The Rhine-Meuse delta at a glance

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Faculty of Geosciences
Department of Physical Geography
Utrecht University

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The author

H.J.A. (Henk) Berendsen studied Physical Geography at Utrecht University, and graduated in 1973. He became a lecturer at the same University in 1973, and associate professor in 1985. In his Ph.D. thesis (1982) he studied the fluvial geomorphology and sedimentology of the Rhine-Meuse delta. He wrote four academic textbooks (in Dutch) on the Physical Geography of the Netherlands, a course he has been teaching since 1973, in addition to geomorphology of rivers and deltas, and Quaternary Geology.
His research is focused on the evolution of the Rhine-Meuse delta (the Netherlands). Over the past decade, he supervised 9 Ph.D. students working in the Rhine-Meuse delta: Torbjörn Törnqvist (1993), Henk Weerts (1996), Hans Middelkoop (1997), Nathalie Asselman (1997), Bart Makaske (1998), Esther Stouthamer (2001), Annika Hesselink (2002), Jeroen Schokker (2003), and Kim Cohen (2003). In addition, he has been involved in the Ph.D. studies of Jaap Kwadijk (1993), Marc Bierkens (1994), Jacob Wallinga (2001), and Giacomo Biserni (2004). Five more Ph.D. studies are in progress.
In 2001 the monumental 'Palaeogeographic development of the Rhine-Meuse delta, The Netherlands' was published, with Esther Stouthamer as a co-author. In 2005 he was awarded the 'Van Waterschoot Van der Gracht' medal, the highest award in Earth Sciences in the Netherlands.

Contents

Preface

This booklet was made for fluvial sedimentologists who want to get an overview of the Rhine-Meuse delta in one day.
The present excursion guide has been completely revised for the International Conference on Fluvial Sedimentology (ICFS) in Delft (2005), and is mainly based on Berendsen & Stouthamer (2001). For references and more details, the reader is referred to their work and the internet site: http://www.geog.uu.nl/fg/palaeogeography.
Figures (in color), animations, and power point presentations can be downloaded from this Internet site.

Dr. H.J.A. Berendsen

1. Introduction

The Rhine-Meuse deltaic plain

The Rhine-Meuse deltaic plain is situated in the southeastern corner of the North Sea Basin (Figure 1), where it forms part of a coastal plain of varying width, which runs from the Strait of Dover in the south to Denmark in the north. The shoreline of the Holocene Rhine-Meuse delta is basically retrograding, and its convex shape is at least partly a Pleistocene relict.

Holocene highstand delta overlies Pleistocene alluvial plain

Because of the morphology of the North Sea Basin the lowstand deltas of the Rhine-Meuse system occur far seaward of the highstand deltas. Consequently, the Holocene coastal prism occurs not on top of the preceding lowstand delta, but overlies alluvial plain sediments of early to late Pleistocene age on top of a delta front sequence of early Pleistocene age. Most of the alluvial plain sediments represent deposition during glacial lowstands, but thin remnants of interglacial highstand coastal prisms occur in the present deltaic plain. The predominance of lowstand deposition is probably largely due to the strong increase in the supply of debris and the discharge of the rivers.

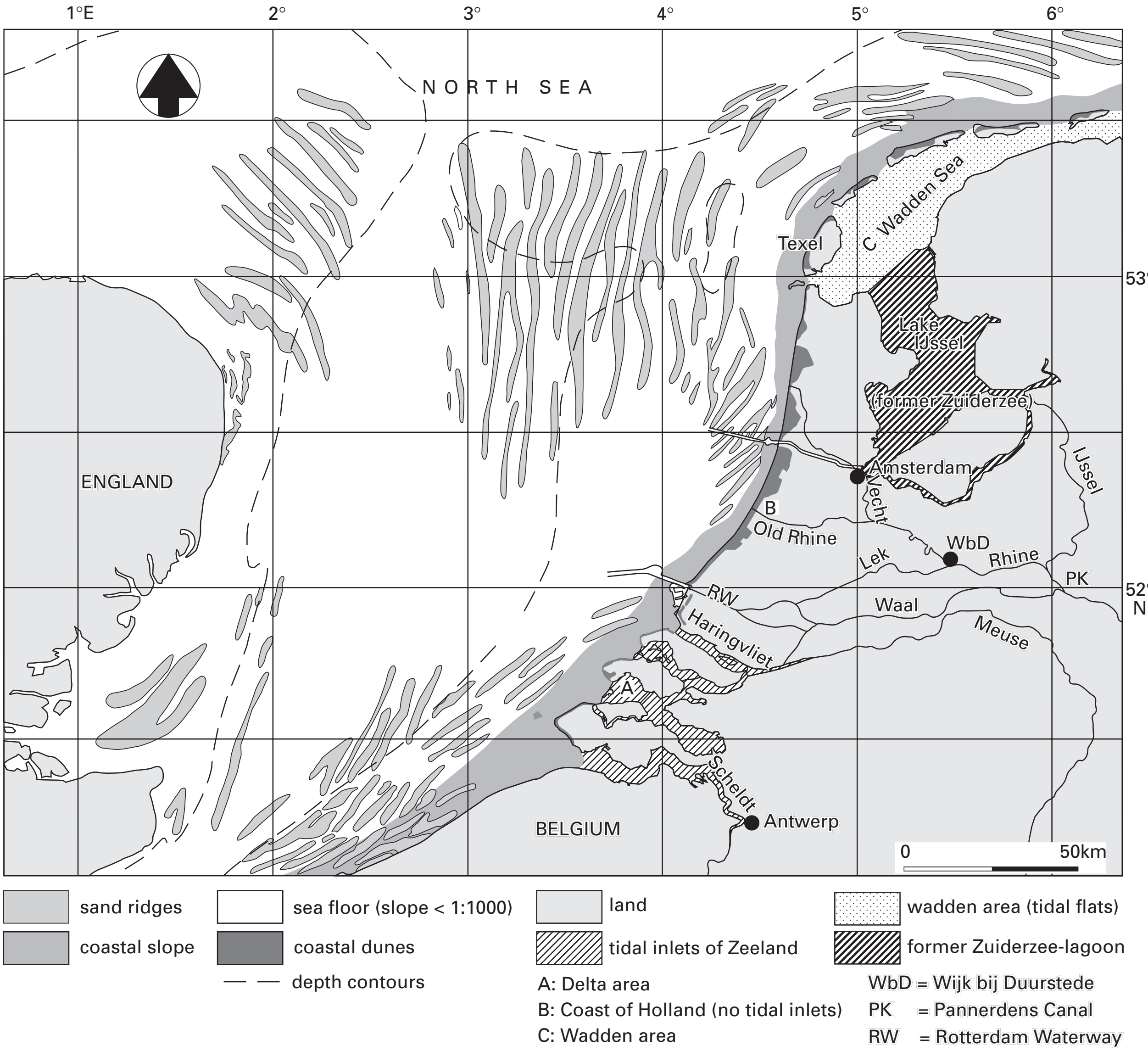

Figure 1 The North Sea basin, and subdivision of the Dutch coast. The delta area (A) is characterized by large tidal inlets. The central coast of Holland (B) consists of barrier beaches with coastal dunes. The northern coast (C) consists of islands and tidal inlets to the 'Wadden Sea', where tidal flats occur. The southern North Sea Basin is shallow. On the sea floor elongated sand ridges occur that appear to be related to tidal currents (after Van de Meene 1994).

During the glacials, the North Sea Basin became the dumping site of huge amounts of glacial debris supplied both by the Scandinavian ice sheet and by the Alpine ice cap by way of the River Rhine.

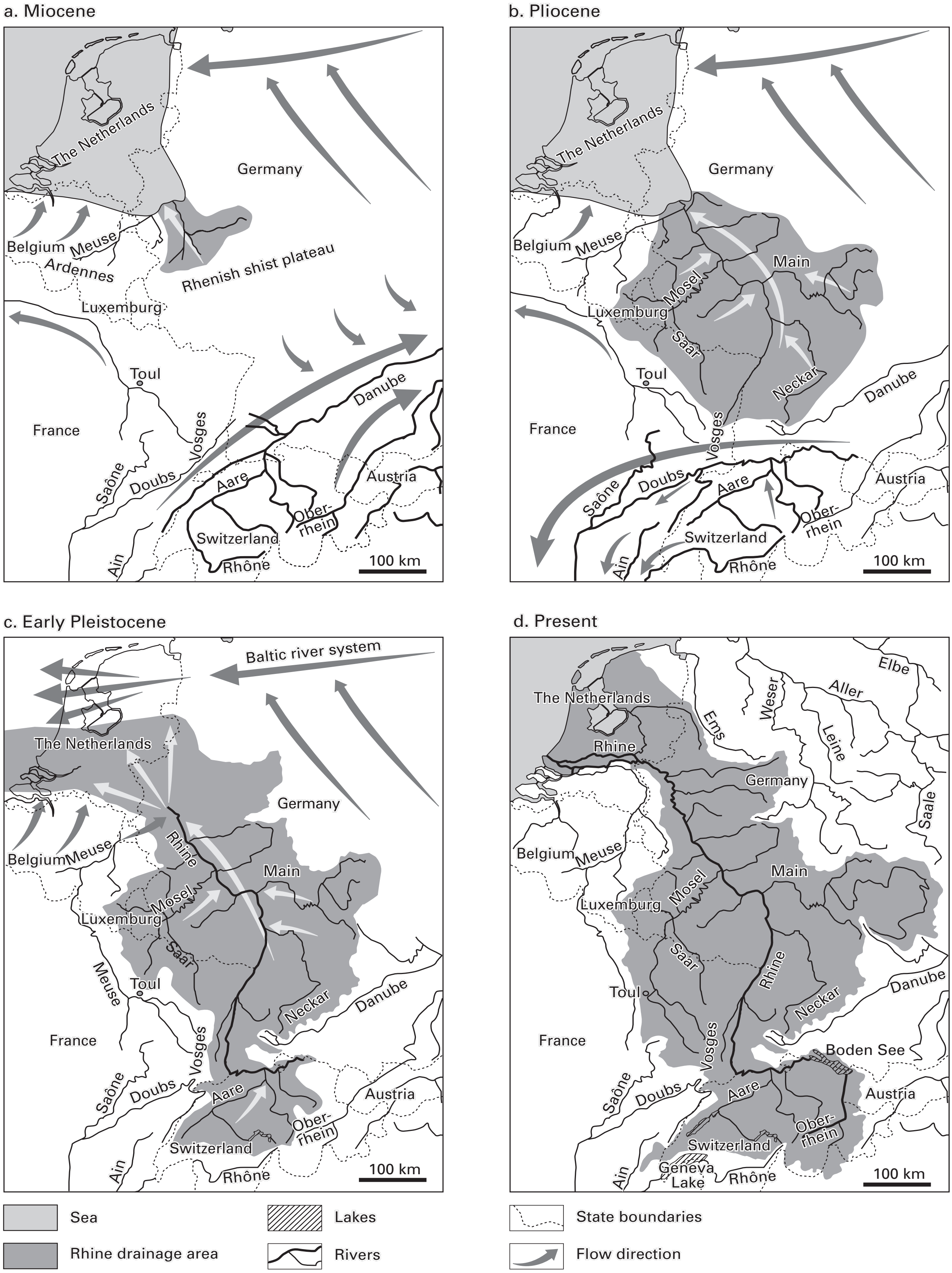

Figure 2 Increase of the drainage area of the Rhine by stream capture. After Berendsen & Stouthamer (2001). Palaeogeographic situation during (a) the Miocene. The Rhine was a small stream, the Danube drained the Alps that were uplifted and folded. (b) Pliocene. The drainage area of the Rhine increased. The Alps drained to the Saône-Rhône. (c) Early Pleistocene. The Rhine extended its drainage area into the Alps. The Meuse extended its drainage into the Vosges mountains. (d) Present. The upstream part of the Meuse drainage was captured by the Mosel during the Saalian glaciation.

It is estimated that peak discharge of the Rhine increased at least tenfold (compared to present annual discharge, Table 1) when draining the Alpine ice cap during cold stages. Hence, peak discharge during glacials was comparable to the present discharge of the Mississippi and Yangtse rivers.

Discharge distribution

The present mean annual discharge of the Rhine is about 2200 m^3/s; that of the Meuse (Dutch: Maas) is smaller by about a factor of ten (Table 1).

Rhine discharge is divided among three distributaries: Waal (6/9 of total discharge), Nederrijn-Lek (2/9) and IJssel (1/9). The Rhine has a mixed snowmelt and rainfall discharge, with a peak discharge in the spring and a smaller peak during the early summer. The Meuse is a typical rainfed river. The Meuse (Maas) has been a tributary of the Rhine during most of geological history.

Table 1 Present discharge in m^3/s of the rivers Rhine, Meuse, Rhone, Mississippi

	Rhine	Meuse	Rhone	Mississippi
Minimum	620	30	360	5600
Mean annual	2200	250	1670	12000
Maximum	13000	3000	13000	56000
Min/max ratio	1:20	1:100	1:36	1:10

Virtually all of the Holocene delta has a subaerial origin, and was formed by river sedimentation in a back-barrier area, or by peat formation. The thickness of the Holocene clayey floodbasin deposits varies from about 2 m near the German border to about 25 m near the Dutch coast, where thick peat layers occur intercalated with clay.

2. Tertiary and Pleistocene evolution

The North Sea Basin formed because of Mesozoic stretching related to the opening of the Atlantic Ocean. Subsidence increased considerably in the Quaternary. A Quaternary sequence of up to 1000 m thickness formed in the central part of the North Sea area.

Miocene origin of the Rhine

The oldest sediments of the Rhine date from the Miocene when it was a small stream (Figure 2a) draining the Graben of the Lower Rhine Embayment. Marine deposition predominated in the Netherlands until the Early Pleistocene (2.5 million years ago, Figure 3).
General regression at the site of the southern North Sea basin set in during the Late Pliocene (Figure 4a). Neogene uplift of the Rhenish Massif (Germany) and the Ardennes (Belgium) led to increasing drainage areas both for the Rhine and Meuse (Figure 2b, c), and during the Pleistocene practically the entire Netherlands became part of a subaerial delta formed by the rivers Rhine, Meuse, Scheldt, Elbe and Weser (Figure 4). From a geological point of view, the apex of the Rhine delta is located near Bonn (Germany), where the Rhine leaves the Rhenish Massif and enters the North Sea basin.

At the Pliocene/Pleistocene boundary the Rhine captured some main tributaries of the Saône-Rhône, e.g. the river Aare (Switzerland, Figure 2b, c), and extended its drainage area in the Alps. This is demonstrated by the heavy

mineral composition (i.e. the presence of saussurite, derived from the Alps) of the Pleistocene Rhine sediments. At the same time the Meuse extended its drainage area (Figure 2c) into the Vosges (France). This is reflected by the presence of granite, porphyr and 'Vosges'-hornblende in the Meuse terraces in the southern Netherlands.
During the Pleistocene Holsteinian (marine isotope stage 11; Figure 3) the Oberrhein (until then a tributary of the Danube) was captured by the Rhine (compare Figure 2c and d). The Saalian glaciation finally diverted the upper course of the Meuse via the Mosel to the Rhine. During the Saalian, the Netherlands became covered by ice from the Scandinavian ice sheets, north from the line Amsterdam-Nijmegen (Figure 5a).

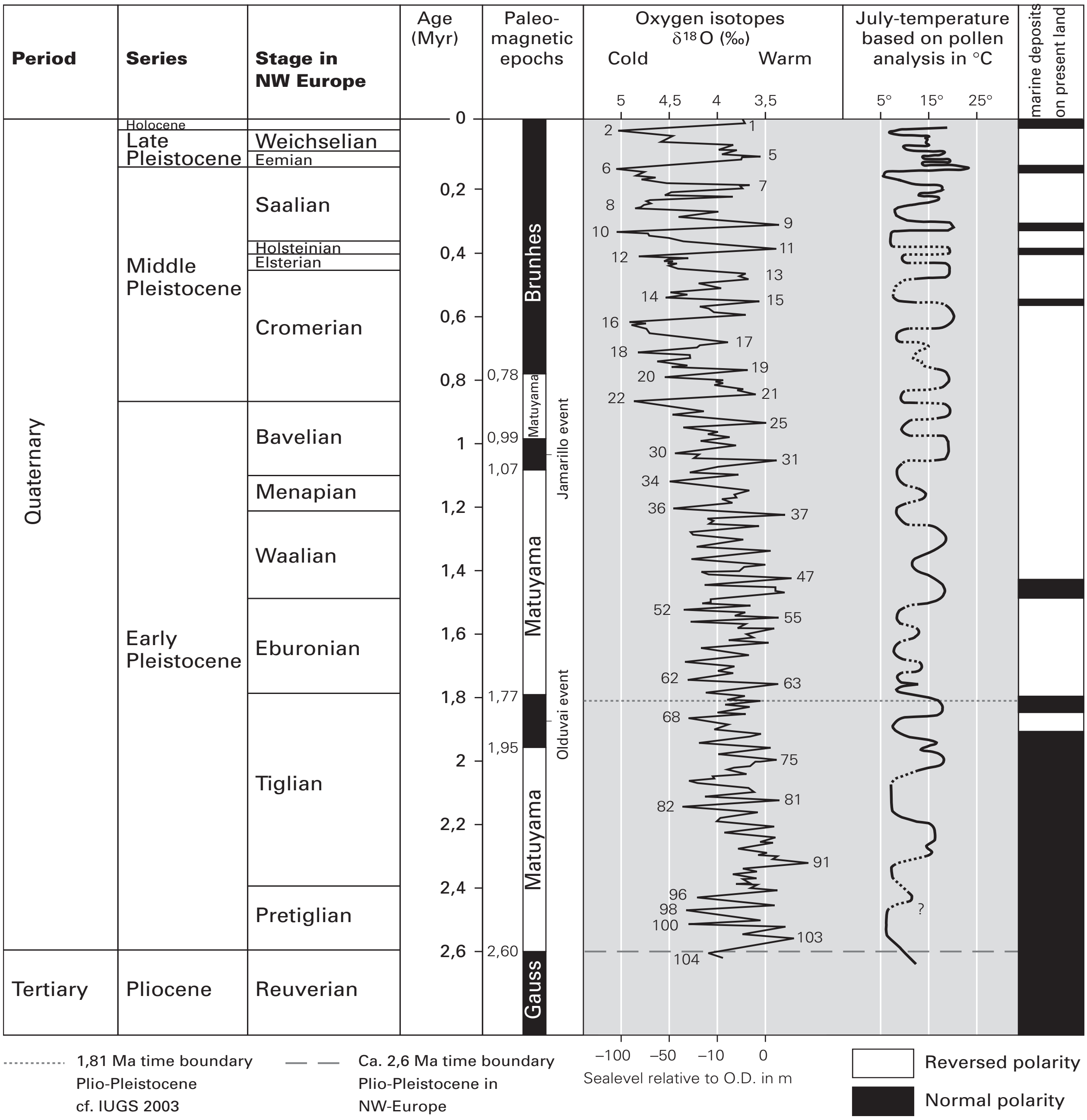

Figure 3 Pleistocene stratigraphy (after Berendsen 2004). Average July temperatures are essentially based on pollen analysis. Correlations with the marine oxygen isotope stages are provisional.

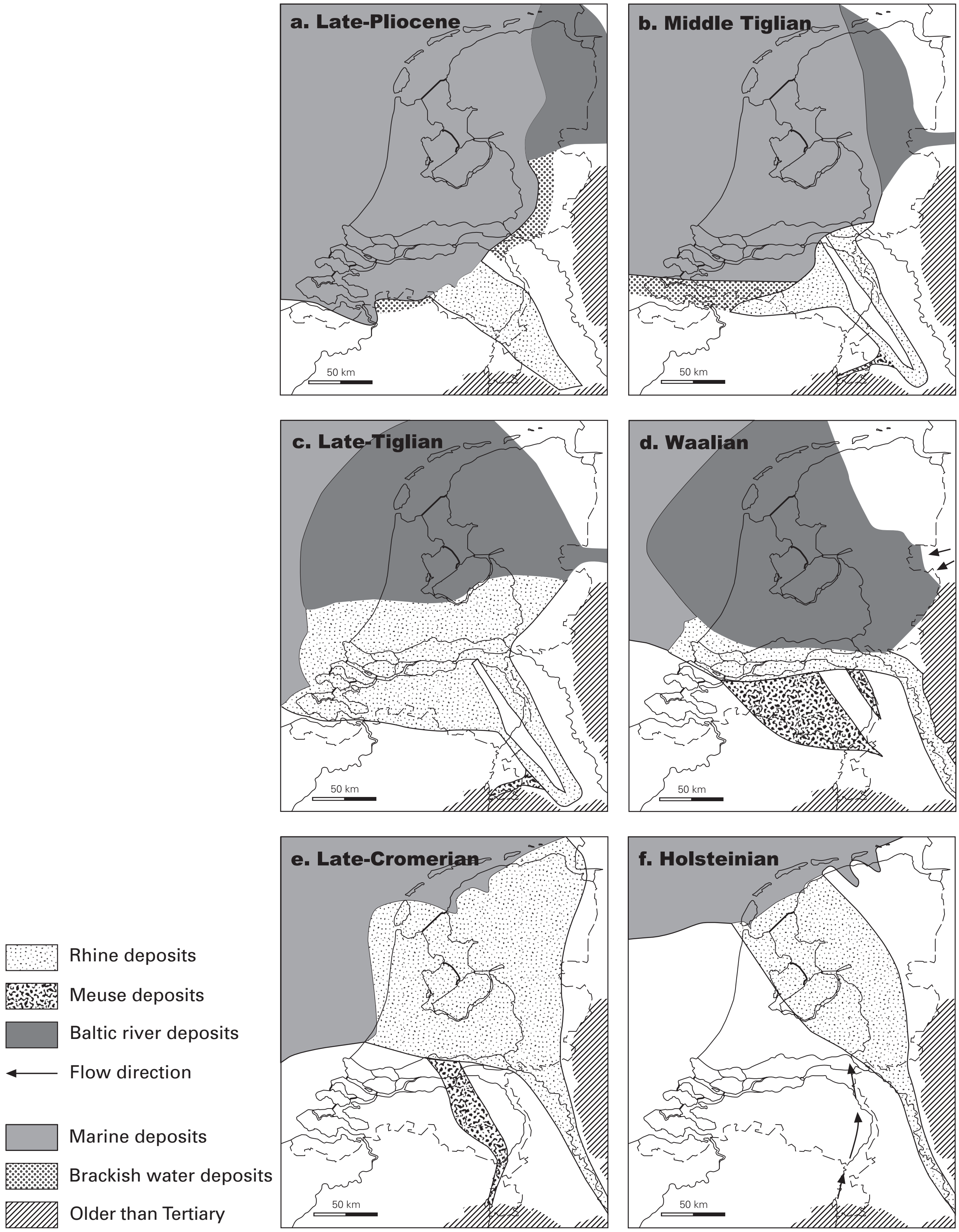

Figure 4 Palaeogeographic situation in the Netherlands during the Early and Middle Pleistocene. After Zagwijn (1974). The coastline changed from concave to convex in the Late-Tiglian as a result of a high supply of debris, even though the Roer Valley Graben and the Lower Rhine area were subsiding. After the Waalian supply by Baltic rivers ceased.

The Saalian ice cap significantly altered the landscape, forming 100 m high glaciotectonic ridges (Figure 5b) that are still important elements in the landscape. At present the ice-pushed ridges still partly control the width of the fluvial plain, thereby determining the areal extent of the Holocene delta.

