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published in
European river activity and climatic change during the Lateglacial and early Holocene
1995

document version
Publisher's PDF, also known as Version of record

Paläoklimaforschung · Band 14

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European river activity and climatic change during the Lateglacial and early Holocene

Special Issue: ESF Project
European Palaeoclimate and Man 9

1995

Akademie der Wissenschaften und der Literatur · Mainz
European Science Foundation · Strasbourg
Gustav Fischer Verlag · Stuttgart · Jena · New York

Climatic change and fluvial dynamics of the Maas during the late Weichselian and early Holocene

Kees Kasse, Jef Vandenberghe & Sjoerd Bohncke

Summary

An outline is presented of the late Weichselian to early Holocene fluvial development of the Maas in the southern Netherlands. The changes in the river regime during the glacial-interglacial transition are related to changes in climate, vegetation, sediment supply and intensity and magnitude of discharge fluctuations. Aggradation prevailed during the late Pleniglacial. In the beginning of the Lateglacial incision occurred by a low-sinuosity river system. Incision continued during the Bølling, Older Dryas and beginning of the Allerød. By this time the Maas changed from a low-sinuosity into a high-sinuosity system and lateral migration became more important. Because of the climatic cooling at the start of the Younger Dryas, the peak discharges of the Maas increased and the river became braided again. This braiding phase continued until the end of the Younger Dryas and was accompanied by widespread river dune formation on the east bank of the valley in the second half of the Younger Dryas. The Holocene warming is reflected by a decrease in peak discharges and sediment supply. As a consequence the Maas incised and a new low-sinuosity channel established during the Preboreal.

Zusammenfassung

Ein Beitrag zur spätweichselzeitlichen und frühholozänen fluviatilen Entwicklung der Maas in den südlichen Niederlanden wird hier vorgestellt. Der Wandel im Abflußregime während des Glazial-Interglazial Überganges ist auf Veränderungen des Klimas, der Vegetation, der Sedimentzufuhr und von Intensität und Ausmaß der Wasserabflußschwankungen zurückzuführen. Während des Spätpleniglazials war die Sedimentaufschüttung der vorherrschende Prozeß. Zu Beginn des Spätglazials haben sich geringfügig mäandrierende Flüsse eingeschnitten. Diese vertikale Erosion dauerte bis zum Anfang des Allerøds an. Zu dieser Zeit wandelte sich die Maas in ein stärker mäandrierendes System mit zunehmender Bedeutung lateral-erosiver Prozesse. Die Temperaturerniedrigung am Anfang der Jüngeren Tundrenzeit hatte erhöhte Abflußspitzen und ein erneutes Verzweigen der Maas zur Folge. Die durch Verzweigen (braiding) gekennzeichnete Phase setzte sich bis zum Ende der Jüngeren Tundrenzeit fort und ging während der zweiten Hälfte der Jüngeren Tundrenzeit mit einer weitverbreiteten Bildung von Binnendünen am Ostrand des Tales einher. Die holozäne Temperaturerhöhung bewirkte eine Verringerung der Abflußspitzen und der Sedi-

mentzufuhr. Deshalb konnte sich die Maas während des Präboreals erneut einschneiden und eine schwach mäandrierende Rinne bilden.

1. Introduction

The inherent morphology of river floodplains, under stable climatic conditions, is determined by such factors as gradient, sediment load and discharge. Climate and tectonism can influence these factors. Tectonic movements are regarded as long-term and gradual processes and uplift is a prerequisite for erosion and terrace formation. The rapid Weichselian Lateglacial and early Holocene climatic evolution and related changes in vegetation, bank stability, sediment load and discharge are responsible for the rapid changes in river pattern and phases of incision and aggradation. The presence of distinct terrace levels indicates that the Lateglacial and early Holocene incision must have occurred in a stepwise fashion. In our opinion this stepwise incision had a climatic cause, superimposed on the general tectonic movements. These climate-induced fluvial developments were clearly established in small lowland valleys in the Netherlands and Belgium (VANDEN-

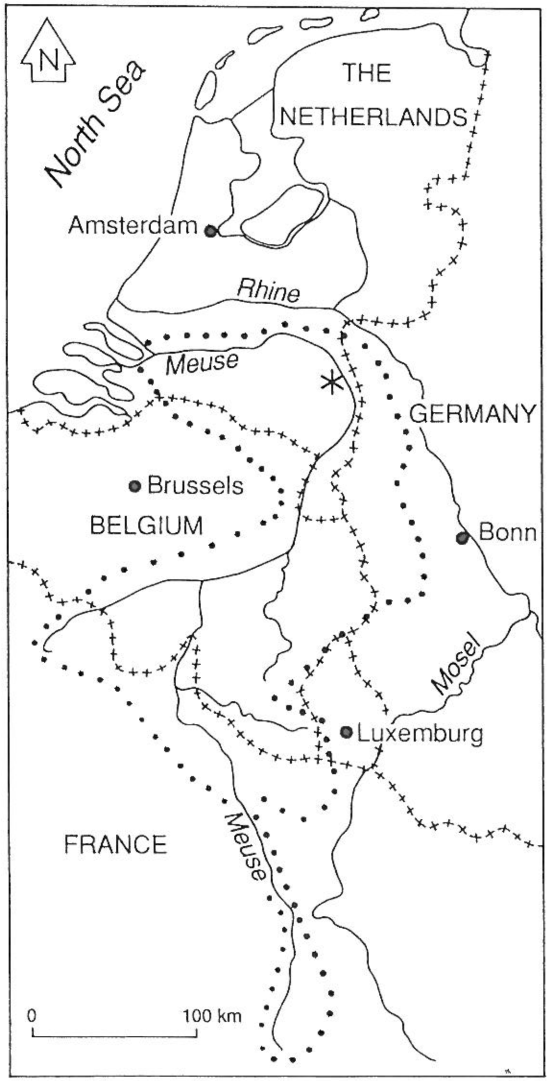

Fig. 1 The Maas (=Meuse) catchment. The study area is indicated by an asterisk

BERGHE et al., 1984, 1987; VANDENBERGHE, 1987; BOHNCKE et al., 1987). The present study aims at reconstructing the fluvial dynamics of a larger river system in relation to climatic change, in order to compare the results from small lowland valleys with those of a larger system like the Maas (=Meuse). This study is part of a project on the development of river systems in western and central Europe in relation to geographical variations in oceanity/continentality and vegetational development.

The late Weichselian to early Holocene development of the Maas has been studied in the southern Netherlands, north of Venlo, where well-developed Lateglacial terraces are present (Fig. 1). Previously, this region has been studied by SCHELLING (1951), PONS & SCHELLING (1951), PONS (1957) and VAN DEN BROEK & MAARLEVELD (1963). The latter authors established a terrace subdivision based on elevation and differences in soil development. Later, MIEDEMA (1988) reinvestigated the pedogenesis of the different terraces and concluded that the Lateglacial terraces are equally characterized by clay illuviation. Palynological investigations of channel scar fillings on different terraces were done on a local scale by TEUNISSEN (1983, 1990). BOHNCKE et al. (1993) investigated the palynology, periglacial structures and micromorphology of a Lateglacial sequence in exposure Bosscherheide. BERENDSEN et al. (this volume) investigated the fluvial development of the Maas immediately downstream of our study area.

2. Regional setting

The Maas rises at approximately 400 m above sea level in eastern France. It cuts through the Ardennes in Belgium, which are up to 700 m high (Fig. 1). North of Maastricht the Maas enters the southern North Sea Basin, which is characterized by a horst and graben structure (Fig. 2). The grabens contain a thick Quaternary fluvial sequence (up to 200 m in the Central Graben), which indicates strong subsidence during the Quaternary (ZAGWIJN, 1989). During the middle Pleistocene (Cromerian) the Rhine and to a lesser extent the Maas occupied the Central Graben (Sterksel Formation). Due to a strong uplift of the Rhenish Plateau in Germany during the late Cromerian (400,000 years ago), the Rhine changed its course to the north and formed the augite-bearing Urk Formation (ZAGWIJN, 1989). In the Central Graben the Rhine was replaced by the Maas, which deposited the Veghel Formation. During the Elsterian, Holsteinian and Saalian the Maas gradually shifted eastwards over the Peel Horst area into the Venlo Graben (VAN DEN TOORN, 1967; ZAGWIJN & VAN STAALDUINEN, 1975). The present-day Maas crosses the Central Graben and the Peel Horst almost at right angles, before it turns to the northwest in the Venlo Graben. The present river morphology in the study area is influenced by these tectonic structures. In the Central Graben the Maas has a strongly meandering course and a broad floodplain. On the Peel Horst the Holocene floodplain is nearly absent along the relatively straight, incised course. In the Venlo Graben the present river has a narrow floodplain and a low-sinuosity meandering course. The river gradient of the Weichselian Lower Terrace in the study area is 22 cm/km (PONS, 1954).

