Climate-controlled fluvial evolution of the Maas and Niers during the Weichselian and Holocene

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Excursion Guide

Vrije Universiteit *Amsterdam* | Universiteit Utrecht | NITG-*National Geological Survey*

Climate-controlled fluvial evolution of the Maas and Niers during the Weichselian and Holocene

DEUQUA-excursion Wednesday 1 September 2004

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Excursion program Maas – Niers valley

The excursion will demonstrate the Late Quaternary fluvial evolution of the Niers and Maas valleys in the Dutch-German border region between Venlo and Nijmegen. Changes in channel pattern, phases of incision and terrace formation, related to climate and vegetation change, will be shown in the field.

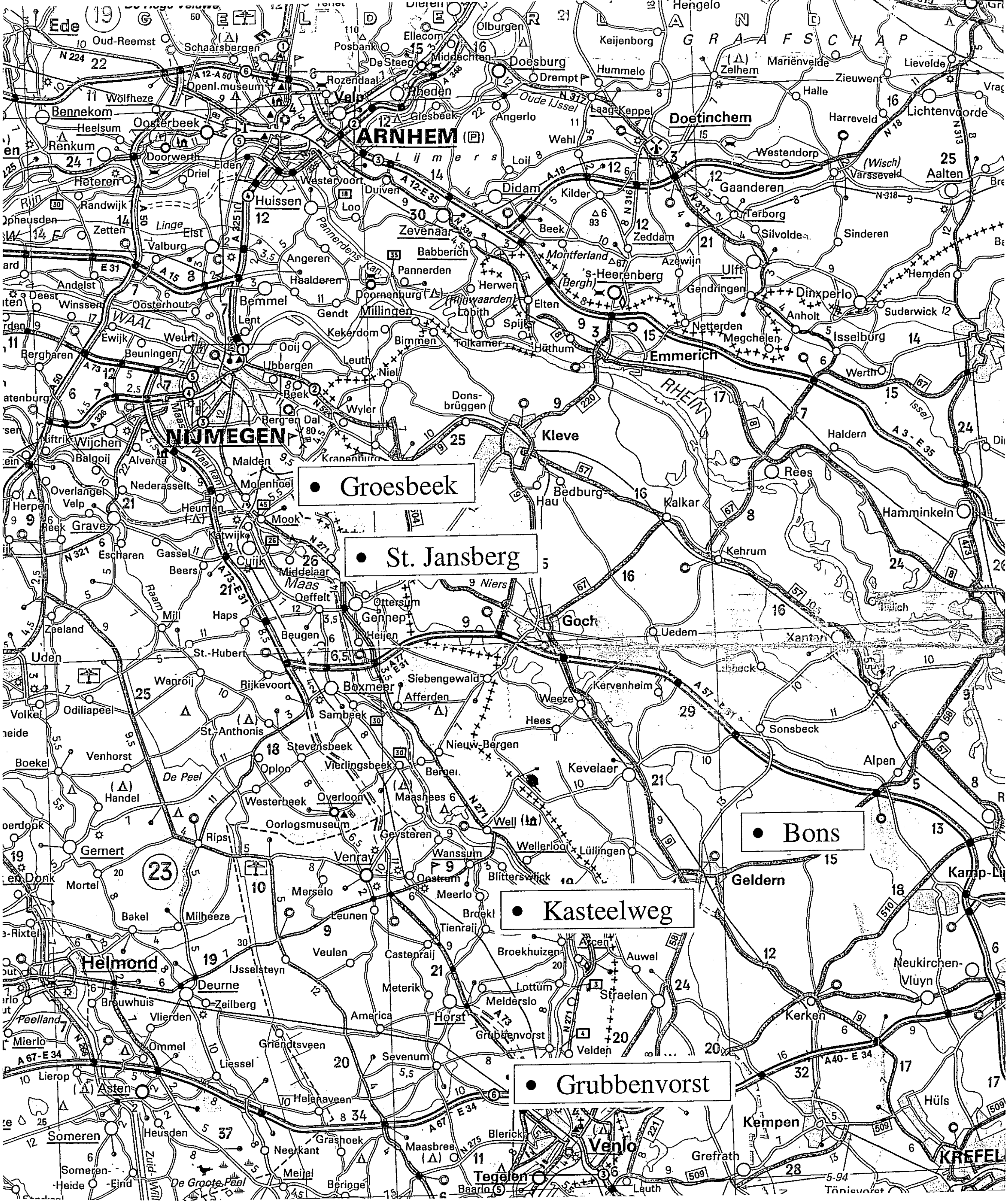
Stop 1: GROESBEEK
- Morphology of Saalian ice-pushed ridges and Weichselian dry valleys
- Local deposition of loess in sheltered areas

Stop 2: St. JANSBERG
- Erosional scarp of the Saalian ice-pushed ridges by the Saalian Niers-Rhine
- Fluvial morphology of the Weichselian Pleniglacial system

Stop 3: BONS MEANDERS
- Morphology of Weichselian Late Glacial large meanders of the Niers-Rhine
- Coring of the fine-grained infill and palynology of the abandoned meanders
- Final abandonment of the Niers-Rhine system

Stop 4: GRUBBENVORST SAND PIT
- Late Pleniglacial fluvial to aeolian facies
- Late Pleniglacial cryogenic stuctures (ice-wedge casts, deformations, vertical platy structures)
- Late Pleniglacial and Late Glacial aeolian sediments

Stop 5: KASTEELWEG - OOIJEN
- Younger Dryas braided floodplain morphology
- Late Younger Dryas and Preboreal fill of channels
- Younger Dryas dune complex
- Early Holocene incision and terrace edge



Location map with the excursion stops in the Maas and Niers valleys

EXCURSION STOP 1: GROESBEEK

- Morphology of Saalian ice-pushed ridges and Weichselian dry valleys
- Local deposition of loess in sheltered areas

The excursion stop is located 4 km southeast of Groesbeek, c. 10 km southeast of Nijmegen (fig. 1). The elevation is c. 40 m a.s.l. The large-scale morphology is related to ice-pushed ridges and sandurs that were formed during the Saalian glaciation (Rheburger Phase). They form the southernmost glacial morphology in the Netherlands and they extend further to the southeast towards Dusseldorf (fig. 2). The ice-pushed ridges surround a small glacial basin (Kranenburger Basin) that opens to the northnortheast, indicating the direction of ice flow to the south (fig. 3).

Like most of the ice-pushed ridges in the central part of the Netherlands (Veluwe), the subsoil of the ice-pushed ridges consists of fluvial (Rhine and Maas) and fluvioglacial sands and gravel of pre-Saalian and Saalian age. Only in the eastern part of the Netherlands the ice-pushed ridges contain fine-grained sediments of Tertiary age. It is striking that arable fields are present on top of the ridges at Groesbeek despite the coarse-grained sediments and associated low ground water table. The reason for this is the local presence of a loess deposit of a few meters thickness. The loess has a high water holding capacity and is well-suited for agriculture. The loess is undated but attributed to the Weichselian Late Pleniglacial. The presence of loess in this part of the Netherlands is unusual because the aeolian coversand to loess boundary is located c. 75 km to the south. The surrounding regions are all covered by Late Pleniglacial and Late Glacial coversands. The local presence of loess in the Kranenburger glacial basin can be explained by the large-scale glacial morphology and its effect on wind speed and aeolian depositional processes during the Weichselian. Most reconstructions of palaeowind direction during the Late Pleniglacial indicate dominant west to northwest sand-transporting winds (Kasse, 2003). The north-south orientation of the basin and surrounding high ridges is therefore at right angles to the palaeowind flow. This will have caused a local increase of wind speed over the ridges resulting in erosion and the formation of gravel lags with ventifacts. On the other hand, wind speed will have been reduced east of the Nijmegen-Groesbeek ridge and loess was deposited in the wind-sheltered position of the low-lying basin. Indeed, a sand to loam grain-size gradient was reported from west east in the area (Schelling, 1949). Similar small patches of loess, related to local wind-sheltered positions were reported east of the Saalian ice-pushed ridges of the Veluwe.

The small-scale morphology of the area is related to so-called dry valleys that 'drain' the flanks of the ice-pushed ridges (fig. 4). The valleys are younger than the Saalian glaciation but older than the Weichselian Late Pleniglacial loess deposits that form a more or less continuous cover over the valleys. Since Eemian deposits or soils are absent on the ice-pushed ridges and in the dry valleys, it can be concluded that the dry valleys were formed later than the Eemian but before the final dry phase of the Weichselian Late Pleniglacial. Often the presence of permafrost, hampering vertical water infiltration and promoting overland flow, is held responsible for the formation of the valleys (Bogaart et al., 2003). However, also deep seasonal frost will have reduced water infiltration and therefore it is likely that the valleys were active during most of the Weichselian glacial.

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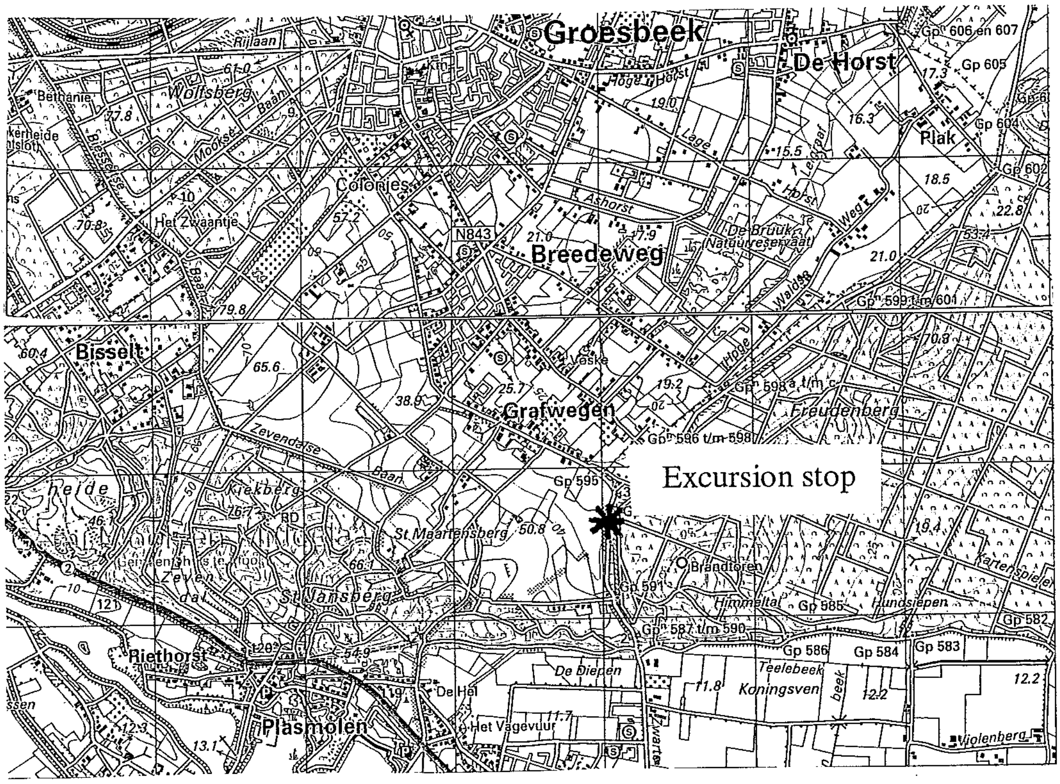

Fig. 1 Location map of excursion stop Groesbeek.

