamics. During the Late Middle Ages human interference in the Kampen area became more active with the construction of several dykes to protect the lands from the sea. The “Zwartedijk” was built in 1302 AD in the southwest of Kampen, giving an impulse to the further construction of more dykes in the area (Fig.16). More dykes were built as the coastline evolved: in 1603 AD a dyke was built north of the city of Kampen along the eastern margin of the Noorderdiep and the western margin of the Gazendiep. Other dykes were built in the 19th century to protect the inner lands from the influence of the sea (Fig. 15). In 1932 AD the “Afsluitdijk was built, closing the former “Zuiderzee” and the freshwater Lake IJssel (“IJsselmeer”) evolved.

4. DATA ANALYSIS

4.1 Transversal cross-sections

4.1.1 Cross-section A-A'

This transversal cross-section is situated near the apex of the IJsseldelta, just north of the city of Kampen, in an area between the River IJssel and the Gazendiep. It extends for 1910 meters, along a general WSW-ENE direction, with the exception of the W-E section between boreholes 200716095 and 200716146 (Fig. 17).

The deposits are mainly uniform fine (150-210 μm) grey sand, well sorted and sub-rounded, rich in calcium carbonate and containing common shell fragments all over the length of the cross-section. Intercalations of fine gravel (Φ 1000-3000 mm) in the fine sand matrix are frequent in the two most western cores (200716094 and 200716095) at depths of -2 meters NAP and 1.40 meters NAP, that may be explained by their location between the IJssel and a dyke structure.

The sand is overlain by several loamy sand overbank deposits, changing to floodplain silty clay loam and silty clay (located between boreholes 200716088 and 200716146) that cover almost the entire profile in a 40 cm to 1 meter thick layer. These clays are thicker in the mid-section of the profile (with a thickness of 1 meter in borehole 200716082), getting thinner to the east, where the sandy deposits reach the surface (borehole 200716080).

There are two residual fluvial channels present in the profile (boreholes 200716088 and 200716076) reaching bottom depths of -2.1 m and -1.7 m NAP respectively and show a typical infill of clay with an alternation of sandy and clay loam bands. Both residual channels are overlain by floodbasin deposits indicating that they stopped activity before the later channel.

The moderate fine calcareous sands are also called "Ramspolzand" and belong to lithostratigraphic unit Bc (Table I), representing the bulk of sandy materials deposited during delta progradation. The coarser sand intercalations within the fine sands are

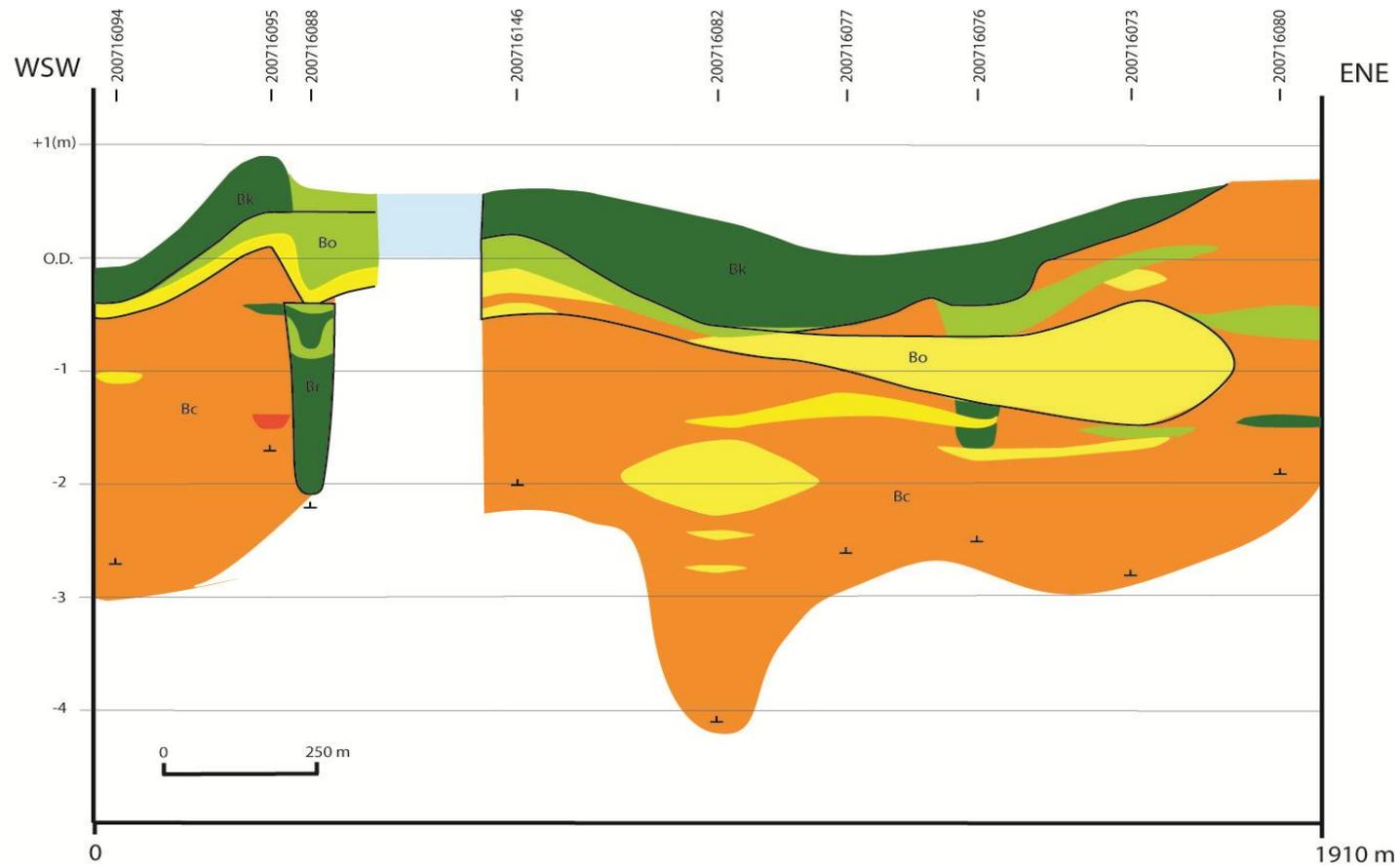


Figure 17 – Cross-section A-A'. For legend see Fig. 18. Lithostratigraphic units described in Table I.

Naaldwijk Formation - shallow marine deposits

-  Clay
-  Clay with sand and silt layers

Nieuwkoop Formation - Autochtoneous organics

-  Peat / clayey peat / peaty clay
-  Sandy peat / peaty sand

Echteld Formation - Estuarine/deltaic Rhine-Meuse deposits

-  Clay, silty clay
-  Silty clay loam / clay loam
-  Clay loam / sandy clay loam
-  Sandy loam / loamy sand
-  Extremely fine sand / very fine sand / fine sand / medium sand
-  Coarse sand / very coarse sand

Kreftenheye Formation - Valley Rhine-Meuse deposits

-  Medium sand
-  Coarse sand / very coarse sand

Boxtel Formation - Aeolian and local deposits

-  Very fine sand / fine sand / medium sand
-  Coarse sand

Miscellaneous

-  Water
-  Anthropogenic
-  Borehole location

Figure 18 – Key to the Kampereiland cross-sections (Fig. 17, 19, 20 and 21).

derived from high flow velocities in small subaqueous channels during delta formation. Using a similar reasoning, the finer clays and silty clays in between the delta sands stand for local low flow velocities.

The characteristics of this cross-section deposits indicate that it represents a Holocene fluvial sequence that belongs to the Echteld Formation. The marked fluvial character of the deposits in this cross-section is related to its proximity to the delta apex. After the embankment of the River IJssel, from about the 11th Century, and respective narrowing of the river bed, the transport of sandy material increased and a subaqueous sandy platform flanked by peat areas started to form into the Almere about the 12th Century (Ente, 1971). From that time on, large quantities of fluvial sediments from the IJssel were discharged through the delta apex as the IJsseldelta was built and prograded into the former Zuiderzee.

4.1.2 Cross-section B-B'

Cross-section B-B' has a SW-NE direction, representing a cross-section of 2220 meters long, ranging from the vicinity of the IJssel to the Gazendiep (Fig. 19). The surface is lowest in the extremes of the cross-section, near the River IJssel and Gazendiep (-0.1 m and 0.0 m NAP), the highest surface point is 0.9 m NAP and corresponds to an anthropogenic mound just SW of the Noorderdiep (borehole 200716005), in the central section of the profile there are also three depressions that correspond to fluvial residual channels. Through the whole extensions of the cross-section there is an upward sequence of peat, sand, loam and clay, with common small crevasse bands and much rarer very low energy condition deposits (clay and gyttja).

The most western boreholes (200716001 to 200716003) represent a section of floodbasin deposits thinner than 50 cm (reaching a depth of -50 cm NAP) between the IJssel and a dyke, resting on top of a fining upwards layers of calcium carbonate rich moderate fine sand with in depth intercalations of coarse sand and fine gravel.