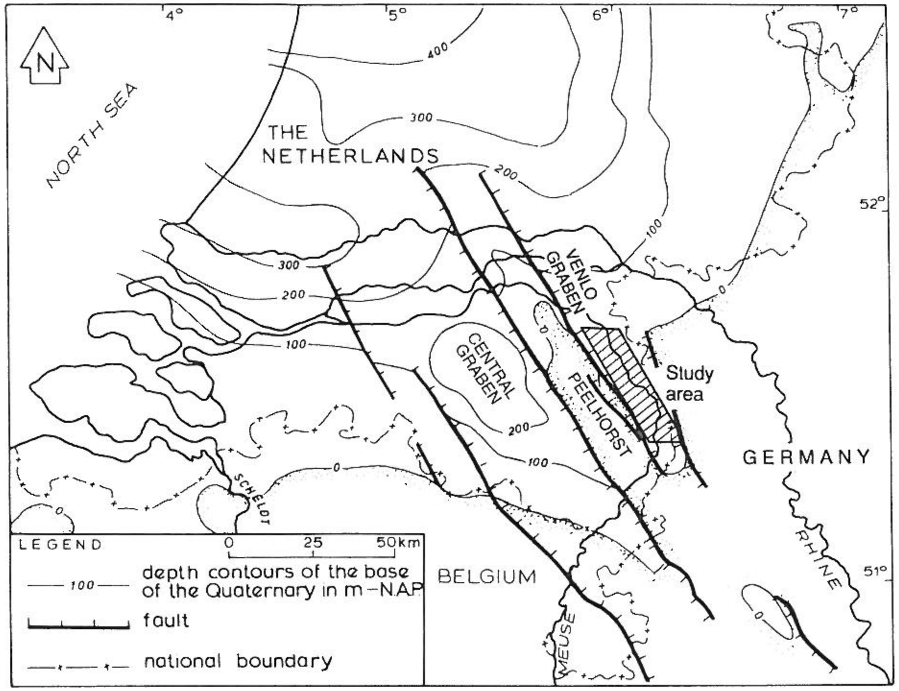

Fig. 2 Tectonic setting of the study area in the southeastern Netherlands (after ZAGWIJN & DOPPERT, 1978)

The Maas catchment is 33,000 km^2. The mean annual July temperature in the catchment area is between 15 and 18°C (17.5°C in the study area). The mean annual temperature in January is between 0 and 2.5°C (1.5°C in the study area). There is little snowfall in winter. Mean annual snow coverage varies from less than 10 days at the coast to 35 days inland. Hence the present-day Maas is a rain-fed river. The mean annual precipitation in the catchment amounts to 700 to 1000 mm (700 mm in the study area) and up to 1300 mm in the highest parts of the Ardennes. The maximum discharge is in January and the minimum discharge between July and September, but interannual variation in the discharge is very large (JONGMAN, 1987).

3. General terrace subdivision

Above the recent Holocene floodplain two late Pleniglacial, three Lateglacial and one early Holocene floodplain level have been distinguished (Fig. 3). The different levels generally have been subdivided on elevation and the presence of terrace scarps using aerial photographs and detailed topographical maps (1:10,000). However, the terrace surface morphology, defined by channels, bars and aeolian forms, has also been used to separate the different palaeofloodplain levels, for instance in those cases where the levels occur at more or less the same elevation. It is assumed then that areas with a similar surface morphology belong to the same genetic phase of river development, while other areas, although at the same topographical level, but with a different morphology, belong to another period.

Furthermore, the vertical sedimentary sequence, derived from exposures and borings, gives information about river type during terrace formation. The sedimentary sequence offers the opportunity to subdivide a terrace level even when height is not discriminating. In addition the depth of palaeochannels and the age of channel fills, obtained from pollen analysis and ^{14}C dates, have been used to distinguish the different levels and to correlate the terraces in a downstream direction.

4. Late Pleniglacial braided system

The late Pleniglacial floodplain has been subdivided in two levels, one with and one without late Pleniglacial aeolian coversands (Fig. 3: Terrace 1 and 2). The fluvial (surface) morphology is characterized by numerous, straight to slightly curved, very shallow channels, typical of a braided river system. The older late Pleniglacial terrace (Terrace 1) is underlain by a peat layer dated at 26,130±180 yr B.P. (peat base) and 25,200±180 yr B.P. (peat top) (WESTERHOFF & BROERTJES, 1990). In a section at Grubbenvorst the basal fluvial sands are characterized by isolated and intersecting, trough-shaped channels and by horizontally bedded fine sand and silt outside the channels. The channels, which are stacked upon each other, are slightly incised and show a concordant, often lateral filling (Photo 2). Towards the top, there is a gradual transition of fluvial to aeolian sands (fluvio-aeolian facies), culminating in the development of a deflation horizon ("Beuningen Gravel Bed") and finally the aeolian deposition of coversands (SCHWAN & VANDENBERGHE, 1991; HUIJZER, 1993; MOL et al., 1993). The Beuningen Gravel Bed serves as a widespread marker horizon (VAN DER HAMMEN et al., 1967), probably dating between 18,000 and 14,000 yr B.P. (KOLSTRUP, 1980), while the coversands at the top date from the end of the Pleniglacial (Older Coversands II).

The presence of syngenetic and intraformational ice-wedge casts (see Photo 2) and large-scale cryoturbations in Terrace 1 points to permafrost conditions in the braided river plain

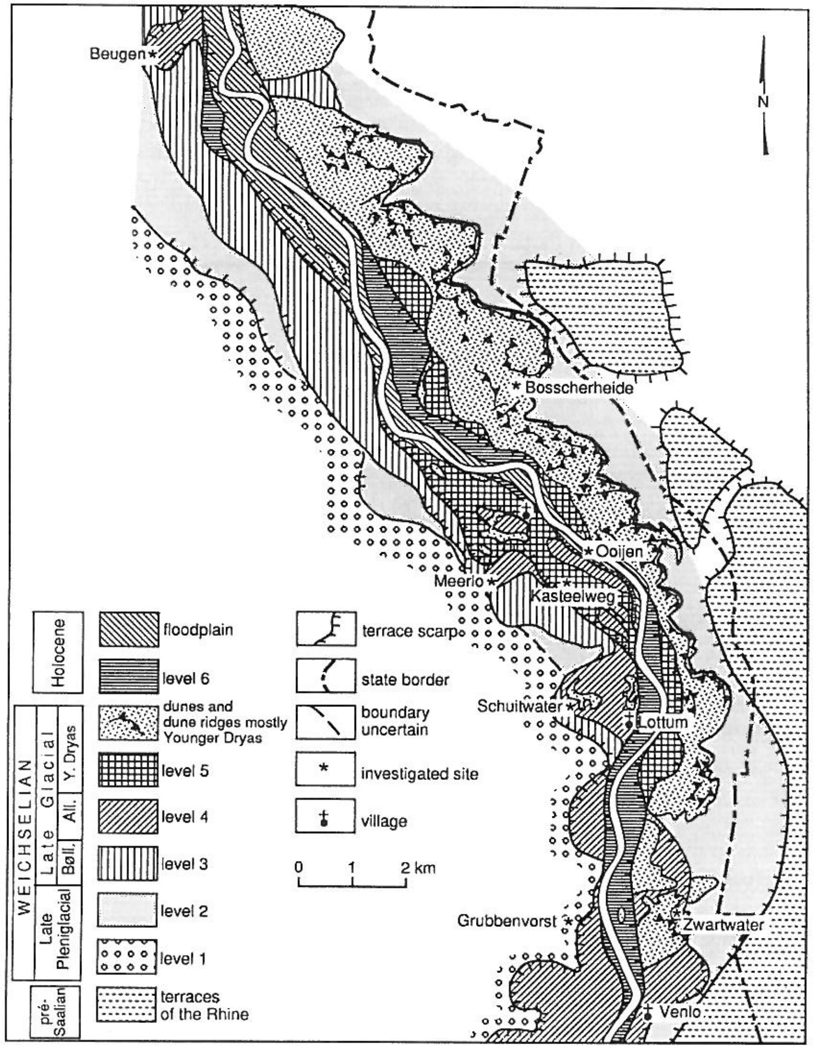

Fig. 3 Morphological map of the Lateglacial and Holocene terraces of the Maas (modified after Buitenhuis & Wolfert, 1988; Wolfert & De Lange, 1990)

and a mean annual temperature below -7°C. The existence of permafrost in the floodplain over a longer period can only be explained by assuming large, seasonal discharge variations. In winter, the floodplain was probably frozen and the discharge was low or even

absent. In spring and summer high discharges probably occurred by the melting of snow and the limited water-storage capacity of the active layer overlying the permafrost. Furthermore, vegetation was virtually absent during that period so that river bank stability was low and sediment supply towards the river was large. Aeolian activity as well contributed to the high sediment load of the river. Consequently, the Maas was a braided and generally aggrading river system. The occasional presence of boulders (not visible in Photo 2) probably also points to a large seasonal difference in discharge. The boulders, embedded in the fluvial sand matrix, are explained by the complete freezing of the channel further upstream leading to the incorporation of boulders from the channel bottom. In the thawing season ice floes with the incorporated boulders floated to the north.