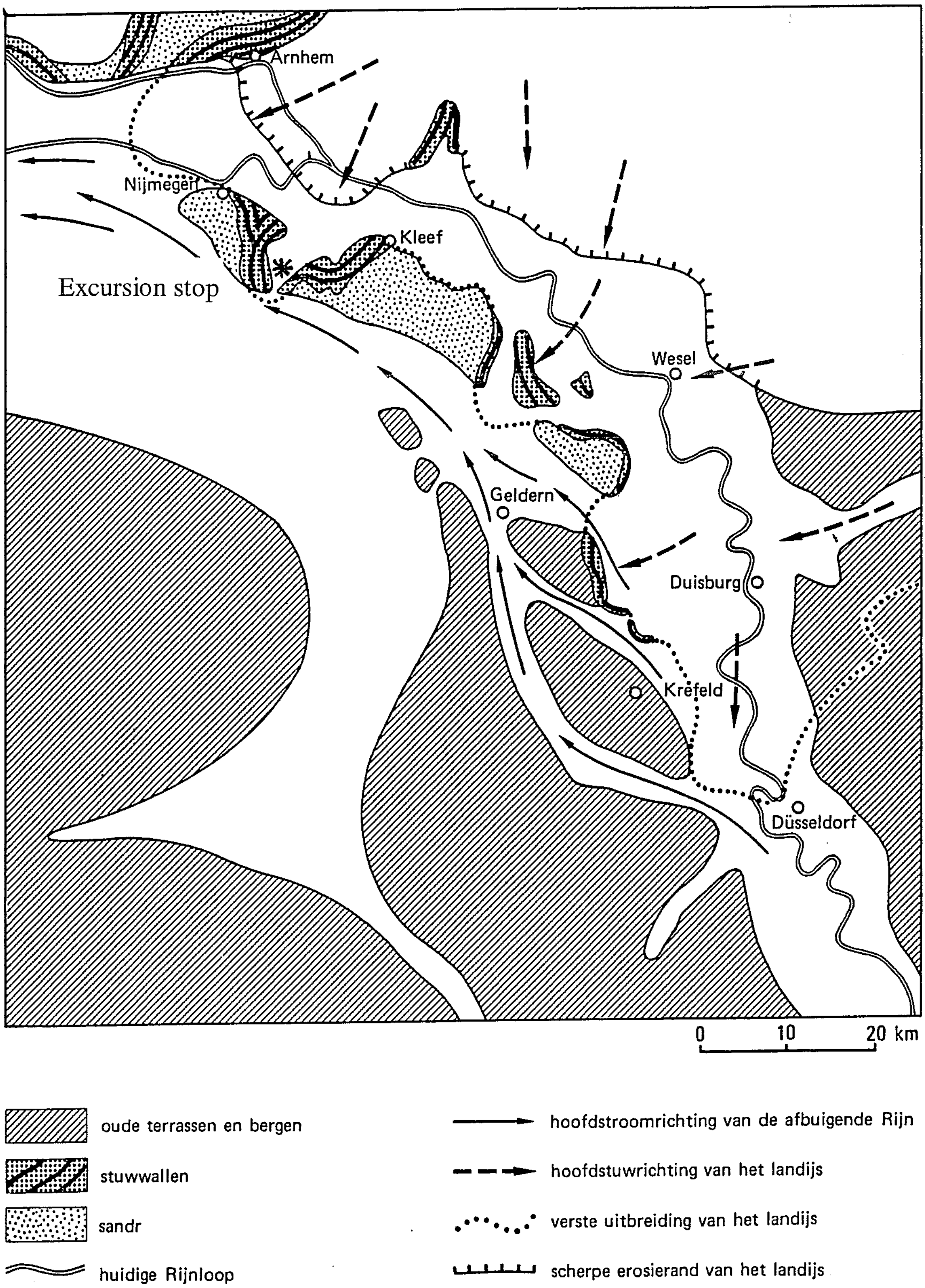

Fig. 2 Deflection of the Rhine to the west because of the advancing Saalian ice sheet (after Stiboka, 1976).

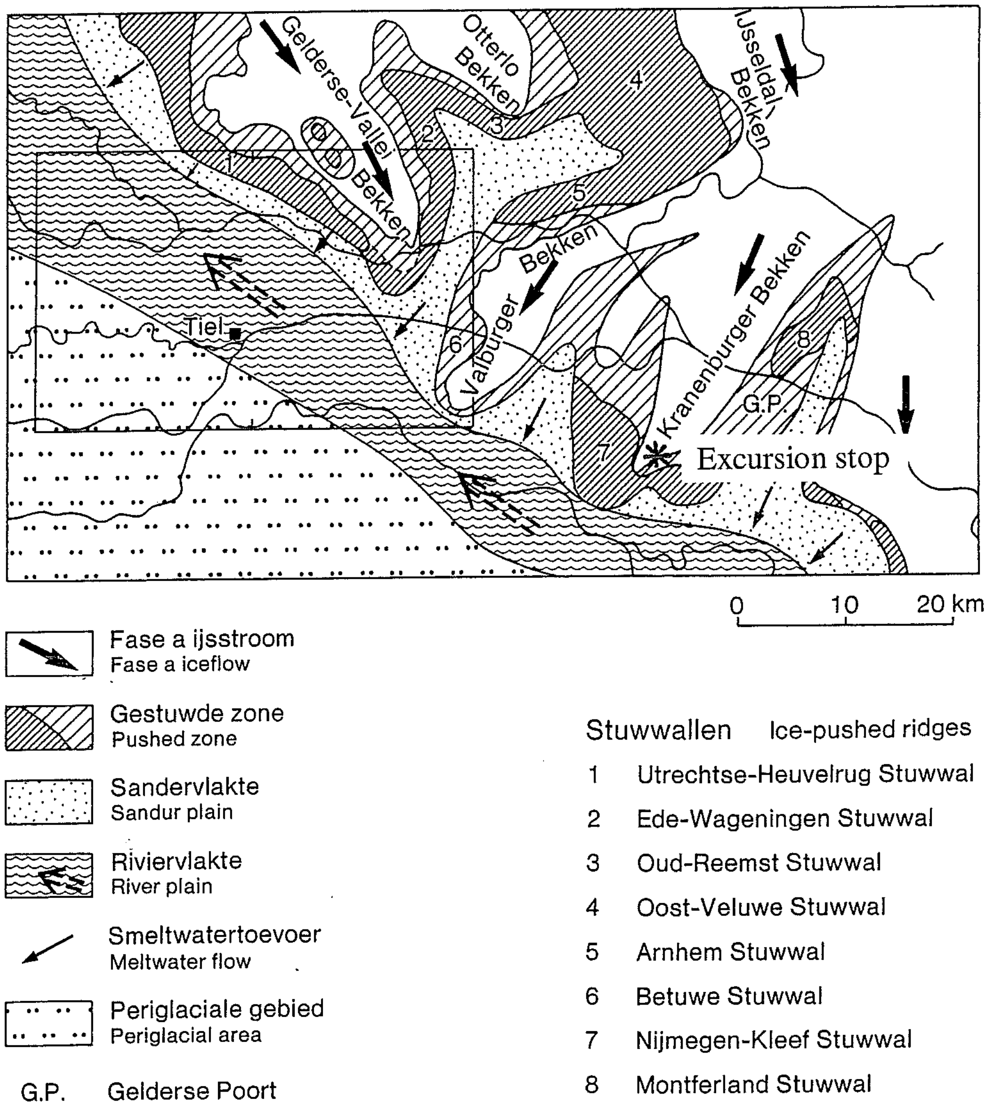

Fig. 3 Palaeogeography of the Saalian glaciation near Nijmegen. The ice-pushed zone is the result of several phases (after Verbraeck, 1984).

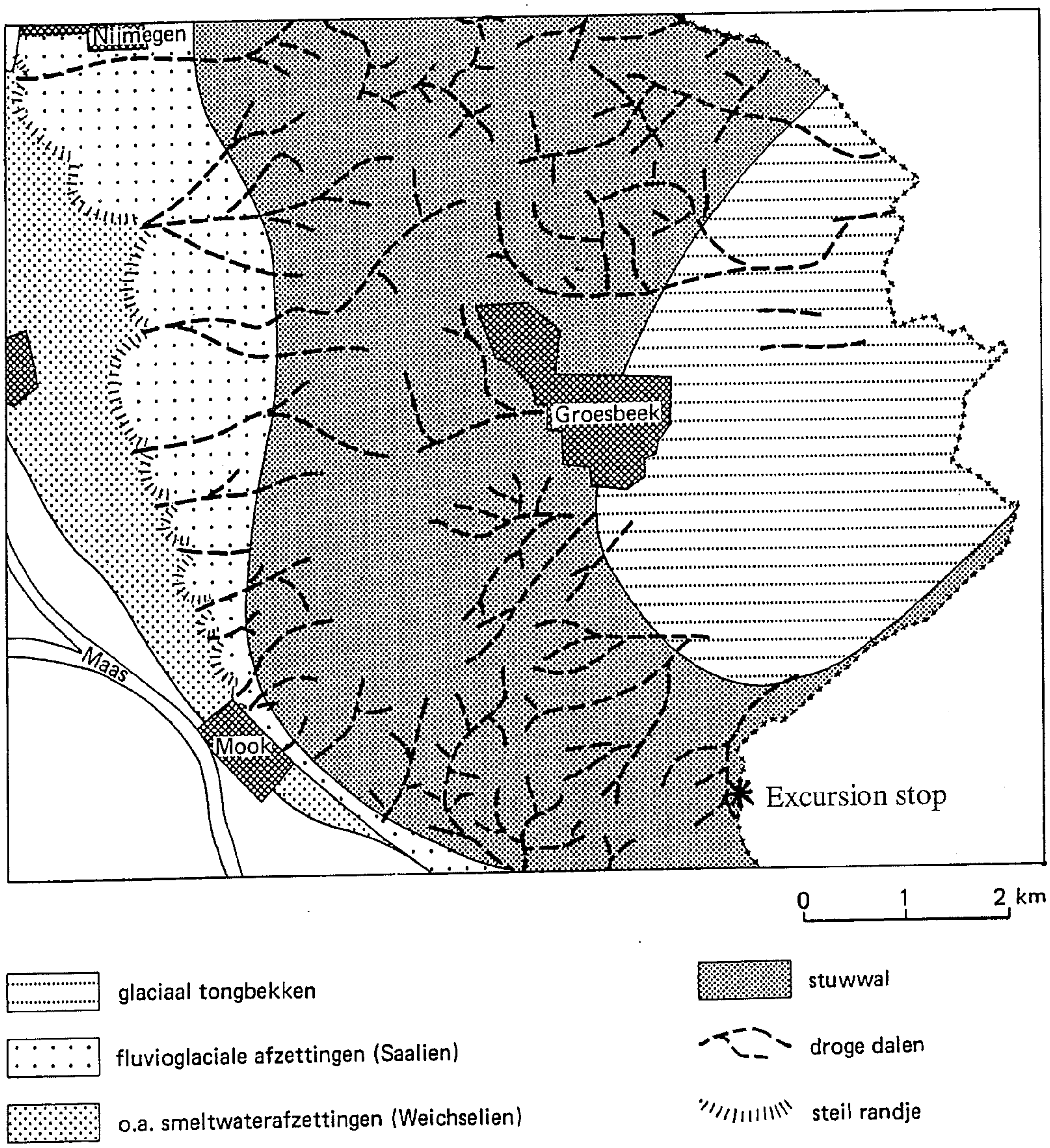

Fig. 4 Glacial and periglacial morphology (dry valleys) near Groesbeek (after Stiboka, 1976).

EXCURSION STOP 2: ST. JANSBERG

- Erosional scarp of the Saalian ice-pushed ridges by the Saalian Niers-Rhine
- Fluvial morphology of the Weichselian Pleniglacial system

The excursion stop is located 2 km east of Milsbeek. The elevation is c. 12 m a.s.l. (fig. 1). This point is located immediately south of the Saalian ice-pushed ridges and sanders of Nijmegen-Groesbeek and the Reichswald in Germany.