The deepest deposits in this cross-section consist of Wood peat (Nieuwkoop Formation, lithostratigraphic unit Vb). The peat is found at a depth of -4.90 m NAP in the SW (borehole 200716011), becoming shallower to the NE (-4.80 m NAP and 4.70 m below NAP in boreholes 200716012 and 200716013 respectively) and reaching a depth of -1.80 m NAP level near the Gazendiep. In borehole 200716019 the peat

layer has a thickness of 1.4 m and is underlain by fine, well sorted, non-calcareous coversands deposited by the wind during the cold Younger Dryas Stadial (10000 to 11000 ¹⁴C yr BP).

Above the peat there is a thick layer of fine to medium sized calcium carbonate rich grayish sand that corresponds to the sandy delta front deposits (Ramspolzand). These deposits are characterized by a poorly defined fining upwards sequence (the deeper sand has an average grain size of 210-300 µm whereas the upper section of the layer has an average grain size of 150-210 µm), with generally moderately sorted sub-rounded to sub-angular grains. The Ramspol sands have an estimated average thickness of 4.50 m in the central part of the cross-section becoming thinner to NE as the peat is shallower and the floodbasin and overbank deposits of the Gazendiep thicker.

On top of the deltaic sands, there are loamy (with sand or clay) or clay loam deposits that correspond to periods when the discharge of the channels was higher and flooding occurred resulting in overbank deposition. Covering the whole length of the cross-section there are floodbasin silty clays and clays (with an average 30-40 cm thickness) that become thicker in the NE, near the Gazendiep (60 cm in boreholes 200716019 and 200716044).

In the central section of the cross-section there are three residual channels (boreholes 200716009, 200716011 and 200716014) characterized by a clear fining upwards section, passing from coarse grained (1000-420 µm) to medium and fine sand with common shell remains (no more information about these shells available in the borehole descriptions), covered by silt and clay infilling. The base of the channel coarse sands in the central residual channel (borehole 200716011) is found at a depth of 4 meters, far deeper than in the other two mentioned channels, but the infilling appears at the same depth which probably means that the three channels stopped activity at the same moment. As it can be seen in the geomorphologic map of the area (Fig. 22), boreholes 200716011 and 200716076 in cross-section A-A' represent the same residual channel.

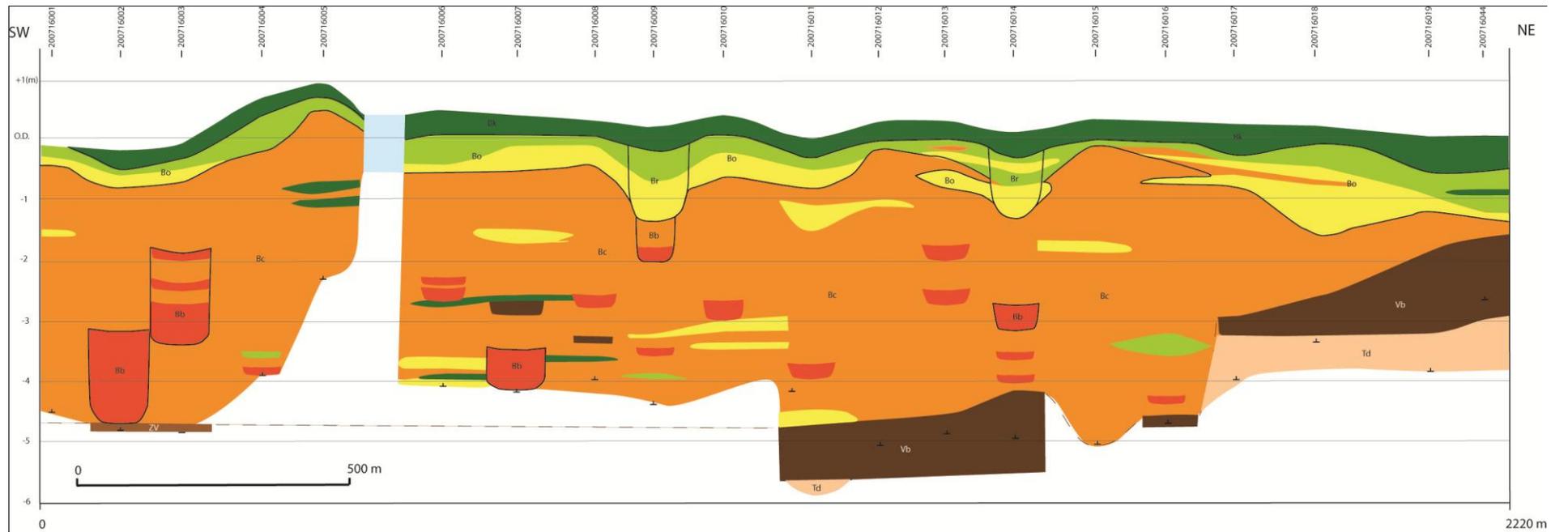


Figure 19 – Cross-section B-B'. For legend see Fig. 18. Lithostratigraphic units described in Table I.

4.1.3 Cross-section C-C'

In cross-section C-C' a transversal generally W-E profile of the delta, with a length of more or less 7200 meters is presented (Fig. 20). It extends from the proximity of the IJssel Lake (IJsselmeer) to the Gazendiep, crossing the whole length of Kampereiland.

Borehole 200711038, with a total depth of -10 m NAP, gives continuous information about the vertical sedimentary sequence showing the traces of environmental changes in the area since the Pleistocene. In the base of the borehole, from -8.60 m to -10 m NAP, the deposits consist of moderately sorted coarse calcium carbonate rich grey sand. These predominantly gray, coarse, weakly sorted sands with high calcium carbonate content and containing many different mineral types are interpreted as Late Pleistocene fluvial deposits of the Rhine and belong to the Kreftenheye Formation (Kb). From a depth of -8.50 m NAP and until -6.50 m NAP, the sand color changes to lighter grey and the calcium carbonate content decreases as the deposits are at shallower depth and less coarse, showing a roughly fining upwards sequence and occasional gravel intercalations. The characteristics of these deposits and the fact that they are overlain by well sorted, fine grained sands without calcium carbonate content suggest a Late Pleistocene periglacial origine (Boxtel Formation, Tp).

On top of the periglacial deposits there is a 50 cm thick layer of well sorted fine (150-210 μm) to very fine (105-150 μm) sand without calcium carbonate. These deposits are aeolian coversands deposited by the wind in a cold, windy climate with sparse vegetation (Boxtel Formation, lithostratigraphic unit Td). These coversands are interpreted to be "older coversands" because they are found in an almost horizontal layer at a depth of around -6 m NAP. In contrast, the so called "younger coversands" (borehole 200713037, lithostratigraphic unit TD) are somewhat coarser and were deposited as aeolian river dunes. Immediately above the coversands there is a 20 cm thick lag deposit of clay loam that can be related to local deposition in creeks that ran across the coversands or an incipient soil formation (Kreftenheye Formation; Wijchen Member), in the case that these deposits cover a broader area, but the available data is inconclusive.

As mentioned above, in boreholes 200713107 and 200710011, directly upon the "older coversands" and reaching a depth of -140 cm NAP there are sub-angular moderately sorted grey sands, with medium grain size (210-420 μm) and no calcium

carbonate. These deposits are aeolian dune deposits formed on the banks of Late Glacial rivers (referred as “river dunes” in Dutch literature) by sand blown out from the channel bed during periods of low discharge and accumulating on the eastern banks of the channels due to the predominant western winds.

On top of the Pleistocene deposits there is a Holocene peat layer (Nieuwkoop Formation). These organic deposits have a thickness of between 50 and 100 cm and are mainly composed by reed peat but can laterally change to wood peat in certain locations (e.g. borehole 200711045). The depth of the top of the peat decreases gently from the western to the central section of the profile (-4.10 m NAP in borehole 200711052 to -5.20 m in borehole 200711044) and is found again at a higher level directly east of the aeolian river dune (-4 m NAP in borehole 200713037) decreasing its depth to the east (-3.30 m NAP in borehole 200714052).

Deltaic deposits appear on top of the peat in a thick layer that can be divided into two different types according to the characteristics of the sediments and processes involved in its deposition. The lower part is deposited directly upon the peat in a band of variable thickness (ranging from 30 to 150 cm in this cross-section) and corresponds to the “pro-delta” deposits (Echteld Formation; delta deposits: laminated sediments). Deposition took place underwater and corresponds to the beginning of the delta progradation. These deposits consist of alternations of laminated bands of clay, fine sand, silt (often bioturbated) and humic layers that reflect variations in river discharge and sediments carried by the river. During periods of considerably low discharge, there wasn't any sediment transport by the river and gyttja formed under very low energy conditions in a quiet and still depositional environment.

On top of the laminated layers there are the delta front sediments that constitute the bulk of delta deposits (Echteld Formation; delta deposits: Ramspolzand). The delta front deposits consist of medium grain sized (150-420 μm), calcium carbonate rich, grey sands. There can be found bands of clay and coarse sand in between the delta front deposits, corresponding to local conditions of lower and higher flow velocities in subaqueous channels. Some fragments of eroded, washed away peat can also be found inside the medium sand deposits.