During high discharges the braided river plain was flooded and medium to coarse sand was deposited by transverse bars. After the flooding, the water level over the braiding floodplain dropped and the water concentrated in small channels that slightly eroded the floodplain (Photo 2). In the next flood transverse bars migrating over the floodplain filled these channels, often laterally, with large-scale cross-bedded sand and climbing ripple cross-laminated fine sand. The large discharge fluctuations are also confirmed by the presence of silt drapings, illustrating standing water conditions, on megaripple structures (Photo 1). Aeolian activity on the seasonally abandoned braided river plain could take place as well (see Photo 1), but the preservation potential of aeolian deposits was low because of subsequent fluvial reworking.

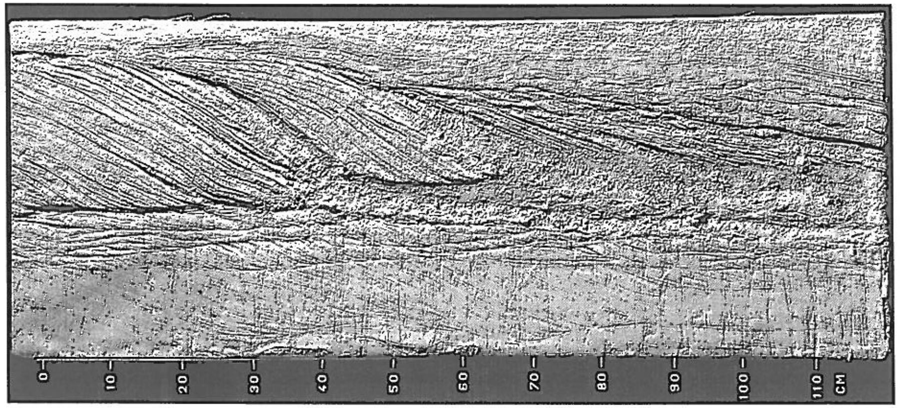

Photo 1 Photograph from lacquer peel illustrating the ephemeral character of the permafrost-dominated, late Pleniglacial Maas at Grubbenvorst. Current direction is to the right (north). The lower part of the peel is dominated by large and small-scale (climbing ripple) cross-bedding. The middle part is characterized by silt drapings, deposited in standing water on the lee side of a temporarily inactive megaripple. The (right) upper part shows a transition from megaripple deposition and standing water silt deposition into small-scale, current ripple, trough cross lamination, upper flow regime plane bedding and wet aeolian, adhesion ripple deposition

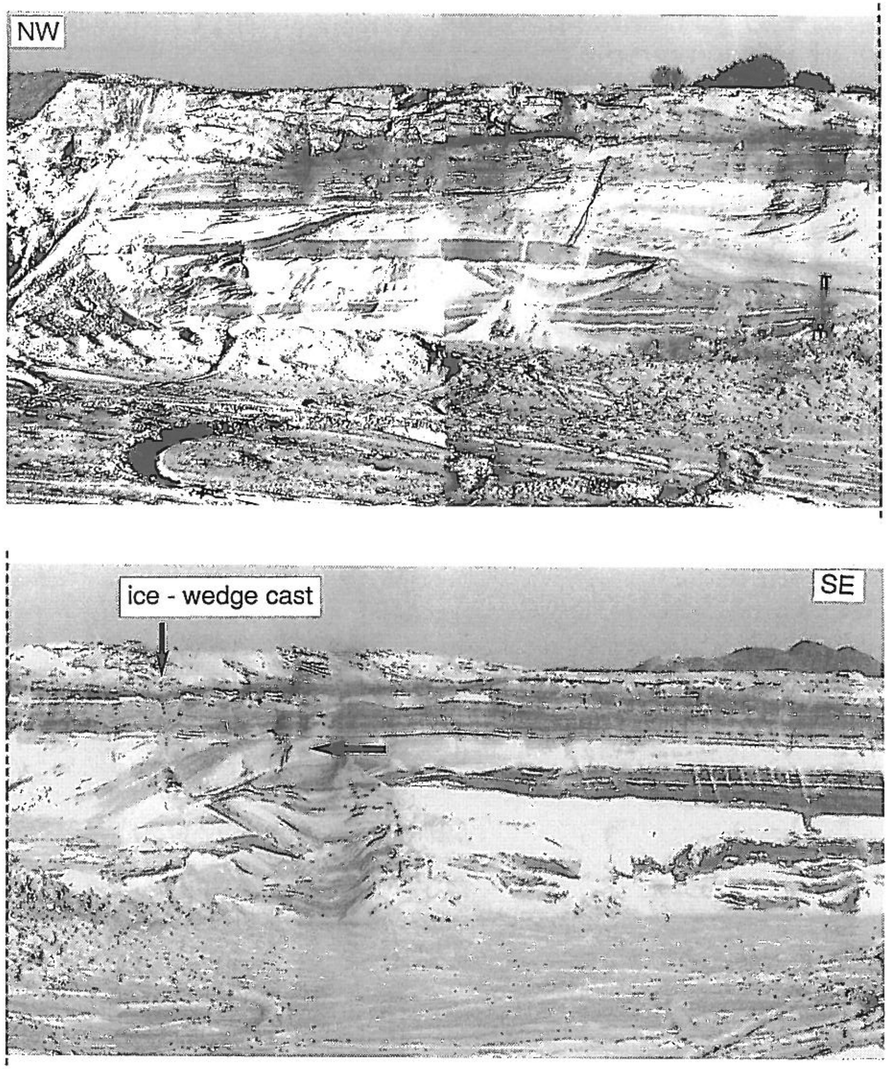

Photo 2 Composite photograph of the late Pleniglacial fluvial to aeolian transition in Grubbenvorst (Terrace 1). In the lower and middle part of the exposure with concave channel lenses, fluvial deposition is dominant. In the upper part with overall horizontal bedding, aeolian deposition is more important (fluvio-aeolian facies). The Beuningen Gravel Bed and overlying aeolian coversands have been removed. Light gray phototones are fine to medium sand; dark gray tones are silty fine sand. Note the erosive base of the channels, the large-scale, lateral channel infilling (left of arrow) and the syngenetic ice-wedge cast. Spade for scale is 1.2 m

The upward decrease of fluvial activity on Terrace 1 is obvious by the diminution of channel depths, the overall fining upward and thinning upward of the sedimentary beds. The upper sediments are probably of aeolian origin, but fluvial reworking in shallow channels and by sheet flow remained important (Photo 2: upper part). This so-called fluvio-aeolian facies is characterized by cm-scale fining-upward sets, concave lenses of coarser sand and local current ripple cross-lamination. The horizontal bedding of the fluvio-aeolian sand is often obscured by cm-scale deformations, which give this unit a fuzzy or crinkly appearance. The crinkliness of the bedding is probably the result of deformation by repeated freezing and thawing or microloading of oversaturated sediment (water escape structures). The general increase in crinkliness in the fluvial to aeolian succession reflects the decreasing accumulation rate and increasing effects of frost. The latter is also demonstrated by the upward increase of so-called vertical platy structures, which are attributed to intense surface cooling (MOL et al., 1993).

The late Pleniglacial fluvial to aeolian succession is a common feature in northwest European river valleys (SCHWAN, 1988). Therefore, it probably reflects a climatic change with a decrease in precipitation, leading to a decrease in discharge and narrowing of the floodplain. The increasing climatic aridity culminated during the formation of the "Beuningen Gravel Bed" (a desert pavement) and the overlying aeolian coversands.

During the widespread coversand deposition at the very end of the late Pleniglacial, the Maas probably shifted towards the east (Fig. 3: Terrace 2). The river maintained its braided character, as has been demonstrated in Bosscherheide (BOHNCKE et al., 1993). The short (1 m) fining-upward sequence at the top of Terrace 2 and its surface morphology point to a braided river system. The braided system was abandoned just before the Bølling period, because peat formation started in the Bølling in the palaeochannels of the braided river plain (TEUNISSEN, 1983). Due to the fluvial activity, aeolian deposits are not or only locally present on this level. The absence of aeolian coversands and the fluvial morphology of floodplain level 2 enables the distinction between the Terraces 1 and 2, since there is only little difference in elevation.