The Saalian glaciation has had an enormous impact on the drainage pattern of the Rhine-Maas system in the Dutch delta plain. Previous to the glaciation the Rhine-Meuse system was flowing through a flat delta plain and had a northnorthwesterly course in the direction of the present-day IJssel Valley and Gelderse Vallei. This direction seems to be tectonically controlled by the more rapid subsidence of the North Sea Basin in that area. The Saalian glaciation totally transformed the landscape by forming up to100 m high ridges especially in the central part of the Netherlands. As a result the original northnorthwesterly course was abandoned and a more westerly flow established which continued with some interruptions until the present day (fig. 2). During the maximum phase of Saalian glaciation Rhine and glacial melt water were deflected to the west along the so-called Niers-Rhine valley (Van de Meene and Zagwijn, 1978; Verbraeck, 1984). This major stream eroded the southern margin of the glacial deposits and a straight and steep scarp was formed. At the excursion point the scarp is up to 50 m high.

As a result of the deglaciation up to 100 m deep glacial basins occurred in the central Netherlands (e.g. IJssel Basin). The Rhine extended its course in a northern direction into the deep IJssel Basin and deposited a sequence of glaciolacustrine and coarse-grained deltaic sediments in the former glacial basin. It is likely that this northerly route into the IJssel Basin had a more favourable gradient (local low base level) and therefore the Niers-Rhine branch was abandoned. This situation persisted in the Eemian during which the Rhine occupied the IJssel Valley while the Niers valley was not in use. However, wet depositional environments did occur, since organic Eemian deposits are known from the subsoil of the Niers valley.

In the course of the Weichselian the IJssel Rhine was abandoned and the Rhine took a westerly route between Arnhem and Nijmegen (Montferland and Gelderse Poort Rhine courses, fig. 2). Also the Niers valley was used again by the Rhine. The reasons for these shifts are not well known but silting-up of the IJssel Basin may have decreased the floodplain gradient hampering the northern flow. The channel belt avulsion to the west took place possibly because of the higher floodplain gradient in that direction. During the Weichselian Pleniglacial the Niers valley was the westernmost Rhine branch. The braided river floodplain of that period is still visible on the present-day surface at the excursion stop. The braided channel morphology is especially preserved along the margins of the Niers valley, because younger Late Glacial systems have eroded the Pleniglacial system in the central part of the valley (see next excursion stop Bons).

References

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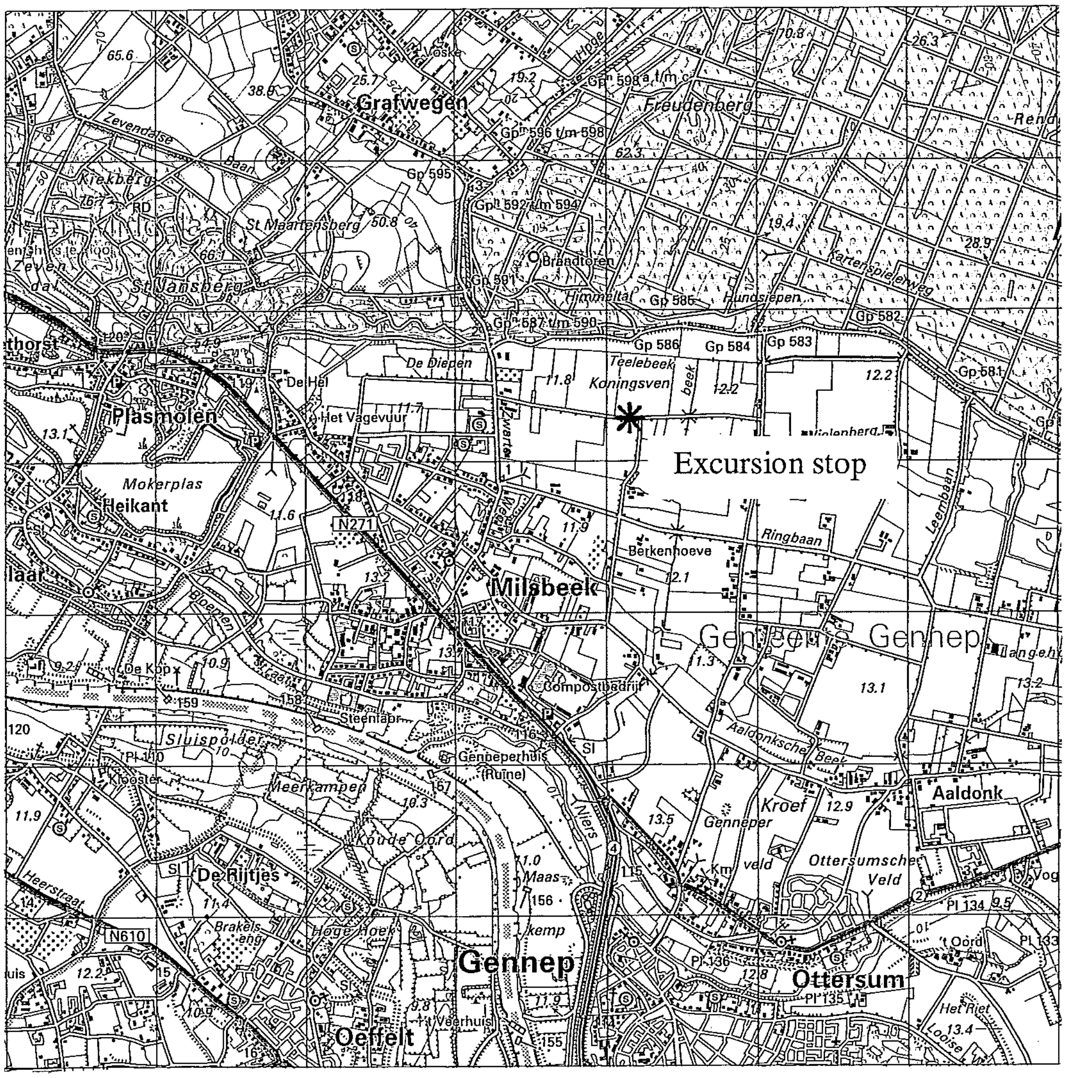
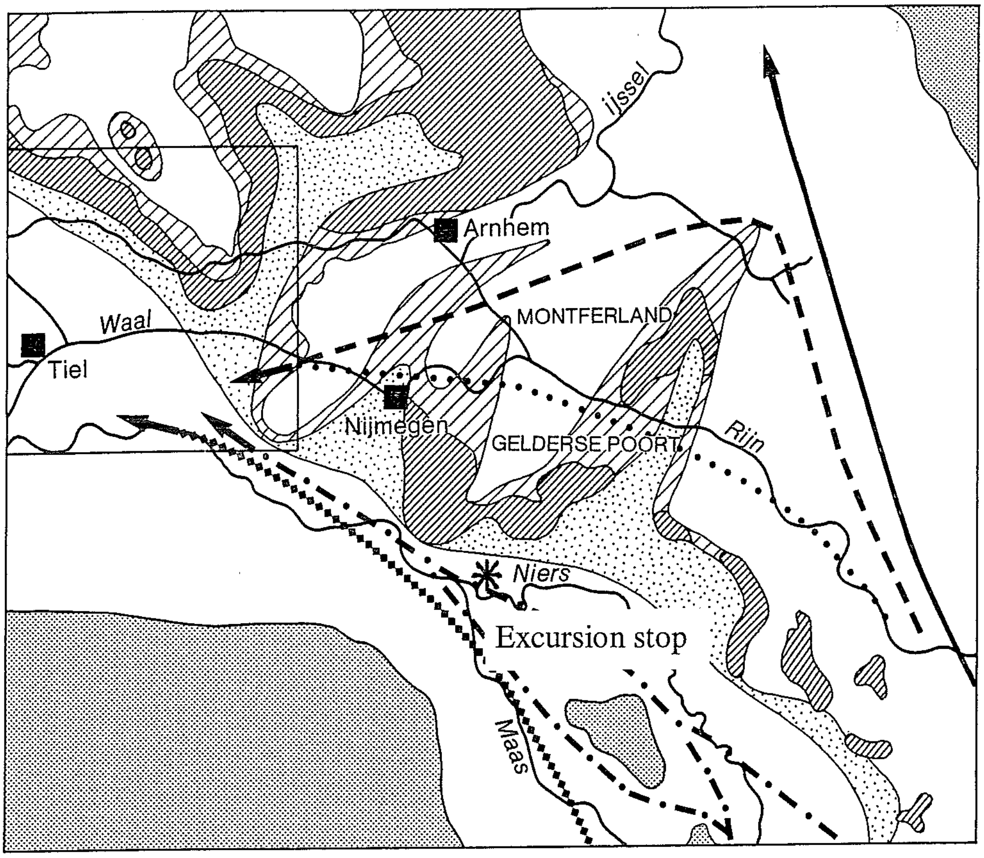


Fig. 1 Location map of excursion stop St. Jansberg.



Rivierlopen	Saalien		Eem.	Weichselien		Hol.
	Midden	Laat		Vroeg	Midden	Laat
	Kr I	Kr II	Kr III	Kr IV	Kr V	Kr VI
Niersdal Rijn	•			•	(•)	
Gelderse-Poort Rijn		•			•	•
Rond-Montferland Rijn		•	•	•		
IJsseldal Rijn		•	•	•		
Maas	•	•	•	•	•	•

Gestuwde zone
Ice-pushed zone

Midden-Saalien sandervlakte
Middle Saalian sandur

Fig. 2 River courses of the Rhine since the Saalian (after Verbraeck, 1984).

EXCURSION STOP 3: BONS

Climate and vegetation change and Late Glacial fluvial response of the Niers-Rhine (western Germany) (after Kasse et al, in prep)

- Morphology of Weichselian Late Glacial large meanders of the Niers-Rhine
- Coring of the fine-grained fill and palynology of the abandoned meanders
- Final abandonment of the Niers-Rhine system

Summary

The Niers valley was part of the Rhine system that came into existence during the maximum Saalian glaciation and was abandoned at the end of the Weichselian. The aim of the study was to explain the Late Pleniglacial and Late Glacial fluvial dynamics and to explore the external forcing factors climate change, tectonics and sea level.

The sedimentary units have been investigated by large-scale coring transects and detailed cross-sections over abandoned channels. The temporal fluvial development has been reconstructed by means of geomorphological relationships, pollen analysis and ^{14}C dating.

The Niers-Rhine experienced a channel pattern change from braided, via a transformational phase to meandering at the Late Pleniglacial to Late Glacial transition. This change in fluvial style is explained by the early Late Glacial climate amelioration (at c. 12,5 ka ^{14}C BP) and climate-related hydrological, lithological and vegetation changes. A delayed fluvial response of c. 400 years (transitional phase) was established. The channel transformations are not related to neotectonic effects and sea-level changes. Successive river systems have similar gradients of c. 35-40 cm/km.

A meandering river system dominated the Allerød and Younger Dryas periods. The threshold towards braiding was not crossed during the Younger Dryas. The final abandonment of the Niers-Rhine was dated at the Younger Dryas to Holocene transition. The first traces of Laacher See pumice have been found in the Niers valley indicating that the Niers-Rhine still existed in the Younger Dryas period.

Introduction

The Niers valley is situated in Germany and the adjacent southeastern Netherlands. The valley was formed during the maximum Saalian glaciation (fig. 1). Then, the Rhine flow was diverted to the west by the southward migration of the Scandinavian ice sheet (Van der Meene & Zagwijn, 1978; Klostermann, 1992). During (parts of) the Weichselian the Niers valley was the westernmost Rhine course. At the end of the Weichselian the Niers-Rhine was finally abandoned.