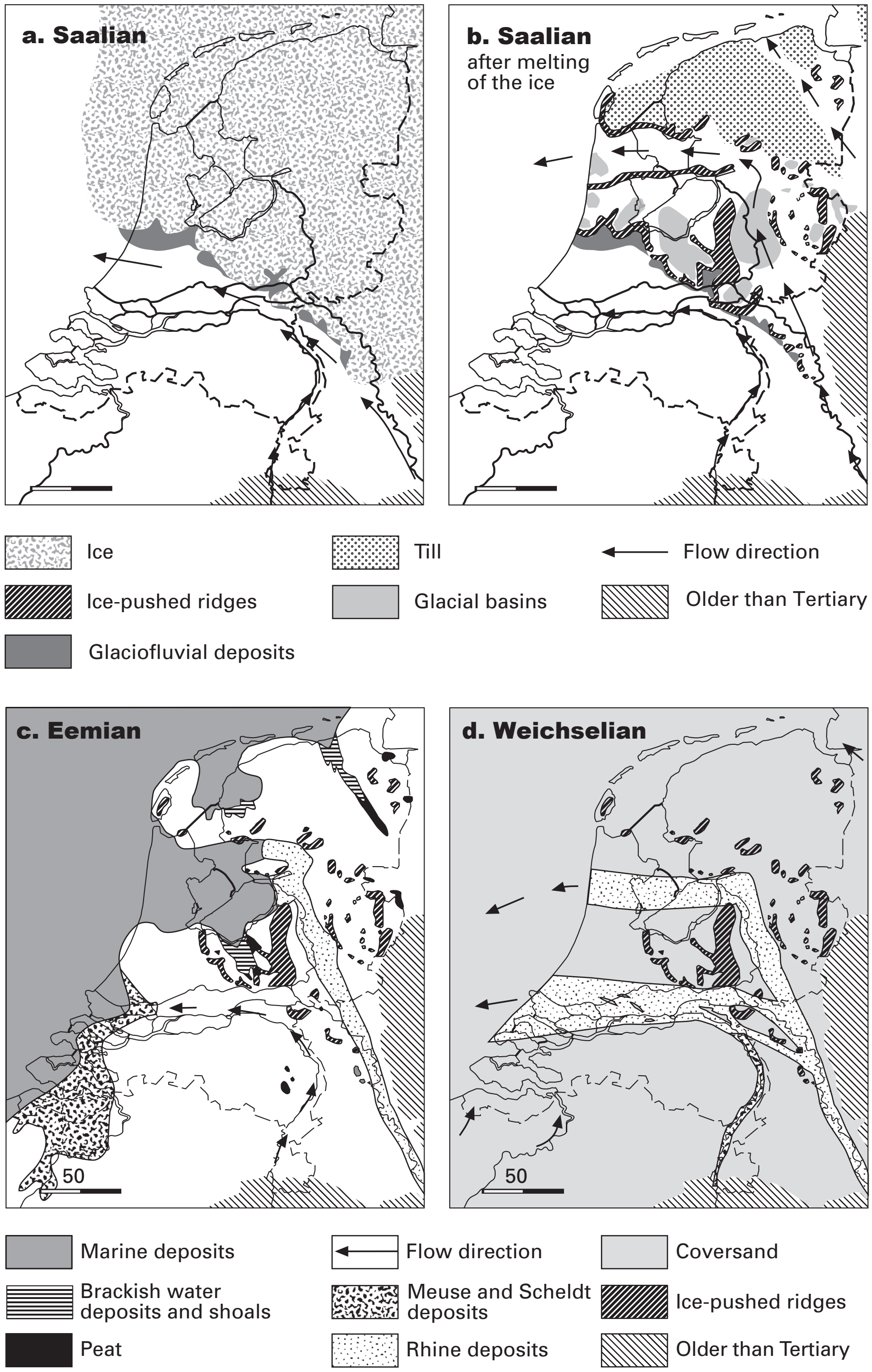

Figure 5 Palaeogeography of the Rhine-Meuse delta during the Late Pleistocene (essentially after Zagwijn 1986). During the Saalian, half of The Netherlands became covered by land ice, and river flow became diverted. The ice formed 100 m high glaciotectonic ridges, that are still visible in the present landscape. During the Eemian interglacial, marine deposits were laid down in former glacial basins, and in the downstream parts of river valleys. During the Weichselian, most of the Netherlands became covered by eolian sands (so-called coversands).

The Elsterian and Saalian ice fronts forced the rivers Rhine and Meuse to shift their northern course westwards, where they joined the Thames and debouched into the Atlantic Ocean through the Strait of Dover and the Channel area (Figure 6). Since that time, this has remained the main course of the rivers, although the Rhine also had a course through the present IJssel valley (Figure 1) during the Late Saalian and Eemian (Figure 5b, c). The Meuse has been a tributary of the Rhine during almost the entire Pleistocene and Holocene.

Weichselian

During the Early Weichselian, the Rhine and Meuse flowed westward through two valleys that are still visible in the morphology of the Late Weichselian surface (Figure 5d). The northern course through the IJssel valley became abandoned during the Weichselian Pleniglacial.
By the end of the Weichselian glaciation (in which the ice sheets did not reach the Netherlands, Figure 6d) most of the Netherlands became covered with eolian sands (Figure 5d).

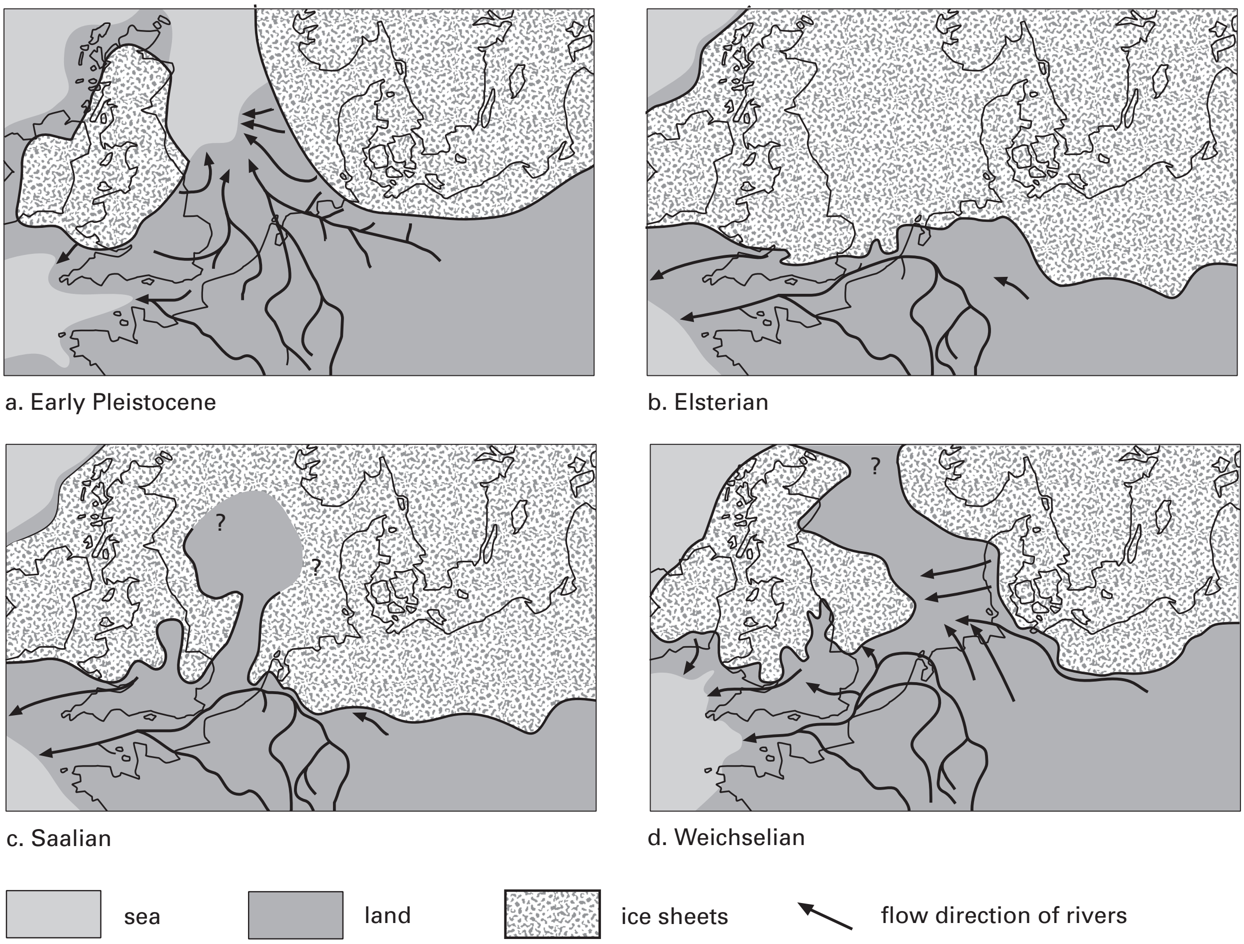

Figure 6 Palaeogeographic situation in the North Sea Basin at four moments during the Pleistocene (Berendsen 2004). (a) Early Pleistocene: ice caps on Scandinavia and the Bristish Isles are not in contact. Rivers drain to the northwest, into the Atlantic Ocean. (b) Elsterien: ice covered the northern part of the Netherlands; rivers were forced to flow to the southwest, through the Strait of Dover. (c) Saalian: ice covered half of the Netherlands. It is uncertain whether the ice caps on Scandinavia and the Bristish Isles were in contact. (d) Weichselian: a large part of the North Sea was ice-free. River drainage through the Strait of Dover was maintained.

During the Weichselian Pleniglacial, sealevel was approximately 120 m lower than today. In the shallow North Sea area, rivers had a comparatively low gradient. In the vicinity of the present coastline downcutting resulted in deep Late Weichselian (= Weichselian Late Glacial) valleys that influence the morphology of the coastline and the course of the rivers until the present. When the glacier ice began melting, around 18,000 yr BP, sealevel rose, but because of the relatively high elevation of the present-day southern North Sea area, this rise started to affect the present coastal area only comparatively late (around 8000 yr BP). Weichselian Pleniglacial braided rivers changed to meandering incising streams in the relatively warm Bølling-Allerød interstadial (13,000 - 11,000 yr BP), and back again to a brief braiding (incising) stage in the Younger Dryas stadial (11,000 - 10,000 yr BP), Figure 7.

Younger Dryas eolian dunes

This, in combination with a dry, windy climate, enhanced eolian reworking of river sand and gave rise to extensive dune formation on the river plain. These eolian dunes are generally referred to as 'river dunes' in the Dutch literature, which is a rather ambiguous term that could easily lead to the misunderstanding that these dunes were formed under water on the river bed. These eolian dunes are up to 20 m high, and often are still visible in the present-day landscape. Because of their relatively high elevation, they became the site of early settlements.

Late Weichselian river terraces

While the North Sea Basin was subsiding, the area between Bonn (Germany) and the Dutch border was uplifted during the Quaternary. Here, alternating glacial and interglacial conditions produced a series of terraces. The hinge line between net erosion and net sedimentation practically coincides with the Dutch-German border, which means that from a geomorphological point of view, almost the entire Netherlands can be seen as a deltaic plain.
At present, the hinge line virtually coincides with the terrace intersection. However, the terrace intersection shifted over considerable horizontal distances in response to sealevel changes that are related to the glacial cycles.

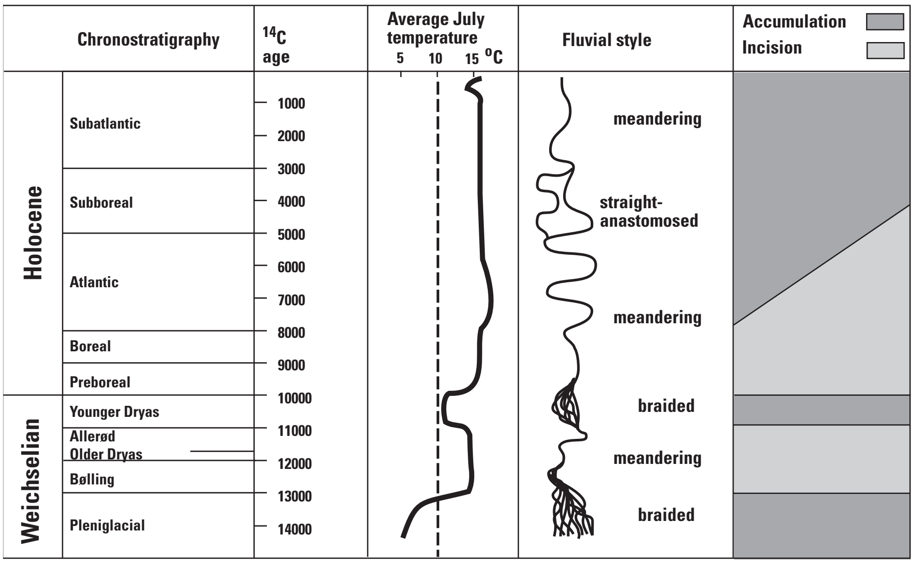

Figure 7 Climate changes and changes in fluvial style during the Late Weichselian and Holocene (after Berendsen, Hoek & Schorn 1995).

3. Holocene evolution

The Holocene palaeogeographic evolution of the Rhine-Meuse delta has been extensively studied by Berendsen & Stouthamer (2001), based on over 200,000 lithological borehole descriptions (collected over 30 years of research by the department of Physical Geography at Utrecht University), 36,000 dated archaeological sites and 1250 ^{14}C dates.

Factors influencing the Holocene evolution

For the Holocene evolution of the fluvial and coastal plain the following factors are of prime importance:

- The *morphology of the Pleistocene (sub)surface*, consisting of a Late Weichselian incised valley, bordered in the north by ice-pushed ridges. This influenced Holocene evolution up to the present.
- *Sealevel rise* and *subsidence*. Relative sealevel rise is the result of eustatic sealevel rise, isostatic subsidence of the land, tectonic subsidence of the North Sea basin and the Roer Valley Graben and compaction of underlying sediments. Subsidence amplified eustatic sealevel rise, leading to preservation of Early Holocene deposits in the western part of the delta. This may be characteristic for deltas in former forebulge areas.
- *Climate change* (resulting in changes in discharge and sediment load). Inundations of the rivers gave rise to floodbasin deposition and peat formation in the back-barrier area of the coastal plain. Peat formation occurred where the accumulation space, created by the rapid rise of the groundwater table, could not be filled with clastic deposits.
- The presence of (Pleistocene) sands in the shallow North Sea bottom, which became available for the building of barrier beaches and coastal dunes. The barriers were important for the preservation of Holocene deposits in the back-barrier lagoon.
- The tidal difference, which is approximately 2 m along the central part of the coast. Tidal differences have not changed much over the last few thousand years, but tidal influence was much more important in the Meuse mouth (where discharge was low, and large estuaries developed), than in the Rhine mouth, which had a greater water and sediment discharge.
- *Human influence*. Since approximately 1000 AD rivers are embanked, small distributaries have been dammed and deposition is limited to the embanked floodplains of only three remaining distributaries (Waal, Nederrijn and IJssel).

Local tectonic influence

The Holocene rivers follow an E-W course and cross the SE-NW trending tectonic structures of the Roer Valley Graben, Peel Horst en Venlo Graben (Figure 8). The Peel Boundary fault, bordering the Roer Valley Graben and Peel Horst is the most active fault. In 1992, it gave rise to an earthquake of magnitude 5.9 on the Richter scale. Differential subsidence at the Peel Boundary fault is approximately 2 m over the last 14.000 years (Cohen 2003). Gradient lines of channel belts have been deformed by post-depositional neotectonic movements (Stouthamer & Berendsen 2000), but syn-depositional effects on the rivers can also be detected (e.g. formation of an asymmetrical channel belt, alignment of meanders to faults, differences in the rate of groundwater level rise). These are local features that are specific to the Rhine-Meuse delta.

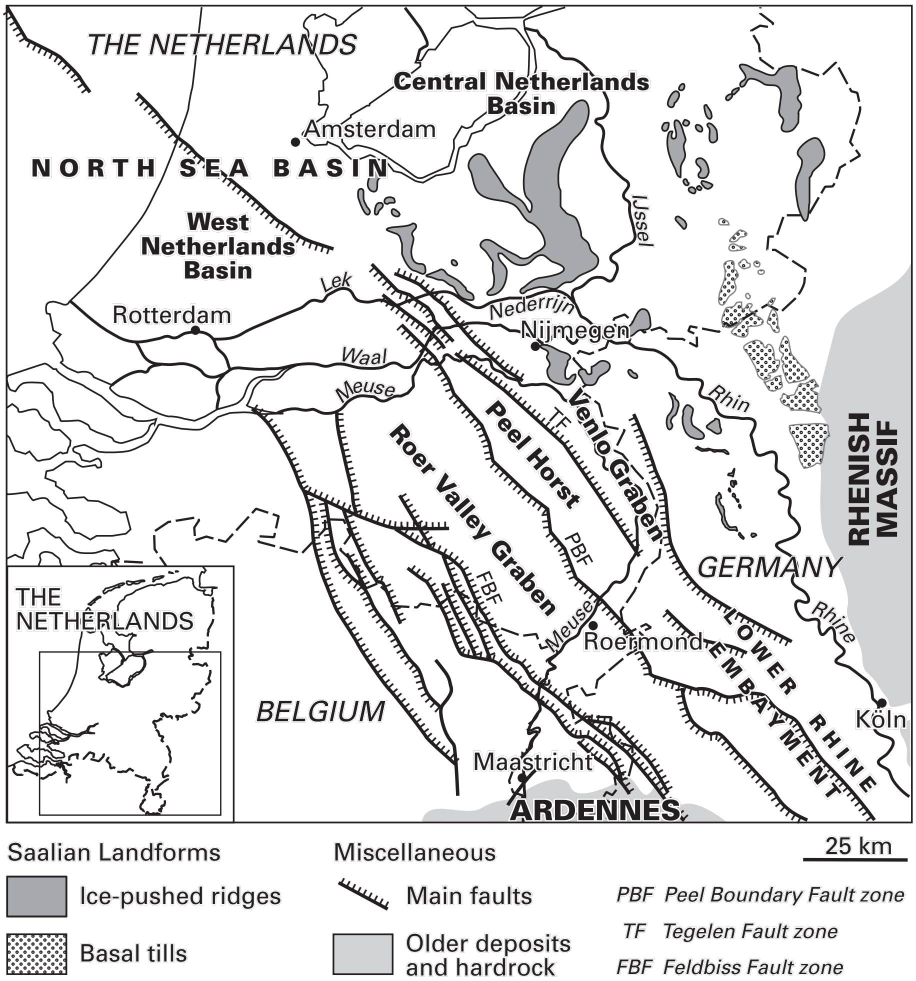

Figure 8 Tectonic elements in the central Netherlands (Cohen 2003). The Peel Horst is relatively uplifted compared to the Roer Valley Graben and the Venlo Graben. The Peel Boundary Fault is the most active fault in the Netherlands.

Subdivision of the Holocene Rhine-Meuse delta

Subdivision of the Holocene delta

The Holocene Rhine-Meuse deltaic plain can be subdivided into (Figure 9):

- The fluvial area, characterized by meandering rivers, 1 - 2 km wide channel belts and relatively small floodbasins.
- The back-barrier coastal plain with strong fluvial influence, sometimes called the ‘perimarine’ area. This area is characterized by narrow, low-sinuosity meandering and straight anastomosing channel belts with large crevasse splays. Floodbasins are large and contain thick layers of peat.
- The estuarine and back-barrier tidally influenced area, with intertidal deposits of Atlantic and Subboreal age, covered with peat.
  Where peat has been excavated for fuel and salt extraction, lakes occur. Since approximately 1450 AD many of these lakes have been pumped dry. In that case Atlantic tidal deposits occur at the surface.
- The barrier beach and coastal dune area. The oldest preserved barrier beaches with low dunes are located farthest inland.

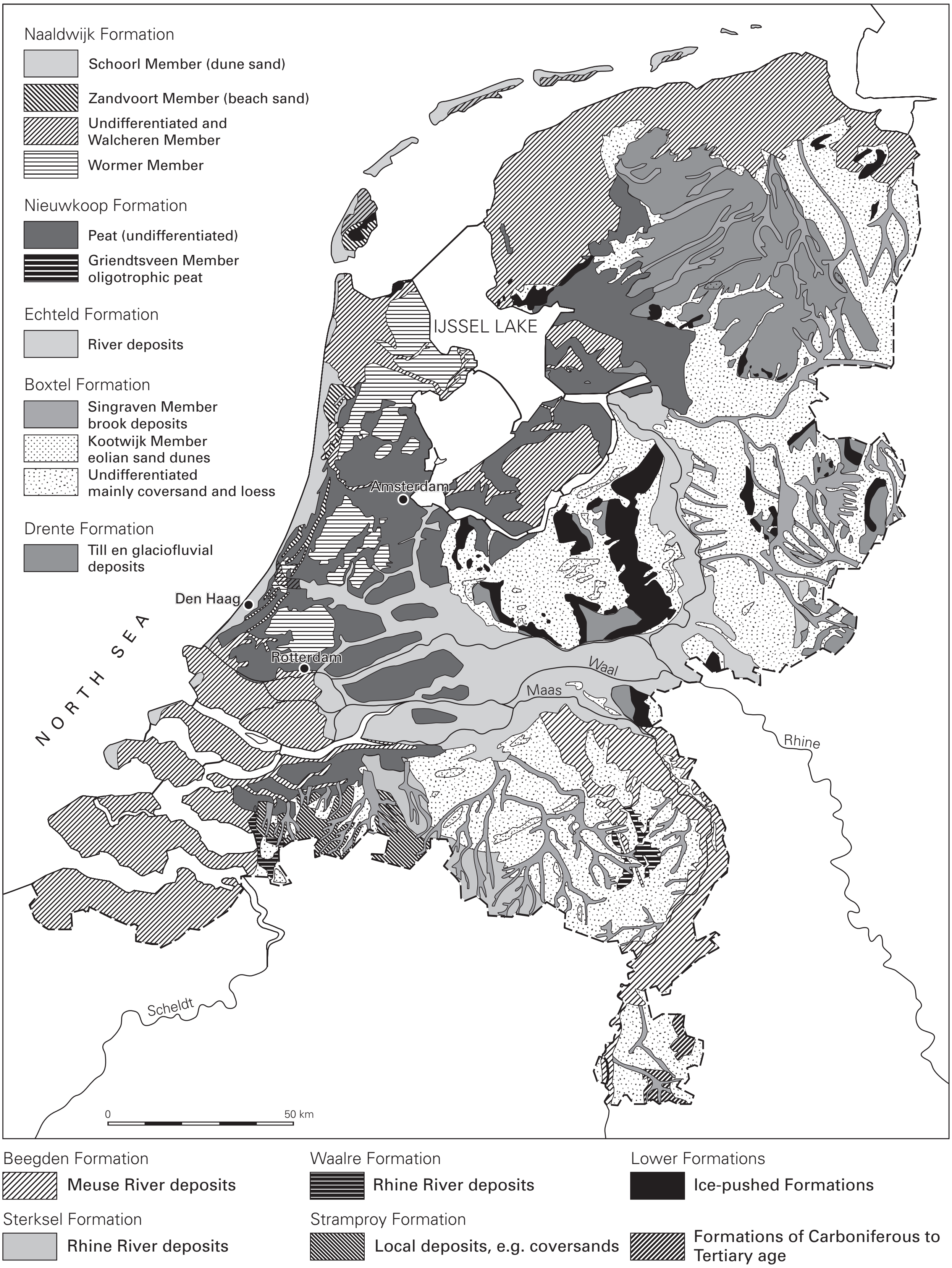

Figure 9 Holocene deposits in the Netherlands (after Berendsen 2004). Lithostratigraphy after De Mulder et al. (2003). The so-called perimarine area is characterized by fresh-water river sediments and extensive peat formation. River channels in this area are of the low-sinuosity meandering type or straight, and have a low width/thickness ratio of the sandbody. River banks consisting of clay and peat resist lateral erosion, and crevasse splays are abundant.