5. Bølling low-sinuosity system

Terrace level 3 occurs at almost the same elevation as the levels 1 and 2, but it differs in channel morphology by its low-sinuosity meandering branches. Also, the palaeochannels on Terrace 3 are generally deeper than on Terraces 1 and 2, which indicates a tendency towards incision. The channel scars are up to a few metres deep and often lack an organic infilling which hampers dating of this level.

The temperature increase at the start of the Lateglacial, probably together with an increase in precipitation after the very dry, desert-like late Pleniglacial, led to the re-establishment of a pioneer vegetation on the barren surface (BOHNCKE et al., 1987). The higher humidity of the soil and the spreading vegetation stabilized the surface and diminished the aeolian deposition. As a consequence river bank stability increased and the sediment load in the river decreased (VANDENBERGHE, 1987). The river morphology changed from braided into low-sinuosity meandering and the river started to incise for the first time after the Pleniglacial aggradation. This incision resulted in the formation of a new floodplain (Fig. 3: Terrace 3) and the abandonment of the former braided plain (Terrace 2) where peat started to accumulate in abandoned channels (e.g. 12,110±70 yr B.P. at Bosscherheide: BOHNCKE et al., 1993). The oldest date derived from a peat layer on Terrace 2 is 12,760±150 yr B.P. (GrN-4478) (TEUNISSEN, 1983) which means that the change from multichannel braided into low-sinuosity meandering occurred just before 12,760 yr B.P. This low-sinuosity system functioned until the *Betula* phase of the Allerød (see below). It thus extended between 12,760 and 11,800 yr B.P.

6. Allerød high-sinuosity river system

Terrace 4 is characterized by its channel scars which exhibit a clear high-sinuosity form (Fig. 3: Beugen, Schuitwater Lottum). The scars are well-preserved as half circular bends separating Terrace 4 from Terrace 3. The channel scars often reveal a well-developed asymmetrical cross section. The depth of the palaeochannels in relation to neighbouring point-bars can be up to 5 m.

The change from a low-sinuosity into a high-sinuosity system was probably gradual and was accompanied by incision of the channels. It is assumed that the low-sinuosity system represents a transitional phase between the Pleniglacial braided system and the Allerød high-sinuosity system. In that case the high-sinuosity system is the delayed response to the very rapid climatic change at the start of the Lateglacial. Only in the beginning of the Allerød period the Maas would then have been in equilibrium with the climatic conditions. However, it cannot be excluded that the change from a low-sinuosity into a high-sinuosity river was caused by climatic or climate-related changes. For instance, the gradual immigration of the vegetation and the increasing soil-profile development in the Bølling-Allerød interstadial complex will have resulted in increasing rainfall interception and an increasing storage capacity of the soil. These factors may be responsible for a decrease of the peak discharges and an increase of the channel sinuosity.

The meander morphology and incision are clearly demonstrated in the cut-off channel at Beugen (Fig. 4). The cross section reveals the asymmetrical shape of a bend in the meandering channel with a steep undercut bank and a gradually rising point-bar slope. Corings in the point-bar east of the channel scar showed that the gravelly channel sediments occur at a higher level in the point-bar than in the channel and on the point-bar slope at the moment of cut-off. If only lateral migration had occurred, the gravelly channel sediments

would be found in a more or less horizontal unit. In the Beugen cross section, however, it means that besides lateral migration of the channel vertical erosion of the channel bed also took place.

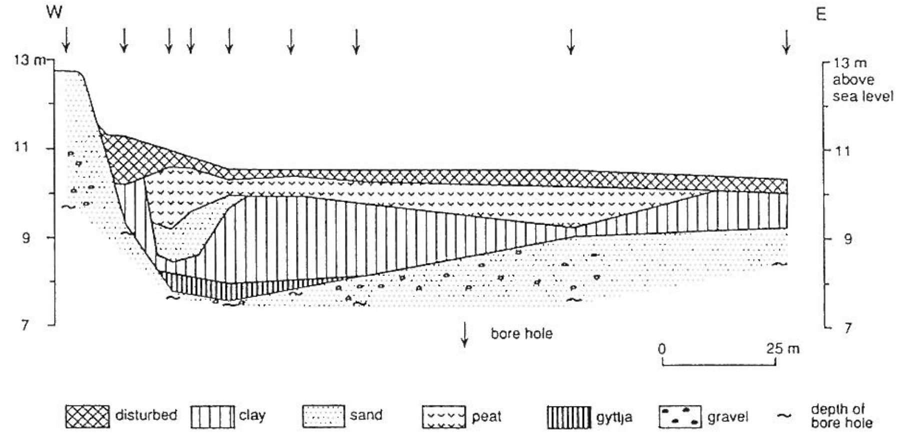

Fig. 4 Cross section over the Allerød oxbow lake at Beugen (Terrace 4). Note the well-developed asymmetry of the palaeomeander. Gyttja at the base dates from the *Betula* and *Pinus* phases of the Allerød; overlying silty clay dates from the Younger Dryas

The meander scar at Beugen is a fine example of a neck-cut off. Such neck cut-offs are caused by the meandering process itself, by lateral and downstream migration of the meander bends. They are not related to climatic changes, in contrast to chute cut-offs which can be formed, besides by normal fluvial processes, by climatic changes also, for instance when swales on the point-bar surface are enlarged during a phase of climatic change with higher discharges. At the moment of neck cut-off (start of the infilling) the Maas continued meandering, only at another place. Therefore, the start of the infilling of the neck cut-off gives an accurate date for the period during which the Maas was a high-sinuosity, meandering river.

The infilling consists of laminated fine-detrital gyttja at the base (Fig. 4). It abruptly overlies coarse sand and gravel of the active channel, which is typical for a neck cut-off sequence (ALLEN, 1965). Neck cut-offs involve the sudden abandonment of the entire meander loop and the entrance to and the exit from the loop are rapidly plugged with sand. Flow diminishes to zero in the distal part of the cut-off and the sequence is dominated by fine-grained vertical accretion deposits (WALKER, 1981).

The oxbow lake of Beugen has been investigated previously by TEUNISSEN (1990: pollen diagram Helbroek). He concluded a Younger Dryas age for the base of the meander fill, probably because of the ^{14}C-date (GrN-14405: 9920±190 yr B.P.) in the lower part of the infilling. However, it is unknown whether the deepest part of the abandoned channel has been sampled and furthermore the date may be too young because of root penetration from an overlying peat layer. More recent palynological analyses of the gyttja at the base of the infilling (see Fig. 4) points to an Allerød age (*Betula* phase) for the start of the infilling (KASSE et al., 1992). It means that the Maas was a large-scale meandering river during the early Allerød. Infillings older than the Allerød have not been found on Terrace 4. In fact, most of the palaeochannels' infillings date from the beginning of the Younger Dryas (KASSE & BOHNCKE, 1991: pollen diagram Schuitwater, base reinterpreted as Younger Dryas; unpublished diagrams Meerlo and Zwartwater), which indicates that the high-sinuosity system was active before the Younger Dryas, this is during the Allerød. In conclusion, the low-sinuosity channels of Terrace 3 had disappeared before the beginning of the Allerød, while the formation of Terrace 4 started during the early Allerød or shortly before and ended at the Allerød-Younger Dryas transition.

In contrast to the Pleniglacial braided system and the Bølling low-sinuosity system, the sedimentary sequence of the high-sinuosity river is characterized by long fining-upward sequences. They have been studied by borings in the point-bars of the large meander west of Lottum (Figs. 3 and 5). The lower part (Fig. 5: 7.65-9.95 m below the surface) of the up to 7.5 m thick fining-upward sequence consists of gravelly, moderately sorted, medium to coarse sand with predominantly horizontal bedding. This coarser-grained, lower part was formed by strong tractional currents on the meander channel bed. The lag deposit is characterized by a high gravel content and the largest gravel size.

The upper part of the fining-up sequence (2.90-7.65 m below the surface) consists of moderately or, more often, well-sorted medium to fine sands. Thin (1-3 cm) sandy silt laminae are present between 2.90 and 5.25 m below the surface. This fine-grained, well-sorted sand was deposited by weaker tractional currents on the point-bar slope. The large-scale cross-bedding and small-scale cross-lamination point to deposition by megaripples and small current ripples. Some smaller fining-up sequences, separated by erosional boundaries, are present within this fine sand unit, probably reflecting reactivation surfaces on the meander point-bar slope during high discharges. The thin silt laminae reflect weak currents and deposition of fines from suspension on the upper point-bar slope during high-water levels. The top of the fining-upward sequence (Fig. 5: 2.40-2.90 m) consists of sandy silt with some sandy intercalations, which is interpreted as a vertical accretion deposit on the point-bar.