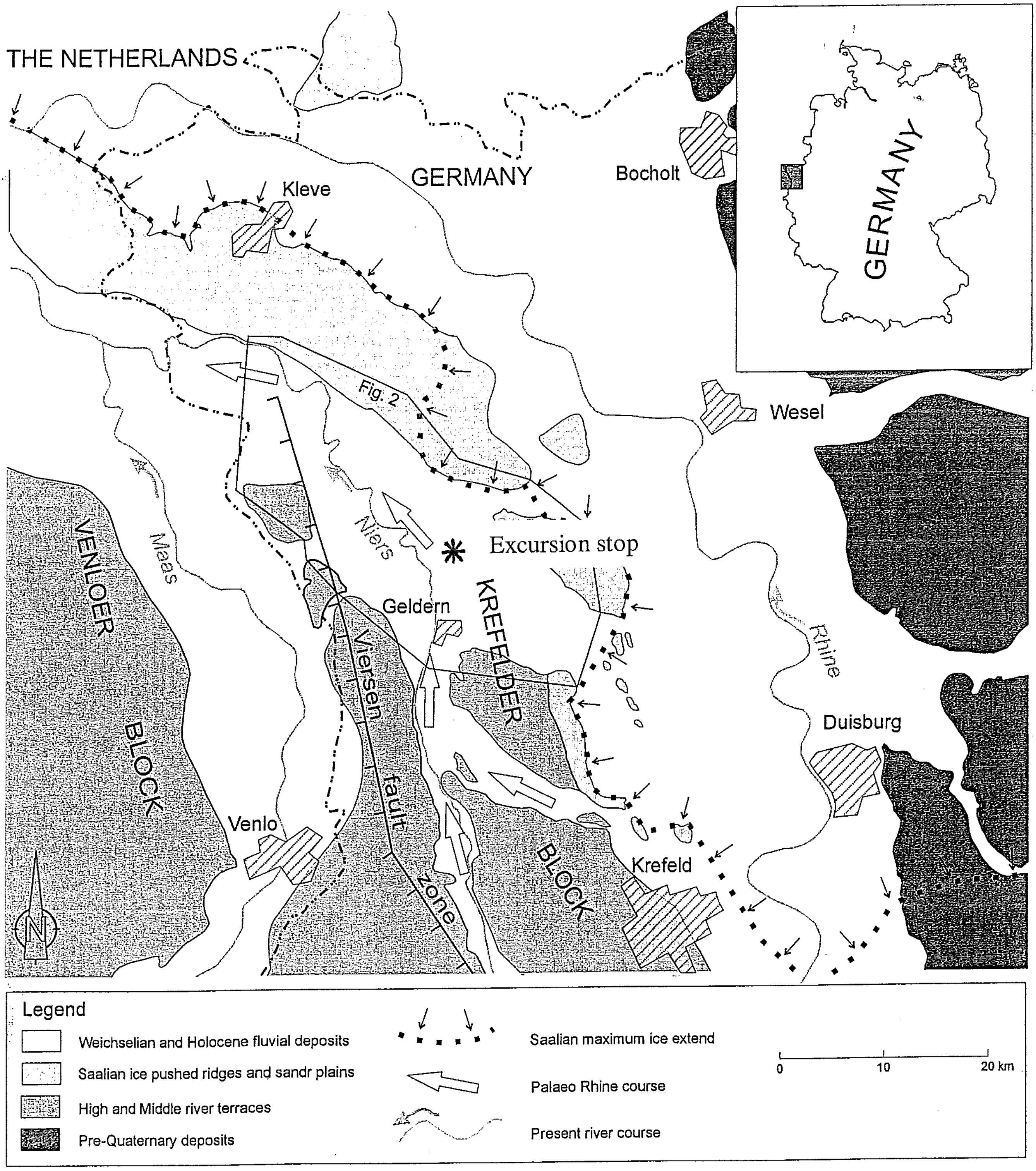


Fig. 1 Geological setting of the Niers Valley.

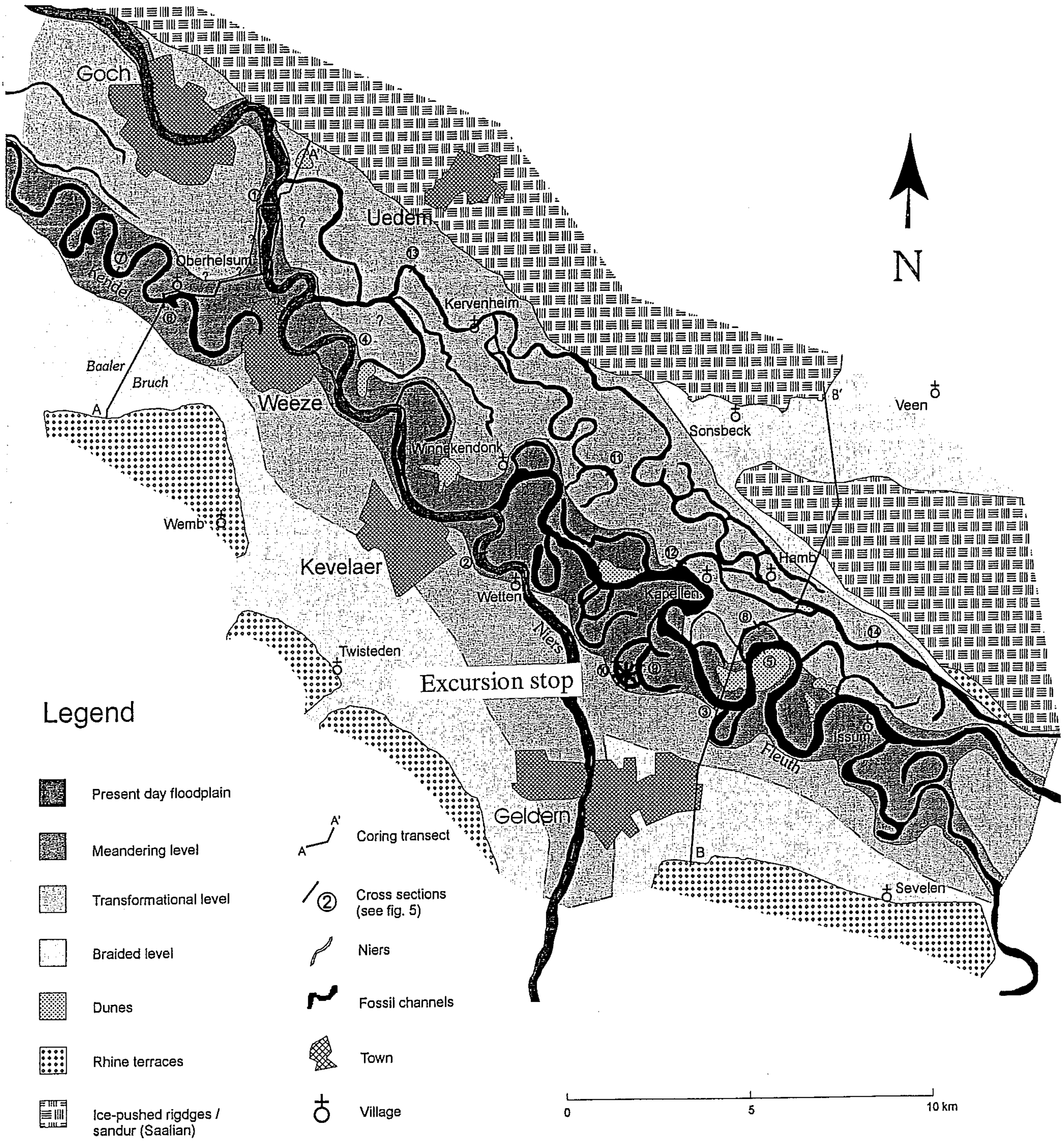


Fig. 2 Late Pleniglacial to Holocene floodplain levels in the Niers valley.

Geomorphology of Late Pleniglacial to Holocene floodplain levels

Three Late Pleniglacial and Late Glacial palaeofloodplain levels have been distinguished above the Holocene floodplain (fig. 2): an oldest level with braiding characteristics at the outer limits of the Niers valley, a transformation level formed during a phase of channel transformation and a level with meandering channel characteristics. It is emphasized that floodplain levels have been distinguished, not terraces, since the different units occur more or less at the same level.

The oldest level has a very subdued morphology and differences in elevation are generally no more than 1 m. The geological maps of the area show slightly elevated gravel bars and connecting and disconnecting channels, suggesting a braided channel pattern during the formation of this level.

The transformation level is well expressed along the northern margin of the valley (fig. 2). It is characterized by straight channels and well-developed small meanders. The channels are not wider than 100 m and the thickness of the fine-grained infill is no more than 1.5 m. Infill of the channels is mostly clastic with an alternation of loam and organic loam beds. These palaeochannels seem to represent the transformation from the multi-channel, braided system, when the full valley width of 5 to 10 km was occupied by the river system, to the younger, single-channel meandering system.

The next younger level is characterized by high-sinuosity meandering channels of several generations. The clear cross-cutting relationships enable to establish a relative chronology of the meander scars, especially east of Kevelaer. The palaeomeanders are generally c. 200 m wide and the fine-grained infill is c. 2 to 3 m thick. Bankfull depth may have been more than 5 m. The infill of the older large palaeomeanders consists mostly of loam or peaty loam and peat. The youngest palaeomeander system of the Issumer Fleuth, however, is organic with lacustrine gyttja and peat. Small dune fields occur in association with this last meander belt (presently occupied by the Issumer Fleuth and Niers, see fig. 2).

The successive palaeofloodplain surfaces (braiding-transformational-meandering) have similar elevations and gradients of c. 35-40 cm/km. The transformational and meandering systems show a (normal) downstream decrease in gradient. The reconstructed gradients reveal some important elements: First, the change in channel pattern of the successive floodplain levels is not related to differences in gradient. Secondly, neotectonic effects on the river gradients could not be established. Thirdly, the impact of sea-level changes was negligible during the formation of the floodplain levels. Fourthly, based on the considerations above, it is postulated that climate change and climate-related changes in water and sediment supply, were the main factors that determined the changes in channel pattern from braiding to meandering (Vandenberghe, 2002).

Sedimentary succession of the palaeofloodplain levels

Two large-scale coring transects perpendicular to the main valley axis (fig. 3) and many detailed cross-sections over abandoned channels have been made (see fig. 2 for location). It is clear that on a large scale the top of the gravel in the cross sections shows a general

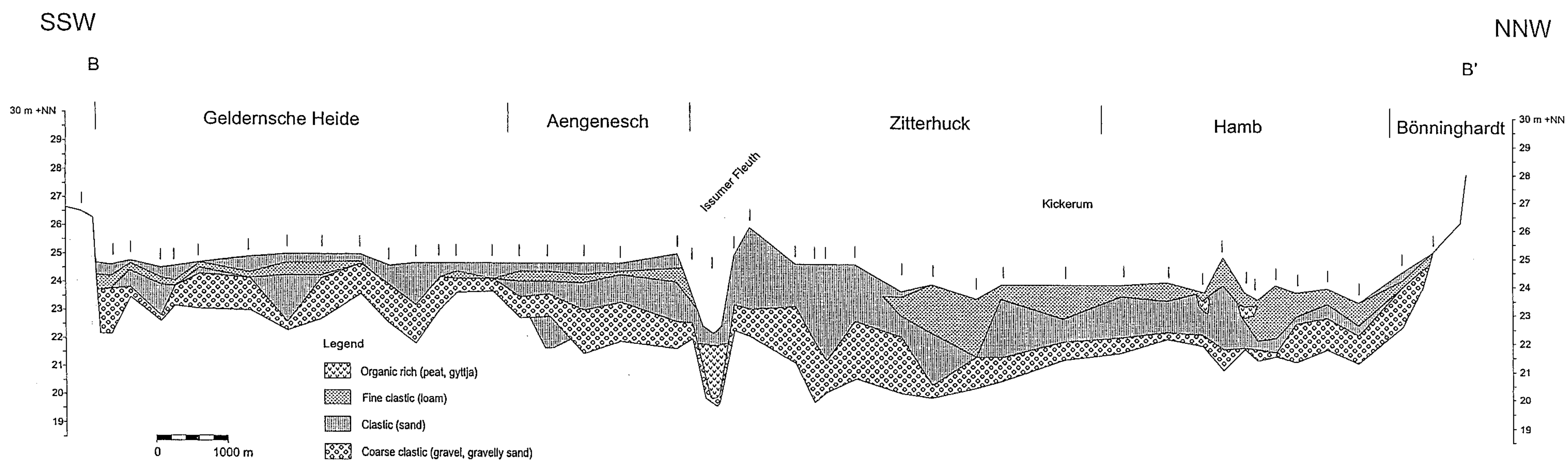


Fig. 3 Lithological cross section over the Niers valley.

dip towards the northeast (fig. 3). In addition to this, in the central part of the profiles the gravel is generally found at a greater depth than at the margins. At the southern margin of the Niers valley, where a braided surface morphology has been found, gravel and gravelly sand is normally found at a shallow depth (average c. 1 m). A loam layer is present at the top of the sequence (Hochflutlehm).