Sandbody geometry and changes in fluvial style

During the Weichselian Late Glacial, incised, braided channel belts were a few km wide, and consisted of sand with some gravel. In the Early Holocene the temperature increased and restoration of the vegetation led to a decrease of peak discharges of the rivers, a general decrease of sediment load and a relatively increased sediment load of fines.

Change in fluvial style

This resulted in a change of fluvial style from braided rivers to incising meandering rivers. In the western part of the present deltaic plain, aggradation started in the early Atlantic (after 8000 yr BP), when sealevel reached the present coastline and started to influence the gradient lines of the rivers (Figure 10). The sea first invaded the mouths of the Pleistocene river valleys. Since the influence of sealevel rise was felt earlier in the lower western part of the country, clayey floodbasin deposits on top of the sandy Pleistocene subsurface are younger in an eastern direction (Figure 10).

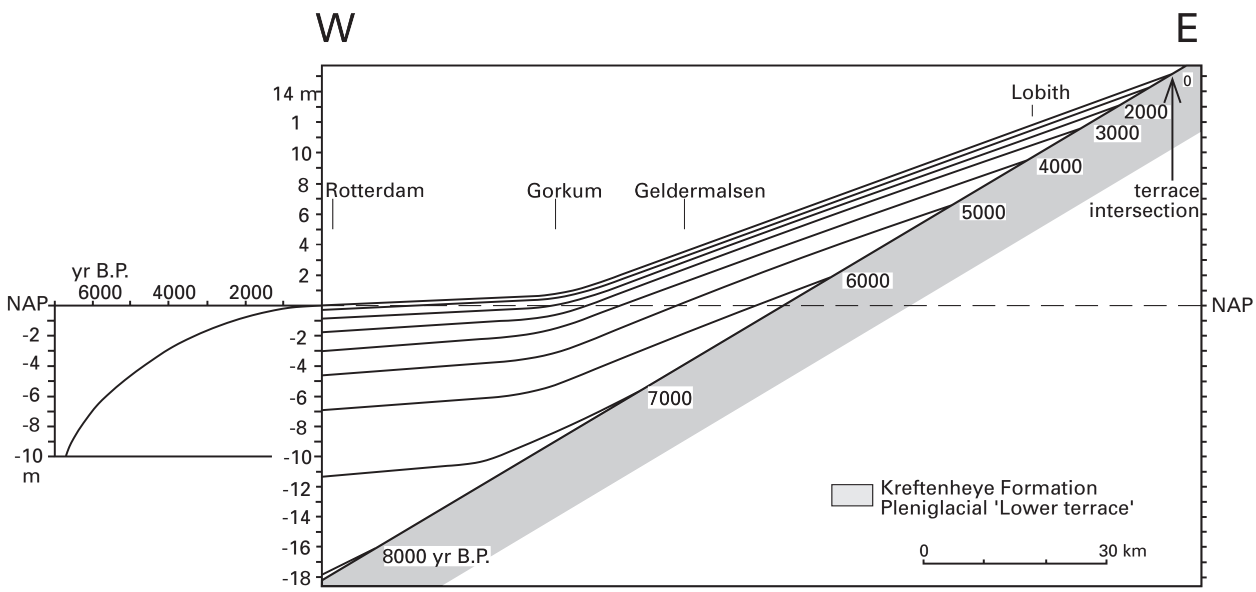

Figure 10 Sealevel rise and groundwater gradient lines (after Van Dijk et al 1991). Groundwater gradient lines are based on the dating of peat layers on the flanks of Younger Dryas eolian dunes, assuming that the peat approximately reflects groundwater levels. East of Gorkum, gradient lines are dominated by river gradients; west of Gorkum tides influenced the rivers. Holocene sealevel rise causing onlap, resulted in an upstream shift of the terrace intersection of Holocene deposits and the Kreftenheye Formation (Lower terrace). The shift of the terrace intersection approximately corresponds to the successive intersections of groundwater gradient lines with the Kreftenheye Formation.

Shifting of terrace intersection

This means, that the terrace intersection of the Pleniglacial terrace and Holocene deposits shifted landward as a result of Holocene sealevel rise. The rate of this onlap is shown in Figure 11, a diagram that is based on ^{14}C-dated peat samples at the base of the Holocene. Sealevel rise slowed continuously, and this would have to result in an ever decreasing landward shift of the terrace intersection. However, it can be seen that the shifting of the terrace intersection first slowed approximately 6000 yr BP, and then increased again. This is a result of the local higher elevated Peel Horst: sedimentation first had to obliterate elevation differences, before a further landward shift of sedimentation could occur. The higher position and relative tectonic uplift of the Peel Horst has also enhanced avulsions in the Peel Boundary Fault zone (Stouthamer & Berendsen 2000), see Figure 16.

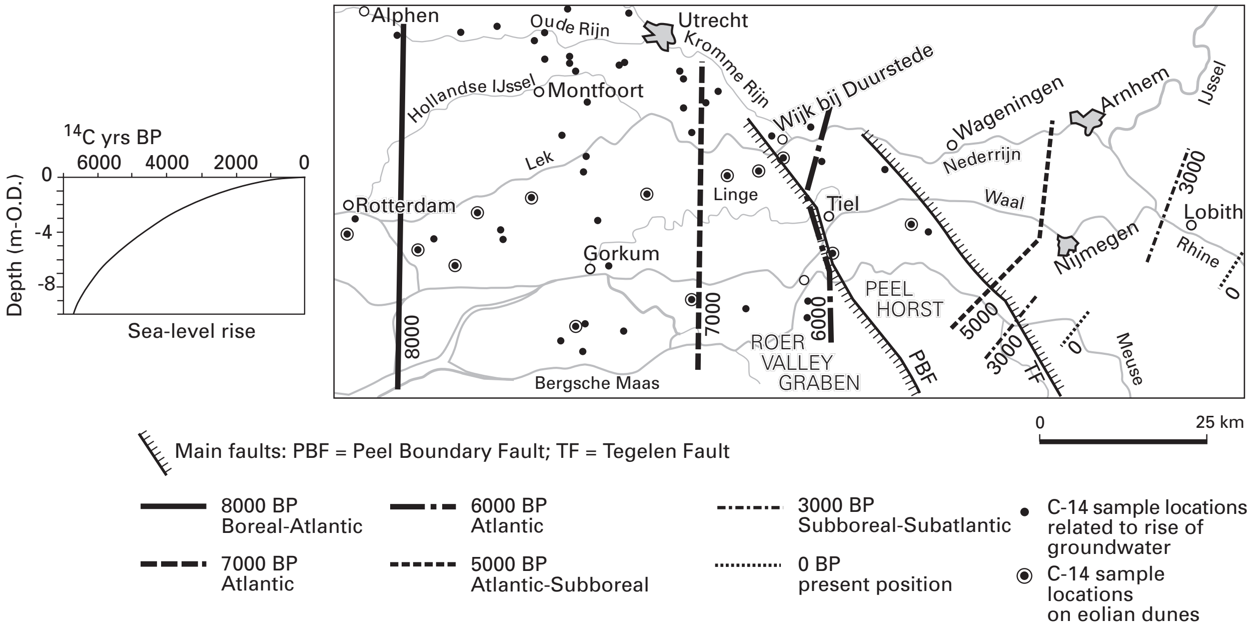

Figure 11 Shifting of the terrace intersection as a result of sealevel rise during the Holocene (Stouthamer & Berendsen 2000). The location of the terrace intersection of the Pleniglacial terrace (Kreftenheye Formation) and Holocene deposits shifted rapidly upstream between 8000 and 7000 yr BP. Then the upstream shift decreased as a result of the topographic high position of the Peel Horst. After 6000 yr BP relief had been leveled and the terrace intersection shifted relatively fast upstream again. Eventually the rate of upstream shift decreased as a result of decreasing sealevel rise.

Straight-anastomosed rivers with large-scale crevassing

In the near-coastal fluvial area, the fast rate of sealevel rise gave rise to a low-energy, narrow, anastomosing river pattern between approximately 7500 and 4000 ^{14}C yr BP (Figure 12). As sealevel continued to rise, this pattern extended further to the east, but it never crossed the Peel Boundary Fault. Individual channel belts consist of straight, ribbon-like sandbodies with a low width/thickness ratio (<15). As sealevel rise decreased, in the younger part of the Holocene, low-sinuosity meandering rivers reached further to the west. This may in part be due to an increase in discharge and sediment load, which can be related to increasing human influence (deforestation of the upstream areas). In a longitudinal direction, spatial changes in fluvial style depend on gradient and substrate erodibility (Figure 13).

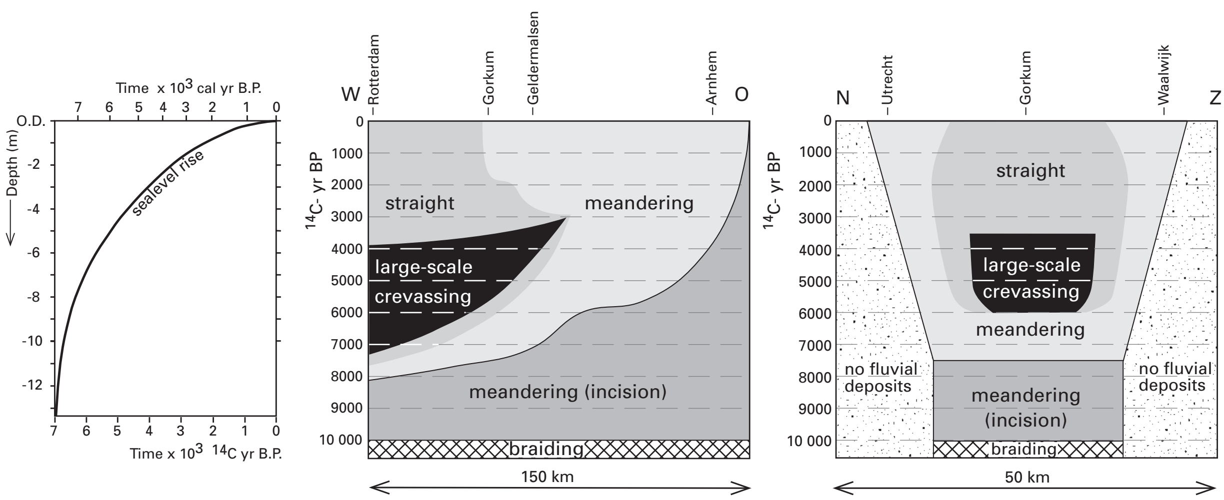

Figure 12 Changes in fluvial style during the Holocene, and their relation to sealevel rise (after Törnqvist 1993, modified by Berendsen & Stouthamer 2001). Rapid sealevel rise, causing high aggradation rates, is believed to be responsible for the formation of a straight river pattern with large-scale crevassing in the west-central part of the delta, between approximately 7500 and 4000 yr BP. When the rate of sealevel rise decreased, meandering rivers migrated westward and large-scale crevassing stopped. Near the northern and southern margins of the delta, straight rivers did not develop, because here the sandy substrate consisting of coversand enhanced lateral migration, and meandering.

In the area upstream of the terrace intersection of the Lower terrace and Holocene deposits, incising high sinuosity meandering rivers occur. Examples are found just across the German border . Downstream of the terrace intersection, the high-sinuosity pattern is maintained, although rivers are aggrading.

High-sinuosity rivers in eastern part of the delta

In the eastern part of the delta, the sandy, easily erodible Pleistocene substrate is still at shallow depth, which enhances lateral (bank) erosion. Further downstream the river pattern changes into low-sinuosity meandering or straight. Here, the Holocene consists mostly of thick clay and peat layers, that resist lateral erosion. This leads to straight channels with a low width/depth ratio.

Upstream shift of fluvial styles

This longitudinal succession of river patterns existed also in the past (at least since the early Atlantic), and shifted upstream as a result of Holocene sealevel rise, which resulted in landward shifting aggradation. This means, that in the western part of the delta, where rivers are now straight, meandering channel belts may be found at greater depth.

This relatively simple model is complicated by differences in substrate composition. In the marginal parts of the delta (the upstream area, and the northern and southern fringes) the substrate consisted of easily erodible Pleistocene fluvial sand or coversand.

Here, lateral erosion was common, and meandering fluvial systems developed, whereas in the central part of the delta, where the substrate consisted of clay, low-sinuosity meandering or straight rivers prevailed (Figure 12).

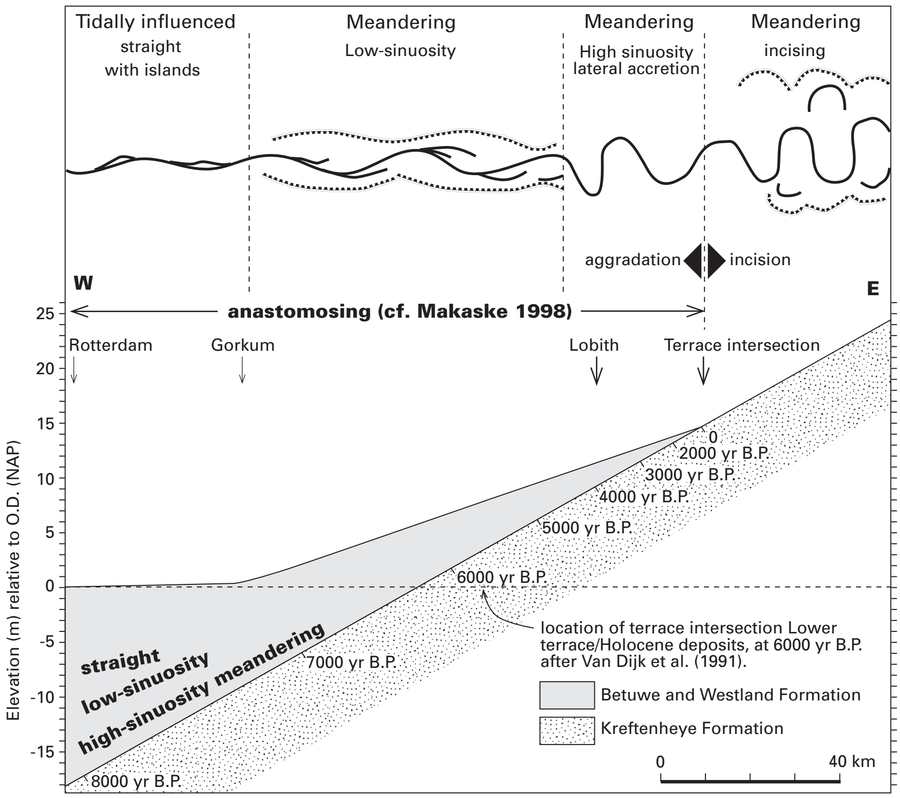

Figure 13 Spatial change in river morphology (after Berendsen & Stouthamer 2001). Meandering rivers occur near the terrace intersection, where easily erodible sandy deposits occur at shallow depth, and river gradient is still relatively high. Westwards, gradients decrease, as well as stream power, and river channels become straight. This longitudinal succession of river channel pattern shifted eastwards during the Holocene as a result of sealevel rise.

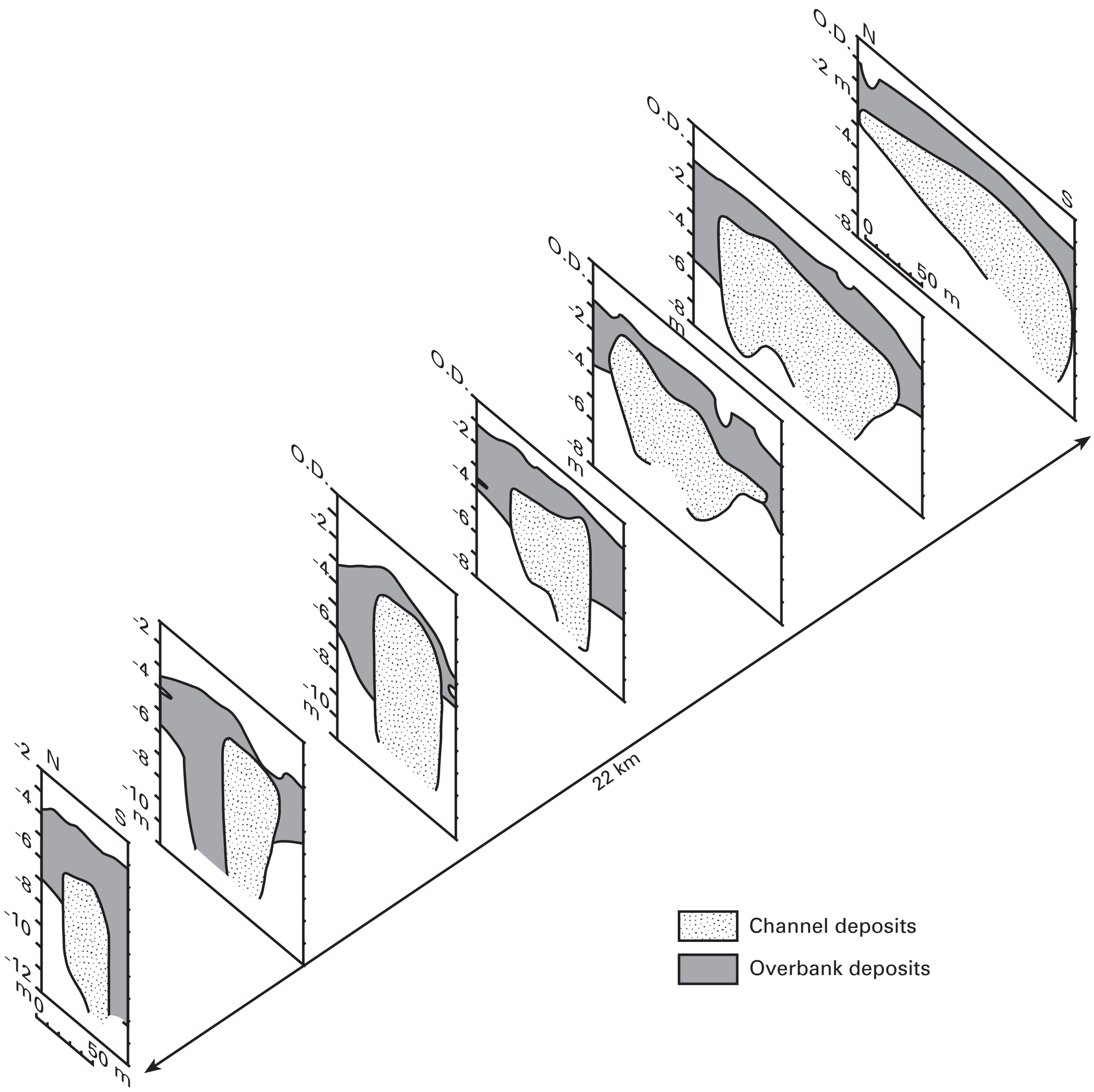

Figure 14 Fence diagram depicting the longitudinal architecture of channel deposits and overbank deposits of the Schaik channel belt (number 150 on the map of Berendsen & Stouthamer 2001). Interpreted from lithological sections by Törnqvist et al. (1993).

Longitudinal changes in width/thickness ratio of channel belt deposits

In general, channel belt width/thickness ratios decrease downstream (Figure 14). This is a result of decreasing width, as well as slightly increasing thickness (or depth of the channel). West of coordinate x=115 channel belt width is virtually equal to channel width (Figure 15). This means that lateral accretion is this area is absent, and the channel belt sandbody perimeter is equal to the channel wet perimeter. Hence, channel dimensions can be used to compare discharge of individual channels.

Channel belt width and channel width are equal in western part of the delta

Figure 15 also shows, that channel belt width of recent channel belts (Waal, Nederrijn-Lek) is larger than channel belt width of older channel belts. This is most likely a result of human influence: the number of coeval channels has decreased as a result of damming of older distributaries. This resulted in an increase of discharge of individual channels, and a greater channel belt width. It has been shown (Berendsen, Gouw & Wonink, unpublished) that channel belt thickness of the recent meandering distributaries (Waal, Nederrijn-Lek) is equal to channel depth. This suggests, that these meandering channel belts are a product of lateral accretion, and that vertical in-channel accretion is not important. Whether this also holds for the straight anastomosed rivers (that existed in the western part of the delta between 8000 yr BP and 4000 yr BP, Figure 12) is presently unknown. According to Makaske (1998, p. 55) the straight anastoming Columbia River in British Columbia (Canada) may show significant vertical in-channel accretion.

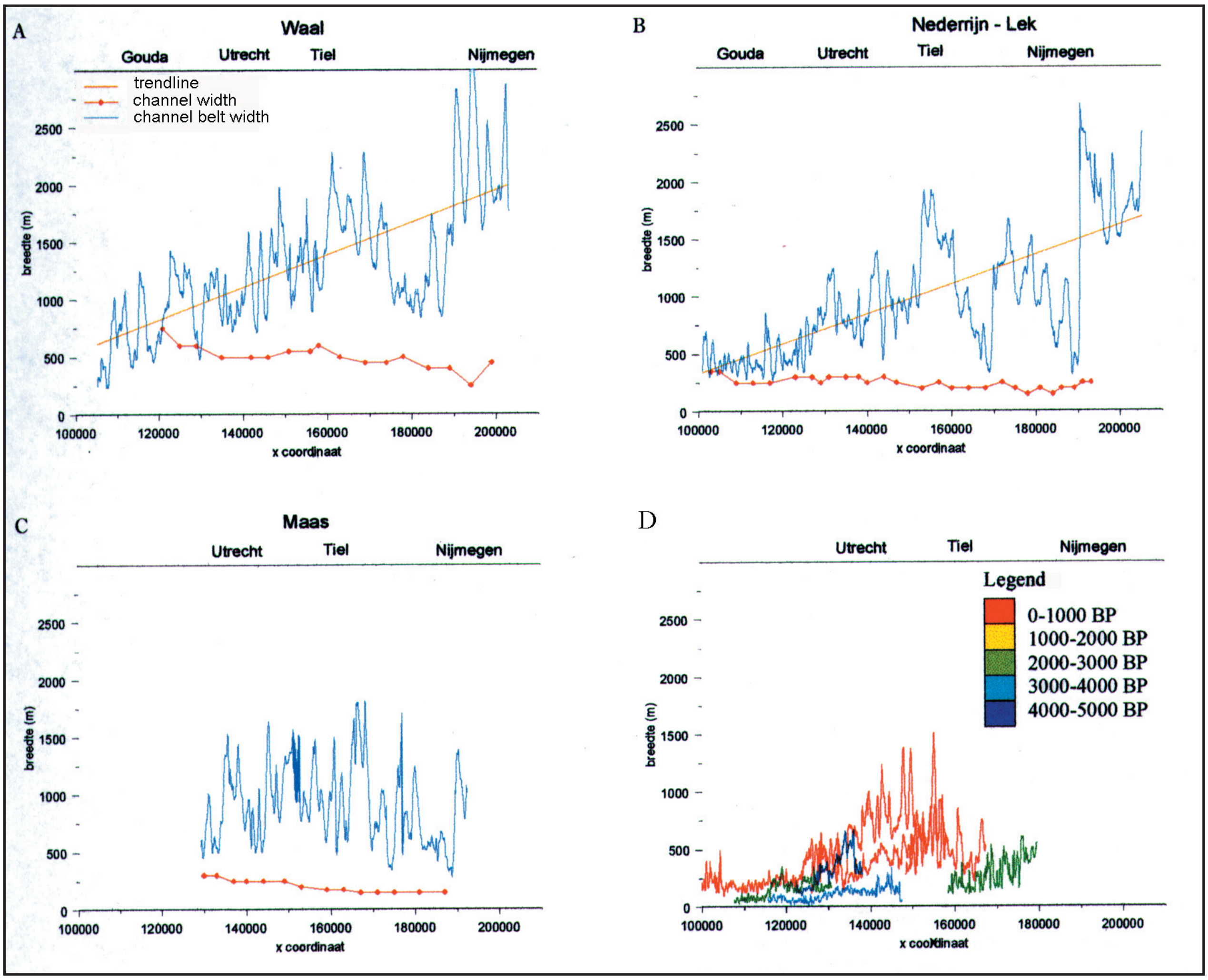

Figure 15 Longitudinal changes in channel belt width and channel width. West of coordinate x=115 (near Rotterdam) channel belt width equals channel width, indicating that lateral accretion is no longer important. The perimeter of the sandbody here is representative for the wet perimeter of the channel, which allows making discharge comparisons.