This well-developed fining-upward sequence was formed by lateral migration of the channel and accretion on the channel point-bar. It reveals, in comparison with the previous late Pleniglacial and early Lateglacial and posterior Younger Dryas river systems, a more constant river discharge. A maximum water depth of 7.5 m during the peak discharges is concluded from the length of the fining-upward sequence.

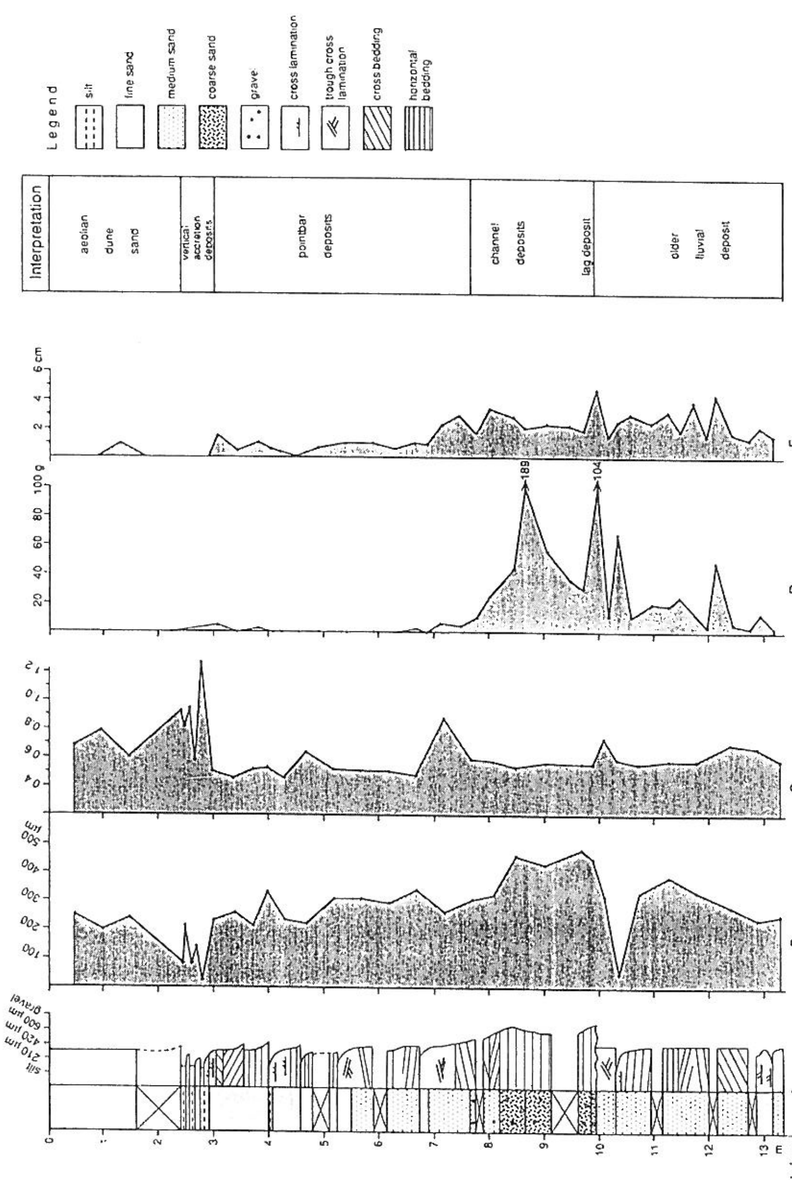

Fig. 5 Sedimentological sequence of the Allerød terrace west of Lottum (Terrace 4). It shows a 7.5 m long fining-upward sequence due to lateral migration of the meandering channel. A=Sedimentary log with lithology and sedimentary structures, B=Mean grain size in μm, C=Standard deviation of the grain size (sorting), D=Gravel (> 3 mm) content recalculated per 10 cm, E=Largest diameter of the gravel. For location, see Figs. 3 and 7 (borehole 2)

On top of the point-bar sediments low dunes, up to 4 m high, are locally present (Fig. 5: 0-2.40 m below the surface). Since there is no evidence for an aeolian supply from outside the area, nor for deflation from the local point-bar subsoil, it is concluded that these dunes were formed during the lateral migration of the meandering river (KASSE & BOHNCKE, 1991). The similar grain-size of the aeolian deposits and the upper point-bar deposits support this hypothesis. It is assumed that during bankfull discharge sediment was deposited high on the point-bar slope. During low discharge these barren sediments were prone to deflation by the prevailing westerly winds (MAARLEVELD, 1960; SCHWAN, 1988). The low dune morphology indicates aeolian deposition on top of the point bar at the edge of a vegetated landscape. Similar conditions exist nowadays along the river Waal (Millingerwaard), where dunes are formed by deflation of the exposed river bank (ISARIN & BERENDSEN, 1992).

The start of the infilling of the meander scar, belonging to the point-bar sequence described in Fig. 5, was dated palynologically to the Younger Dryas (KASSE & BOHNCKE, 1991: diagram Schuitwater, reinterpreted by new ^{14}C dates). This means that the lateral channel migration forming the long fining-upward sequence and the connected deflation and dune accumulation took place just before the Younger Dryas, i.e. the Allerød period.

7. Younger Dryas climatic and fluvial changes

Terrace 5 is separated from Terrace 4 by its straight terrace edges (Fig. 3). The surface fluvial morphology of this terrace is characterized by rather straight to slightly curved palaeochannels which give the impression of a braided river system.

Fine examples of the Younger Dryas braided floodplain can be found to the northwest and southeast of the village of Blitterswijk (Fig. 6). Islands or bars in this floodplain are locally covered by river dune sand, which was blown out of the multi-channel plain during periodic low-water levels. The lithology of Terrace 5 reveals a larger lateral and vertical variability than terrace level 4. The bars in between the channels consist of (gravelly) sand and fining-upward sequences are less pronounced and normally shorter than in Terrace 4. The channels are shallow (up to 2 m deep) and often have a clastic infilling. They are underlain by coarse sand and gravel of the active braiding phase. A two-metre organic fill was found locally in scour pools in the abandoned braided channels. Palynological investigations reveal a late Younger Dryas age of the base of the infilling, indicating that the river system that formed terrace level 5 was active later than the Allerød, but before the end of the Younger Dryas (KASSE et al., 1992: pollen diagram Kasteelweg). This age has previously been determined as well by WESTERHOFF & BROERTJES (1990: pollen diagram Linkstraat).

At the start of the Younger Dryas a strong decline in summer temperature occurred (BOHNCKE et al., 1987), which was also demonstrated at Bosscherheide (BOHNCKE et al., 1993). According to the latter authors, the presence of initial ice-wedge casts and other periglacial features point to, at least local, permafrost conditions and a mean annual temperature between -2 to -5°C. The degradation of the Younger Dryas permafrost at Bosscherheide has been dated very accurately. In this exposure peat and humic soil material, which formed between 12,110±70 yr B.P. and 10,880±50 yr B.P. (BOHNCKE et al., 1993) during the Bølling, Older Dryas and Allerød periods, are disturbed by initial ice-wedge casts and periglacial involutions (Photo 3). The involutions are discordantly overlain by undisturbed fluvial silts, moss-peat laminae and aeolian dune sand. The moss-peat laminae were dated at 10,500±60 BP, which indicates that permafrost degradation and involution took place between 10,880 and 10,500 yr B.P. The establishment of the permafrost may have occurred also between 10,880 yr B.P. and 10,500 yr B.P., but it cannot be excluded that local permafrost already penetrated the soil before 10,880 yr B.P., this is during the later part of the Allerød. BOHNCKE et al. (1993) present palaeobotanical evidence from which it is concluded that more intense frost action in the soil occurred after 11,300 yr B.P.