Overlying the delta sand deposits (Ramspolzand) there are clay loam and sandy loam overbank deposits and floodplain clays, in a layer that varies in thickness from 30 to 100 cm across the cross-section. In the western part of the cross-section, from the IJsselmeer to the western bank of the Noorderdiep, the superficial cover is made

of marine clays (Naaldwijk Formation; Walcheren Member deposits) deposited by the former Zuiderzee during marine transgressive periods. To distinguish between marine and fluvial sediments is complicated, the main differences (besides a salinity test) is that river clays have incorporated coarser sand particles and brown colour whereas the marine clays have a blue-gray colour, usually less organic matter and more calcium carbonate content from the marine shells.

Several residual channels can be observed along the length of the cross-section. In the westernmost third of this cross-section, two residual channels (boreholes 200711038 - 200711045 and borehole 200711044) have an infilling of blue-grey marine silty clays and clays (Naaldwijk Formation; Walcheren Member deposits), with frequent organic bands in the upper layers, that were deposited during marine transgressive periods, when salt water from the former Zuiderzee penetrated upstream through the abandoned channels. The Rechterdiep channel presents deposits of a point bar (borehole 200711045) with several consecutive sequences of fining upwards coarse (300-420 μm) to medium (420-600 μm) sand (Echteld Formation).

Between the Noorderdiep and the Late Glacial "river dune" area, the boreholes are much shallower and do not reach the "pro-delta" deposits but, even so, many features can be seen. On the eastern bank of the Noorderdiep there are natural levee deposits and marine deposits that penetrated inland from the north through an abandoned channel (Naaldwijk Formation). There is evidence of the former Garste channel (boreholes 200712091-92), again with the presence of point bar deposits, and another residual channel (borehole 200713067) just west of the "river dune" location.

To the eastern section of the cross-section the floodbasin deposits become generally thicker with the proximity of the Ganzendiep. In borehole 200714056, from -0.9 m to 3.50 m below NAP, there are thick deposits of poorly sorted medium to coarse channel sands forming a channel belt, naturally with an older age than the Ganzendiep, that can be traced in boreholes further north (200714126) as well as south of the Ganzendiep (borehole 200716131).

4.1.4 Cross-section D-D'

Cross-section D-D' represents an 7325 meters transversal W-E profile of Kampereiland (Fig. 21). Many of the features, as well as the same lithologic sequence, that were observed in the previous cross-section are repeated in this one. The different main deposits (e.g., fluvioperiglacial; "pro-delta"; and Ramspolzand deposits) have the same characteristics as in cross-section C-C' but differ in features like thickness, location and depth at which they are found. There are also several residual channels and former channel belts that are a continuation of the ones mentioned in the previous cross-section.

From -4 meters NAP downwards there is mainly wood peat (Nieuwkoop Formation) with a thickness of 1 meter to 50 cm. In the western section of the cross-section the peat is not reached due to a lacking of drills deeper than 4 meters -NAP in this area and also because here the peat was eroded and is probably found at a lower depth if it wasn't completely eroded during the formation of the delta. Below the peat there is Late Pleistocene fine grained well sorted sand (150-210 μm).

Overlying the peat are the laminated "pro-delta" deposits that mark the start of deltaic progradation. Again the laminated package is made of thin clay, fine to very fine sand, organic materials, gyttja and loamy bands. In the eastern section of the cross-section the "pro-delta" deposits are found at a depth of -2 meters NAP and have an average thickness of 150 cm. To the central and western sections of the cross-section the laminated deposits become deeper and thinner, maybe because they were eroded during the building of the prograding delta. Above the laminated "pro-delta" there is the fine to medium grain size (150-420 μm), calcium carbonate rich, delta sand of the Ramspol package (Echteld Formation; delta deposits: Ramspolzand).

The delta deposits are covered by overbank loamy deposits and floodbasin clays (Echteld Formation) that change laterally to marine clays (Naaldwijk Formation; Walcheren Member deposits) in the western and eastern sections of the profile, defining the range of tidal influence in this area (see marine clays description for cross-section C-C'). In the extremities of the cross-section, the 50 cm to 1 meter thick lagoonal brackish water loams and clays form coastal ridges that appear as a higher ridge in the surface. Boreholes 200712009-12, in the central section of the cross-section, represent a former delta channel that was penetrated by tidal deposits after loosing discharge capacity.

Again, in the western part of the cross-section, there are traces of the former Rechterdiep channel but these deposits cannot be interpreted because materials were dumped here during construction works. The Noorderdiep appears in this cross-section, there is still water in the channel but it is no longer active. As was mentioned above, in the central part of the cross-section there are traces of a former distributary channel of the Noorderdiep that was infilled by marine tidal loams and clays; one plausible explanation may be that this channel acted as a route for marine influence penetration further inland, when the channel lost discharge competence. The Garste channel deposits appear three times in this cross-section (boreholes 200713003, 200713008 and 200713009), with a tidal deposits infill further downstream (boreholes 200713008 and 200713009). Further east, along the cross-section, there is evidence of another channel belt (borehole 200713011 and again in boreholes 200714113 and 200714003-5). The most eastern borehole of the cross-section represents a residual channel with clay infill and laminated bands of clay, loam and gyttja, that stands for an abandoned, or temporary, loop of the Gazendiep channel. With the exception of the Noorderdiep, all the other channel belts and residual channels mentioned above are covered, and sometimes partially filled, by marine tidal deposits showing that their channels were influenced by tides after the end of their activity period.

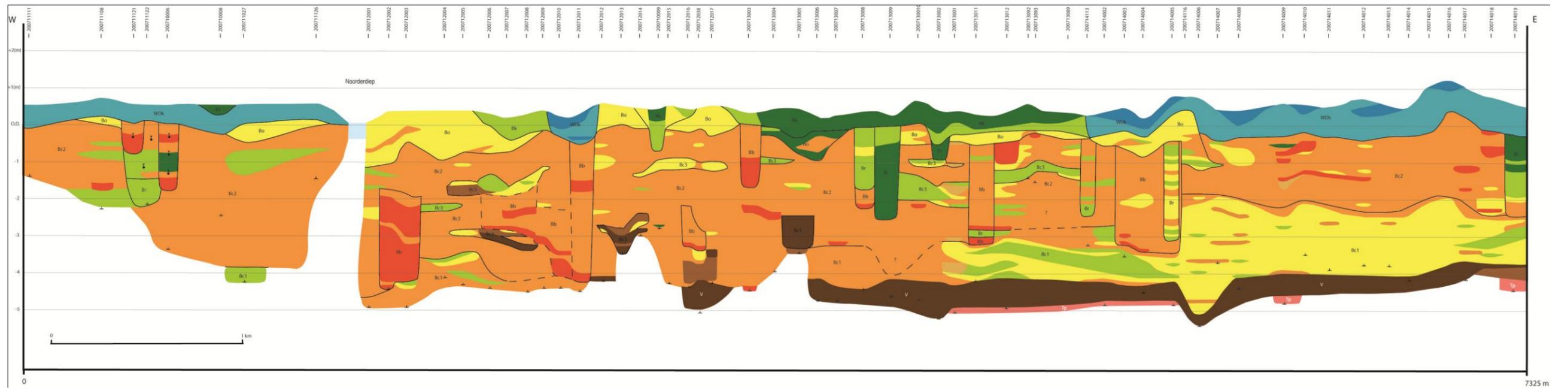


Figure 21 – Cross-section D-D'. For legend see Fig. 18. Lithostratigraphic units described in Table I.

5. SEDIMENTARY ARCHITECTURE AND PALEOGEOGRAPHIC EVOLUTION

5.1 Geomorphogenetic map

5.1.1 Lithogenetic units

Based on the Berendsen (1982) classification, the fluvial deposits (F) in the study area are divided into “geomorphogenetic units” that include channel belt (Fs1-Fs4), residual channel (Fs7; Fs8), alluvial ridge and mouth bar (Fs3), floodbasin (Fk), crevasse-splay (Fc) and dike-breach deposits (Fo1), with a special “unit” for embanked floodbasin sediments deposited between the river and a dike structure (Fu). In the case of a residual channel (Fs7 and Fs8), after discharge stops, the remaining channel is filled in with sand, silt, clay (in the study area usually marine clays) and even peat. The alluvial ridge deposits include fluvial channel sediments (Fs1), as well as deltaic overbank deposits like lateral accretion ridge and natural levee (Fs31; Fs32; Fs3). Crevasse-splay deposits (Fc) have a lobate elongated shape and are formed when the natural levee is breached during floods and coarser materials are transported into the floodplain. During periods of high discharge, the floodbasin is inundated and clays, poor in calcium carbonate, often intercalated by peat, humic materials or plant remains, and sometimes palaeosols, are deposited forming deltaic floodbasin deposits (Fk21; Fk22; Fk23; Fk24; Fk25).

The tidal deposits in Kampereiland are lagunar clays (Lk) deposited in a brackish water perimarine environment that cover deltaic sediments (fluvial loams and sands). On the outer limit of fluvial influence, there are coastal ridges (Lw) that consist of lagoonal clays and loams that emerge as a higher surface topography.

In the southeast of the study area, there are in depth eolian dune deposits formed on the banks of Late Glacial Rivers that influence the lithostratigraphy of the profile. These areas are represented in the geomorphogenetic map as aeolian “river dunes” (Eo).

5.1.2 Geomorphogenetic overview

Near the apex of the delta, the fluvial processes dominate, deposits derived from the action of the river IJssel and its branches in the area during the progradation of the delta mark the upper sections of the lithogenetic profile. In this area, several residual channels are observed, as well as the Gazendiep and the Noorderdiep, which still contain water in sections of the channel. To the east of the Gazendiep the geomorphology of the landscape is related to the presence of Younger Dryas eolian river dunes in the subsurface.