In the area with the transformational river system near Hamb and Oberhelsum (fig. 3) gravelly sediment occurs at a larger depth (c. 1-2 m) and the overlying sand and loam units are more important. In the central part of the valley, with clear large-scale meandering surface patterns, gravel occurs at a depth of 2-4 m. The overlying sandy unit (c. 1-3 m thick) often reveals a clear fining-upwards tendency formed by lateral migration of the meandering channels. Locally, small dune fields are found in association with the last meander generation of the meandering system (fig. 3: Finkenhorst).

Pollen diagram Reinders (figs. 4 and 5)

This pollen diagram is from the transformation level. Channels are straight to sinuous, small (50 m wide) and shallow (1.5 m) (fig. 4). They represent the last stage and the transformation of the multi-channel, braided system, when the full valley width of 5 to 10 km was occupied by the river system, to the younger, single-channel meandering system. River water flow became more and more concentrated in fewer channels. The fill of the channel is mostly loamy because of clastic influxes from younger channel generations.

The frequent occurrence of *Picea, Abies, Corylus, Alnus* and high values of *Pinus* throughout the diagram indicate an allochthonous component in the whole pollen diagram. This hampers the interpretation of the pollen diagram.

The lower part of the diagram, from 155 to 117 cm, shows a gradual *Pinus* decrease and increase of the Gramineae values. The high *Pinus* values at the base coincide with a sandy loam lithology and can probably be attributed to reworking. The continuous presence of *Hippophae* is a strong indication for the Early Dryas biozone (c. 12,1-11,9 ka ^{14}C BP; fig. 11). The interval from 137 to 117 cm below the surface reveals an increase in *Betula* that could correspond to the *Betula* phase of the Allerød (11.9-11.3 ka ^{14}C BP; Hoek, 1997).

The upper part from 117 to 80 cm is strongly dominated by high *Pinus* values. Theoretically, these high Pine values could be part of the Preboreal *Pinus* phase, however, typical elements for the early Holocene like *Monoleet psilate* and *Typha latifolia* are missing and therefore this part of the diagram predates the Holocene. Most likely, the interval from c. 95 to 117 cm represents the *Pinus* phase of the late Allerød (c. 11,2-10,9 ka ^{14}C BP; fig. 11).

From c. 95 cm upwards there is a gradual increase in *Artemisia, Pediastrum* and *Salix*, probably reflecting the climatic cooling of the Younger Dryas period.

Four ^{14}C-dates have been obtained from the channel fill. The lowermost date (1.28-1.30 m below surface: 11,710 ± 80 BP) indicates an early Allerød age and is in agreement with the Allerød *Betula* phase biostratigraphic interpretation. The sample at 1.11-1.13 m was taken to discriminate between the Allerød pine phase or the early Holocene pine phase. The date obtained (11,300 ± 60 BP) supports the biostratigraphic interpretation of the Allerød pine zone. The uppermost dates (0.88-0.90 m) were obtained

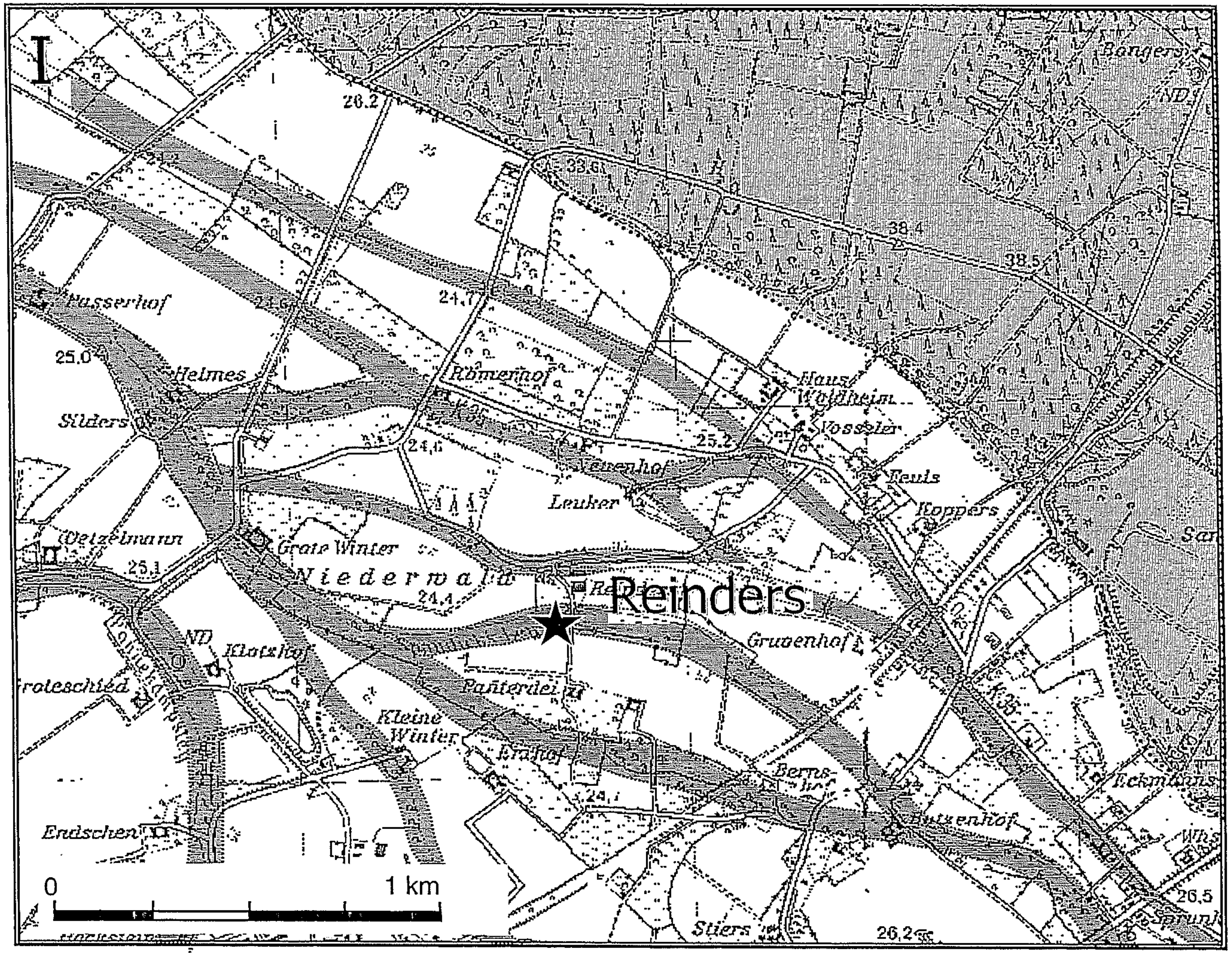


Fig. 4 Channel pattern of the final braided phase near Reinders.

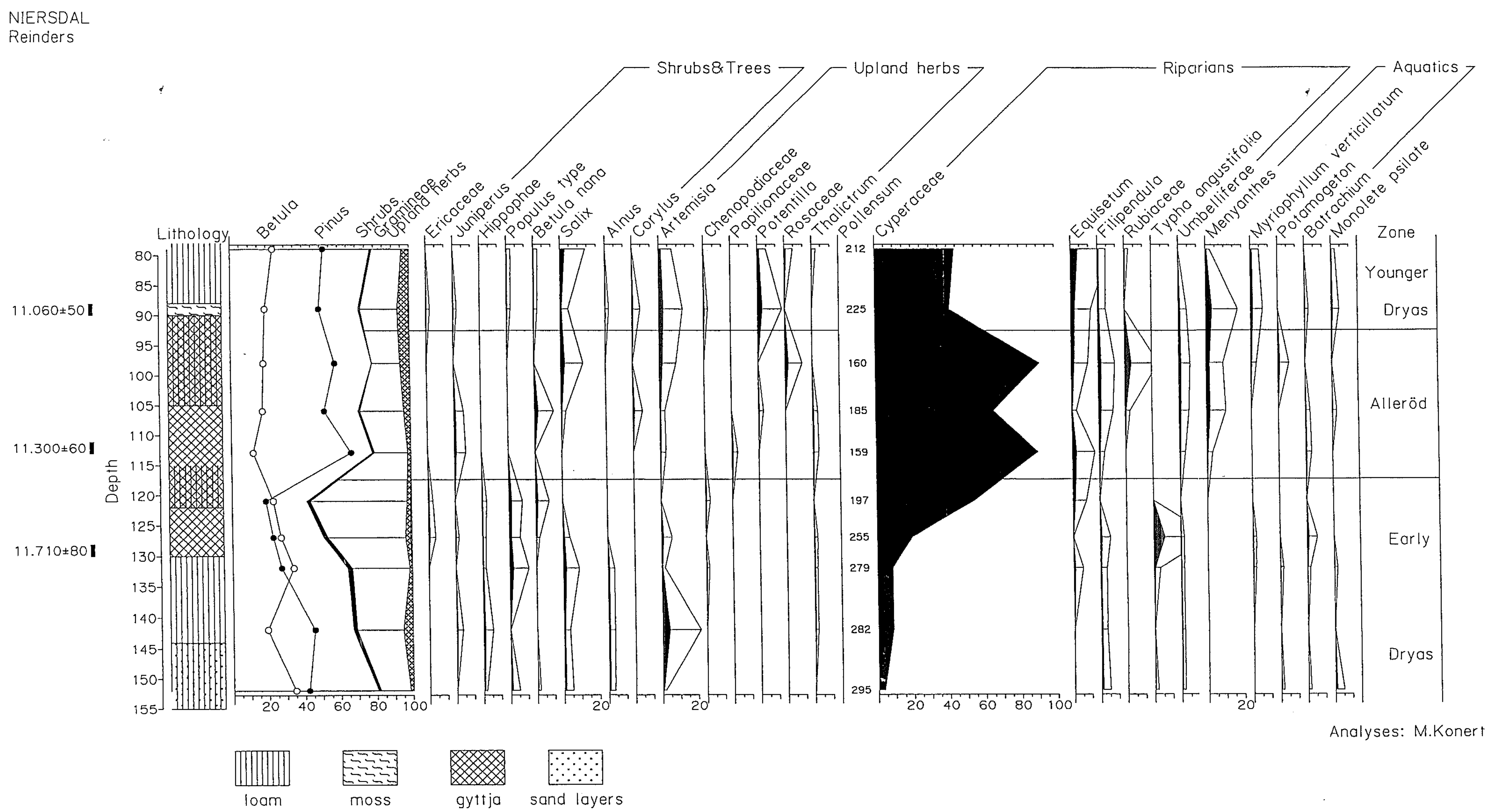


Fig. 5 Pollen diagram Reinders.

from a well-developed moss mat. The ^{14}C-date on the mosses (11,750 ± 60 BP) is older than the underlying ^{14}C-dates. Most likely, the date on the mosses has suffered from the uptake of old carbon from the water body by the floating mosses. Therefore, the date on the mosses is rejected. The ^{14}C-date on the terrestrial macro remains from the same level (11,060 ± 50 BP) is in agreement with the biostratigraphic interpretation of the onset of the Younger Dryas.