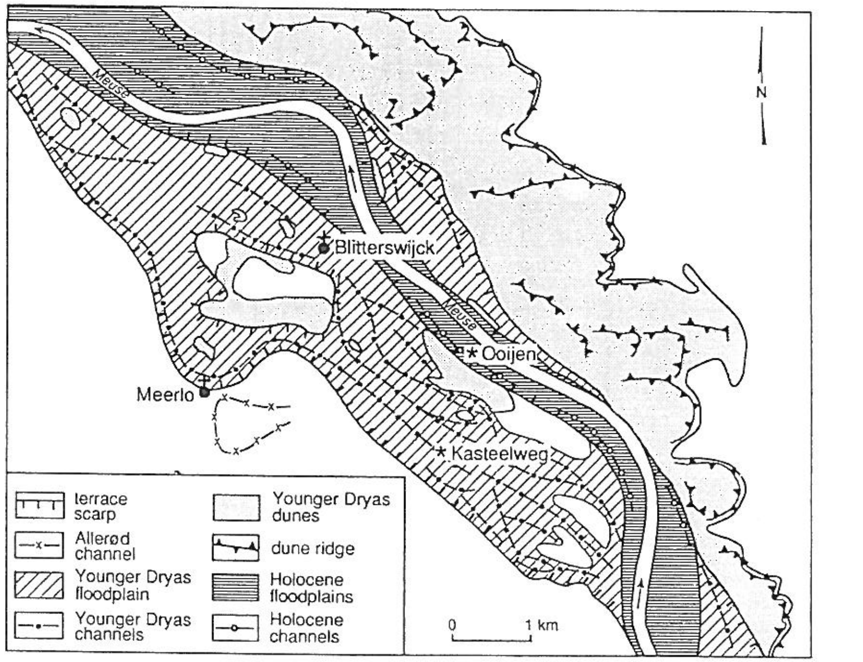

Fig. 6 Morphological map of the Younger Dryas braided river plain (Terrace 5) at Blitterswijk. Note the straight palaeochannels, the remnants (=islands) of older terraces in the braided river plain and the Younger Dryas dune belt on the east bank of the Maas

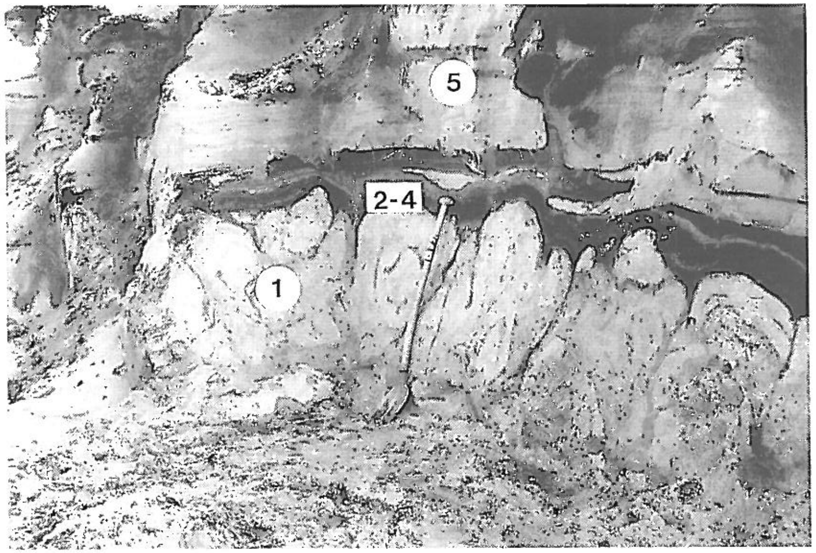

Photo 3a For legend see page 139

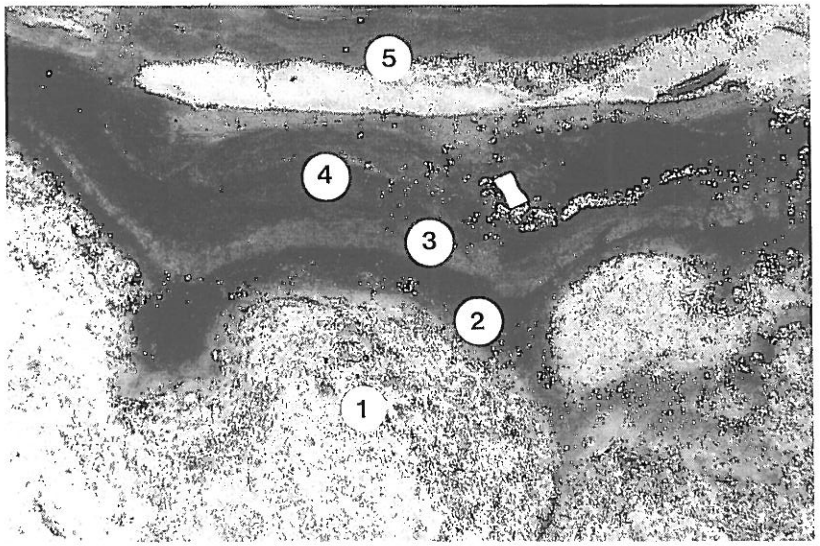

Photo 3b For legend see page 139

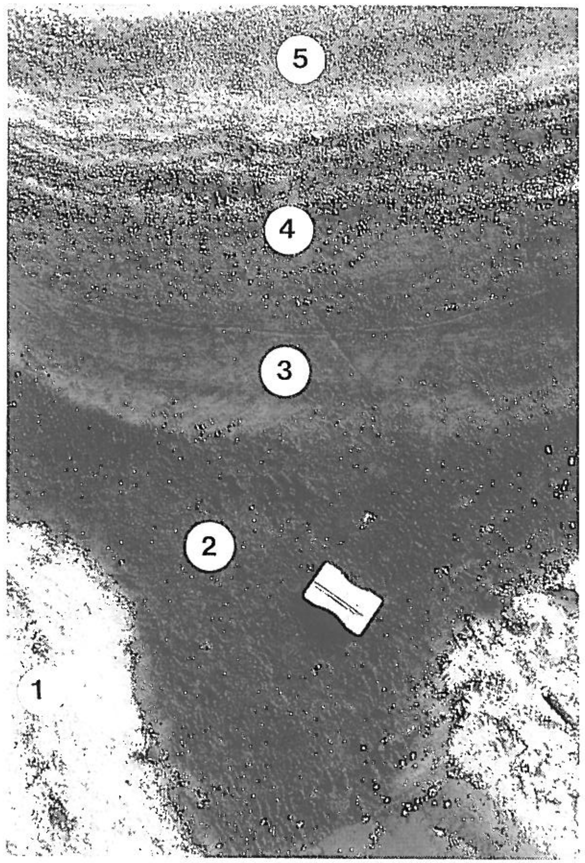

Photo 3a-c Large-scale Younger Dryas cryoturbations at Bosscherheide. The involuted late Pleniglacial fluvial sediments (1) and the Lateglacial Allerød soil (2: top is 10,880 yr B.P.) are overlain by non-involuted Younger Dryas fluvial silt (3), peat (4: 10,500 yr B.P.) and aeolian sand (5). The dates have been derived from BOHNCKE et al. (1993). The black layer in the lower part of unit 5 (Photo 3a-b) is due to humus illuviation from the overlying Holocene podzol. Spade is 1.2 m; pencil sharpener is 2.5 cm

The strong summer temperature decline at the start of the Younger Dryas resulted in a drastic change of the river pattern. The lower temperature and the deterioration of the vegetation cover led to a lower evapotranspiration. In combination with a possible increase in precipitation larger amounts of meltwater during spring will have been the result. Furthermore, the deeper frost penetration and local development of permafrost reduced the water-storage capacity of the soil. These factors resulted in higher and more frequent peak discharges. Wetter conditions were also registered by high lake levels in pingo scars in the northern Netherlands (BOHNCKE & WIJMSTRA, 1988). The effects of the Younger Dryas climatic cooling are registered in the following phenomena:

(1) *Flooding of older terraces.* A fluvial loam bed was deposited over the Allerød humic soil and the Younger Dryas involutions at Bosscherheide (see Photo 3). The presence of *Classopollis*, derived from Lower Cretaceous rocks in the upper reaches of the Maas, in this thin loamy bed clearly demonstrates the regional character of the flooding. This flooding phase can be dated between 10,880 and 10,500 yr B.P. (BOHNCKE et al., 1993). A time-equivalent start of a flooding phase in the Mark valley was found by BOHNCKE et al. (1987).

(2) *Change in the infilling of the meander neck cut-off at Beugen.* The gyttja of Allerød age at the base of the oxbow lake at Beugen is erosively overlain by a gray, fine sandy, calcareous clay (Fig. 4). This clay, which also contains *Classopollis*, represents a phase of renewed activity of the Maas in the meander scar. It means that fluvial inundations resulting in clay deposition could reach the channel after a period of organic accumulation. However, current velocity in the oxbow lake was low and only sandy clay was deposited from suspension. The Allerød channel morphology was not modified. The erosional contact between the gyttja and the clay correlates pollen-analytically with the start of the Younger Dryas (KASSE et al., 1992). This implies that at the Allerød-Younger Dryas transition, peak discharges in the Maas valley became so high that they could overflow the sand plugs at the entrance and exit from the abandoned meander.