In the mid-section of the map, the lateral accretion deposits, mouth bars (Fs3) and channel deposits (Fs1) are located near the former channels, with the presence of floodbasin deposits (Fk) in the topographic depressions in-between. The extension of fluvial influence is limited by coastal ramparts (Lw) that appear as higher ridges in the landscape topography. This is an area where perimarine influence is also evident: there are brackish water tidal deposits (Lk) near the coastal ridges and, in the western and northeastern extremes of Kampereiland, the marine deposits penetrated deeper inland through the inlets in-between the coastal ramparts and into the floodbasin area (Lk3). The residual channels in this area are filled by tidal sediments deposited when the former Zuiderzee waters penetrated upstream through the depressions of residual channels during high water stages.

On the outer reaches of the coastal ridges the upper profile deposits reflect a lagoonal environment, where marine clays were deposited by the tides of the Zuiderzee in shallow waters (Lk1; Lk2).

One remarkable feature that can be seen in this map is that the former branches of the IJssel in Kampereiland had a tendency to change or slightly divert their course direction to the east, this is explained by the effect of the main western winds that can have a very high intensity in this area, pushing the water to the east and also influencing the main direction of the storm directions in this area. Another explanation may be related to the fact that, during delta evolution, the development of mouth bars and coastal ramparts made higher elevations in the surface inducing the diversion of the channels to a more eastern course due to a more favourable topographic gradient.

5.2 Landscape development

5.2.1 Late Pleistocene

According to Berendsen & Stouthamer (2001), the oldest Rhine sediment deposits date from the Miocene when it was still only a river of relative small dimensions. In the Late Pliocene there was a general regression of the North Sea Basin, the uplift of the Rhenish Massif (Germany) and the Ardennes (Belgium) increased the drainage areas of the Rhine and Meuse. The Pleistocene marked a period in which the whole Netherlands was an extended subaerial delta formed by the rivers Rhine, Meuse, Scheldt, Elbe and Weser, with the apex of the Rhine delta located where the river Rhine leaves the Rhenish schist plateau.

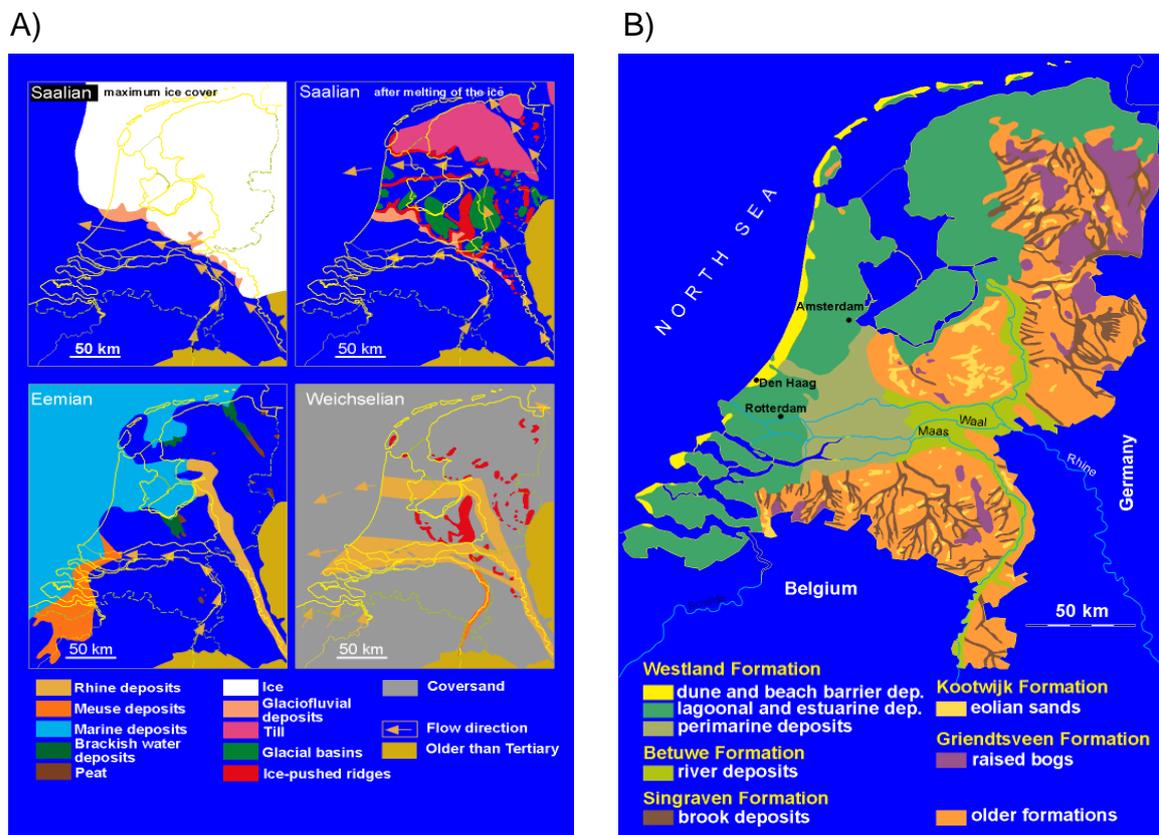


Figure 23 - A) Late Pleistocene palaeogeography of the Rhine-Meuse delta and B) Holocene deposits in the Netherlands (Berendsen & Stouthamer, 2001).

During the Saalian glaciation (~370-130 kyr ago) the ice sheets from Scandinavia reached the Netherlands, north from the line Amsterdam-Nijmegen (Fig. 23). The ice margin consisted of several lobated tongues that created deep scour basins and

pushed coarse-grained sediments, that were deposited in an initial phase of the ice advance, into up to 100 meters high ice-pushed ridges. The advance of the ice mass forced the Rhine-Meuse to abandon its former SSE-NNW course and to shift into a fully proglacial east-west direction that run close to the ice margin, forming a delta that drained to a large proglacial lake delimited by the coalescent British Isles and Scandinavian ice masses (Busschers et al., 2008), joining the River Thames and discharging into the Atlantic Ocean through the Strait of Dover and the Channel area. The Rhine and Meuse maintained the main westward flow direction until the present and another course of the Rhine, through the actual IJssel valley, was also active during the Late Saalian and Eemian (~130-115 kyr ago) warmer period.

As the glaciation reached its terminal phase, the desintegrating ice mass left a network of subglacial scoured basins behind. These basins became lakes and shallow pools as they were infilled by postglacial sediments. After melting of the ice, the IJssel tongue basin acted as a large sediment trap in which the Rhine occupied a northern drainage course. With the Rhine prograding into the IJssel basin, thick sequences of pro-deltaic fine-grained and fluvial medium- to coarse-grained sediments were deposited (Busschers et al., 2008).

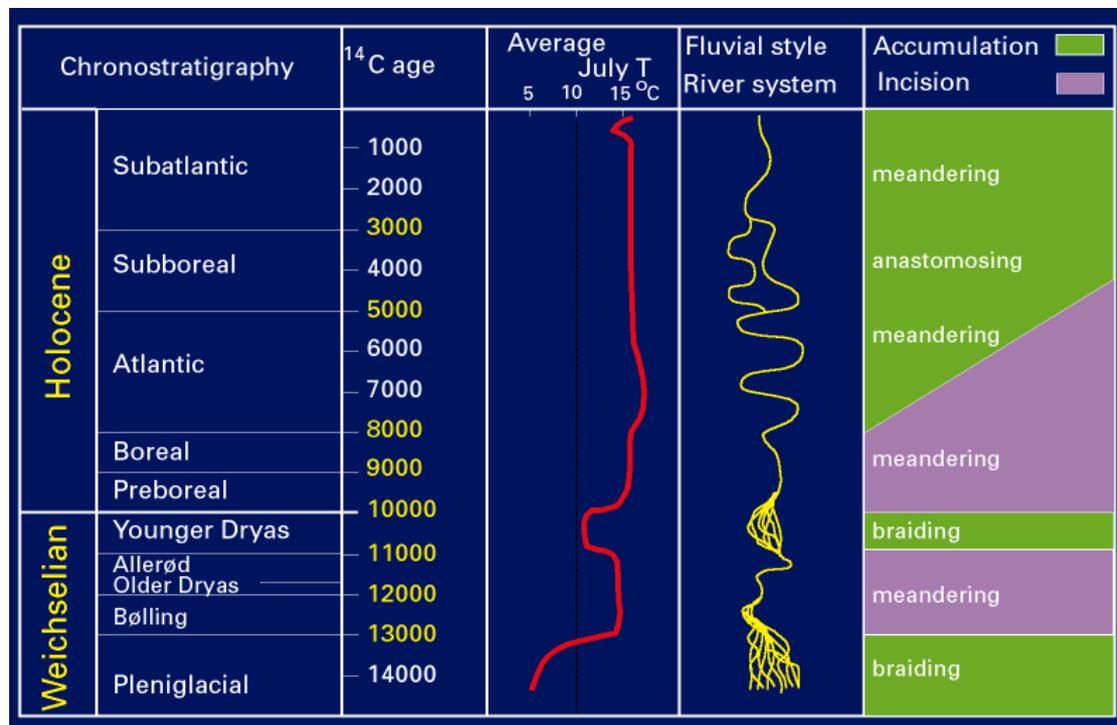


Figure 24 Climate changes and changes in fluvial style during the Late Weichselian and Holocene (after Berendsen et al., 1995).