Avulsion

Avulsion, meander cutoff, and lateral erosion are processes that are critical to alluvial architecture, because they determine channel recurrence interval on the floodplain and, as a result, channel density and interconnectedness. These characteristics from modern analogs may have important practical applications in the study of ancient rocks, for example in the oil industry, where fluvial architecture often has to be determined from borehole information.

Avulsion was an important process in the evolution of the Holocene Rhine-Meuse delta. The avulsion history has been extensively studied by Stouthamer & Berendsen (2000).

Approximately 90 avulsions occurred over the last 7000 ^{14}C yr, but only 30 of them can be considered as major avulsions.

Avulsion locations are influenced by sealevel rise, neotectonics and changes in discharge and sediment load

The location of avulsion sites (on a time scale of millennia) is influenced by the same factors that influenced the palaeogeographic evolution. The main factors are: sealevel rise, neotectonics and changes in discharge and sediment load. All avulsions presumably start with the formation of a crevasse splay. In many cases, large crevasse splays occur near avulsion sites. Especially between 4000 yr BP and 2000 yr BP, many avulsions seem to be related to differential tectonic movements of the Peel Horst and the Roer Valley Graben (Figure 16). The relatively fast subsidence in the area west of the Peel Boundary fault created new accommodation space, in an area where rivers still had a significant gradient. This enhanced the occurrence of nodal avulsions near the Peel Boundary Fault.

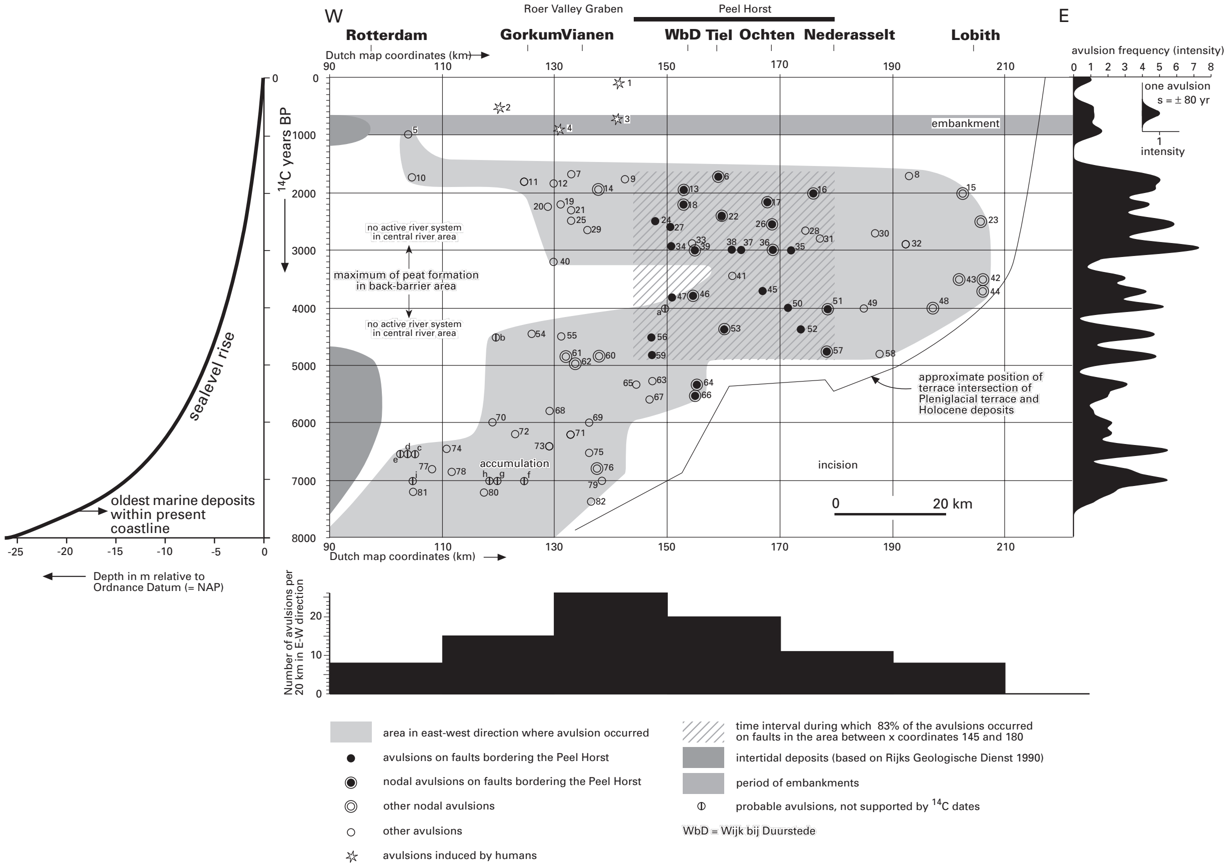

Figure 16 Sealevel rise, location of the avulsion sites in the Rhine-Meuse delta plotted on an E-W axis, and avulsion frequency during the Holocene (Stouthamer & Berendsen 2000). The spatial distribution of avulsion locations is mainly determined by: sealevel rise (7500-3700 yr BP), neotectonics (4900-1700 yr BP), increased discharge and/or within-channel sedimentation (2800-1000 yr BP), and human influence (after 1000 yr BP). Note that the tectonic elements cut this section at an angle. Most avulsions occurred in the central part of the delta and around the western fault zone of the Peel Horst (between coordinates 130 and 170). The avulsion frequency reached a maximum from 3000 to 1700 yr BP; lowest avulsion frequencies occurred from 5300 to 5000 yr BP, 3500 to 3000 yr BP, and after 1500 yr BP (from Stouthamer & Berendsen 2000).

Average period of existence of channel belts is 1280 yr

Avulsion frequency initially seems to have been determined by rapid sealevel rise (during the Atlantic). A second maximum of the avulsion frequency was reached between 3000 and 1700 BP, which may be related to increased discharge and/or within-channel sedimentation, or both. The period of existence of channel belts varied widely throughout the Holocene (Figure 17), but shows no significant trend over time and space. It remained on average constant at approximately 1000 ^{14}C yr, with a large standard deviation of 700 ^{14}C yr. When only the best dated channel belts are taken into account, the average period of existence of channel belts is 1280 ± 820 cal yr.

The average avulsion duration is ~325 yr. It seems to be constant over time, although it varied widely, from 'instantaneous' avulsion (< 200 yr) to gradual avulsion with a duration of over 1000 ^{14}C yr. This implies that the average interavulsion period can be estimated to be ~600-700 ^{14}C yr, or ~800-900 cal yr, but variation is considerable.

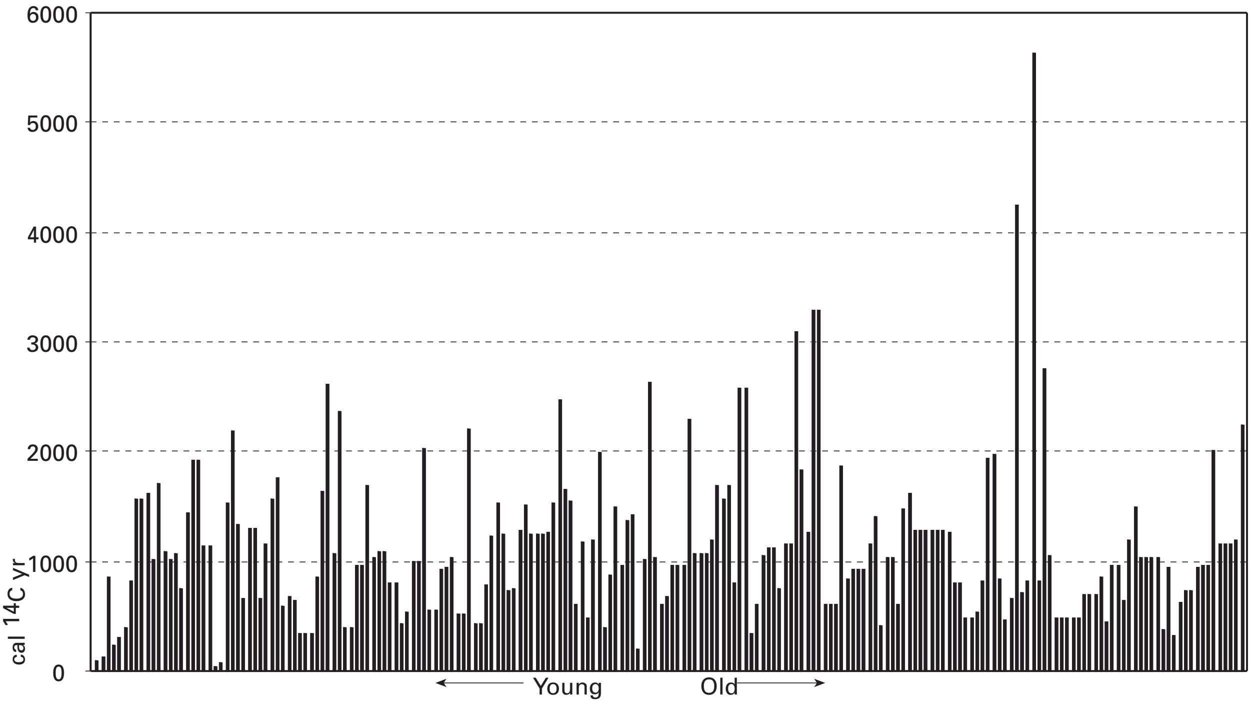

Figure 17 Period of existence of channel belts in the Rhine-Meuse delta (Berendsen & Stouthamer 2001). The diagram shows no significant trend in the length of the period of existence of the channel belts. The average period of existence is approximately 1000 ^{14}C yr, or 1100 cal yr.

Number of coeval channels determined by avulsion frequency

If avulsion duration and interavulsion period are constant, then the number of coeval channels in the Rhine-Meuse delta was essentially determined by avulsion frequency. A lack of data from other deltas in the world presently prohibits to determine whether this a general characteristic, or a phenomenon that only applies to the Rhine-Meuse delta.

Sealevel rise

Sealevel rise (Figure 10) has had a significant influence on the development of the Holocene stratigraphy in the estuarine area. The rate of sealevel rise during the Early Holocene was very high: about 1 m/100 ^{14}C years, but gradually decreased during the Holocene. The rapid rise of sealevel before 6000 yr BP is caused by the melting of the Weichselian ice sheets; the slower rise of sealevel during the younger part of the Holocene is mainly due to isostatic subsidence. The Holocene transgression caused an overall stack in backstepping pattern (Figure 10), and peat formation moving inland. Figure 10 is based on peat samples on the flanks of eolian dunes, and gives a detailed picture of groundwater rise in an E-W section of the delta. More recent observations (Cohen 2003) show, that groundwater level rise is also influenced by local differential tectonic movements of the Roer Valley Graben and the Peel Horst (Figure 18).

The back-barrier intertidal and estuarine deposits were formed during two main periods of transgression: during the Atlantic (8000 - 5000 yr BP) and during the Subatlantic (3000 yr BP to the present). The deposits were previously known as ‘Calais Deposits’ and ‘Duinkerke (=Dunkirk) Deposits’ respectively. In recent years, this subdivison has been abandoned, because of difficulties in recognizing these units. Now, a lower clastic Member (the Wormer deposits) and an upper clastic Member (the Walcheren deposits) are recognised (de Mulder et al. 2003), based purely on lithostratigraphic criteria. The period in between is characterized by large-scale peat formation, which is related to the closure of the beach barrier coast, separating the back-barrier lagoon from the intertidal area.

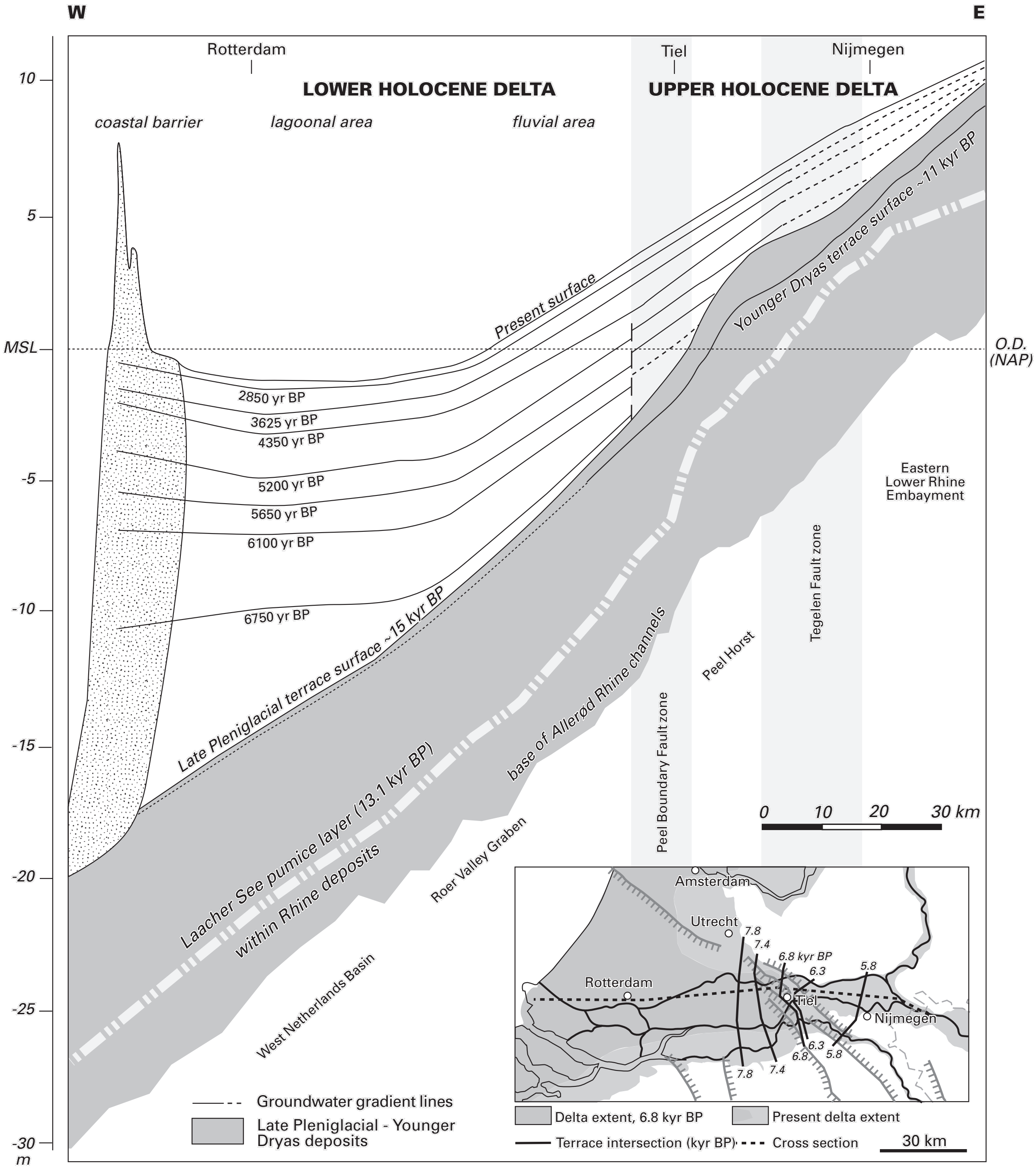

Figure 18 Groundwater level rise around the Peel Boundary Fault. Local groundwater curves show the effect of differential subsidence (Cohen 2003).

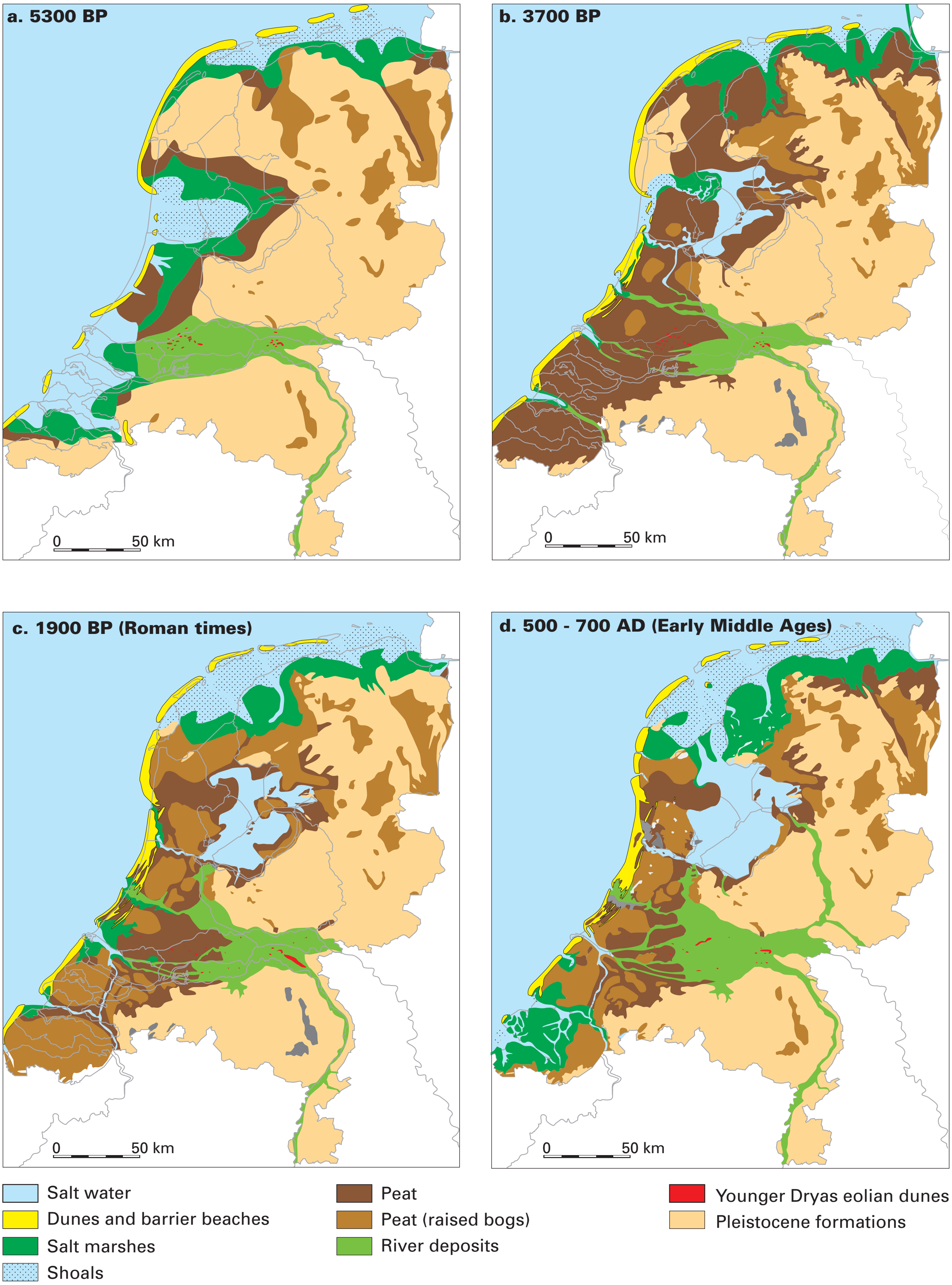

Figure 19 Holocene palaeogeographic development of the Netherlands (after Zagwijn 1986). Approximately 5300 yr BP the oldest barrier beaches were formed, east of the present coastline. Approximately 3700 yr BP the coast was closed and extensive peat formation occurred in the back-barrier area. Approximately 1900 yr BP (Roman times) coastal erosion started in the north and in the southwest. At that time the Oude Rijn (=Old Rhine) was still the main distributary of the Rhine. Approximately 500-700 AD coastal erosion increased, and main drainage of the Rhine shifted to the southwest. Rhine and Meuse (=Maas) both emptied into the 'Meuse estuary' near Rotterdam. The Oude Rijn was finally dammed in 1122 AD.

Coastal barriers and dunes

Coastal retrogradation until 5300 yr BP, progradation until 2000 yr BP

In the coastal dune area, the transgressional phase lasted until approximately 5300 yr BP, when barrier beaches were formed just east of the present coastline (Figure 19a). Sand used for coastal build-up is mainly reworked Pleistocene sand, supplemented by a minor Rhine provenance (an estimated 10 %). As sealevel rise slowed, new barrier beaches were formed westward from the older ones (Figure 19b). From that time on coastal progradation occurred, especially in the central part of the coastal area, although sealevel continued to rise. The rates of deposition were greater than the rate of sealevel rise; apparently, wave action in the shallow, sandy North Sea was able to transport large amounts of sand towards the coast. An important source of sand probably was the protruding northwestern coast (Figure 19a), where Pleistocene sands occur at relatively high elevations.

Extensive peat formation after closure of the coast

The resulting wide barrier beaches offered better protection against marine ingressions. In the back-barrier area a large swamp existed. River inundations and precipitation gave rise to large-scale peat formation during the Subboreal and early Subatlantic (Figure 19b and c), from approximately 4000 yr BP - 3000 ^{14}C yr BP. Peat formation continued up to 2000 yr BP, but gradually decreased.

Coastal erosion since 2000 yr BP

During the last two thousand years, differentiated coastal development is seen (Figure 19c and d) because of the dimunition of sand sources available for progradation (ebb tidal deltas and rivers), and locally different current and wave patterns.

4. Historical evolution

Humans have been present in the Rhine-Meuse delta for at least 200,000 years; however, human influence on the landscape was small until the Neolithic (6400 - 3650 yr BP), when forests were cleared and agriculture started on the natural levees, on the Younger Dryas eolian dunes and on higher Pleistocene sands.

Old Rhine was the northern boundary of the Roman Empire

Approximately 2000 yr BP, the northernmost distributary of the Rhine (Vecht, Figure 1) silted up and the Old Rhine (which had been the most important Rhine distributary since 5000 yr BP) became the northern border of the Roman Empire during most of Roman times (50 BC - 400 AD). Roman occupation left many archaeological traces in the substrate, and many villages were founded. The Romans even influenced locally the course of rivers by digging canals.

Embankments completed 1000 AD

Human influence increased enormously during the Middle Ages, especially from 1100 AD onwards. A first step was the embankment of the rivers, which was completed around 1300 AD. During the same time period the Old Rhine (Kromme Rijn - Oude Rijn) was dammed near Wijk bij Duurstede in 1122 AD (for the location see Figure 1) and the river mouth of the Old Rhine degraded. The ebb-tide delta was eroded and the sand became available for transport to the coast. This stimulated the formation of the so-called Younger Dunes, that form dune ridges up to 40 m high. The damming of the Old Rhine distributary and some other channels, like the Hollandse IJssel (1285 AD) and the Linge (1307 AD), reduced the number of Rhine distributary to the present three, and there was a tendency for the main flow to shift southwestwards. The Waal gradually became more important, and both the Meuse (=Maas) and the Lower Rhine-Lek joined the Waal before debouching through a tidal inlet just south of the present Nieuwe Maas-Rotterdam Waterway.