The age of the basal part of the fill indicates that the transformation from a braided to a meandering system occurred during or shortly before the Early Dryas period (c. 12,1 ka ^{14}C BP). After abandonment the channel scar was frequently inundated and filled with clastic suspension material from younger meandering channel generations during the Allerød and Younger Dryas.

The dates indicate that the change in fluvial style does not coincide with the early Late Glacial climate amelioration (at c. 12,5 ka ^{14}C BP) and climate-related hydrological changes. The fluvial response seems to be delayed by c. 400 years (transitional phase) which can be related to the gradual Late Glacial development of the vegetation cover and related changes in water discharge (more regular) and sediment supply (decreasing input) (Hoek, 2001; Hoek & Bohncke, 2002).

Cross section Bons A (figs. 6 and 7)

This cross section is from the high-sinuosity meandering system with clear cross-cutting relationships that enable relative dating of the meanders scars (fig. 6). The palaeomeander is circa 200 m wide. The cross section is clearly asymmetric with the stoss side at the left (fig. 7). The fill is 2 to 2.5 m thick and consists of loam at the base overlain by organic material.

Palaeodischarge calculations reveal that the palaeoNiers-Rhine discharge was c. 10 to 20 % of the present-day Rhine discharge, indicating that during the Late Glacial probably a more northerly course co-existed with the Niers-Rhine.

Pollen diagram Bons A (fig. 8)

The lower part of the fill below 162 cm is clastic in nature and characterized by high *Pinus* values (fig. 8). In addition, some *Alnus* and *Carpinus* pollen is present and therefore it is likely that part of the *Pinus* pollen has been reworked. The presence of *Empetrum*, A*rtemisia, Helianthenum* and *Pediastrum* indicates the (final phase of the) Younger Dryas period (c. 10,9-10,1 ka ^{14}C BP; fig. 11). *Juniperus* attains maximum values towards the end of this zone. It is succeeded by an increase of *Betula* that represents the Preboreal *Betula* phase (147-162 cm). The increase of *Filipendula* and *Typha latifolia* is also characteristic for the start of the Holocene. The Younger Dryas to Holocene boundary fully coincides with a change in lithology from loam to peat at 162 cm, indicating the total stop of fluvial, suspension-rich, inundation due to the final abandonment of the Niers-Rhine system.

From 137-147 cm *Betula* has declined and Gramineae values reach a maximum. This zone is equivalent with the Rammelbeek phase (c. 9,9-9,7 ka ^{14}C BP; fig. 11).

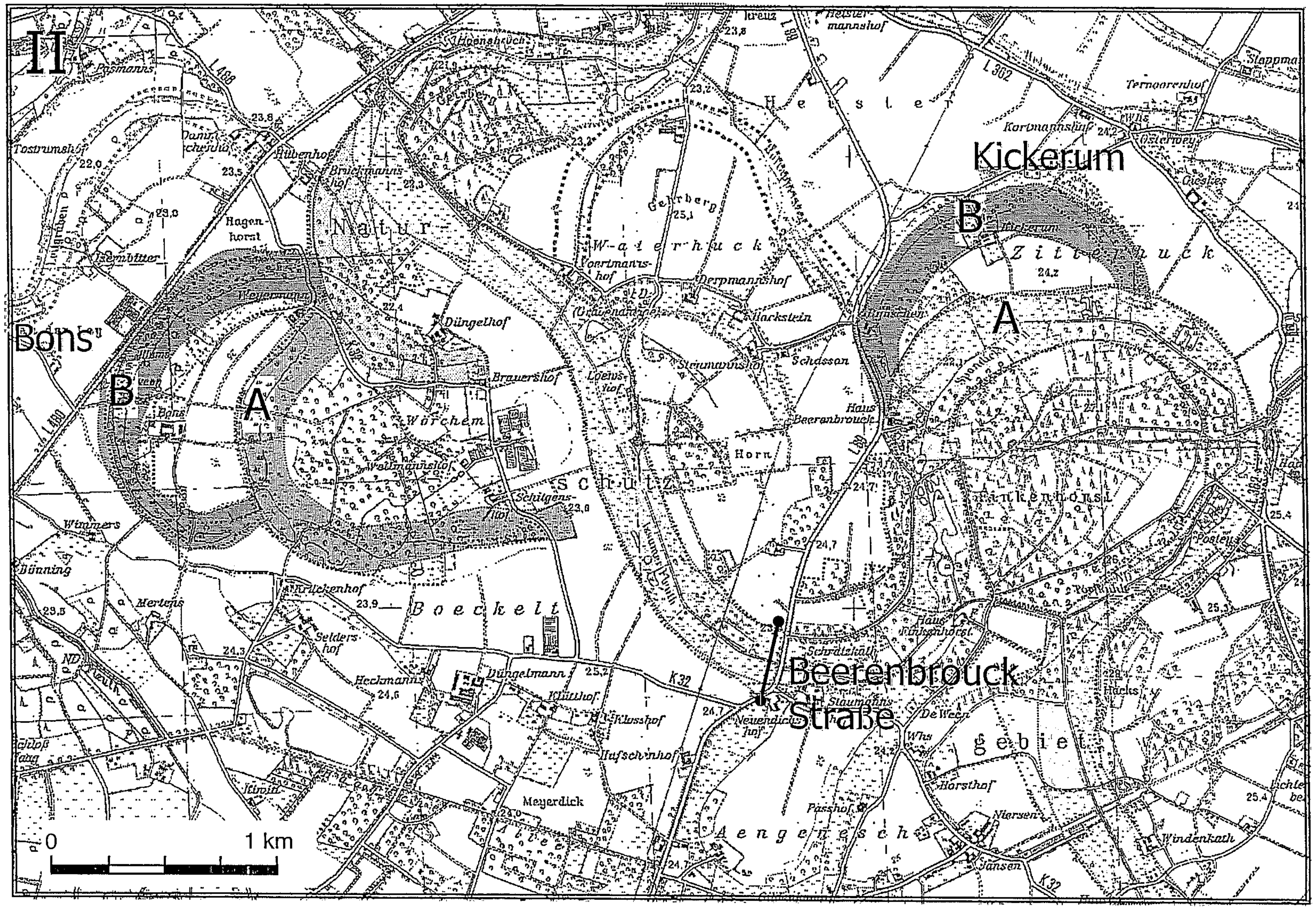


Fig. 6 Cross-cutting relationships of the meandering system near Bons.

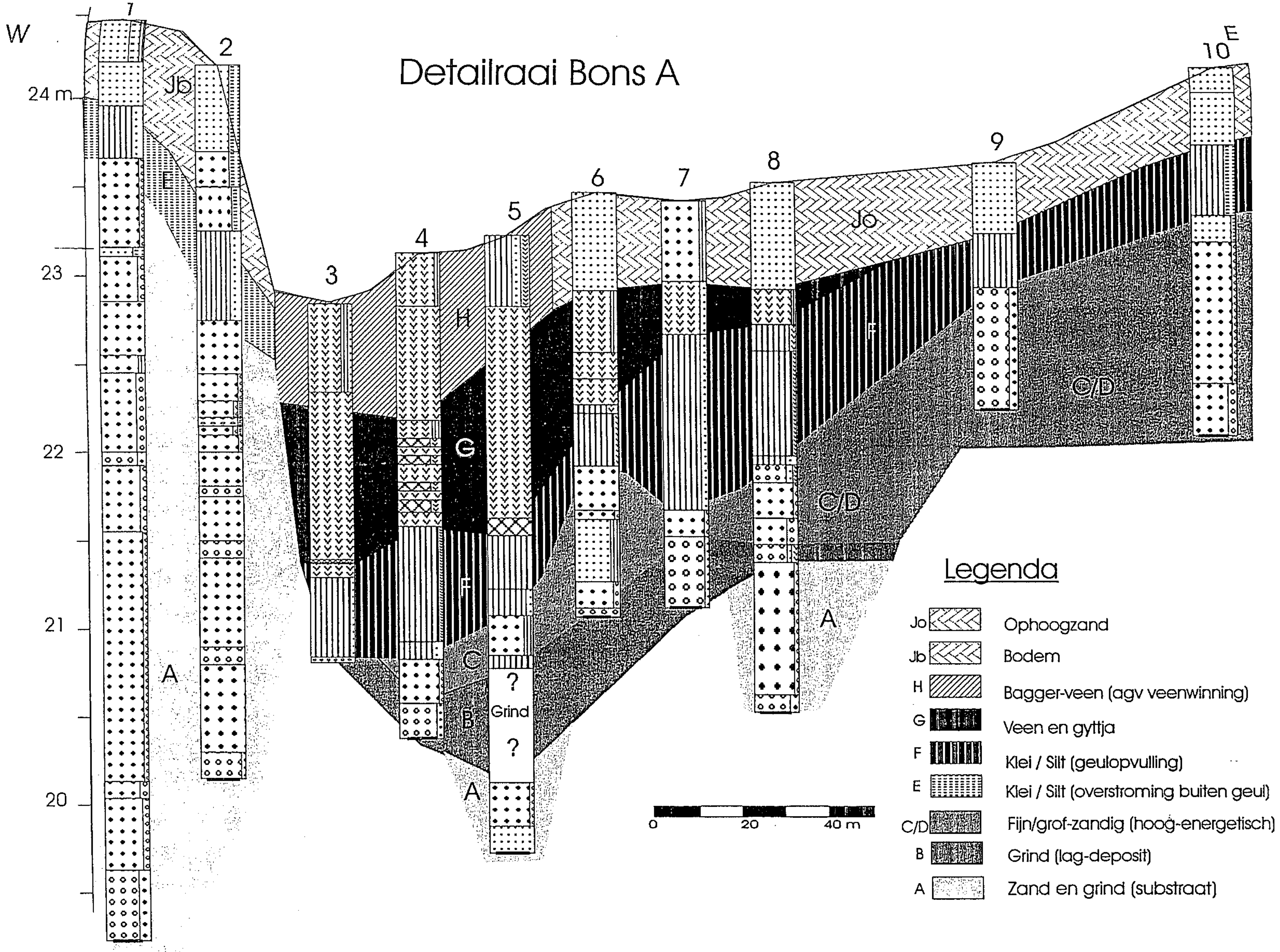


Fig. 7 Cross section Bons A.