The early stage of the Eemian is marked by a change to a meandering fluvial system as a response to warming climate conditions. Disappearance of permafrost and vegetation development favoured infiltration upstream, whereas more moderate snowmelt events resulted in a more evenly distributed discharge over the year (Busschers et al., 2007). The consequence was a lower supply of coarser materials and fining of the sediments during this phase. Sea level-rise (RSL) promoted an increase of accommodation space and derived aggradation of the river system and valley-wide sedimentation. The end of the Eemian stage and Early Glacial period were characterized by alternations between cooling stadials with open vegetation and warmer periods of boreal forest development (Busschers et al., 2007).

During the Early Pleniglacial erosion of the underlying Eemian sandy sequences occurred. Channels incised and extensive lateral channel shifting incorporated older sediments and marine mollusc shells into the channel lag deposits. The extensive lateral channel shifting and erosion is probably a result of higher peak discharges associated with spring-snowmelt and an increase of runoff that derived from a reduction of infiltration as widespread permafrost layer was formed in the catchment area. According to Busschers et al. (2007), the accentuated lateral erosion and mobilization of older deposits that occurred during the first phase of the Early Pleniglacial reflects a sediment supply shortage by the Rhine to the downstream areas.

In the first phase of the Middle Pleniglacial the climate was more temperate than during the Early Pleniglacial and, as a result, vegetation partly recovered and soils developed, although not reaching the warmer conditions present during the Eemian-Early Glacial interstadial periods. As a result, sediment supply increased during the early Middle Pleniglacial because sediments were available for mobilization after the intense phase of vegetation and soil degradation that occurred in the Early Pleniglacial cold period.

As a response to an accentuated cooling of climate conditions and increased sediment supply during the second half of the Middle Pleniglacial (after ~50-45 ka BP), the Rhine fluvial system aggraded and shifted. Eventually the Rhine partly avulsed, reacting to a more favorable gradient and concentrated part of its discharge from the IJssel valley to a western course. Eventually, during the Late Middle Pleniglacial to Late Pleniglacial transition (after ~40-35 ka BP), the effect of a differential glacio-isostatic upwarping rebound provoked complete avulsion of the Rhine, which abandoned the northern course through the study area. From that time

on, during the Late Pleniglacial and Late Glacial, sedimentation in the area was controlled by local channel belts, periglacial and eolian processes.

As a response to climate change operating factors, the braided river patterns of the Weichselian Pleniglacial changed to meandering incising systems in the Bølling-Allerød interstadial warmer period (13,000-11,000 ^{14}C BP) and back to a braided pattern in the Younger Dryas stadial (11,000-10,000 ^{14}C BP), when a dry and windy climate reworked river sand and built aeolian dunes (Fig.24). In the Late Weichselian the low-lying areas of the Netherlands acted as a deltaic plain and terraces were formed due to alternating glacial and interglacial periods, with the separating line between net erosion and net sedimentation situated along the present Dutch-German border.

The Weichselian deposits in Kampereiland are evidence of a landscape of tundra vegetation settled on Pleistocene sands (cross-section C-C', Fig. 20). Periglacial deposits are observed in the central area of Kampereiland, deposited in a SE-NW direction, and were transported by braided style inland creeks sustained by ice-melt water (cross-section C-C', lithostratigraphic unit Tp). Aeolian coversands that were blown by the wind during the cold and dry Younger Dryas Stadial are observed on top of the periglacial fluvial deposits or in adjacent areas (cross-section C-C', lithostratigraphic unit Td). During periods of draught, sands were blown from the Late Weichselian channel beds and deposited on the banks (usually in the eastern margins of the streams due to predominant western winds) forming "river dunes" that were imposed as a more elevated topography in the surface, like in the case of IJsselmuiden and Grafhorst (cross-section C-C', lithostratigraphic unit TD).

5.2.2 Holocene

The onset of the Holocene interstadial period (around 10,000 ^{14}C BP) marks a change to a warmer climate, river patterns in the area changed from braided to meandering and aeolian activity decreased, soils formed and vegetation cover became denser. Due to the warming conditions, sea level rise (SLR) led to an increase in groundwater levels that provoked the development of peat when the groundwater level reached the surface during the Late Atlantic (cross-sections B-B', C-C' and D-D', lithostratigraphic unit V).

Hente (1971) stresses that until the Subboreal (around 3950-3450 ^{14}C BP), the sedimentation in the Lake IJssel region is similar to what occurred in the western

Netherlands, marking a marine-perimarine border that shifted eastwards due to a relative sea level rise, with the marine influence reaching the eastern Lake IJssel region and advancing north through the drainage channels established in the former Pleistocene gullies of the peaty perimarine area. The rate of SLR decreased in the Subboreal, resulting in an increase of sandy materials transport by the rivers and leading to the stabilization of the beach ridge system in the Dutch coast and closing of the gap (called “Zeegat van Bergen”) that maintained the connection between the inner lands and the sea (Zagwijn, 1986). The beginning of the Subatlantic is marked by a further decrease in the SLR rate, the peat in the north of the Netherlands was vastly eroded and the Wadden Sea started to develop. In the area of the Zuiderzee Basin the formation of Lake Flevo took place which remained without an opening to the sea during Roman times.

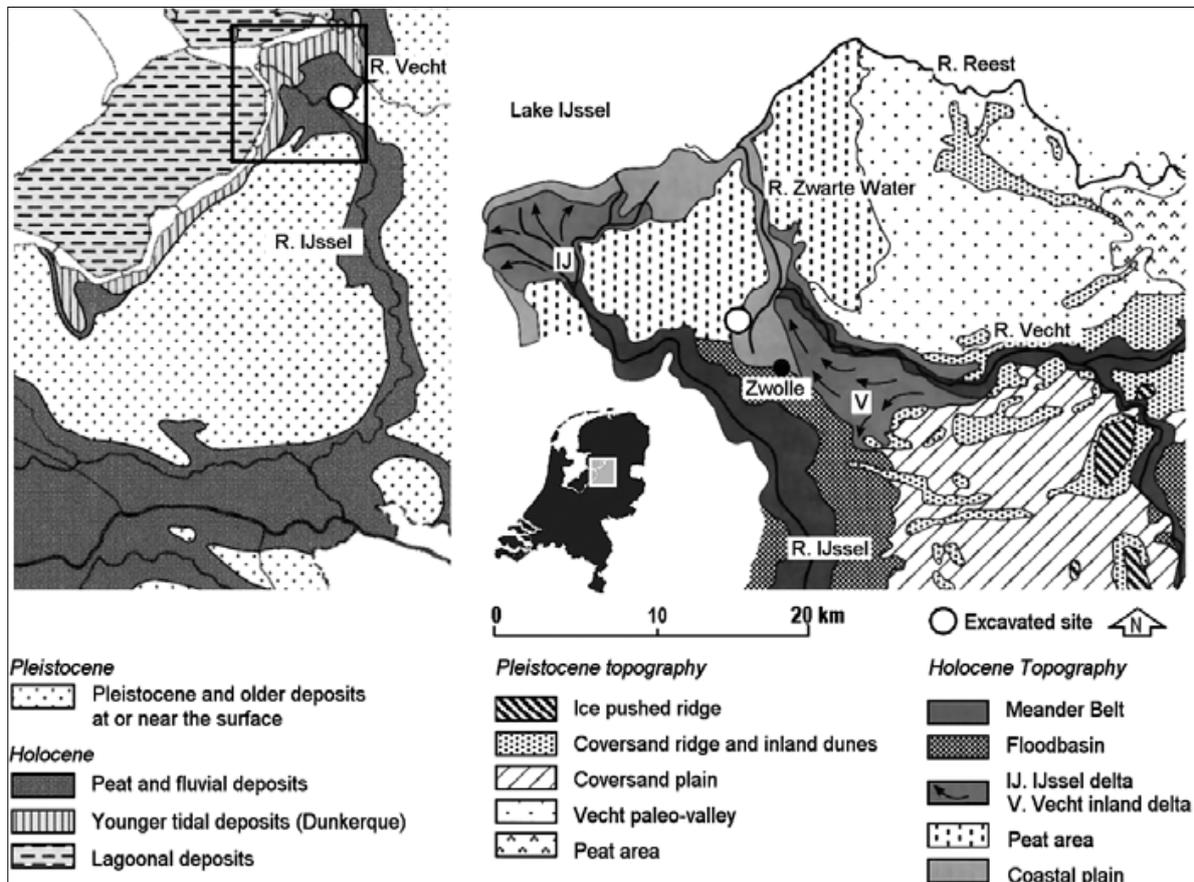


Figure 25 Maps of a) The geological distribution of Pleistocene and Holocene deposits in the centre of the Netherlands after Zagwijn (1991); and b) The geomorphological setting of the northwestern part of the Province of Overijssel (Kooistra et al., 2006).