Human-influenced discharge distribution

The digging of the Pannendens Canal (PK in Figure 1) in 1707 AD significantly altered the discharge distribution over the Rhine distributaries. From then on the discharge distribution has remained the same. The Haringvliet (Figure 1) soon became the main outlet. In recent years this tidal inlet has been closed off by a dam, forcing the main flow through the Rotterdam Waterway (Figure 1), which was dug in 1872 AD.

The embanking of the rivers opened the possibility of draining and reclaiming the peat bogs in the western part of the deltaic plain. In the Late Middle Ages this peat was excavated and used for fuel or to extract salt for the population of the growing cities. Since groundwater levels were high, these peat excavations rapidly turned into lakes. Storms enlarged the lakes by wave action. In the long run, this created a dangerous situation, and many lakes were pumped dry successively, especially between 1600 and 1900 AD. One of the best known examples is the Haarlemmermeerpolder (the location of Amsterdam airport), the first lake that was pumped dry in 1852 AD using three steam engines.

The construction of groynes, especially since 1850 AD, has caused narrowing and deepening of the river beds, thus enhancing the shipping industry. When the entrance to Rotterdam harbour was widened, river water levels were lowered, and tidal influence is now experienced about 20 km further upstream (Figure 20) compared to a century ago.

In recent years, pressure on the embanked floodplains has increased: agriculture, industrial activity, nature conservation and recreation all demand space, resulting in conflicting interests.

Plans are now being made for digging clay and sand and thus lower the level of the embanked floodplains, and increase storage capacity during floods. At the same time ‘side-channels’ are dug to recreate a more natural and more diverse environment for plants and animals.

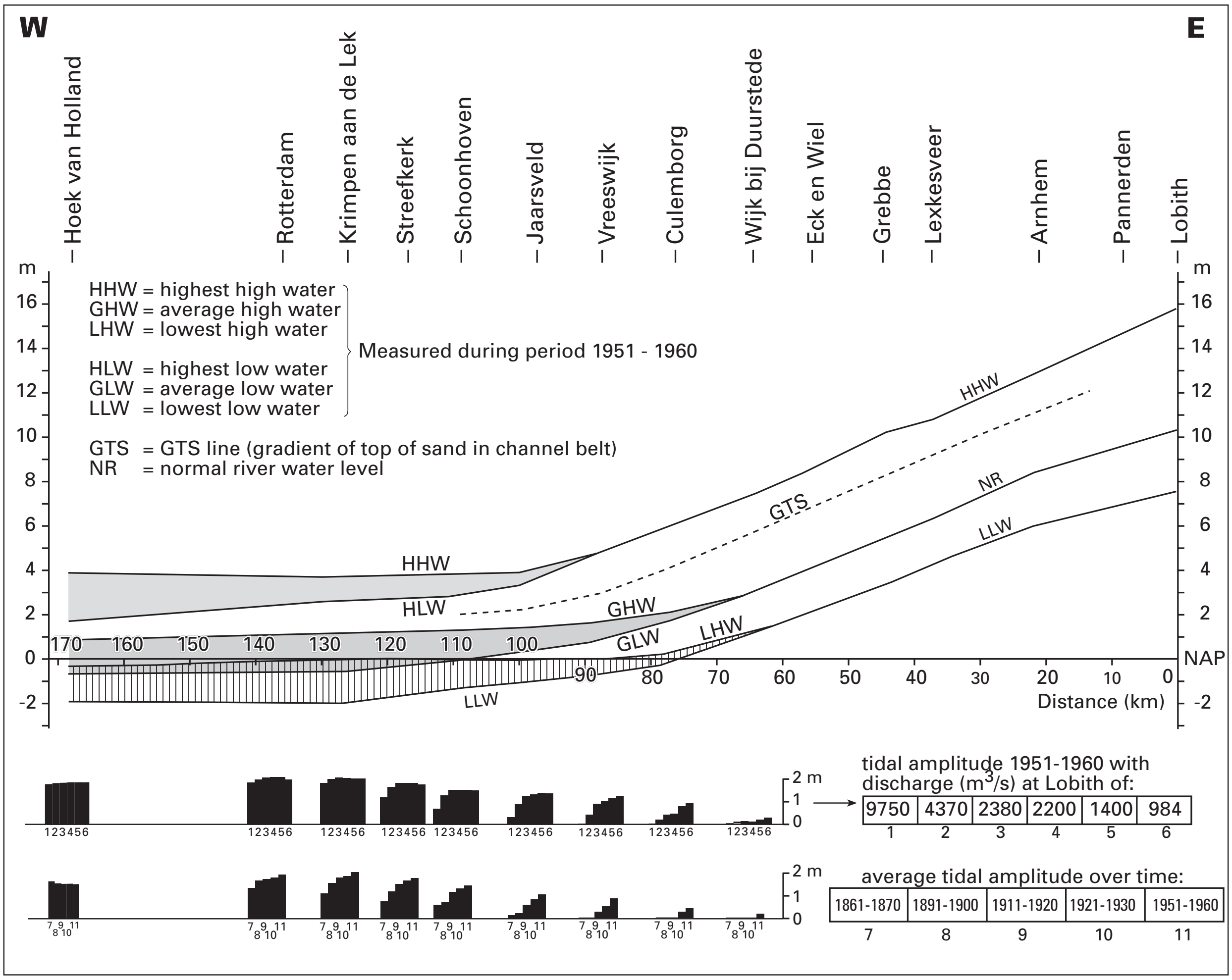

Figure 20 Water levels and tidal amplitude on the Lower Rhine-Lek (Berendsen 1982). The gradient lines represent highest high water, normal river level and lowest low water, measured over the period 1951-1960. The GTS line (*G*radient of the *T*op of *S*and) corresponds approximately to bankfull discharge water levels. Tidal influence at high discharge reaches 70 km upstream, to Jaarsveld. At low discharge tides reach 90 km inland. The upper histograms show tidal amplitude at different locations, with varying discharge. The influence of tides reaches further inland at low discharge than at high discharge. The lower histograms show that tidal amplitude on the river has increased over the last century. This is a result of the widening and deepening of the Rotterdam Waterway.

Preservation potential

Since most Pleistocene highstand (interglacial) deposits were completely eroded during subsequent glacial lowstands, the preservation potential of the Holocene deposits seems to be low. However, anastomosing river deposits also occur in rapidly subsiding basins, e.g. Molasse basins, where the preservation potential is regarded very high (Smith, 1983). The Holocene of the Netherlands therefore can serve as a sedimentary model for other (ancient) environments and for fossil subsurface deposits.

5. Excursion

Maps: ANWB Midden-Nederland
Topographical map 1: 50.000

Table 3 Time schedule, route and approximate time schedule (accuracy depending on traffic):

Departure	Stop	Route	Arrival
8:30 AM	1	Delft – The Hague – Utrecht – Maarsbergen – Amerongen - Rhenen	10:00 AM
10:45	2	Rhenen - Bergharen	11:15
11:25	3	Bergharen - Appeltern - Blauwe Sluis - Maasbommel	11:45
11:45	4	Maasbommel - Nieuwe Schans - Dreumelse Berg	12:00
12:00	5	Dreumelse Berg - Dreumel - Heerewaarden	12:15
12:20	6	Heerewaarden - Rossum - Hurwenen	12:30
12:30		Hurwenen - Waardenburg - Restaurant 'De Waalbrug'	1:00 PM
1:00		LUNCH	2:00
2:00	7	Waardenburg - Beesd - Rhenoy - Leerdam - Diefdijk - Wiel van Bassa	2:45
2:45	8	Wiel van Bassa - Zijderveld - A2 - IJsselstein - Montfoort	3:45
4:45		Montfoort - Linschoten - Woerden – Rotterdam - Delft	6:00

Excursion stops

1. **Ice-pushed ridge at Rhenen.** Maximum extent of Saalian ice-sheet.
2. **Younger Dryas eolian dune** of Bergharen, overlying Bølling-Allerød-interstadial floodbasin clay. Morphology and cross section.
3. **Residual channel**: View and morphology of a residual channel and point bar of the recent river Meuse at Maasbommel.
4. **Younger Dryas dune.** The isolated Younger Dryas eolian dune 'Dreumelse Berg' is on the Peel Boundary Fault. Morphology.
5. **Former confluence of Waal and Maas** near Heerewaarden. Hydrological problems; view of the rivers.
6. **Dike reconstruction** at Hurwenen: Dike reconstruction, various dike levels between Rossum and Zaltbommel. Water and flood control problems.
7. **Dike breach scour hole.** The 'Wiel van Bassa' is the largest dike-breach scour hole in the Netherlands. It is located where a sandy channel belt crosses the dike.
8. **Late Holocene meandering channel belt** near Montfoort. The meandering Stuivenberg channel belt that existed from 3960 – 3180 ^{14}C yr BP shows a clear point bar morphology. The sedimentology of the point bar, residual channel and natural levee will be shown. Examples of drilled cores, and demonstration of the Van der Staay suction corer.

Figure 21 Excursion route

STOP 1: Ice-pushed ridge at Rhenen

Observations

The ice-pushed ridge at Rhenen (Figure 22) is approximately 70 m high and offers an overview of the Weichselian Late Glacial valley, that is now filled with Holocene sediments of the Rhine and Meuse. The lakes to the southeast were formed by excavation of sand. The area has now become part of a nature conservation project.

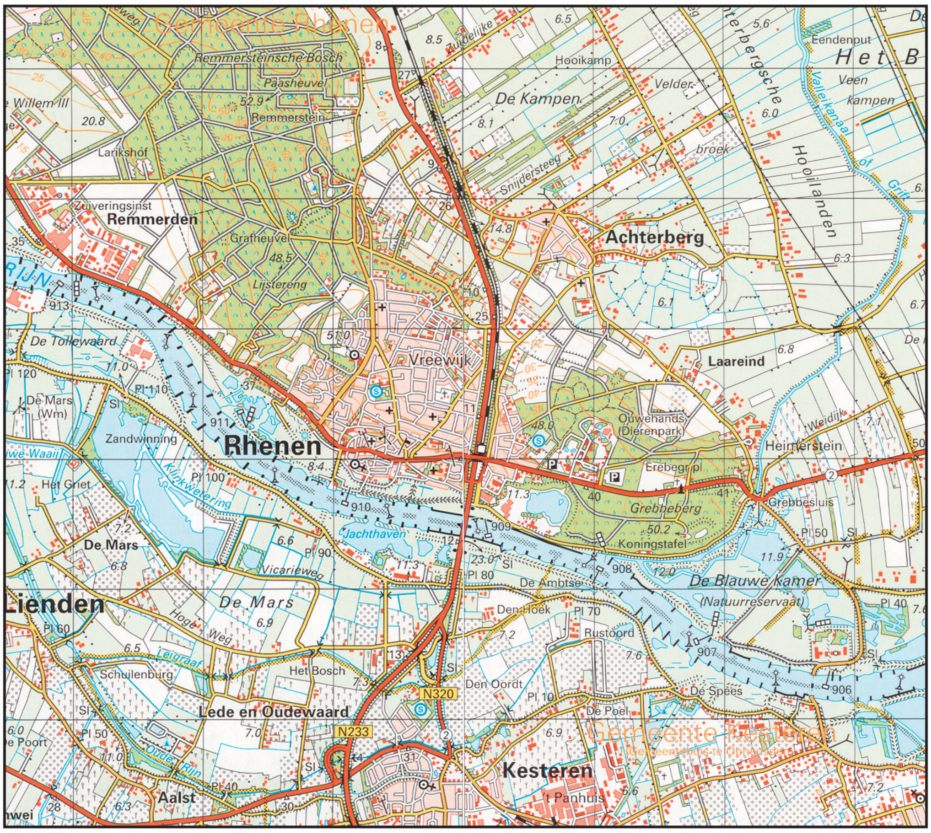

Figure 22 The ice-pushed ridge at Rhenen (topographic map 1: 50.000).

Geology/geomorphology

The ice-pushed ridges were formed during the Saalian glaciation, when the Scandinavian ice sheet reached its maximum extent in the Netherlands. Measurements of strike and dip of glacially deformed layers have shown, that the ice-pushed ridges were formed in five successive stages during the Saalian glaciation. The ridge of Rhenen originally was connected to the ice-pushed ridge of Wageningen (Figure 23), but the southernmost tip has been eroded by the Rhine. The ridges enclose the ‘Gelderse Vallei’ a 100 m deep glacial basin, that is largely filled with glaciofluvial deposits and coversand.

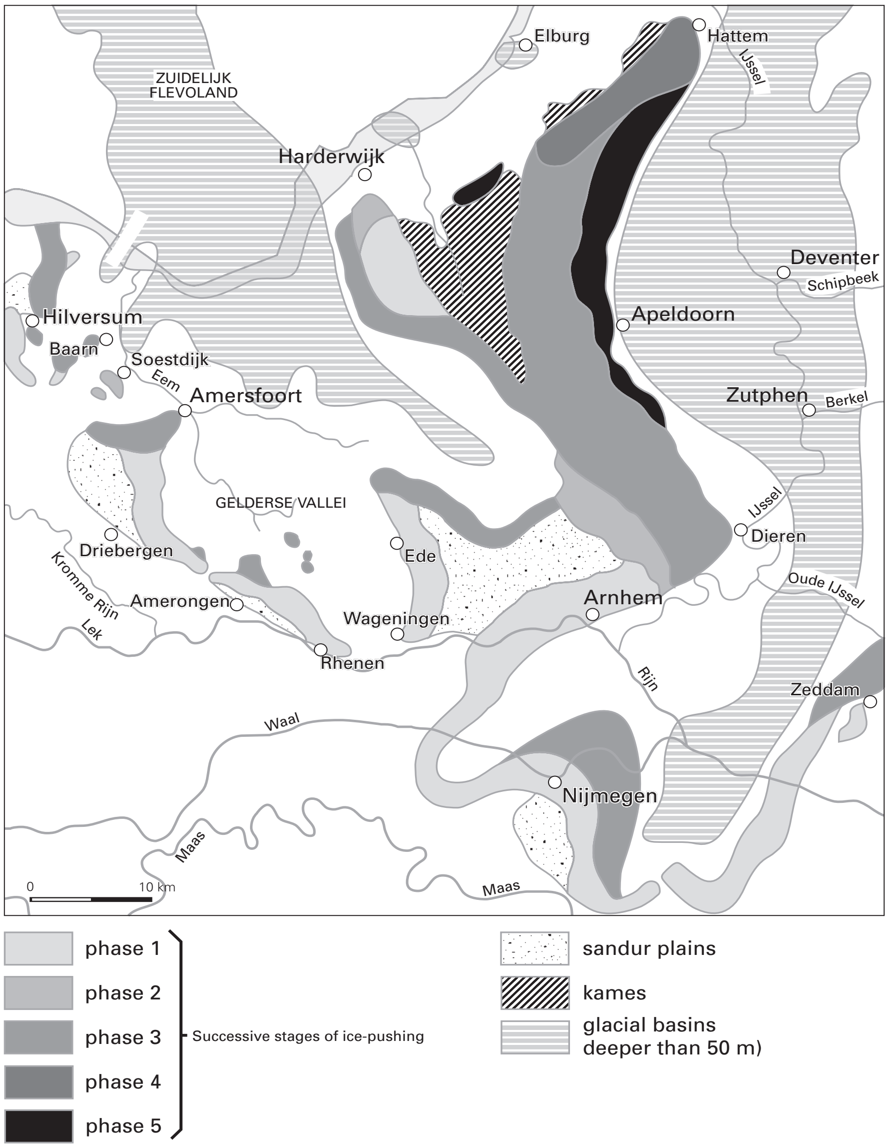

Figure 23 Ice-pushed ridges of Saalian age in the central Netherlands (Berendsen 2004). The Gelderse Vallei is a glacial basin that is mainly filled with glaciofluvial deposits and coversand.

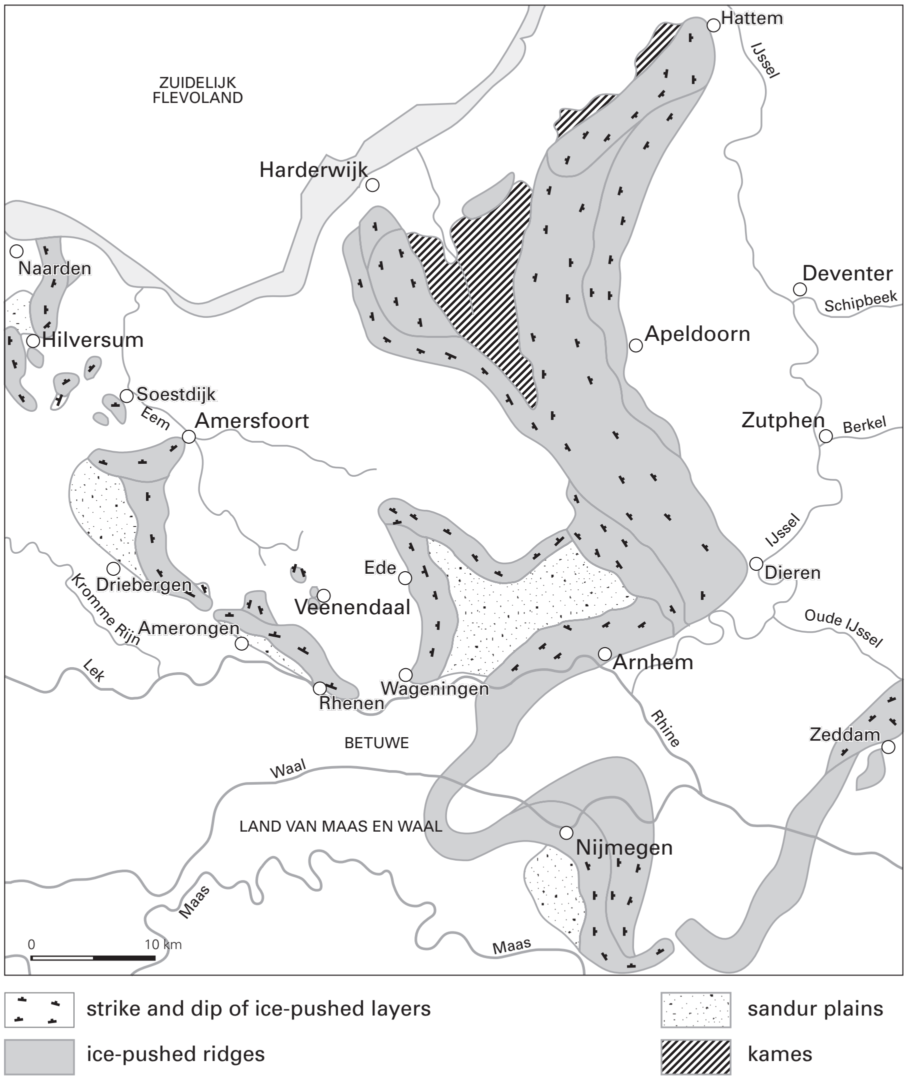

Figure 24 Strike and dip of ice-pushed layers (after Maarleveld 1953, 1981).

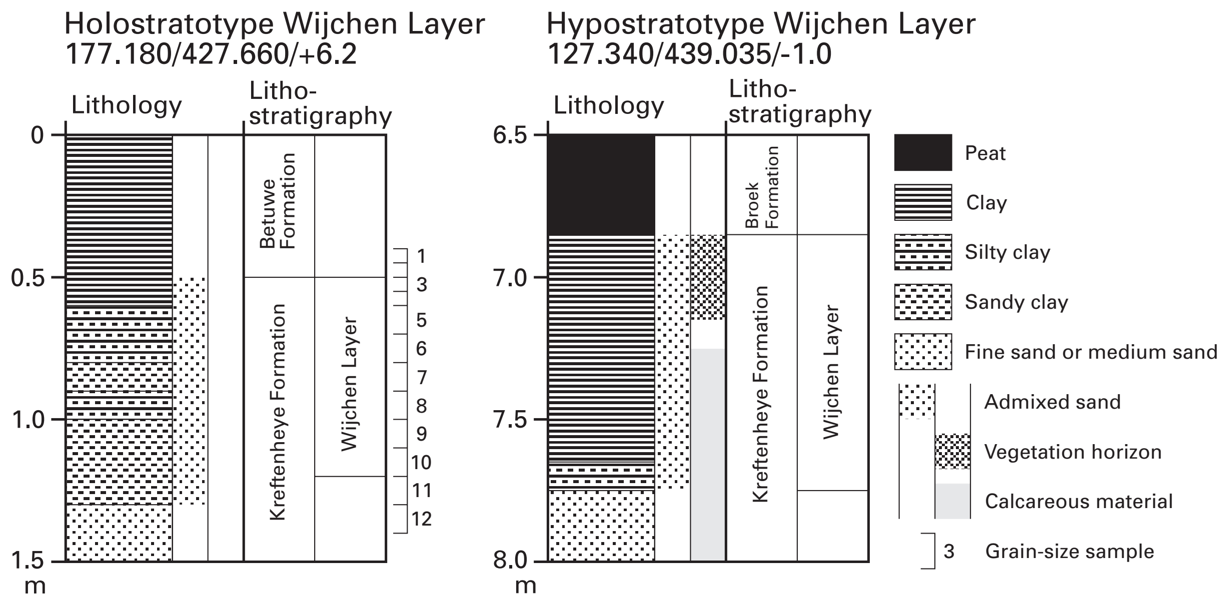

Figure 25 Holostratotype of the Wijchen Layer (after Törnqvist, Weerts & Berendsen 1993). The Wijchen Layer overlies the Late Weichselian terraces.