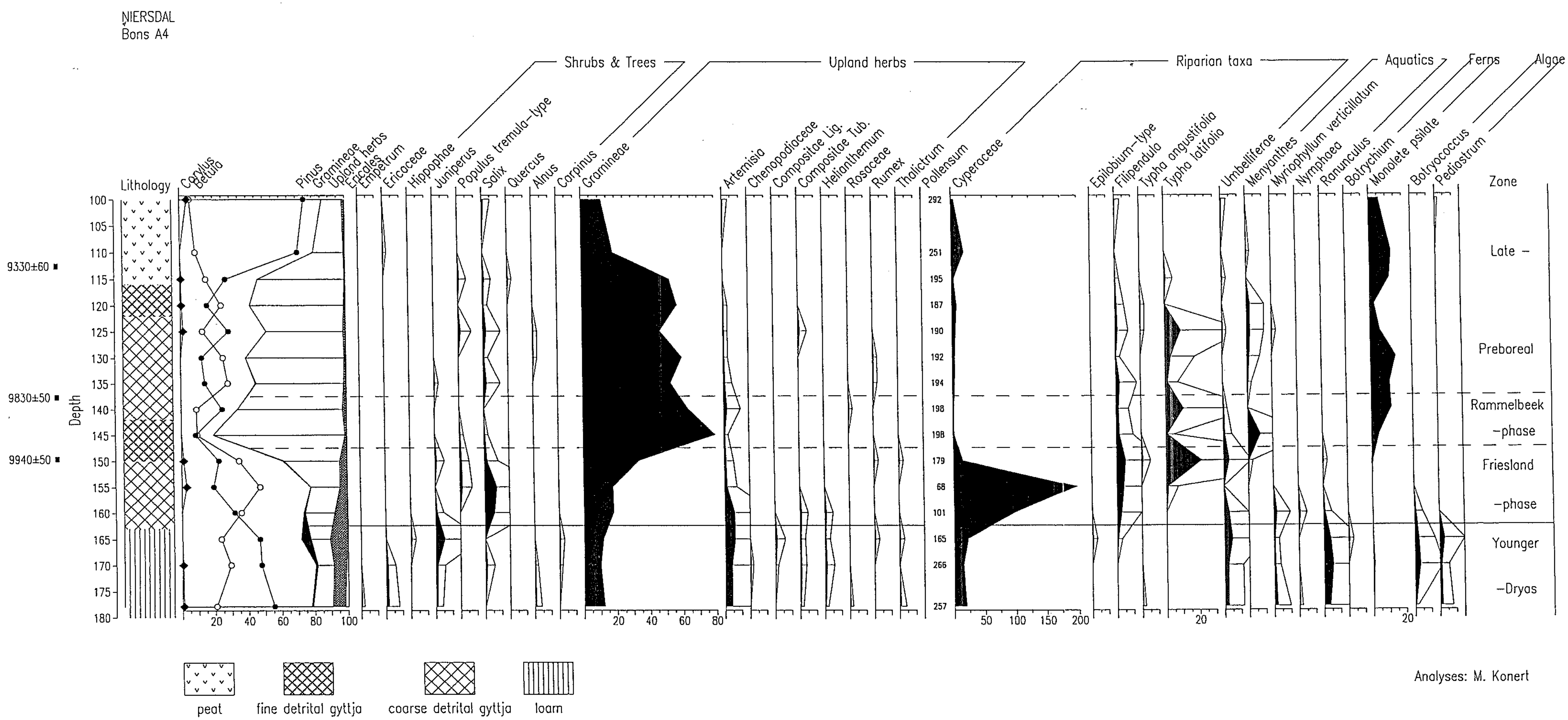


Fig. 8 Pollen diagram Bons A.

Although the Gramineae values remain high, *Betula* shows an increase again at 137 cm, which is interpreted as the start of the Preboreal *Betula* phase. The strong increase of *Pinus* at 112 cm is the start of the late Preboreal *Pinus* phase (100-112 cm).

Four ^{14}C-dates from this site fully confirm the biostratigraphical interpretation. The samples taken from the base and top of the Rammelbeek phase gave ages of 9940 ± 50 and 9830 ± 50 BP respectively. The sample at 112 cm, dating the start of the Holocene *Pinus* phase, gave an age of 9330 ± 60 BP. Abundant *Menyanthus* seeds from the same level gave an age of 9280 ± 50 BP indicating that a hard-water effect and consequently older ages, frequently reported for aquatic plants, did not occur.

The lowermost spectra of the infill are from the Younger Dryas period. This biostratigraphical interpretation indicates that the Bons infill is older than the Kickerum A infill which is also supported by the geomorphological evidence (fig. 6). This finding indicates that the Bons meander was abandoned during the Younger Dryas and that a meandering system was active in the Niers valley during or before the Younger Dryas stadial (Allerød).

Pollen diagram Kickerum A (fig. 9)

The pollen diagram is from the last generation of high-sinuosity meanders, formed just before the final abandonment of the Niers-Rhine valley (fig. 6). The fine-grained infill shows a symmetric channel form. The palaeomeander is over 250 m wide and the organic infill is circa 2 m thick. In the deepest part of the channel the fill consists of calcareous gyttja changing upward into peat. Small dune fields occur in association with this last meander belt (see fig. 2), indicating deflation of temporarily dry channels or pointbar slopes, possibly related to increased discharge fluctuations. Klostermann (1986, p. 78-81) published a palynological study from the same palaeomeander circa 1 km southwest of Kickerum A.

The lower part, below 172 cm, of the diagram is characterized by *Pinus* and *Betula* and shows a strong dominance of the Gramineae (fig. 9). This zone is indicative for the early Preboreal Rammelbeek phase (c. 9,9-9,7 ka ^{14}C BP; fig. 11). It could be argued, because of the low arboreal pollen values, that this zone is part of the Younger Dryas biozone, However, the absence of *Empetrum* and low value of *Juniperus* exclude a Younger Dryas age. The overlying zones are dominated by *Betula* (142-172 cm) and *Pinus* (117-142 cm) which are correlated with the Preboreal *Betula* and *Pinus* phases. The upper part of the diagram, with increasing *Corylus*, and low *Quercus* and *Ulmus* values, is indicative for the start of the Boreal (c. 9,1 ka ^{14}C BP; fig. 11).

The biostratigraphic interpretation is supported by a ^{14}C-date of 10,070 ± 60 BP from the base of the channel fill. The date indicates that the infilling started at the Younger Dryas to Holocene transition or shortly afterwards in the early Holocene. Therefore, the moment of abandonment of the large-scale meandering Niers-Rhine system is situated close to the Weichselian to Holocene boundary. It means that the Niers-Rhine course was still active during the Younger Dryas period. This contrasts with previous findings in which the final abandonment of the Niers-Rhine course was placed before the end of the Allerød (Verbraeck, 1984, p. 107). Furthermore, it shows that the

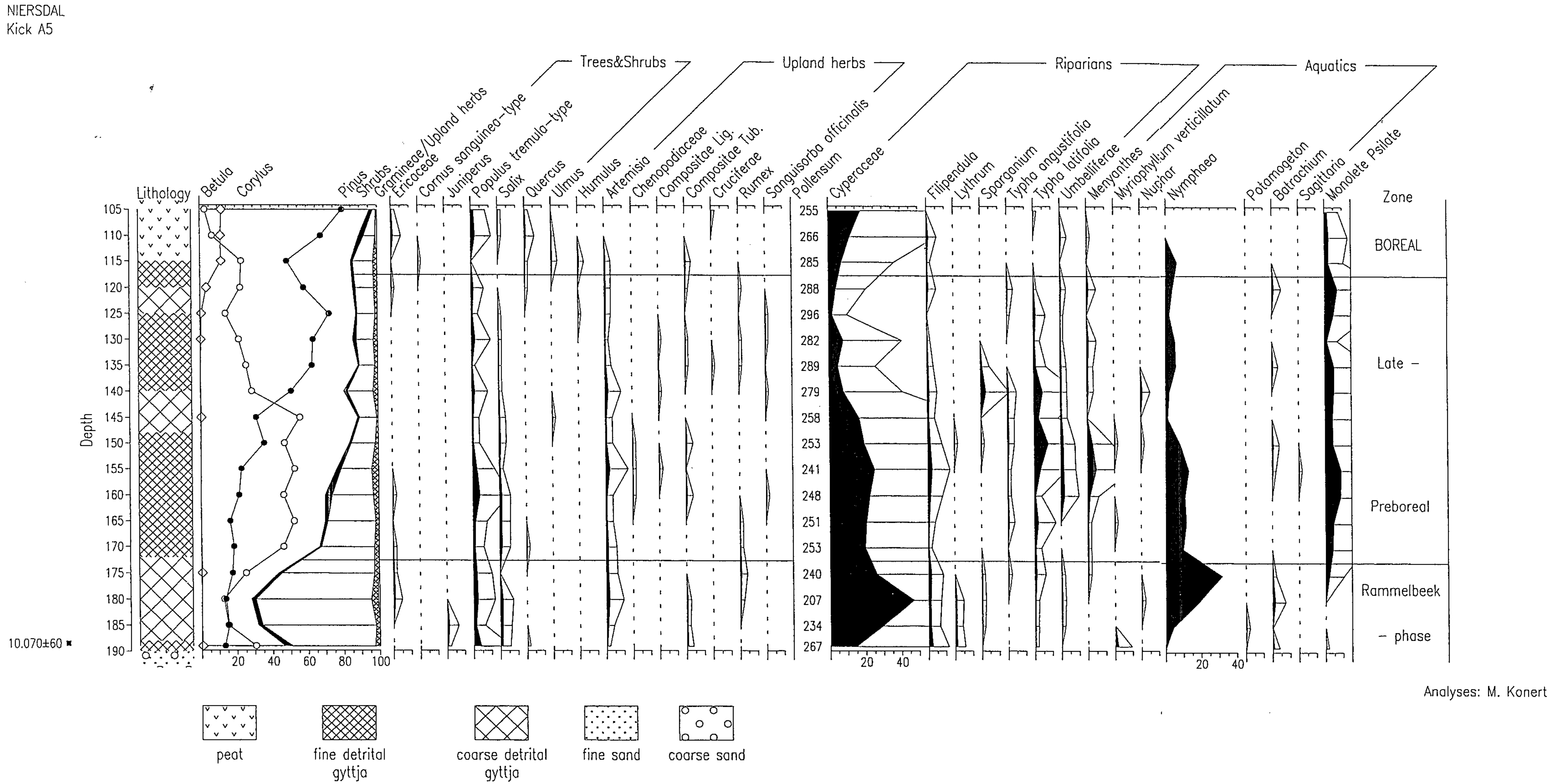


Fig. 9 Pollen diagram Kickerum A.

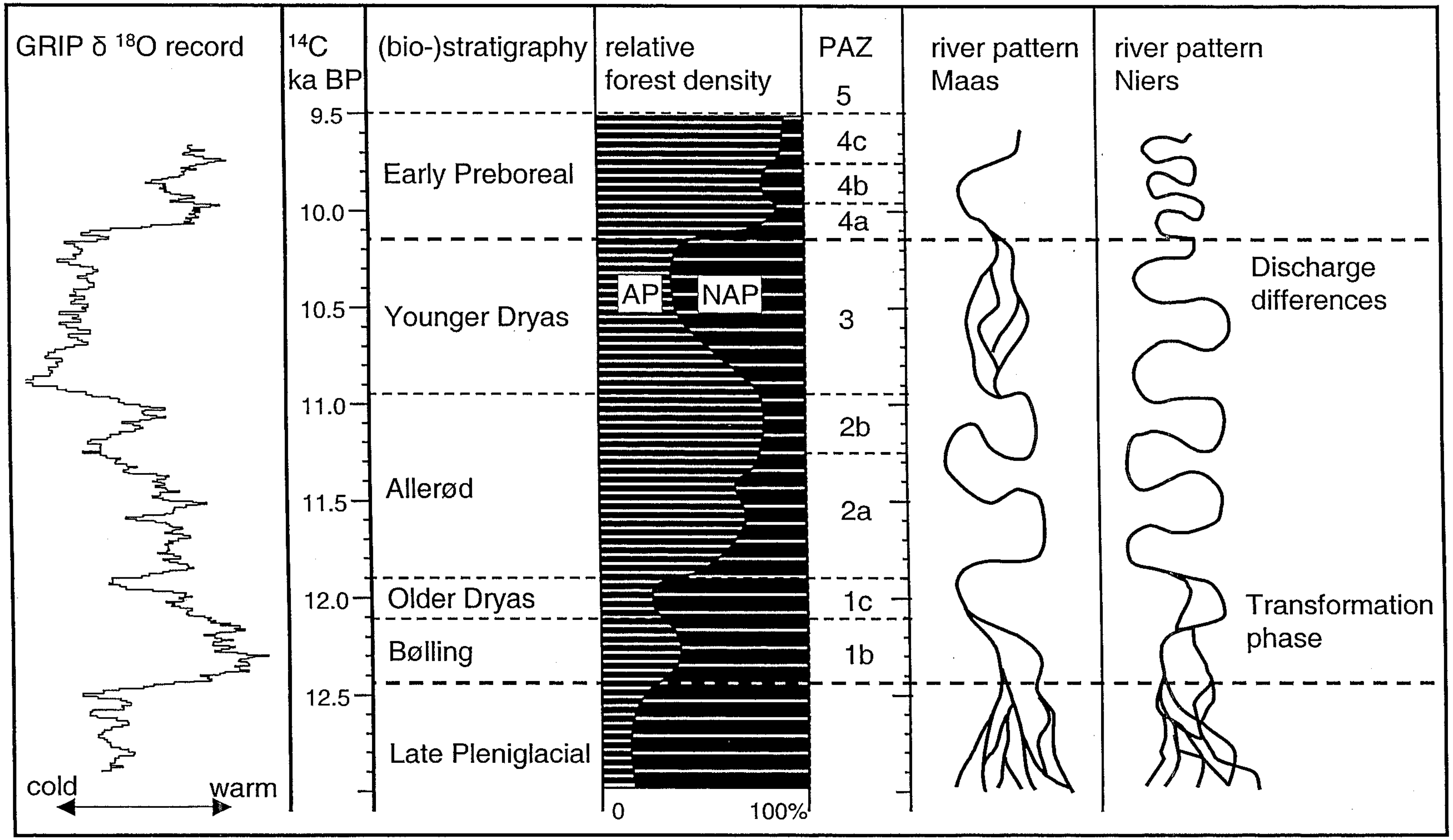


Fig. 10 Synthesis of channel pattern changes in relation to climate and vegetation change.