Thus, during the Subboreal and Subatlantic, there was the establishment of a complex network of creeks and lakes in the Pleistocene gullies of the peat land area, with the Overijsselse Vecht and IJssel channels following the same course of the former Pleistocene rivers (Fig. 25). After the Subatlantic period (2900 ¹⁴C BP) sea level rise and the water flow in the channels on the peat surface led to erosion and displacement of the peat in the central area of Kampereiland that explains the limited thickness of the peat deposits in this area. In the IJsseldelta area, the peat deposits consisted of wood and reed (*Phragmites*) peat that formed under eutrophic freshwater conditions (although *Phragmites* peat can also form under mesohaline conditions) with a water depth of 0 to 0.5 meters (Chambers et al., 1999). Locally, the coversands below the peat were reworked and deposited as sand bars.

From around 1700 ¹⁴C BP, the River IJssel, that still followed the course of the Pleistocene rivers with only a local discharge function, was captured by the Rhine near Arnhem and became a distributary of this river. This resulted in an increased discharge and amount of sediments transported by the IJssel from that period on that enabled the formation of the IJsseldelta after the 12th Century .

5.2.3 Delta formation and development

The IJssel delta was formed and evolved in a relatively very short period of time. In more or less 500 years the delta reached its maximum extent, followed by a dwindling of the delta dynamics until 1932 AD, with the construction of the *Afsluitdijk*, when the *Zuiderzee* was closed and became the Lake IJssel. The development of the IJsseldelta shows a succession and quite fast and dynamic changes in environmental conditions that affected the mouth of the River IJssel, shaping the delta and eventually ending its active role in the landscape evolution.

In a first phase the rivers IJssel and Vecht drained slowly through low gradient marshlands and sedimentation associated to these rivers was mainly composed by clays and some fine sands. After 1000 AD, discharge of the River IJssel increased considerably, changing the volumes of water flowing through the mouth of the IJssel (then situated nearby Kampen) and the characteristics of the sediments transported downstream. The most plausible explanation for this fact is the completion of the avulsion process that fully connected the river IJssel with the Rhine (Makaske et al., 2008). With the increased discharge the IJssel transported a higher quantity of sediments that were also much more sandy and coarser than before. Linked to a

higher discharge of the IJssel, there are historic records that show an embankment of the river from about the 11th. Century, resulting in a narrowing of the river bed and increasing even more the transport of sandy materials (Ente, 1971).

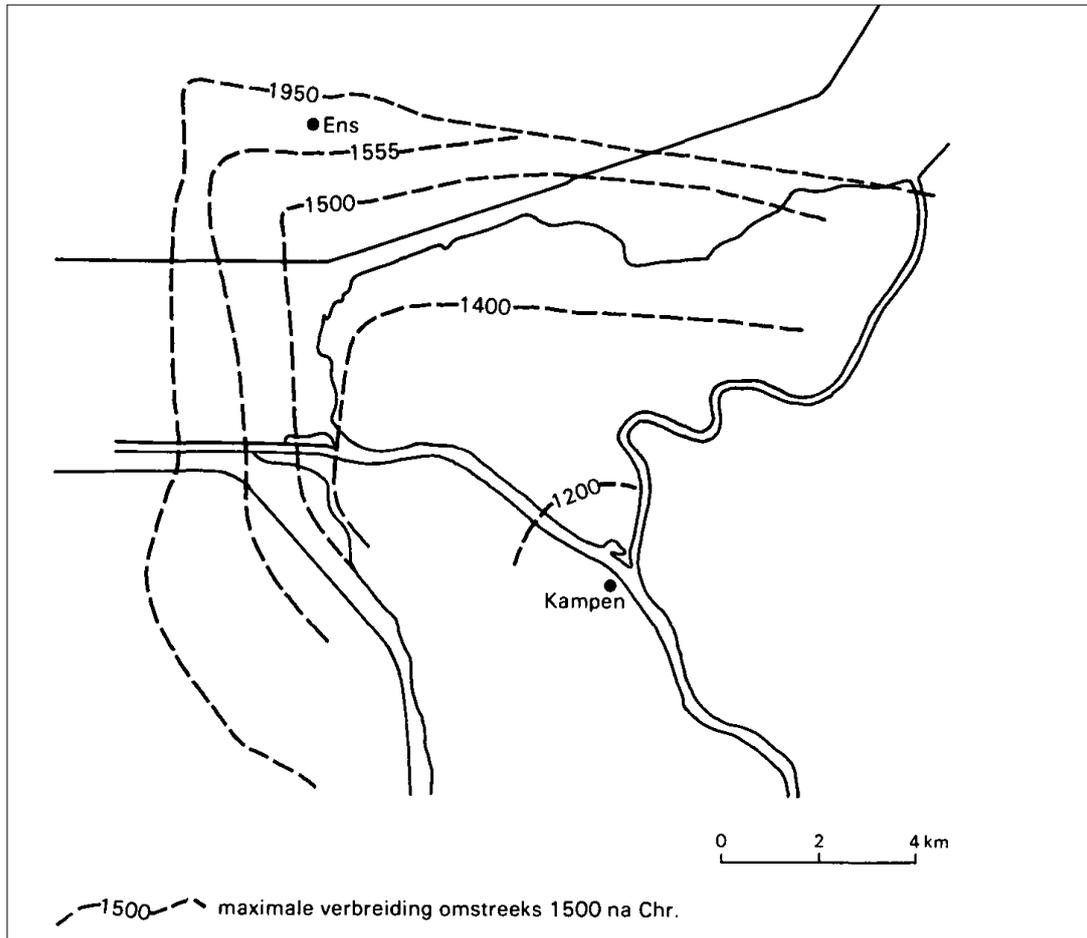


Figure 26 - Maximum extent of river sands in the IJsseldelta, age is given in AD. (Eilander et al. 1990, in Ente, 1973/1974).

The beginning of delta formation is marked by a layer of *Valvata* shells (freshwater gastropod) that was formed in the second half of the 12th Century with the sediments that were deposited above being younger than 1200 AD (Ente, 1974). The environmental conditions for the development of this shell layer report to somewhat still water in an environment dominated by mud, silt and fine sand, where there is some influence of the tides in the Zuiderzee but still dominated by freshwater from the river, where the reworking of the bottom materials was processed mainly by very slow mudflows. It was under these circumstances that a laminated “pro-delta” deposit was formed, representing the underwater vanguard of delta progradation. Among the laminated layers of clays, silts and very fine sand, there are occasional *gyttja* muds (also called nekron mud) that can only be formed in stagnant anaerobic freshwater,

giving a further insight to the environmental conditions during pro-delta formation. The lithostratigraphic unit Bc1 is clearly represented in cross-section D-D' (Fig. 21) and corresponds to the pro-delta deposits, in the cross-sections nearer to the delta apex these deposits are not found because they were displaced and overrun during delta progradation.

During the 13th and 14th centuries (Ente, 1971) a thick package of fine to medium grain sized, calcium carbonate rich sand, called "Ramspolzand" (literally Ramspol sands in English) was deposited on top of the pro-delta sediments by the branches of the IJssel, showing a clear coarsening upwards vertical sequence of the sediments. These delta front sands are the main constituent of the delta, building up the main bulk of delta sediments and being thicker in the center area of the delta than in its margins, as is shown in the constructed cross-sections. The Ramspol sands are referred to as lithostratigraphic unit Bc and Bc2 in cross-section D-D'.

Approximately 1364 AD, a considerably large area of the delta had silted up north of Kampen and it is historical documented that it was given to the city of Kampen by the Archbishop of Utrecht (Ente, 1971), this important record is used as a time mark for delta progradation.

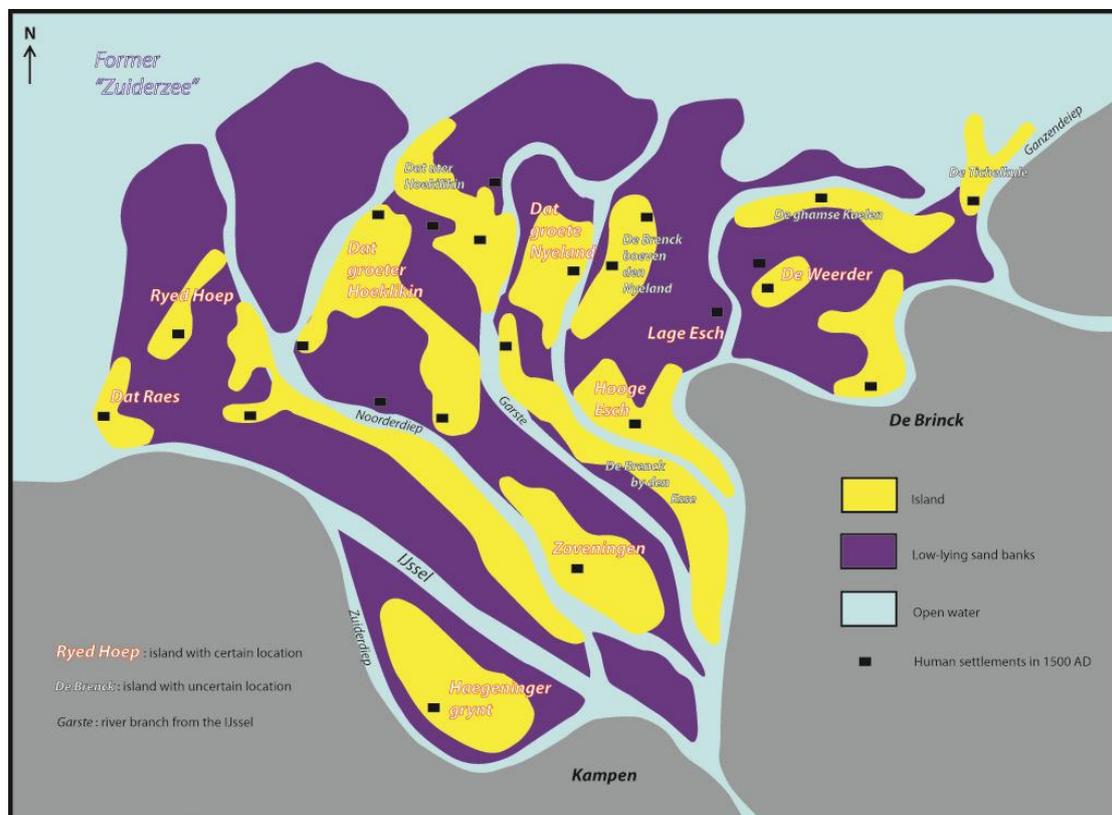


Fig. 27 - Kampereiland in the year 1500 AD. Adapted from Dirx (1996).