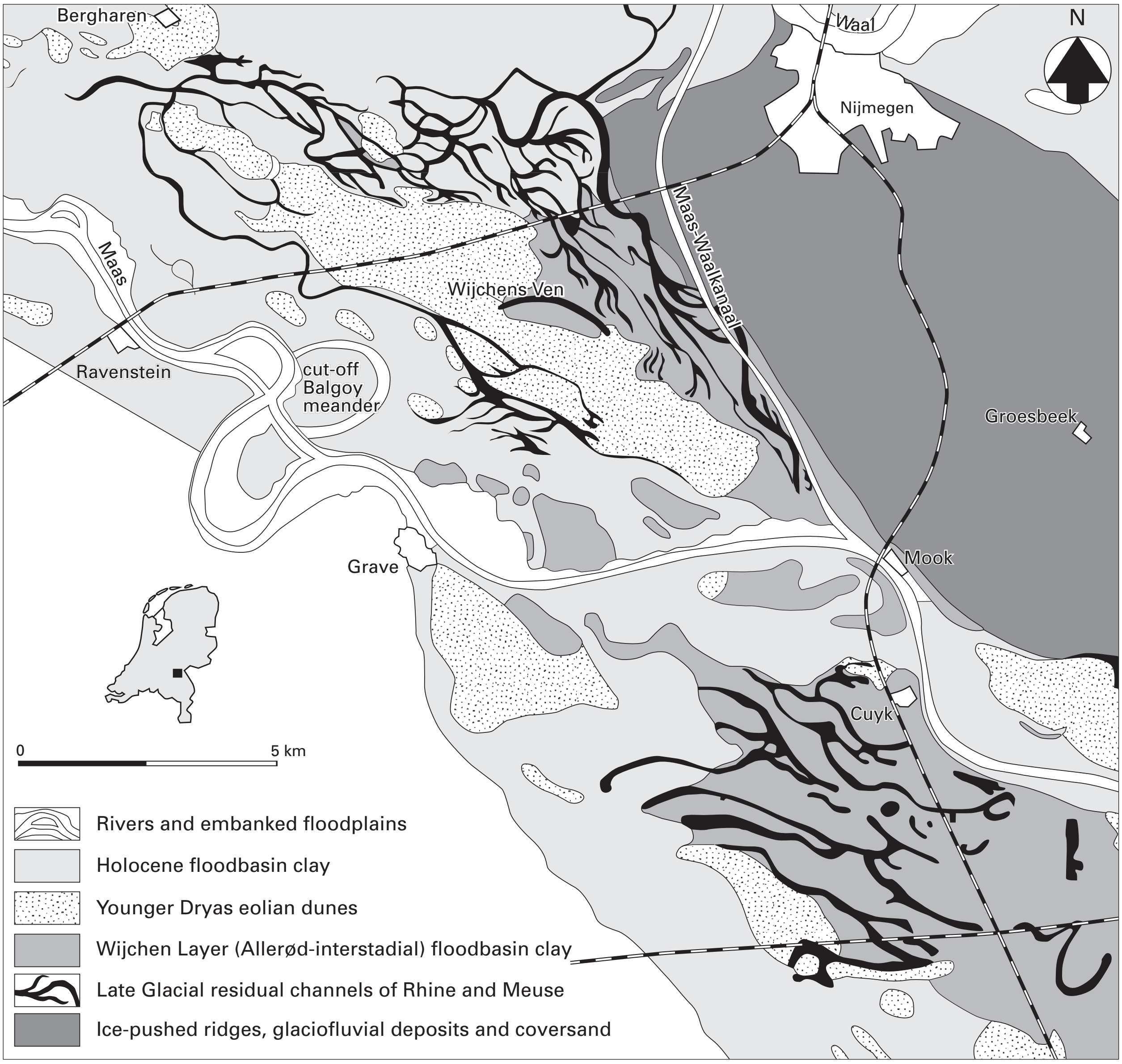

Figure 26 Braided river pattern in the Pleniglacial Lower terrace near Nijmegen (after Pons 1957). The braided channels mostly became filled with clay or peat during the Bølling-interstadial. However, some channels, like the meandering channel in the north continued to exist into the Allerød-interstadial. The base of the residual channel fill was ^{14}C dated as Younger Dryas (10700 yr B.P.).

Channels north of the Younger Dryas eolian dunes and west of the Maas-Waal kanaal belong to the so-called Lower Terrace (of Weichselian Pleniglacial age). Some of these residual channels in the Lower Terrace are still visible in today's morphology and pattern of parcelling. During the Younger Dryas rivers were braided again. Eolian dunes were formed on top of the Wijchen Layer on the Pleniglacial terrace. The sand was blown out of the Younger Dryas braided river deposits, that occur southwest of the dune complex. Wind direction was from the southwest. Dunes may be up to 15 m high, and often are parabolic. They also occur in the western part of the river area. In the western Netherlands they are almost entirely covered by peat.

STOP 2: Younger Dryas eolian dune

Observations

The landscape is an almost featureless floodplain. The dune of Bergharen stands out with an elevation of 21 m above sea level (Figure 27), this is approximately 15 m higher than the surrounding area. The dune of Bergharen is covered with coniferous trees, which is unusual for the fluvial part of the delta: coniferous trees only occur on sandy, high substrate.
A cross section of the dune and buried terraces will be discussed.

Figure 27 Location of the Bergharen Younger Dryas eolian dune (topographical map 1: 50.000).

Geology/geomorphology/sandbody geometry

The eolian dune of Bergharen is of Younger Dryas age (Figure 28). All the Younger Dryas dunes overlie the Wijchen Layer (Figure 28), a floodbasin deposit of Bølling-Allerød-interstadial age on top of the Pleniglacial terrace. The Pleniglacial (braided river) terrace deposits consist of sand and gravel.

Further to the north, the sand is excavated. The Wijchen Layer forms a permeability barrier between the eolian dune sands and the underlying sand sheets. This makes the dunes into *isolated sand bodies* if they become fully encased in floodbasin clays.

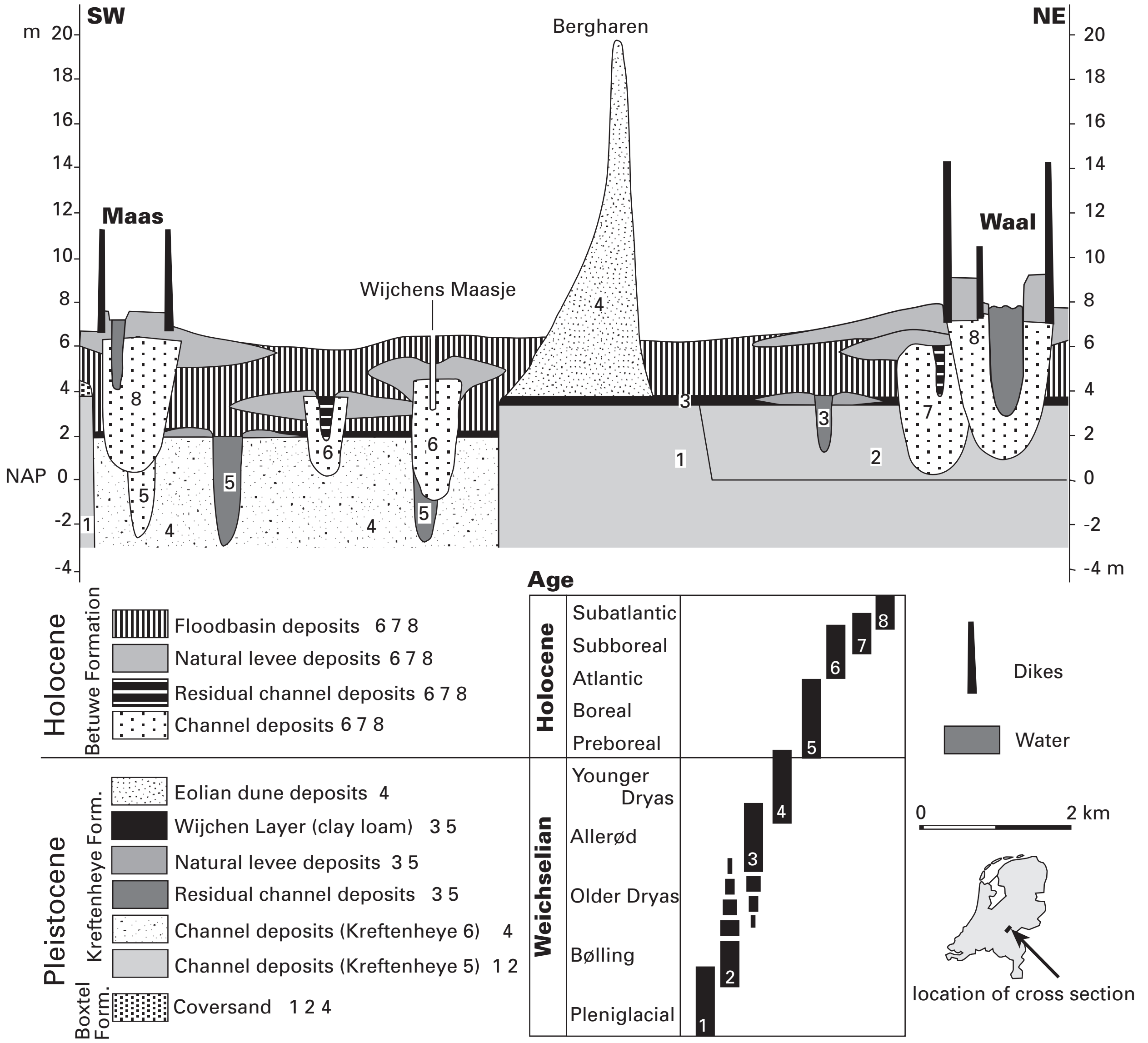

Figure 28 Schematic geological cross-section from the Meuse to the Waal near Bergharen (after Berendsen et al. 1995). The difference in elevation between the Lower terrace (Kreftenheye-5) and terrace X (Kreftenheye-6) here is about 1.5 m. The Younger Dryas eolian dunes of Bergharen overlie the Wijchen Layer of Allerød-interstadial age, that occurs on top of the Lower terrace.
SW of the river Maas the Wijchen Layer on top of the Lower terrace is virtually lacking. This is a result of southwards increasing uplift of the Peel Horst. The embanked floodplain of the Waal is at a higher level than that of the Maas, as a result of higher discharge and sediment load of the Waal.
The Weichselian Late Glacial river deposits in the substrate consist of coarse grained sand with some gravel. They were formed by braided rivers in at least two stages: the Pleniglacial Terrace (1/2), and at an approximately 1.5 m lower level the Younger Dryas Terrace X (4). In between a meandering stage occurred (3). The dune of Bergharen was formed during the Younger Dryas. It consists of sands blown out of the periodically dry braided-river plain to the southwest (Younger Dryas terrace, known as Terrace X). During the Preboreal and Boreal rivers were deeply incised, meandering streams (generation 5). Some of these channels have been completely filled up with gyttja, clay and peat. The sequence in these channel fills covers almost the entire Holocene. Elsewhere, active river channels became gradually filled with sand. In the middle-Atlantic the river channels were filled up to the level of the Late-Weichselian terrace level, and accumulation by meandering rivers started on top of Terrace X. Sedimentation on top of the Lower Terrace began at an even later stage. Sea level rise caused the 'terrace-intersection' between Lower Terrace and Holocene deposits to move progressively upstream (eastwards).

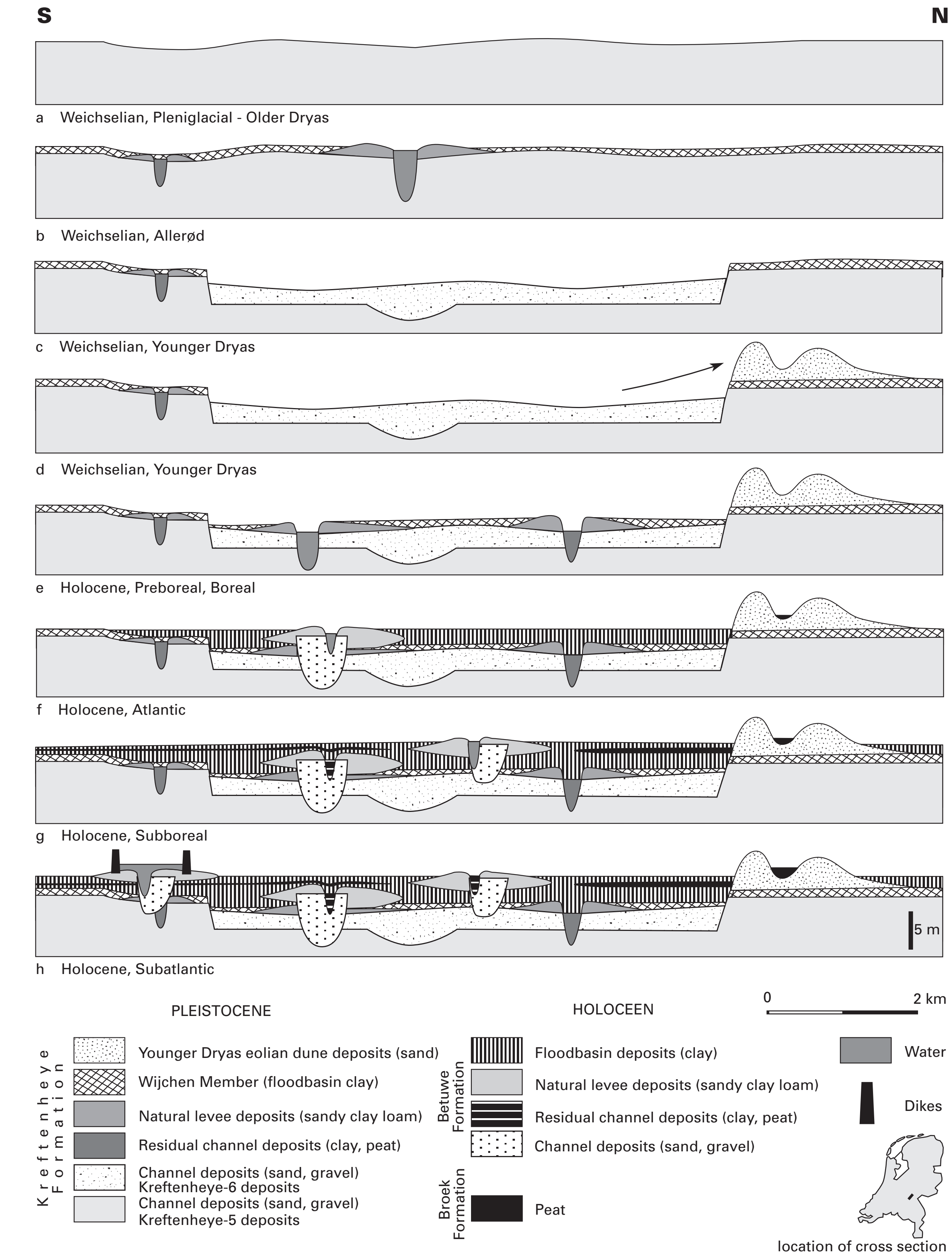

Figure 29 Schematic diagram of the evolution of the river Maas during the Weichselian Late Glacial and the Holocene (after Berendsen et al. 1995). During the Allerød-interstadial the floodbasin clay of the Wijchen Layer was deposited on the Pleniglacial braidplain. In the Younger Dryas incision took place by braided rivers. During dry periods eolian dunes were formed along the channel. In the Early Holocene the river pattern changed to meandering again, and a floodbasin clay (Wijchen Layer) was deposited on top of the Younger Dryas braidplain. During the Atlantic aggradation started, indicating that the section was located downstream of the terrace intersection of Holocene deposits and the Pleniglacial Lower terrace. The pleistocene surface slopes from approximately 10 m + NAP (NAP = Dutch Ordnance Datum) near the German border to 25 m - NAP near Rotterdam (= approximately 22 cm/km). The slope of the present river plain is 10 cm/km. Pumice granules were deposited by the Rhine in the Younger Dryas terrace after the Laacher See volcanic eruption approximately 11000 yr B.P.

The Younger Dryas (buried) terrace that occurs to the SW of the dune is incised relative to the Pleniglacial terrace level. The Younger Dryas terrace was the source of the sand for dune formation, especially during the very dry last 500 years of this period. The Younger Dryas terrace contains pumice, derived from the Laacher See volcanic eruption in the Eifel region (Germany). The pumice occurs in an approximately 1 m thick layer, and even shows tectonic deformation where the terrace overlies the Peel Horst (Figure 18). The eruption occurred 11,063 ± 12 ^{14}C yr BP (Friedrich et al. 1999).
The evolution of the Late Glacial terraces is shown schematically in Figure 29.
The terrace intersection between the Younger Dryas and the Pleniglacial terrace presumably occurs near Rotterdam (Figure 30). Both terraces consist of *continuous sands,* containing gravel, of the Kreftenheye Formation.
In the floodbasin to the NW of the dune, young Holocene flood basin deposits are approximately 1.5 m thick and consist of clay directly overlying the Wijchen Layer of Bølling-Allerød-interstadial age (Figure 28). The Holocene floodbasin clay has an admixture of wind-blown sand. Holocene channel belts all incised into the Pleistocene substrate. They act as water bearing corridors within the Holocene deposits.

The Holocene onlaps over the Pleistocene, and the stratigraphic hiatus at the Pleistocene/Holocene boundary here represents a period of approximately 5000 years of parasequential progradation in the basin. The hiatus postdates the Wijchen Layer of Bølling-Allerød-interstadial age, and it predates the Holocene Betuwe Formation, which has virtually the same facies as the Wijchen Layer. The top of the Wijchen Layer forms the transition from a lowstand (LST) to a transgressive phase (TST).

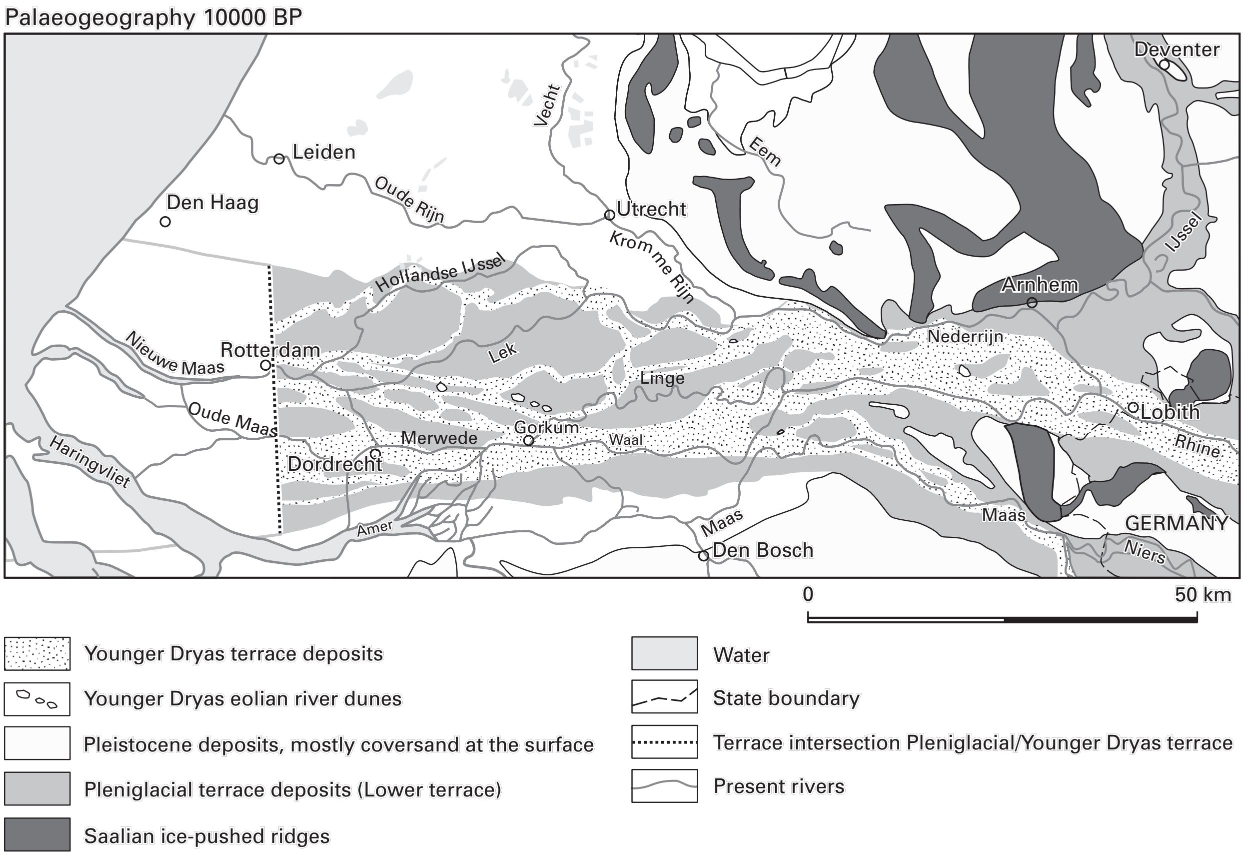

Figure 30 Palaeogeography around 10,000 BP, showing Pleniglacial and Younger Dryas braidplains (after Berendsen & Stouthamer 2000).

STOP 3: Residual channel

Observations

At Maasbommel (Figure 31) a residual channel and point bar of the recent embanked river Meuse (=Maas) is visible from the dike. The view is from the north to Ooijen. The topography of point bars and swales is seldom visible in the Netherlands, because the terrain is almost always leveled for agriculture.

Figure 31 Map of the embanked floodplain of the river Maas around 1850 near Maasbommel and present location of the Maas residual channel (topographical maps 1: 50.000).

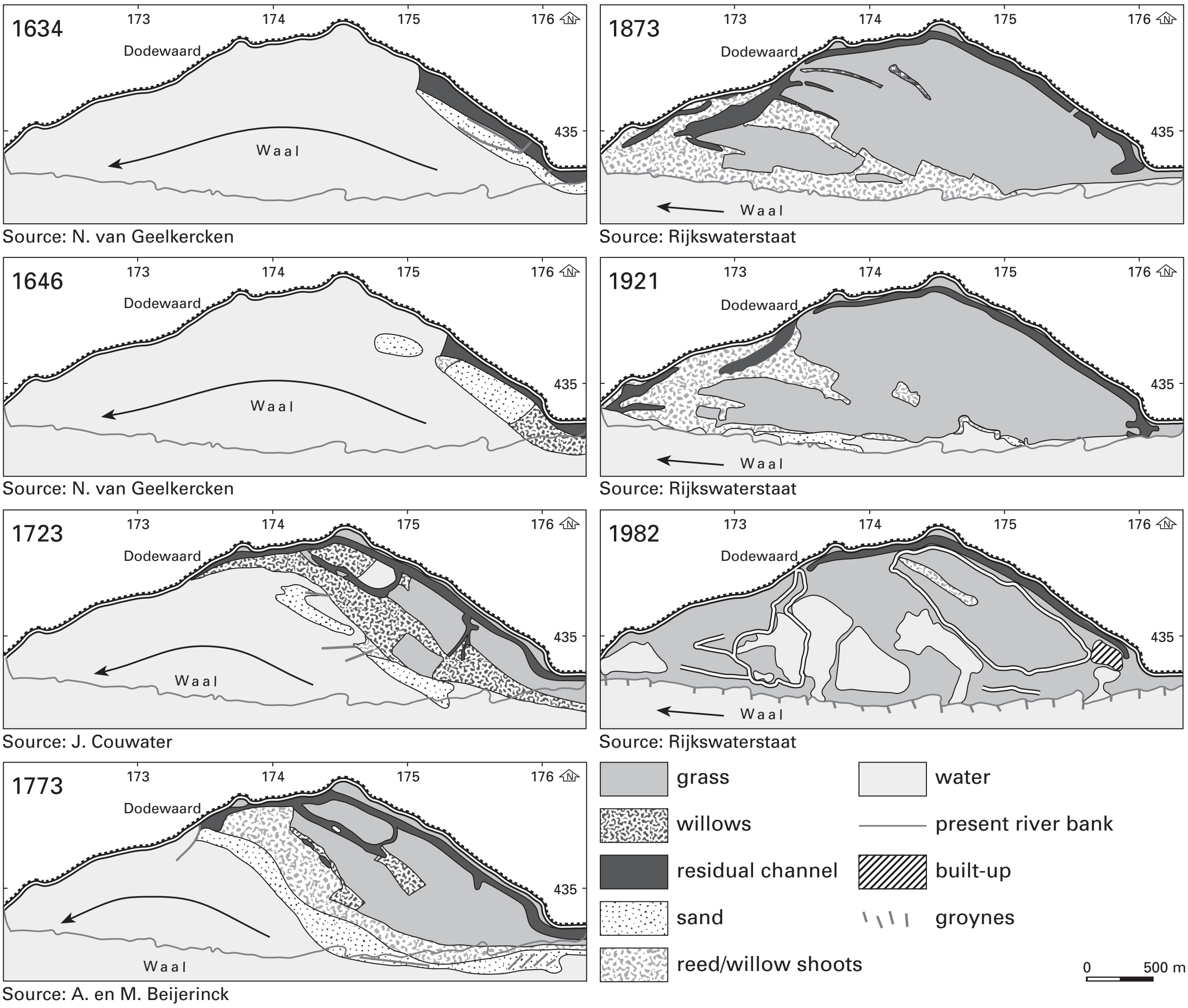

Figure 32 Formation of embanked floodplains according to Middelkoop (1997). The most recent fluvial sediments along the major rivers in The Netherlands are found in the embanked floodplains.