(3) *Chute cut-off of existing meanders.* The high-sinuosity meandering channel pattern that existed during the Allerød (Terrace 4) was abandoned by the process of chute cut-off. During a chute cut-off an old swale on the meandering river plain is reactivated and eroded. The flow in the main channel diminishes and eventually the meander loop is abandoned (WALKER, 1981). The higher peak discharges at the beginning of the Younger Dryas are probably responsible for these chute cut-offs. Sand plugs were deposited in the entrance and exit of the meander loops and oxbow lakes developed, as can be seen at the Allerød meander of Lottum (Fig. 7). In the oxbow lakes fine-clastic deposition prevailed during the Younger Dryas, since fine sediment entered the lakes by overbank flooding (KASSE & BOHNCKE, 1991: pollen diagram Schuitwater, base reinterpreted as Younger Dryas; unpublished diagrams Meerlo and Zwartwater).

(4) *Change in the channel pattern.* The increase in discharge and discharge fluctuations resulted in the adaptation of the fluvial system to the new situation. The fluvial pattern changed from single-channel meandering into multi-channel braided. Since Terrace 5 is generally found 2-3 m below Terrace 4, considerable (lateral) erosion in the valley occurred by these higher peak discharges. This rapid erosion is explained by the restricted sediment supply because of the presence of a vegetation cover. Although the canopy had become somewhat opener in comparison with the Allerød period, most of the land surface outside the river floodplain remained vegetated with herbs and shrubs (BOHNCKE et al., 1988). The formation of parabolic dunes in the second part of the Younger Dryas (see below) points to the presence of a vegetation cover as well.

The change from meandering into braided during the Younger Dryas has been reported previously by PONS (1957), who defined a braided X-terrace of the Maas of Younger Dryas age north of our study area. A similar change to braiding has been described in

other north-central European rivers (Vistula river in Poland: KALICKI, 1991: 29 and references cited there). However, this braided phase is not omni-present, since other rivers do not exhibit a Younger Dryas braided phase, but instead continue their meandering pattern. For example, the relatively small Mark river in the Netherlands (VANDENBERGHE et al., 1984, 1987) and the Warta river in Poland (KOZARSKI, 1983), which is comparable in size with the Maas, do not have a braided phase. This implies that, besides climate, other threshold values determined the resulting channel pattern. Sediment load of the river and river gradient are important factors in this respect (VANDENBERGHE, 1993, this volume).

Fig. 7 The chute cut-off of the Allerød meander at Lottum formed by higher peak discharges because of the strong climatic cooling during the start of the Younger Dryas. Sand plugs block the entrance and exit of the palaeomeander

(5) *Widespread aeolian deposition*. On the east bank of the Maas valley an extensive, up to 4 km wide, sand-sheet belt occurs with parabolic dunes on top, lying on Lateglacial or older terraces (Figs. 3 and 6). The dune morphology is characterized by parabolic forms, especially at the eastern (downwind) border of the sand sheet, where the dunes are up to 10 m high. The parabolic morphology indicates a west-southwesterly wind. This direction is in close agreement with previous results (MAARLEVELD, 1960).

The internal sedimentary structures in the sand sheet are dominated by horizontal bedding and low-angle cross-bedding. Large-scale foresets, which are formed at the lee side of dunes are very rare, perhaps due to a lack of exposures in the parabolic ridges. However, locally some large-scale foreset beds were found. Surprisingly, the wind direction appeared to be north instead of west-southwest (Photo 4). It indicates that, although the dominant sand-transporting wind was from the west-southwest during the Younger Dryas, short-term periods occurred with a northern wind, strong enough to transport fine to medium sand.

Photo 4 Large-scale cross-bedded slip face in the Younger Dryas dune complex at Bosscherheide. The foresets were formed by a subordinate northerly wind, which contrasts with the dominant sand-transporting wind from the west-southwest. The white and black layers at the top are the eluviation and humus illuviation horizons of the Holocene podzol. Spade is 1.2 m

The peat at the base of the aeolian deposit allows to date the start of the aeolian deposition just after 10,500±60 yr B.P. (BOHNCKE et al., 1993: exposure Bosscherheide). The top of the peat layer is characterized by an alternation of moss laminae and aeolian sand laminae,

which indicates a drowning of the vegetation in the aeolian sand (Photo 3c). This gradual transition implies the possibility of a certain diachronism for the start of the aeolian accumulation, for instance between the western, upwind part and the eastern, downwind part of the sand sheet.

The aeolian sedimentation phase along the Maas was probably restricted to the second half of the Younger Dryas. In a geomorphologically comparable situation on the eastern bank of the Schelde valley, 125 km west of the Maas valley, SCHWAN (1991) gives a date of 9050±45 yr B.P. for the base of a peat bed overlying a Younger Dryas dune complex. VAN GEEL et al. (1980/1981) present a date of 10,150±90 yr B.P. for the base of a peat layer overlying Younger Dryas aeolian sand in the eastern Netherlands. The aeolian sediment of this phase is fine to medium grained. Due to the prevailing west-southwestern winds, the source area must have been the Younger Dryas palaeofloodplain west of the dune field. It is likely that the aeolian sediment has been blown out from the Younger Dryas floodplain on the older adjacent terraces during periods of low discharges. The aeolian sediments within the Younger Dryas floodplain itself had a low preservation potential, since they were easily eroded during subsequent periods of high discharges. Only at a few locations small Younger Dryas dunes were found on bars between the braided channels (see Fig. 6). The accumulation of large amounts of sand in the extensive dune field on the eastern bank of the Maas is closely connected with the presence of a braided system. Due to the large width (ca. 1-1.5 km) and the generally north-south orientation of the braided river plain (see Fig. 3), the westerly wind was able to deflate large quantities of sand from the river plain. Since gravel pavements have never been found in the braided valley, we conclude that repeated fluvial reworking of the braided floodplain constantly provided fresh material for deflation. These alternating processes of fluvial sedimentation and aeolian deflation indicate that during the late Younger Dryas the Maas had a strong intermittent character with large discharge fluctuations.

It is argued below that the large-scale deflation from the floodplain and dune formation must have stopped during the Younger Dryas - Preboreal transition because of a change in river pattern from braided into low-sinuosity meandering and revegetation of the former braided floodplain. Locally, dunes are present on Terrace 5 which lie on top of the scarp between the Terraces 5 and 6 and which are arranged perfectly parallel with this scarp (see Fig. 6: east bank Maas, east of Blitterswijck). This position suggests that deflation on a more local scale still occurred during the early Preboreal incision when the terrace scarp was formed between the Preboreal floodplain and the inactive Younger Dryas level 5.

8. Holocene warming and fluvial changes

Terrace 6 is especially well developed in the southern part of the study area (Fig. 3). The fluvial morphology is characterized by large-scale point-bars or channel side-attached bars at the inner bends of the Maas (Fig. 7, east of Lottum). These bars were formed by the lateral migration of a low-sinuosity river. The present-day Maas is found in an incised, narrow channel in Terrace 6. Towards the north Terrace 6, as well as Terrace 5, disappear under the recent sediments of the Holocene floodplain. The clear fluvial morphology of Terrace 6 indicates that aeolian activity, typical of the Younger Dryas Terrace 5, had ceased, probably because the floodplain became revegetated and the river had changed from a multi-channel braided into a single-channel, low-sinuosity system.

Terrace 6 is separated from Terrace 5 by an erosional scarp. This incision has been dated by pollen analysis of the abandoned channels on Terraces 5 and 6. In the abandoned channels on Terrace 5 organic deposition started at the end of the Younger Dryas (KASSE et al., 1992: pollen diagram Kasteelweg; WESTERHOFF & BROERTJES, 1990: pollen diagram Linkstraat). The oldest fill in a Holocene channel on Terrace 6 dates from the early Boreal (KASSE et al., 1992: pollen diagram Ooijen), which indicates that the incision and the formation of floodplain level 6 occurred between the Younger Dryas and the Boreal, i.e. the Preboreal. The well-developed point-bars on Terrace 6 date from the Preboreal as well, since the interpoint-bar fine-grained sediments show a Boreal pollen spectrum (diagram Aastbroek, east of Lottum, unpublished). Furthermore, Terrace 6 is characterized by soils with well-developed clay-illuviation horizons (FAO, 1988: Argillic-B horizon of Luvisols). This type of soil is also characteristic for the older Lateglacial terraces and it is absent in the younger Holocene floodplain (MIEDEMA et al., 1983; JONGMANS & MIEDEMA, 1986; MIEDEMA, 1988). The latter authors suppose a late Weichselian to early Preboreal age for this type of soil formation, which is in close agreement with the pollen-analytical results of Terrace 6.