During the 13th century, marine influence was very intense in the area southwest of Kampen, there are abundant records of storm surges and floods for this time period. As a result a salt marsh with higher elongated ridges along the coastline evolved in that area. To protect the hinterland near Kampen from the marine influence several dyke were built in 1302 AD just southwest of the Kampen (Fig. 16). During the 13th and 14th century the “Zuiderdiep” (Fig. 10) was the main distributary of the river IJssel (Eilander, 1990).

At 1400 AD the coastline of the IJssel delta was situated more or less half-way north between the city of Kampen and the actual Kampereiland (Fig. 26), and it is believed to have been stable for the next 100 years (Ente, 1971).

Due to the low gradients in the delta, flooding was common and sandy/loamy alluvial accretion ridges were formed in the banks of the channels, representing a higher topography that was seldom used to place human settlements (Fig. 27). It was during this period that the interaction between discharge of the river branches in the delta and action of the waves from the Zuiderzee started to build “breaker zone” coastal ridges in the former shoreline of Kampereiland that represent a higher surface topography than the alluvial accretion ridges (Fig. 22, geomorphogenetic unit Lw). Between the higher coastal ridges, especially in the west and northeast of Kampereiland, there were inlets through which the salt water of the Zuiderzee could penetrate during high tide, creating local brackish water lagoons (Fig. 22, geomorphogenetic units Lk2 and Lk3). In the outer side of the coastal ridges marine clays were deposited and the dynamic processes where induced by the Zuiderzee (Fig. 22, geomorphogenetic units Lk1 and Lk2).

In 1603 AD a v-shaped dyke was built north of Kampen, following the eastern margin of the Noorderdiep and the western margin of the Ganzendiep (Fig. 16). The location of this dyke shows that at this time the Noorderdiep and Gazendiep were active branches of the IJssel and that the channels in between, like the Garste, were now abandoned channels (Fig. 22).

During the 17th. Century the River IJssel lost influence. The comparison between the human settlements in the delta in the year 1500 AD and 1682 AD indicates that the delta was prograding at a slower rate than before (Fig. 27 and Fig. 28). The “Kaart van Overijssel” by N. ten Have (Fig. 14) shows that in 1690 AD the Zuiderdiep,

Ganzendiep, Noorderdiep and the Keteldiep were still active branches of the IJssel but that the main discharge flowed through the Rechterdiep channel.

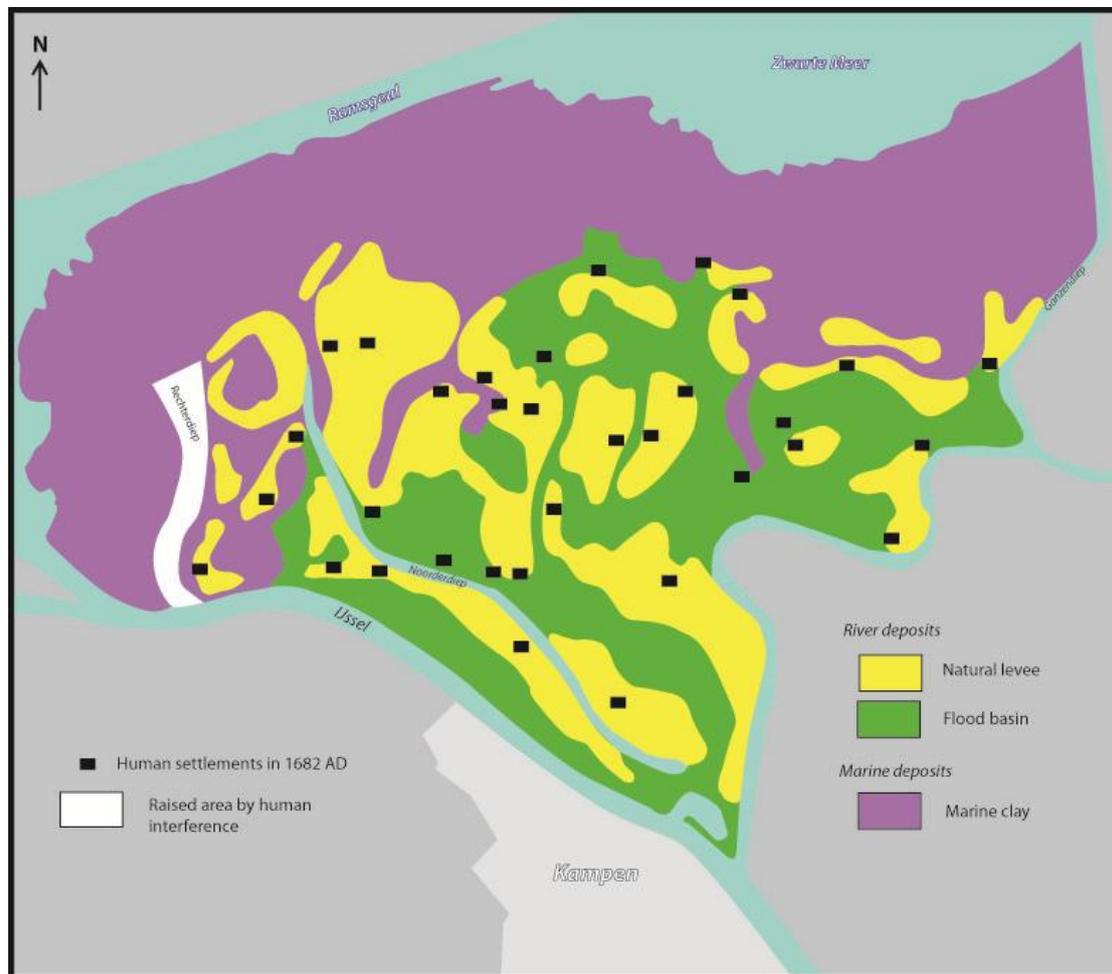


Fig.28 - Human settlements in Kampereiland in the year 1682 AD. Adapted from Dirx (1996).

Approximately 1700 AD the IJssel delta reached its maximum extension, the coastal ridges were fully developed and, in the northeastern extreme of Kampereiland, sediments were deposited by wave action in an already above water sandy platform, extending the coastal ridge further inland (Fig. 22). At this moment the area at the inside of the coastal ridges consisted of an above water delta whereas in the outer reaches river mouth sand banks developed. The only river channels that were still active at this time were the Rechterdiep and the Ganzendiep, the Noorderdiep was inactive although still containing water in the channel, and the residual channels of the other branches were often penetrated by marine waters that filled the former channels with clays.

After 1700 AD more dykes were built in the area and there was an accentuated decrease and eventually an end of delta dynamics in Kampereiland. The most

probable cause for the end of the delta dynamics was the building of the Pannerdens canal in 1701 (finished in 1707), which diverted to other Rhine branches water that before discharged to the IJssel, producing a lower discharge and transport capacity in the River IJssel.

Several marine dykes were built during the 19th century to protect Kamepereiland from the sea (Fig. 16). In 1932, with the construction of the “Afsluitdijk” (meaning Enclosure Dam in English) the former saline “Zuiderzee” became the freshwater Lake IJssel (“IJsselmeer”) and no more marine influence reached the IJsseldelta area. In 1980 a major project, under the supervision of Cornelis Lely, that reclaimed large areas from Lake IJssel, creating the Noordoostpolder and Flevopolder was finished.

6. CONCLUSIONS

The IJssel delta formed in a short period of time, with the time range from the beginning of delta progradation until reaching a final extension only being 500 years. The onset of delta development is set after 1000 AD, when discharge of the IJssel increased substantially as a result of the completion of the Rhine avulsion into the IJssel valley, and more sediment was transported by the river, including coarser materials and more sandy sediments than before. During the 13th and 14th centuries the bulk of delta front sands ("Ramspolzand") was deposited on top of the pro-delta sediments in the area of Kampereiland as the IJsseldelta prograded from the apex near Kampen to the north.