The lithology of these floodplains mainly consists of sandy clays (overbank deposits) covering coarse sand (channel deposits). The geomorphology shows a typical point bar geomorphology of scroll bars and swales, and some residual channels. Analysis of historic river maps (the oldest from 1600 AD) has given a detailed insight in the genesis of a series of floodplains along the rivers Waal and Meuse. Sand bars, formed in the river beds, were planted with willows, which enhanced the trapping of sediments. New sand bars were successively formed in a downstream direction. Although this was initially described as an example of human-induced counter-point accretion, the formation of the sand bars may also be explained as the tail of a normal point bar.
The development began when a high sand bar or an island in the river bed was formed close to the river dike (1634). Dams were built to connect the island to the land. The channel between the former island and the dike was closed off and gradually filled up with clay. During several phases (1723-1921) the floodplain became larger when new islands were reclaimed and merged with the floodplain. The vegetation shows a sequence, related to the geomorphological development of the floodplains: in the beginning, the island was covered with a natural vegetation of reed and willow shoots (1723, 1773). After a few decades this vegetation was replaced by willows, while reed and young willows were still growing in the low and wet parts. Finally, the willows were cut and were replaced by grasslands (1873, 1921). This succession of vegetation was repeated for each new phase of embanked floodplain growth. The resulting morphological pattern of these floodplains therefore was formed during subsequent stages of floodplain growth under the influence of humans. It is not reflecting a former braided river or a set of parallel active channels in the past.

Geology/geomorphology/sandbody geometry

The residual channel fill is 6 m thick and consists of clay and peat. The channel was artificially cut off approximately 1860 (Figure 31). The natural levee deposits (sandy and silty clay) typically are 0.5 to 1 m thick and overly channel deposits, consisting of (coarse) sand.
Reconstructions from historical maps (dating back as far as 1600 AD) show that the evolution of the embanked floodplains is strongly influenced by humans. However, this evolution also led to more or less parallel swales (Figure 32) on the point bars. Many embanked floodplains were formed only during the last 3 or 4 centuries. Their development has been strongly influenced by humans: in fact they can be seen as the result of land reclamations from the river bed. Figure 32 schematically shows the step-wise development of such a floodplain of the river Waal (Middelkoop 1998).

STOP 4: Younger Dryas dune

Observations

The ‘Dreumelse Berg’ eolian dune stands out as a solitary ‘hill’ in the otherwise featureless floodplain.

Geology/geomorphology/sandbody geometry

The eolian dune ‘Dreumelse Berg’ is of Younger Dryas age. It is partly covered by Holocene floodbasin clays, and stands out as a solitary dune. The dune overlies the Wijchen Layer, a clay layer, interpreted as a floodbasin deposit of Bølling-Allerød-interstadial age (Figures 28 and 29). These floodbasin deposits are at the top of the Pleniglacial (braided river) terrace, that consists of sand and gravel. The dune is actually located on the Peel Boundary Fault. Sand was blown out of the Younger Dryas terrace, located to the SW of the dune. See also description at STOP 2.

STOP 5: Former confluence of Waal and Maas

Observations

Heerewaarden is near the former confluence of the rivers Waal and Maas (Figure 31). Three natural channels connected the rivers approximately 300 years ago (Berendsen 1986). Some of the old channels that connected the two rivers are still visible in the terrain. The fortress near the 'Nieuwe Kanaal van St. Andries' was part of the 'Nieuwe Hollandse Waterlinie', a defense structure consisting of fortresses and polders that could be inundated. The fortress was built in 1816 AD. The defense structure, that roughly runs from Amsterdam to 's-Hertogenbosch, became obsolete after the invention of the airplane.

Hydrology/geomorphology/human influence

Because water levels of the Waal are on average approximately 2 m higher than those of the Maas, the Waal often lost water to the Maas. However, further downstream there was a confluence of Waal and Meuse. This caused serious water control problems in the Bommelerwaard-polder, in between the two rivers, resulting in numerous dike breaches. The problems were solved in 1904, when a dike was constructed that separated the two rivers near Heerewaarden. At the same time a new channel was dug from the Meuse to the Haringvliet (the Bergsche Maas, giving the Meuse its own outlet to the sea) and the old channel leading to the confluence with the Waal (the Afgedamde Maas) was dammed. Also, a few large meanders were cut off. This led to a dramatic amelioration of drainage control in the polder Bommelerwaard: average water levels in the Meuse were lowered by 2 m.

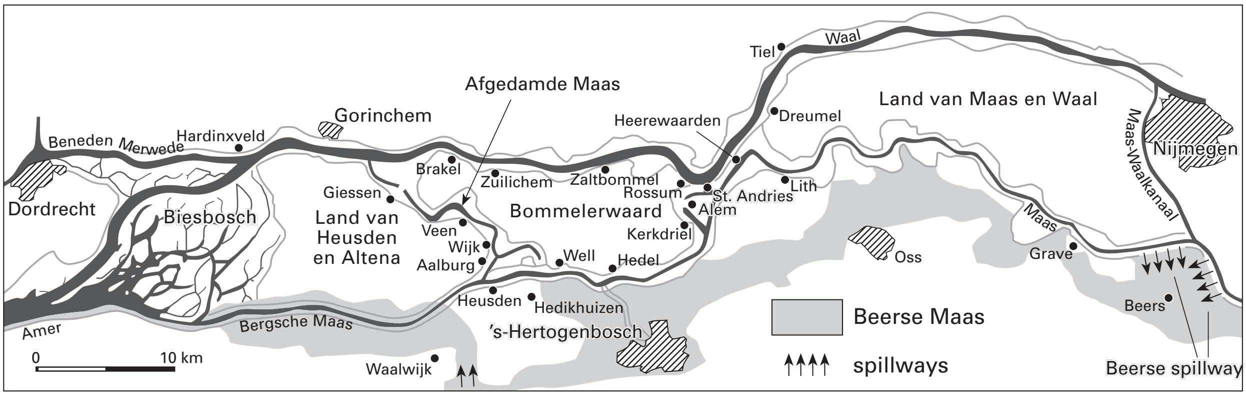

Figure 33 The former confluence of Waal and Maas near Heerewaarden (from Berendsen ed. 1986). Originally three natural connections between Waal and Maas existed near Heerewaarden. This caused serious water problems in the Bommelerwaard polder (see text). In 1904 the rivers were separated by a large embankment (dike), and a new course was dug for the Maas (the Bergsche Maas) to the Amer (a tidal creek). At the same time the Afgedamde Maas was dammed. This finally solved the problem of water control in the Polder Bommelerwaard.

STOP 6: Dike reconstruction

Observations

At many places, different dike levels can be seen, e.g. near Restaurant 'De Oude Molen', which is at the dike level of 1861. A higher dike lies in front of the restaurant, and the latest reinforcement is visible at the riverside (3 different dike levels in one location!). Many houses still have their front door at the dike level of 1861 (the last time the Bommelerwaard polder was flooded). Now the dikes are almost everywhere at the desired 'delta' level. At some places a new dike has been constructed in front of the old dike, to preserve houses.

Geology/Geomorphology/Hydrology

The dike between Rossum and Zaltbommel (Figure 33) is on a sandy substrate of a subrecent channel belt. During floods, seepage was excessive, leading to dangerous situations. Therefore the dike between Rossum and Zaltbommel was reinforced between 1990 and 1995.
Dike reinforcement in the Netherlands has for long been a controversial subject, because the original reinforcements involved removing houses that were built on top of the dikes, straightening and widening of the dikes, removal of trees, ponds, etc. In many places, this resulted in a dull landscape. In the 1980's it was agreed to maintain the original situation whenever possible, and to use 'keen designs' at places where old farmhouses, churches etc. have to be preserved. The 1993 and 1995 floods of the Rhine and Meuse (which almost resulted in disaster) also helped to diminish public opposition to reinforcements.

As a result of the increased greenhouse effect, discharge of the Rhine is expected to increase by approximately 15 % during the winter, and to decrease during the summer by the same amount. To accommodate higher floods, new plans are being developed. These do not involve the construction of higher dikes, because that would also increase risks of dike bursts. Instead, the embanked floodplains will be excavated to a depth of 1-2 m, and obstacles in the embanked floodplains will be removed. In this way a peak discharge of 17,000 m^3 at Lobith can be accommodated (the peak discharge measured so far is 12,000 m^3 at Lobith). It may be necessary to take additional measures, and increase the capacity to a flood of 18,000 to 20,000 m^3 at Lobith. In that case large temporary storage basins or spillways into the embanked areas would have to be constructed. In this densely populated area these measures will meet great resistance from the local population.

LUNCH

STOP 7: Dike-breach scour hole

Observations

The 'Wiel van Bassa' (Figure 34) is the largest dike-breach scour hole in the Netherlands.

Geology/Hydrology

Originally the dike breach scour hole was 22 m deep. The Diefdijk, which runs N-S from the river Lek to the river Linge, was meant to protect the area west of it from inundation. However, during inundations of the upstream area, the dike was breached here several times. The reason for this is the presence of a sandy channel belt in the substrate, that carried sub-surface water flow, thereby undermining the dike. The scour hole follows a meander in the channel belt, because the sandy channel deposits could be eroded more easily than the tough floodbasin clays. Some typical permeability data for architectural units in the Rhine-Meuse delta are given in Table 2.

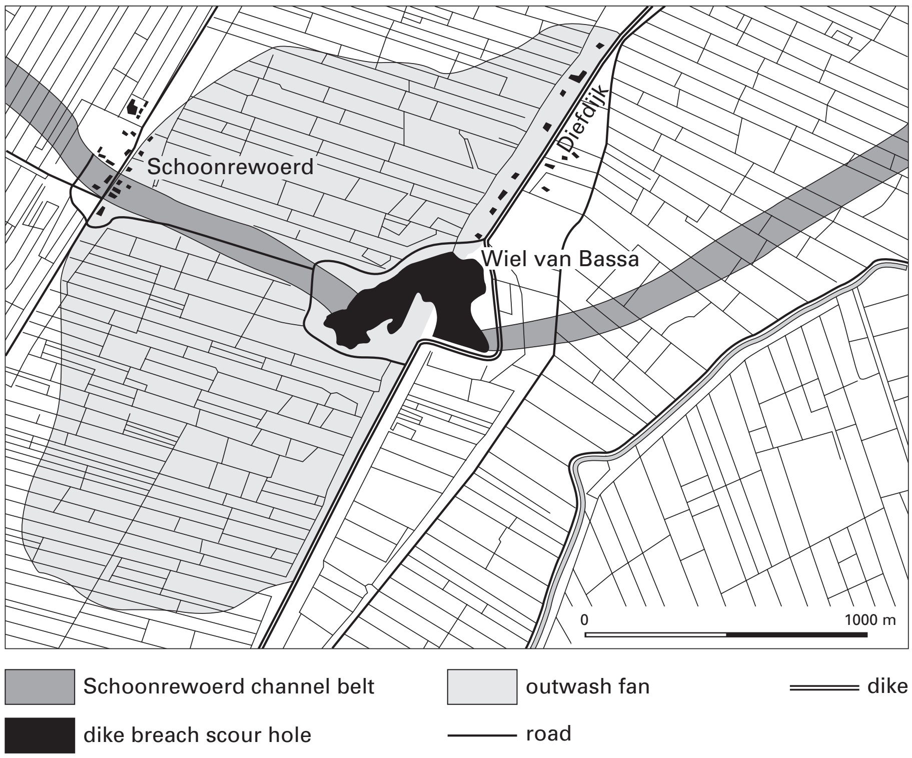

Figure 34 The 'Wiel van Bassa' is a 22 m deep dike breach scour hole, formed in the Schoonrewoerd channel belt (after Verbraeck 1984, from Berendsen 2004). The permeable channel deposits generally cause seepage water to flow from the rivers into the reclaimed farmland. In some areas, 75 % of the excess water is caused by seepage. Permeability of various deposits (k-value, in mm/day) is given in Table 2 (after Weerts & Bierkens, 1994).

Table 2 Permeability of various deposits (k-value, in mm/day, Weerts & Bierkens 1994).

Lithogenetic unit	**k-value**
Floodbasin deposits (clay)	< 0,1-10
Floodbasin deposits (peat/peaty clay)	5-100
Residual channel deposits (clay)	0,1-100
Natural levee deposits (silty/sandy clay)	10-100
Channel deposits (sand)	1000-50000
Pleistocene eolian dunes	500-20000
Pleistocene (braided) river deposits	> 20000

STOP 8: Late Holocene meandering channel belt

Observations

The 'Stuivenberg' channel belt near the town of Montfoort has a point bar, swales, and a residual channel that are still visible in the terrain.

The 'Van der Staay suction corer' will be demonstrated on the sandy point bar. Cores shown:

1. Point bar deposits of meandering Stuivenberg river system.
2. Residual channel deposits of Stuivenberg river system.
3. Natural levee deposits overlying flood basin deposits.

Geology/Geomorphology/Sandbody geometry

The Stuivenberg channel belt (Figure 35 and 41) that existed from 3960 – 3180 ^{14}C yr BP, was a meandering river system, with a clear point bar morphology (for terminology see Figures 36 and 37). The residual channel is approximately 3.6 m deep and is filled with clay and thin peat layers. A cross section of the residual channel (although at a different location) is shown in Figure 38. At the coring location, the residual channel base overlies overbank deposits of an underlying channel belt.
Figure 39 shows a typical fining upwards sequence of this channel belt and the overlying natural levee.
Just west of the site, a large crevasse-splay complex is present, that was interpreted as a failed avulsion (Stouthamer 2001). The final phase of the crevasse-splay complex was radiocarbon dated at 3500 yr BP (Berendsen 1982). The principle of radiocarbon dating channel belts is illustrated in Figure 40 (after Berendsen 1982 and Berendsen & Stouthamer 2001).
South of this area, older channel belts of the Graaf river system (Figure 41) are found that have a distinctly different channel type. They are considered as 'straight anastomosing' channel belts, and are characterized by extensive crevassing. The difference between meandering channel belts and straight anastomosing channel belts is illustrated by Figures 41 and 42. The straight anastomosed channel belts have a low (< 15) width/thickness ratio of the sandbody, and hardly any lateral accretion. This means that the sandbody perimeter matches the channel perimeter. Therefore, straight anastomosed channel belts allow to make discharge estimates (Makaske 1998), compare Figures 13 and 15.
Overbank deposits of the straight anastomosed channel belts are extremely complicated in planform (Figure 43) and cross section (Figure 44), and include extensive crevasse splays. Contrary to Törnqvist (1993), avulsions of these channel belts were not very common (Stouthamer 2001). This may be explained by low downstream and cross valley gradients (no gradient advantage existed), and cohesive river banks, consisting of clay and wood peat. It is believed that the straight anastomosed river systems developed when eustatic sealevel rise neared its highstand (Figure 13); this coincided with a minimum in the avulsion frequency (Figure 16).
Many avulsions in the Rhine-Meuse delta seem to be associated with the formation of a large crevasse-splay complex (Figure 45), that develops over time and space into a singular channel (Smith 1989, Makaske 1998, Stouthamer 2001, Berendsen & Stouthamer 2001). However, if the avulsion

succeeds, much of the original splay is eroded away, and in some cases (especially when discharge of the avulsed channel increased), hardly any splay deposits are preserved. Avulsion locations and frequencies in the Rhine-Meuse delta are related to sealevel rise, tectonics, discharge and sediment load (or sedimentation).

After 3000 yr BP avulsion frequency increases as a result of the combined effect on increased discharge and within-channel sedimentation (Stouthamer & Berendsen 2001), which may indirectly be related to the decreased rate of selevel rise (Figure 46).

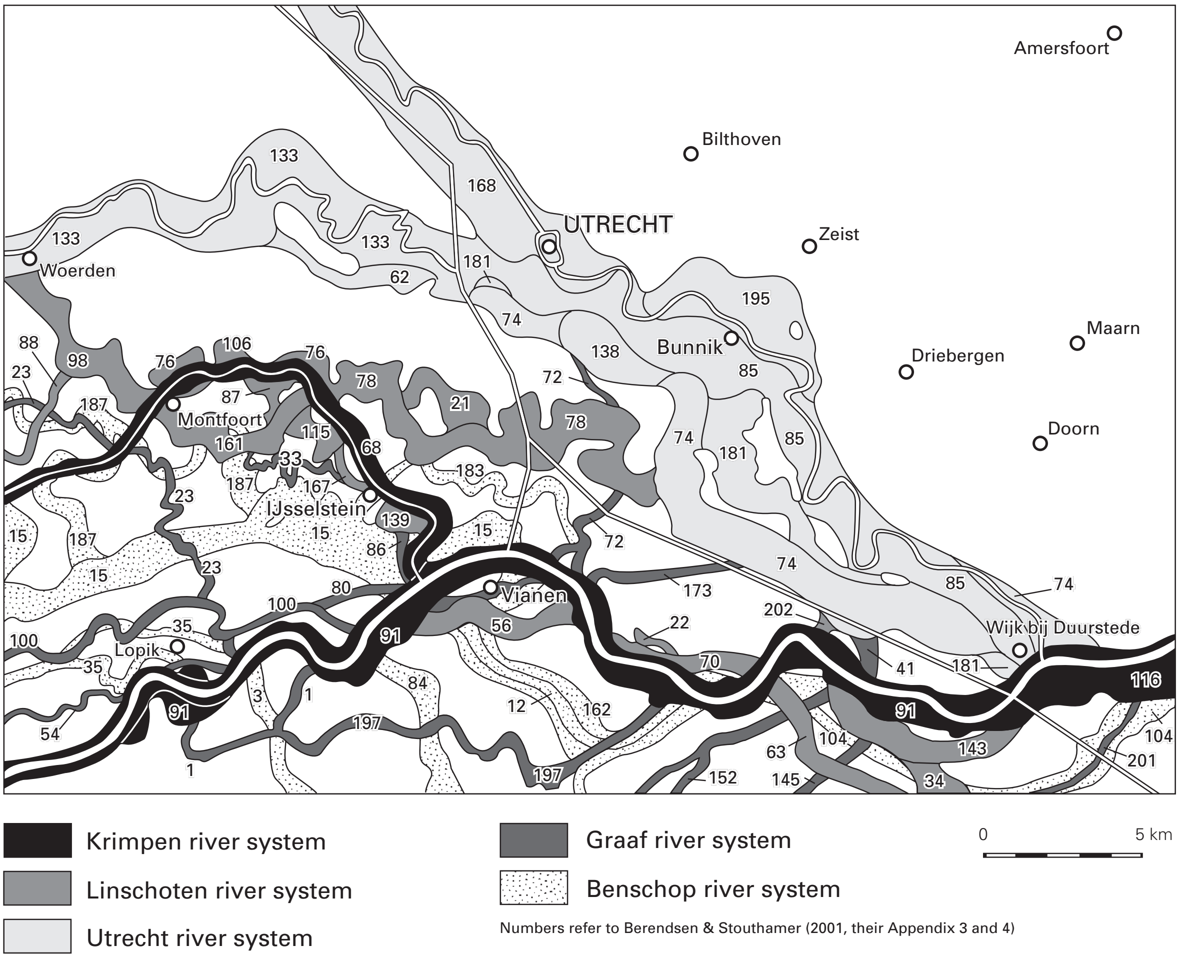

Figure 35 Map of different generations of Holocene channel belts in the vicinity of Utrecht (Berendsen 1982). The Benschop river system consists of channel belts that are older than 5300 ^{14}C yr BP. The Stuivenberg channel belt (161) near the city of Montfoort is shown in detail during the excursion.

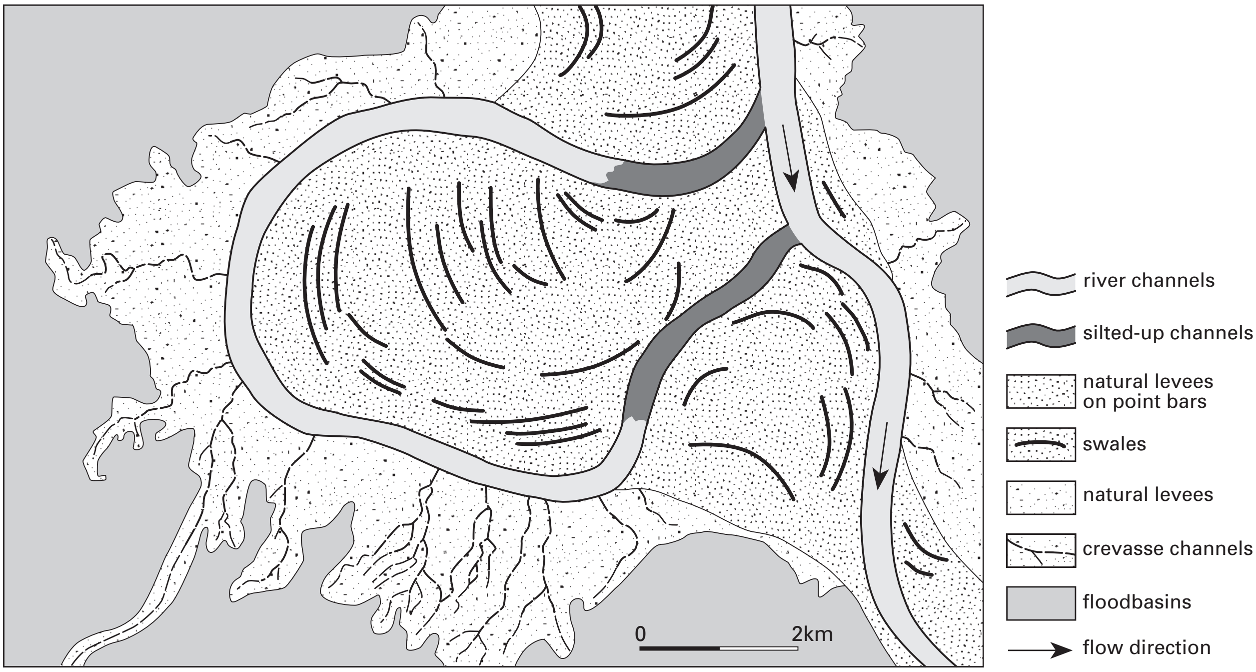

Figure 36 Schematic map of a meandering river with terminology (After Reineck & Singh 1971, from Berendsen & Beukenkamp 1983).

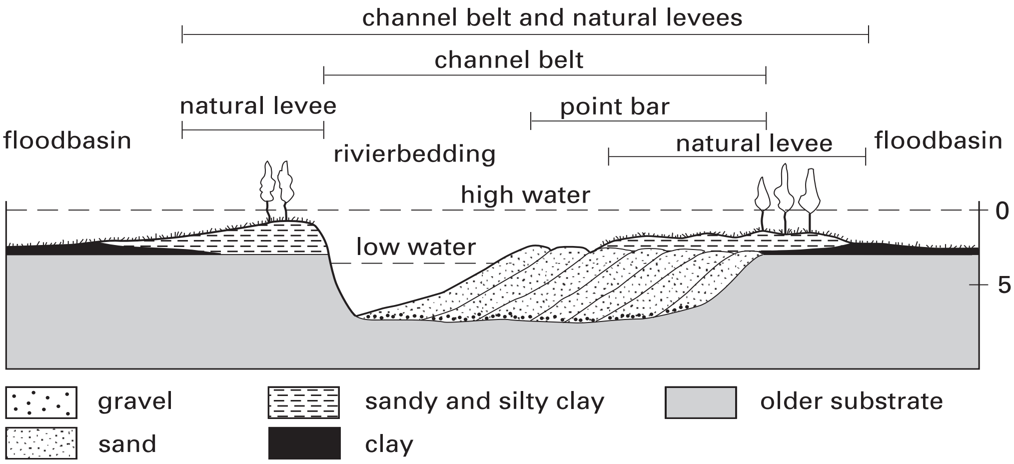

Figure 37 Schematic cross section of a meandering river with terminology (After Reineck & Singh 1971, from Berendsen & Beukenkamp 1983).