The fluvial changes in the beginning of the Holocene can be explained by climatic factors. The rapid temperature rise at the start of the Holocene probably led to lower peak discharges, which caused the transition from a braided to a low-sinuosity, meandering river. According to VANDENBERGHE (1992) especially the steep increase in the winter temperature from -20°C in the Younger Dryas to 2°C at the end of the Preboreal is responsible for less meltwater and therefore, for lower peak discharges. However, according to VAN GEEL et al. (1980/1981) the Preboreal is still characterized by its continentality. BECKER & KROMER (1986) stated that during the Preboreal the winter temperatures were low, as is witnessed by the frost damage of pine trees in Preboreal deposits in Germany. This indicates a continuation of continental climatic conditions in which snow meltwater is relatively important. Another factor, which may have been relevant for the fluvial dynamics is the water-storage capacity of the soil. The deep seasonal frost or permafrost during the Younger Dryas will have limited the water-storage capacity and meltwater was released directly towards the river, causing extreme peak discharges. The rise in winter temperature at the start of the Preboreal resulted in the disappearance of the local permafrost and a decreasing depth of the seasonal frost. This will have restored the infiltration capacity of the soil leading to lower peak discharges and a higher base flow in comparison with the Younger Dryas. Probably for this reason and because of a reduced sediment supply by the

restoration of the canopy cover after the Younger Dryas opening of the vegetation, the river incised and changed from braided into meandering. The braided plain of Terrace 5 was abandoned and one of the braided channels developed into a low-sinuosity meander.

Although the peak discharges had become lower in the Preboreal in comparison with the Younger Dryas, river activity was still important. This is illustrated by the presence of the well-developed side bars or point-bars on Terrace 6. The rapid lateral migration and point-bar formation during the Preboreal can be explained by the relatively important peak discharges due to snow meltwater in the continental climate. The abandonment of floodplain level 6, by incision of the Maas, probably occurred at the end of the Preboreal or start of the Boreal. VAN GEEL et al. (1980/1981: pollen diagram "De Borchert") stated that higher winter temperatures and a moister climate established at the end of the Preboreal. The first occurrences of *Hedera helix* and *Viscum* around 8600 yr B.P. (VAN GEEL et al., 1980/1981) indicate mild winter temperatures (higher than -1.5°C) and relatively high summer temperatures (higher than 17°C) respectively (IVERSEN, 1944). The presence of *Viscum* in the second part of the Boreal shows that the mean July temperature in the eastern Netherlands was slightly higher than at present (17°C), since *Viscum* is nowadays only found in the southernmost part of the Netherlands (17.5°C)(VAN GEEL et al., 1980/1981: 412). It is possible that the more oceanic climatic conditions in the Boreal were caused by the establishment of the warm Gulf Stream circulation pattern in the North Atlantic Ocean (RUDDIMAN & MCINTYRE, 1981) and in the North Sea by the flooding of the Strait of Dover (JELGERSMA, 1979). The higher winter temperature decreased the amount of snow meltwater, leading to a further decrease of the peak discharges. Lateral migration, due to high discharges, decreased. The more regular discharges and perhaps a gradual uplift of the area resulted in the incision of the Maas in Terrace 6. In the northern part of the study area this incision is masked by younger Holocene alluvial aggradation, related to the sea-level rise.

9. Conclusions

The late Pleniglacial, Lateglacial and early Holocene development of the Maas is summarized in Fig. 8. The major conclusions are:

(1) Climatic changes during the Lateglacial and early Holocene were reflected by changes in channel pattern and channel morphology.
(2) Vegetation, sediment supply, water-storage capacity of the soil and the magnitude and frequency of peak discharges were the principal factors determining the river response.
(3) Phases of incision or intensified erosion occurred during both warming and cooling of the climate.
(4) During the late Pleniglacial, until 12,760 yr B.P., the Maas was aggrading in a braided plain (Terraces 1 and 2). Aggradation was caused by a low water-storage capacity (permafrost), a high sand supply (no vegetation), a decrease in precipitation and large discharge fluctuations.

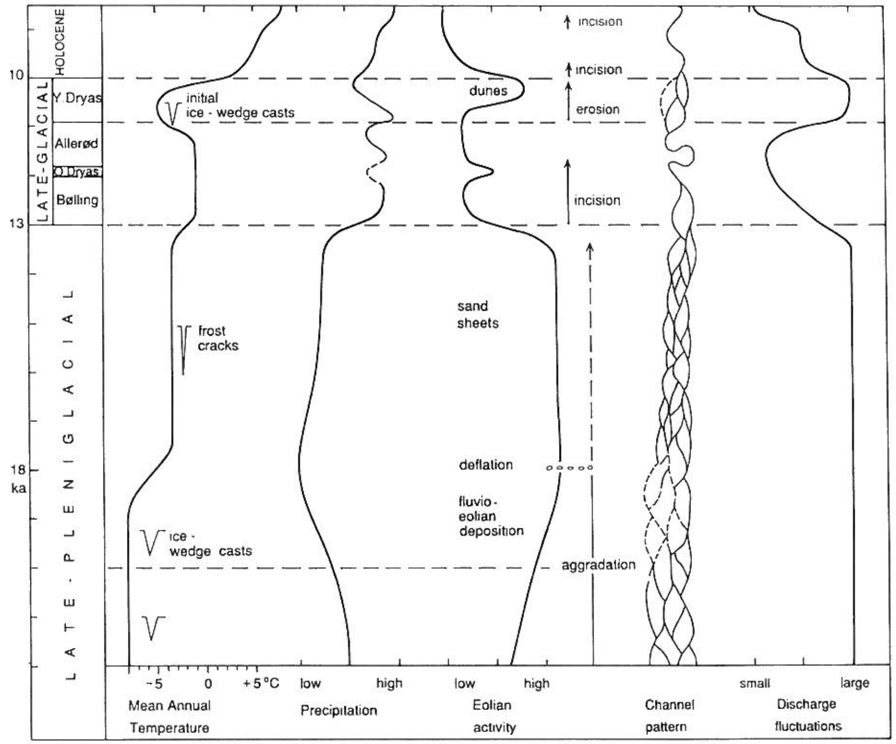

Fig. 8 Synthesis of the late Weichselian and early Holocene fluvial development of the Maas. Mean annual temperature is based on periglacial structures and vegetation. Precipitation and aeolian activity are qualitative estimates based on the sedimentary record and partly on lake level changes (BOHNCKE & WIJMSTRA, 1988). Discharge fluctuations are inferred from climate, channel pattern and the sedimentary record

(5) The Lateglacial fluvial development is characterized by rapid stepwise erosion.

(6) The start of the Lateglacial shows a chain reaction: because of the higher temperature and precipitation, the land surface became protected by vegetation. This more stable land surface reduced the sediment supply and incision started by a low-sinuosity meandering system (Terrace 3), which was active from ca. 12,760 yr B.P. till the *Betula* phase of the Allerød.

(7) The low-sinuosity meandering pattern of the Bølling changed into a high-sinuosity pattern (Terrace 4) during or just before the *Betula* phase of the early Allerød. This change is interpreted as a delayed response of the river to the rapid climatic development at the start of the Lateglacial. The high-sinuosity system was active until the start of the Younger Dryas.

(8) The Older Dryas event is not reflected by a change in river pattern.

(9) The Younger Dryas climatic cooling around 10,880 yr B.P. resulted in a lower evaporation and a decline of the forest vegetation. Together with a decrease in the water-storage capacity of the soil by deeper frost penetration and permafrost development this resulted in higher peak discharges. The river pattern changed from meandering into braided (Terrace 5). The floodplain was lowered by (lateral) erosion, because of the higher discharges in connection with a restricted sediment supply. During the second half of the Younger Dryas (after 10,500 yr B.P.) deflation of the braided floodplain became important and large river dune complexes were formed on the east bank of the Maas. The braiding phase ended at the start of the Preboreal.

(10) The Preboreal temperature rise led to the restoration of the water-storage capacity of the soil and a decrease in the amount of meltwater. Peak discharges, therefore, decreased, but remained relatively important because of the continental climate. The relatively more regular discharge and lower sediment supply resulted in incision and a change from a braided into a low-sinuosity meandering river (Terrace 6). The large-scale deflation from the floodplain stopped. The low-sinuosity system was abandoned already before the early Boreal.

(11) The change from a continental to a more oceanic climate at the end of the Preboreal or early Boreal led to a further decrease in peak discharges. The rapid lateral migration of the Maas during the Preboreal stopped. Incision became dominant and the low-sinuosity floodplain was fossilized.

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