Approximately 1400 AD the coastline was situated more or less half-way between the city of Kampen and the actual Kampereiland, natural levees were formed alongside the channel banks and were often used to build human settlements. The interface action between fluvial processes and wave action start to build up coastal ramparts that stood as a higher topography with an increasing elevation in the "Zuiderzee" direction. In the west and northeast of Kampereiland, during high tides the salt water penetrated inland through tidal inlets creating small brackish water lagoons in the back of the coastal ridges. In the outer side of the coastal ridges mainly marine clays were deposited.

After 15th century and more evidently during the 17th century, the IJssel started to lose relevance in the sedimentation of Kampereiland and progradation of the delta was much slower. In 1700 AD the IJsseldelta was fully formed and reached its maximum extension, the area inside of the coastal ridges became an above water delta and in the outer areas river mouth sand banks developed and marine deposits were formed by wave action silting up abandoned channels and extending the coastal ridges further inland. With the construction of the Pannerdens Canal in 1707 the IJssel lost part of its discharge and delta dynamics ended in the area.

The major asset of this research is the broad scale overview of the IJsseldelta. Conjugating historical data, results from previous researches and the data provided by the BSc students of Physical Geography (Utrecht University) from the 2007 fieldwork campaign, it was possible to build four transversal cross-sections and build a geomorphogenetic map for the whole of Kamperiland, giving a further insight to the lithology, geology, genesis and main processes involved in the IJsseldelta formation and evolution.

The major setback was the difficulty to establish a reliable chronostratigraphy for rates of sedimentation, relative sea-level oscillations and landscape development. AMD ^{14}C radiocarbon dating may be a reliable method to obtain dates for the onset of delta formation when applied to basal peat or to date the beginning of individual channels activity and overbank sedimentation if applied to peat sampled for the base of former channels or from the base and top of natural levees. If more resolution dates are obtained in the future, it will be possible to built more defined time-lines for delta evolution and upgrade the geomorphogenetic map.

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APPENDICES

Appendix 1 – ¹⁴C Analysis

Dating with ¹⁴C is the most common and used method for Holocene fluvial systems. The activity of a fluvial system is set as the time period of sedimentation in the channel belt, with the formation of the respective sandy channel deposits, and the overbank deposits, which include natural levee (clay), crevasse splay and floodbasin deposits (Berendsen & Stouthamer, 2001). Cross-sections are usually used to correctly correlate the overbank and channel deposits.

Berendsen & Stouthamer (2001) set five types of radiocarbon dates that can be used to date fluvial systems:

1 – dates from the top of organic beds underlying clay overbank deposits give the onset of river sedimentation;

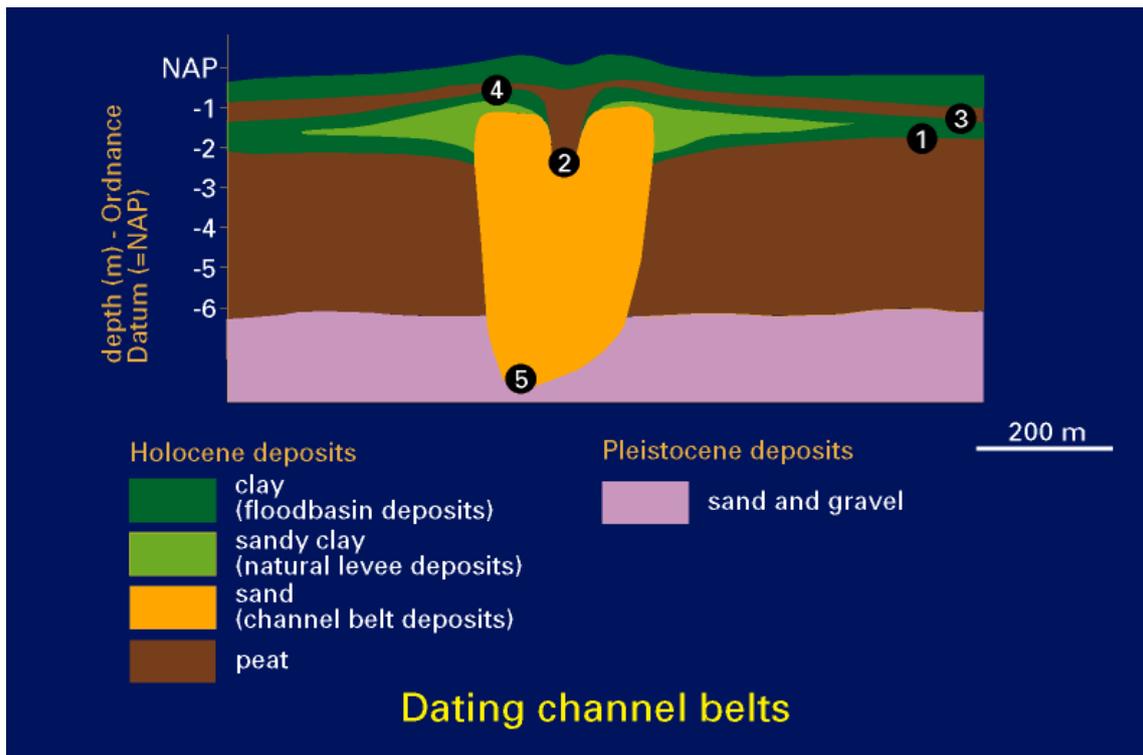
2 – dating the base of organic beds in abandoned residual channels provides the end of within-channel sedimentation;

3 – the date of the base of an organic bed overlying the overbank deposits gives the end of river sedimentation;

4 – peat on top of natural levees is used to date the minimum age of the end of activity of the channel belt (usually with a time hiatus between the formation of the levee and peat formation);

5 – on the other hand, peat from channel lag deposits will provide the maximum age for the beginning of river activity (this peat is often reworked from layers older than the river system)

When this dating method is used, special attention must be taken for the fact that the base of the peat can be diachronous, with the similar material deposit varying in age for different locations, because peat formation starts earlier in the centre of the floodbasin reaching the natural levees. This time hiatus also increases upstream, due to the decrease of upstream aggradation rate, and over time because of a decrease of sea-level rise and floodbasin aggradation rates. As a result, samples from the top of peat layers (with the condition that important peat erosion did not occur in that spot) seem to be more suitable to ¹⁴C dating (Berendsen & Stouthamer, 2001).



Schematic cross-section of a channel belt illustrating the method of radiocarbon dating channel belts (Berendsen & Stouthamer, 2001).

- 1 - ^{14}C sample top of peat in floodbasin: beginning of river sedimentation;
- 2 - ^{14}C sample base of peat in abandoned channel: end of within-channel sedimentation;
- 3 - ^{14}C sample base of peat in the floodbasin: end of river sedimentation;
- 4 - ^{14}C sample of peat on natural levee: minimum age of for end of river activity;
- 5 - ^{14}C sample of peat from channel lag: maximum age for beginning of river activity;

It is also very important when sampling takes place, to avoid locations where erosional contacts were spotted. Using this method, the dates of sedimentary deposits are obtained indirectly, through the dating of organic beds, this means that a temporal hiatus can occur between the end of peat formation and the start of sedimentation or/and between the end of the sedimentation period and start of peat formation.

Radiocarbon dated basal peat can also be used as a proxy to *reconstruct groundwater-level rise* (Cohen, 2005). A series of dated Holocene basal peat samples, with each individual sample directly indicating a paleogroundwater level, can be plotted as groundwater index points in an age-depth graph to build a groundwater-rise curve. Again, the location of the basal peat sampling is a major factor, samples collected on the flanks of Late Glacial aeolian dunes do not suffer the

effects of compaction with the depth of the basal peat representing the regional groundwater-table level at the time of its formation, dates of basal peat overlying the Late Glacial Maximum (LGM) and other terraces can also be used as groundwater index points. The loam of the *Wijchen Member* on top of the LGM terrace indicates a period of active pedogenesis during the Bølling-Allerød and has the same average thickness (40 to 60 cm) all over the Rhine-Meuse delta independently of the overlying deposits, so compaction of this layer can be neglected and the basal peats can be used as groundwater index points like in the case of samples collected in the aeolian dune flanks. The *Wijchen Member* layer, because it represents a pedogenetic phase, also is an indicator of a temporary situation when the groundwater levels were in average 50 cm below the surface (Cohen, 2005). The rise of groundwater levels in the Holocene resulted in the increase of accommodation space, it was faster at the onset of the extensive basal peat formation around 7200 cal. Yr BP and then continued slower with a decreasing rate after 5500 cal. yr BP (Cohen, 2005).

¹⁴C dating would be a major asset to reconstruct the evolution of the IJsseldelta area, giving precise dates to the start and end of activity in the different river branches of the IJssel in Kampereiland. Several key locations for ¹⁴C dating were spotted and can be used in further researches.

Proposed borehole locations for radiocarbon dating

Cross-section	Borehole	Coordinates (XCO; YCO)		Peat type	Peat top depth (meters -mv)
B-B'	200716002	188953	510035	Vo	4.6
	200716011	189739	510702	Vb	4.9
	200716014	189953	510895	Vb	4.4
C-C'	200711038	186404	511665	Vr	4.9
	200711044	187536	511805	Vr	5.1
D-D'	200713012	190411	512450	Vo	4.7
	200714004	191431	512808	Vo	4.6
	200714009	192156	512977	Vb	4.7