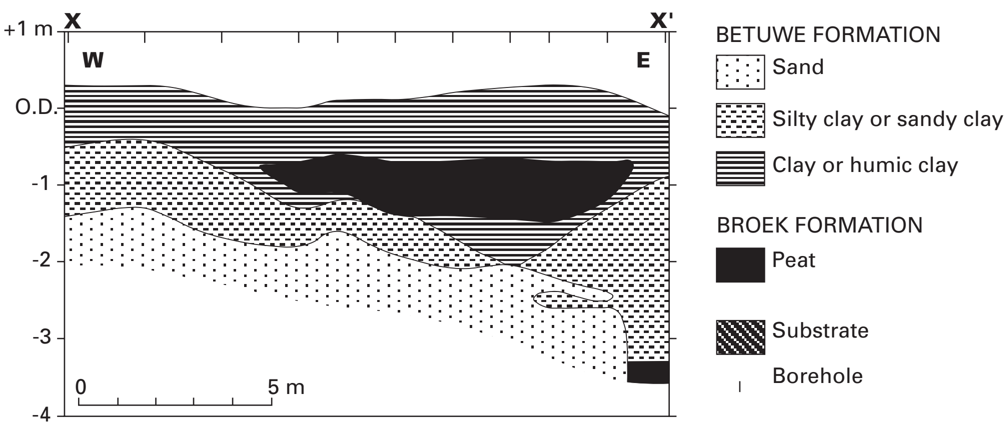

Figure 38 Cross section of the residual channel of the Stuivenberg channel belt (after Törnqvist 1993).

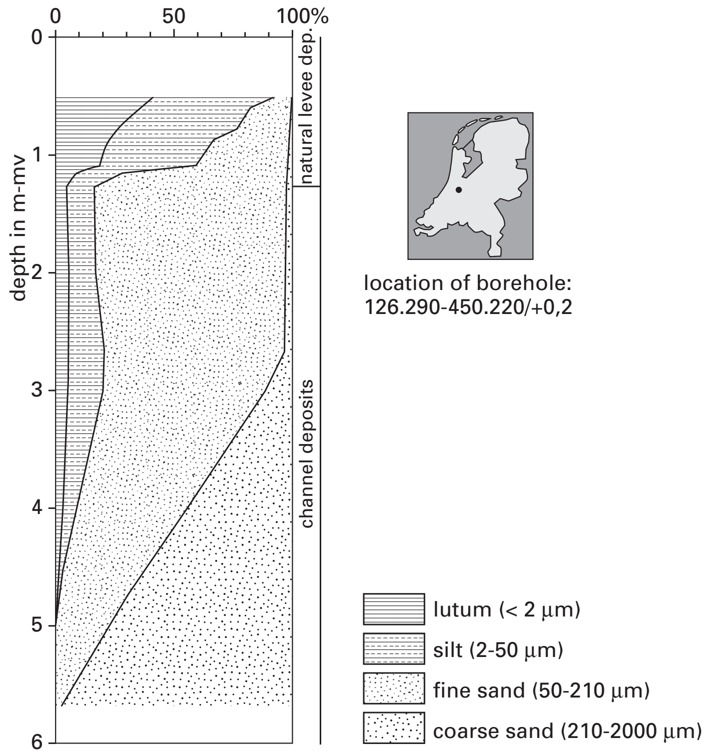

Figure 39 Grain size distribution of natural levee and point bar deposits of the Stuivenberg river system (Berendsen 1982). The curves show a fining upwards sequence. Depth is indicated in m below the surface (surface = mv). At a depth of 4 m below the surface median grain size is 420 μm; gravel content is 0.1 %.

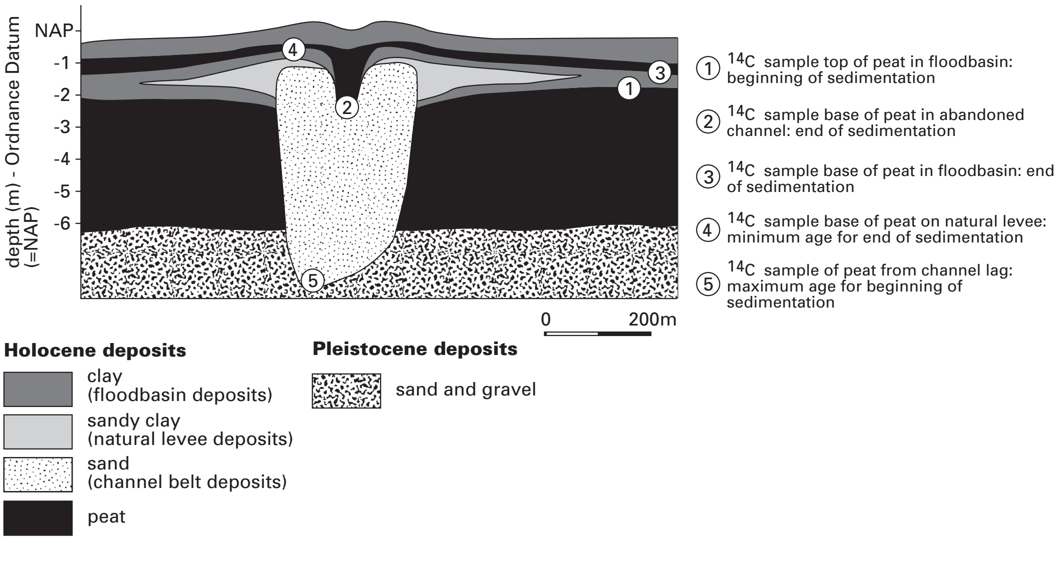

Figure 40 Simplified cross-section of a channel belt, illustrating the method of radiocarbon dating of channel belts (after Berendsen 1982). Sample 1 yields the beginning of sedimentation, samples 2 and 3 the end of sedimentation. Radiocarbon dates from the base of abandoned channels, sample type 2, have been extensively used for this study. Type 2 and 3 samples yield comparable ages, if type 3 samples are collected in the floodbasin and not on the natural levees. Sample type 4 gives a minimum age for the end of activity of the channel belt, sample type 5 (peat from channel lag deposits) yields a maximum age for the beginning of river activity.

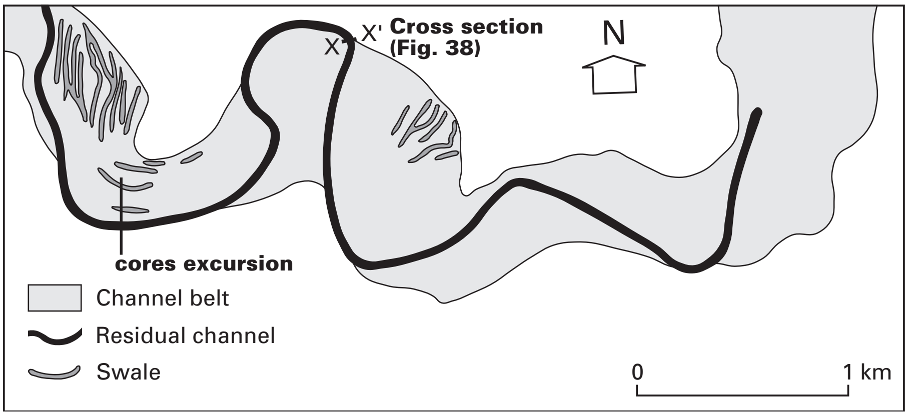

Figure 41 Channel belt planform of the meandering Stuivenberg channel belt in the Rhine-Meuse delta, after Berendsen (1982). The point bar and swale topography is locally still visible in the field. The beginning of peat formation in the residual channel was ^{14}C dated at 3150 yr BP.

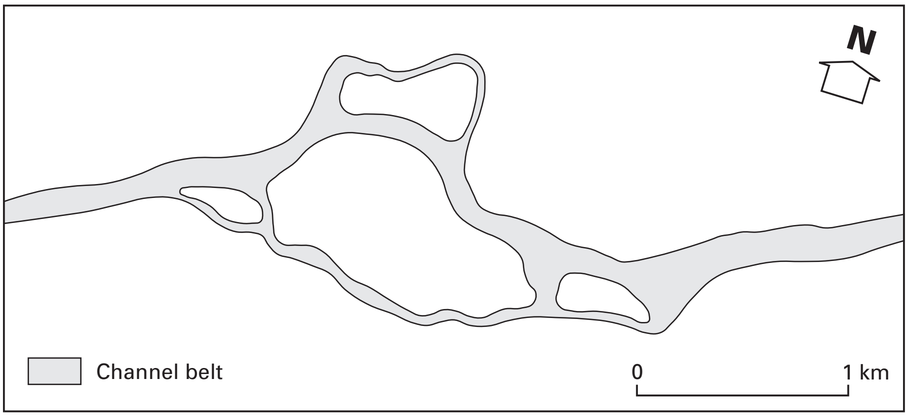

Figure 42 Channel belt planform of the straight anastomosing Schoonrewoerd channel belt in the Alblasserwaard (Törnqvist 1993

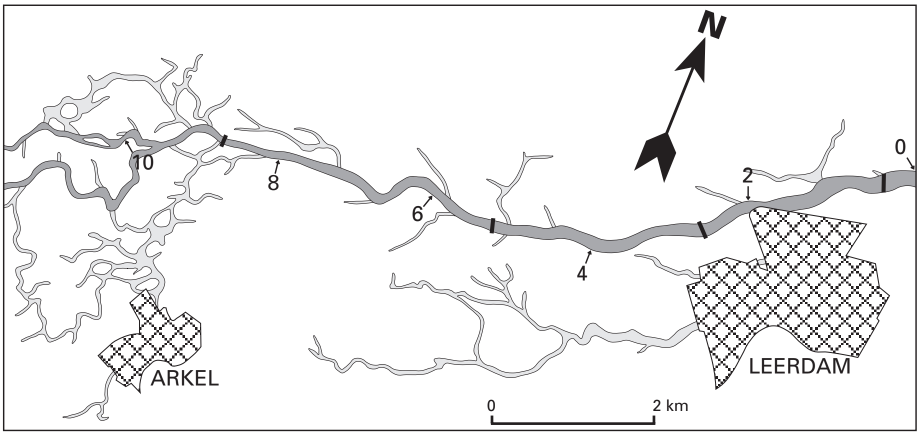

Figure 43 Straight anastomosing channel pattern in the Alblasserwaard-Vijfheerenlanden (east of Rotterdam), after Törnqvist (1993). This map shows channel deposits and crevasse-splay deposits of the Schaik system, a Middle Holocene anastomosing distributary system in the Rhine-Meuse delta. The map shows sediments of the uppermost 2-3 m and is based on approximately 6000 boreholes (60 boreholes km^{-2}). Note the significant increase of the number of crevasse splays downstream of the 8-km point.

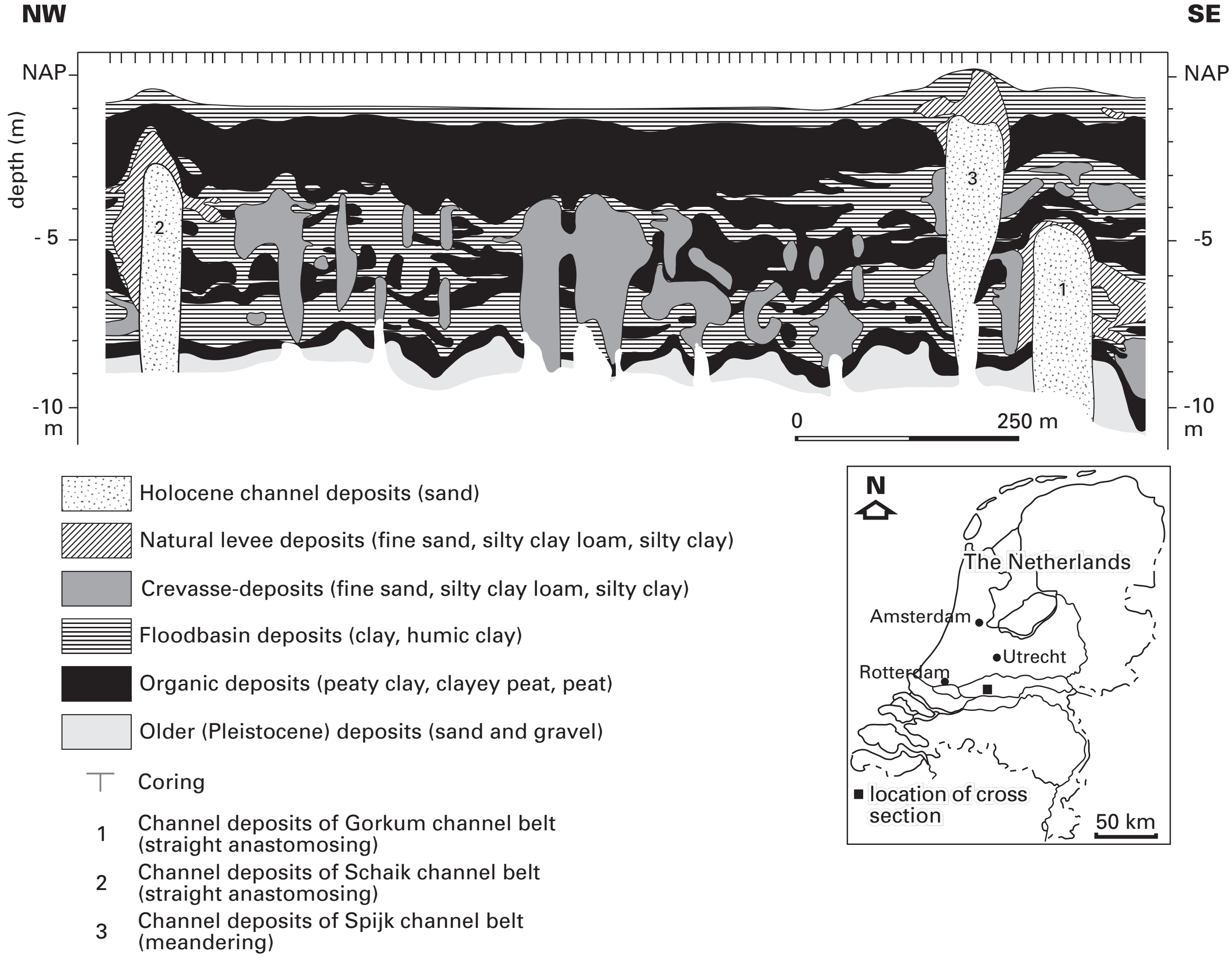

Figure 44 Cross-section of the anastomosed Gorkum and Schaik channel belts, showing complicated overbank deposits. In comparison the overbank deposits of the much younger meandering Spijk channel belt are simple; they wedge out laterally. The width/thickness ratio of the sandbody of all three channel belts is <15. After Weerts & Bierkens (1993).

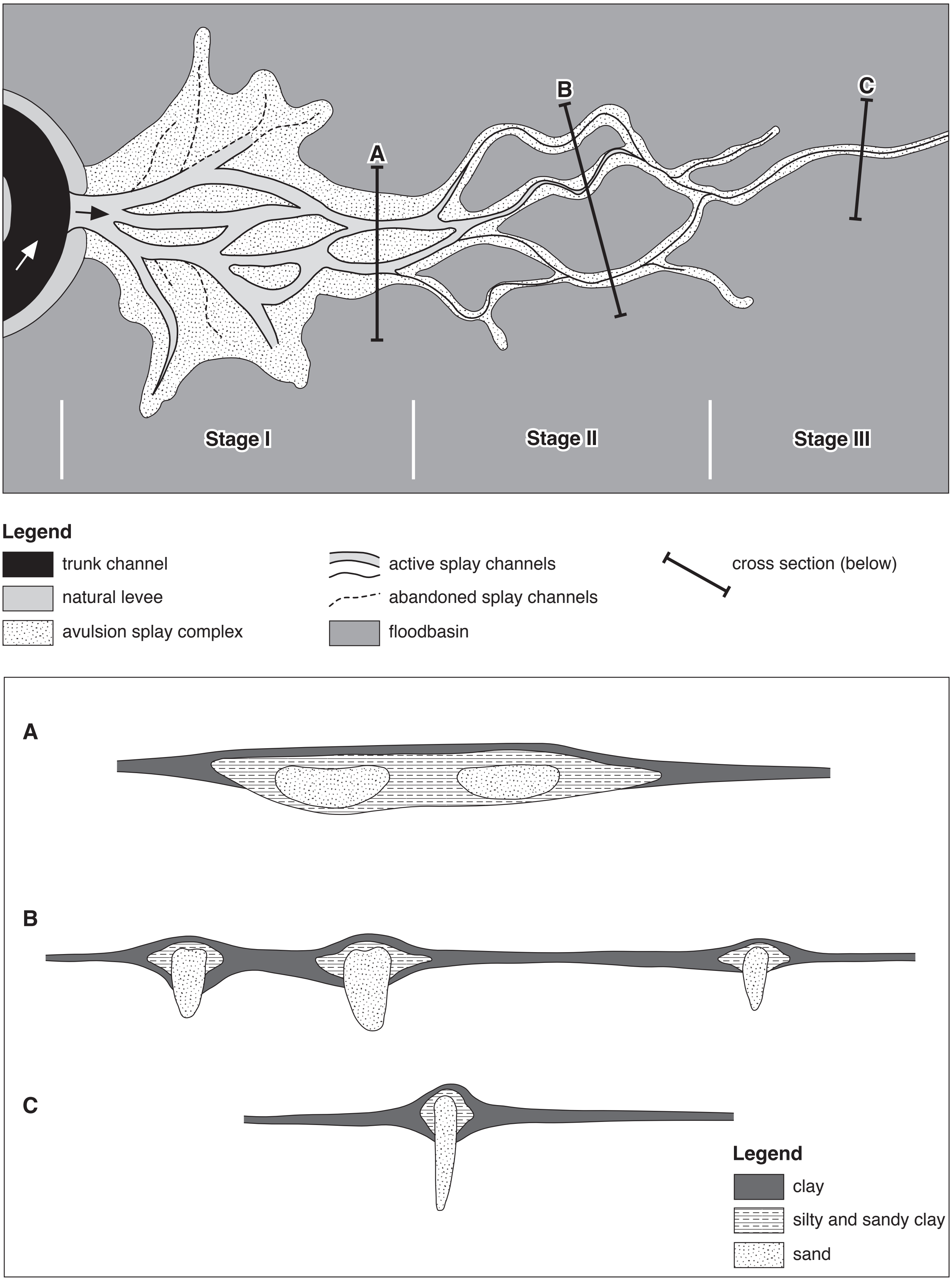

Figure 45 Schematic representation of the avulsion belt of the Schoonrewoerd channel belt (Makaske 1998). The cross-sections show, that in a longitudinal direction, the width/thickness ratio of the channel belt sandbody decreases. A similar situation is observed near the 'failed avulsion' of the Stuivenberg channel belt (Stouthamer 2001). When the avulsion is succesful, the crevasse-splay deposits near the avulsion site are often partly eroded by the new channel.

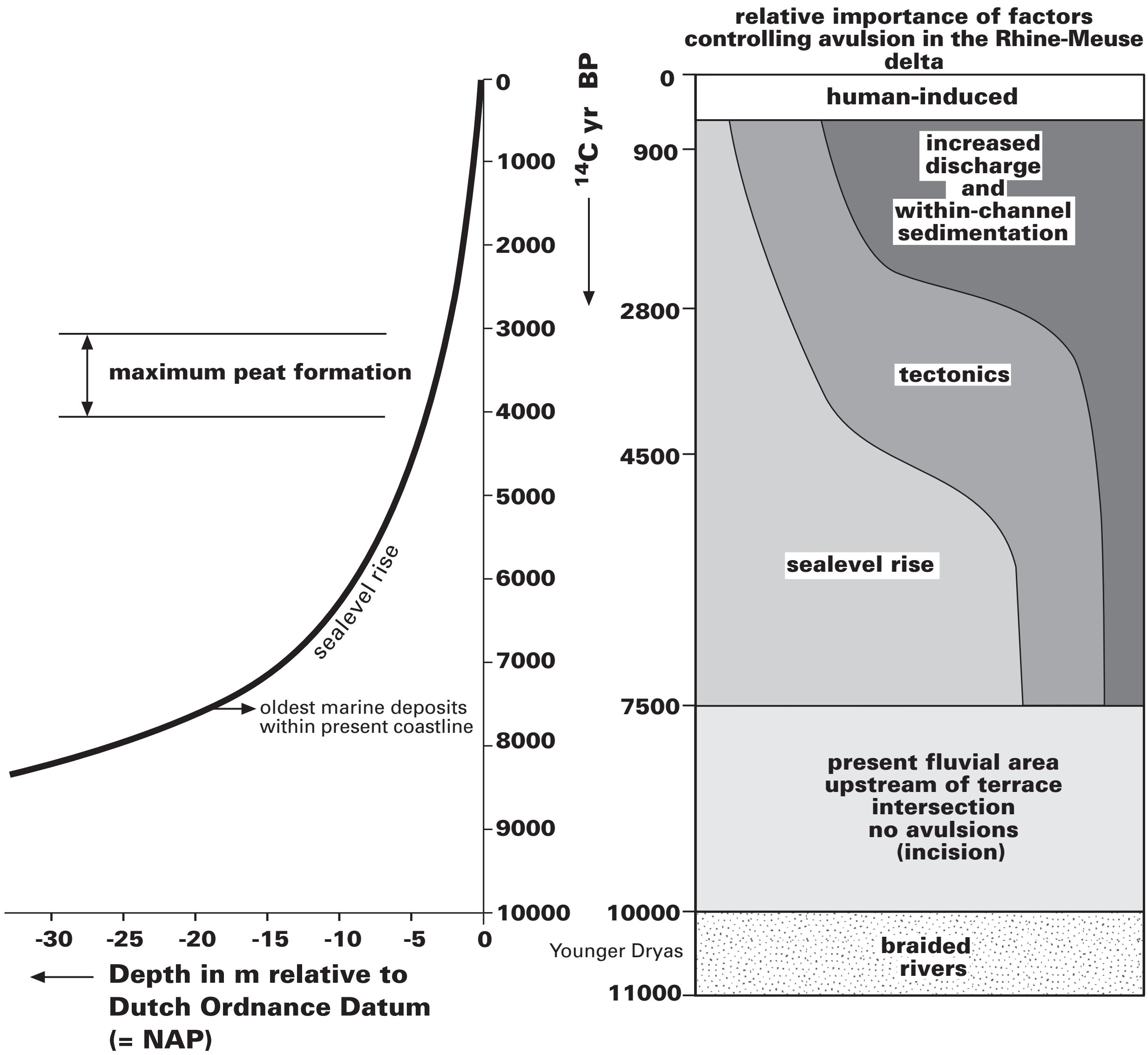

Figure 46 Summary of main factors controlling avulsions in the Holocene Rhine-Meuse delta (from Stouthamer & Berendsen 2000a). On a timescale of years and decennia, avulsions are thought to occur as a result of stochastic events, like high discharge, the presence of beaver and ice dams, log jams, hippo trails etc. On a longer time scale of the Holocene, Stouthamer (2001) showed that avulsion locations are determined essentially by sealevel rise, tectonic movements, and increased discharge and within-channel sedimentation.

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