age BP	Biostratigraphy Netherlands (Hoek, 1997)		Biostratigraphy north-western Germany (Menke, 1985; Behre, 1996)	
9,150				
	5	Late Preboreal	Boreal	
9,500				
	4c		Präboreal c	
9,750				
	4b	Early Preboreal	Präboreal b	
9,950				
	4a		Präboreal a	
10,150				
	3b			
10,550		Late Dryas	Jüngere Dryas	
	3a			
10,950				
	2b		Alleröd	
11,250				
	2a2	Allerød	Mitteler Dryas	
11,500				
	2a1		Bölling	
11,900				
	1c	Earlier Dryas	Älteste Dryas	
12,100				
	1b	Bølling		
12,450			Meiendorf	
	1a	Earliest Dryas		
12,900				Artemisia rise
	LP	Late Pleniglacial	Pleniglazial	

Fig. 11 Comparison between the biostratigraphical subdivision of the Late Glacial and Early Holocene in the Netherlands and northwestern Germany.

Niers-Rhine was a meandering system during the Younger Dryas, in contrast to the nearby Maas river (Kasse et al., 1995; Huisink, 1997).

The results of the geomorphological and palynological study in the Niers valley have been summarized in figure 10. The channel pattern changes of the Niers-Rhine have been compared with the Maas river and correlated with the northwest European vegetation development and Greenland ice-core record.

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

(After Kasse et al, 1995; Huisink, 1997; Van Huissteden & Kasse, 2001)

Introduction

The late Weichselian and Holocene evolution of the Maas (or Meuse) valley in northern Limburg, north of Venlo, is controlled by tectonic and climatic factors. Changes in climate, vegetation and river discharge resulted in changes in the fluvial depositional environment and in terrrace formation.

The Maas river takes its rise at approximately 400 m above sea level in the Mesozoic rocks of the Paris Basin in eastern France. In its course to the north it cuts through the Paleozoic rocks of the Ardennes Massif (up to 700 m high) in Belgium. North of Maastricht it enters the southern North Sea Basin (fig. 1).

The Maas catchment is 33.000 km^2. The mean annual July temperature in the catchment area is between 15 and 18° C. The mean annual temperature in January is approximately between 0 and 2.5° C. 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 Maas is a rain-fed river. The mean annual precipitation amounts 700 to 1000 mm 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).

In the following table the mean, maximum and minimum discharges in m^3/sec are given of the rain-fed Maas at Borgharen (Netherlands-Belgian border) and of the meltwater-fed Rhine at Lobith (Netherlands-German border) (Jongman, 1987).

	Maas	Rhine
period	1911-1960	1901-1975
catchment	33.000 km^2	185.000 km^2
mean summer discharge	130	1850
minimum summer discharge	0	640
maximum summer discharge	1150	7150
mean winter discharge	390	2540
minimum winter discharge	0	620
maximum winter discharge	2800	13000

These figures clearly show the larger difference between the mean winter and mean summer discharge (Q mean winter/Q mean summer) of the Maas in comparison with the Rhine. Furthermore, the fluctuations in discharge of the Maas during the winter (Q max winter/Q mean winter) and especially during the summer (Q max summer/Q mean summer) are larger than those of the Rhine, illustrating the rain-fed character of the Maas.

In Decembre 1993-January 1994 and in January 1995 the Maas valley was flooded by the highest floods ever recorded. At Borgharen discharges of 3120 m^3/sec and 2870 m^3/sec were measured during the two floods. The floods were caused by exceptionally high rainfall in the Ardennes and northern France.

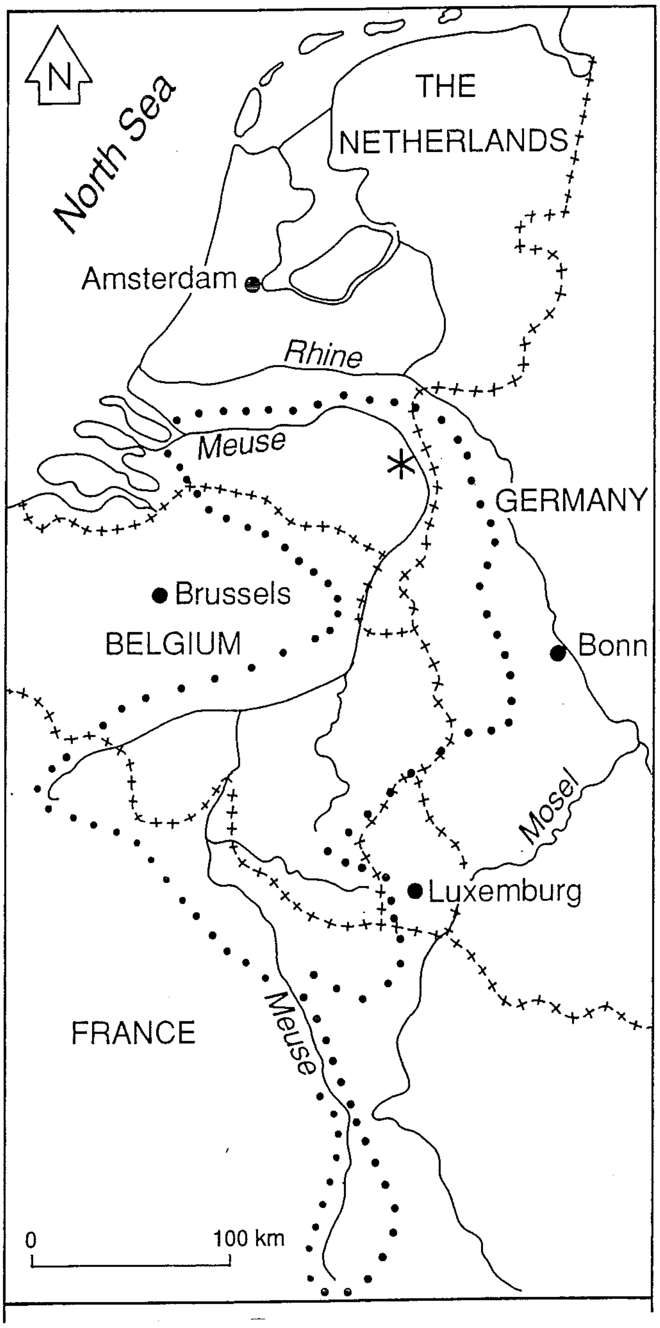

Fig. 1 Catchment area of the Maas (from Kasse et al., 1995)

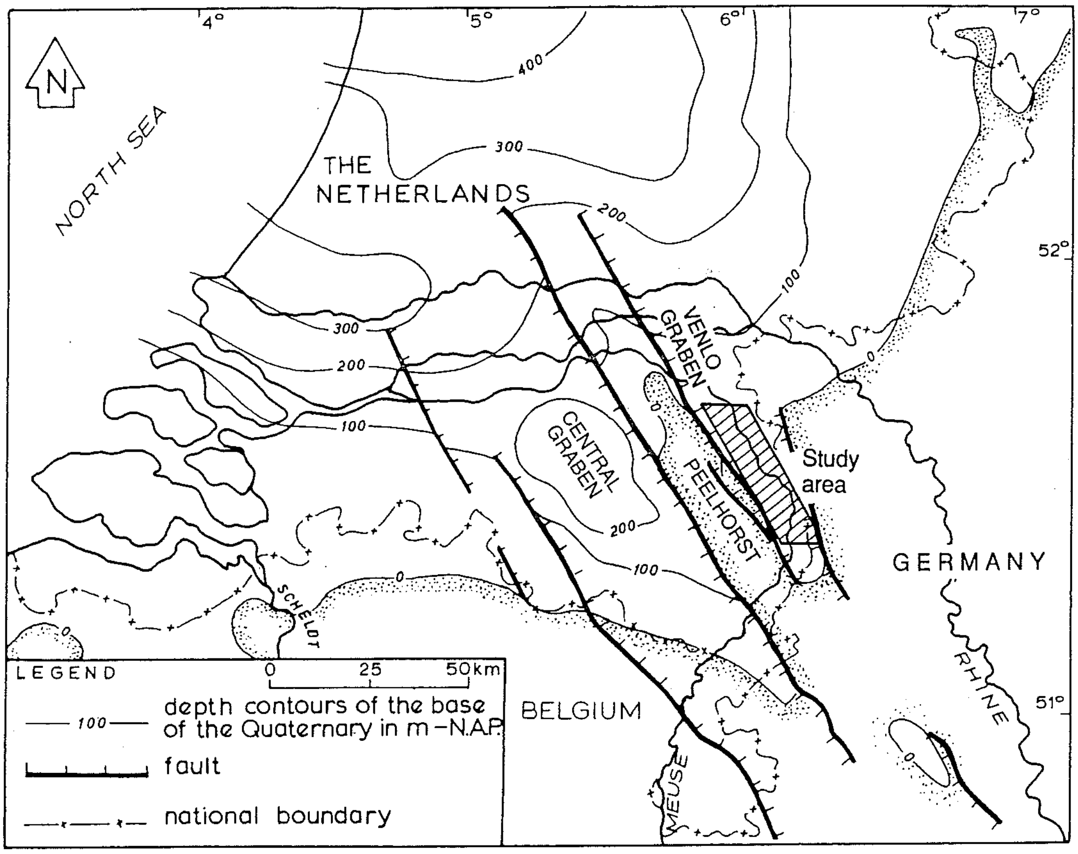

Fig. 2 Major tectonic units of the Southern North Sea basin and depth contours of the base of the Quaternary deposits (after Zagwijn & Doppert, 1978).

Tectonic setting

North of Maastricht the Maas enters the southeastern part of the North Sea Basin, which is characterized by the occurrence of southeast-northwest oriented faults (fig. 2). This fault system, which is the northwestern continuation of the Lower Rhine Graben, forms structural lows (Central Graben and Venlo Graben) and highs (Peel Horst) in the excursion area.

The Grabens contain a thick Quaternary sequence (up to 200 m in the Central Graben), which indicates the continuous 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 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).

At present the Maas crosses the Central Graben and the Peel Horst almost at right angles, before it bends to the northwest in the Venlo Graben. The actual river morphology reflects the tectonic movements. In the Central Graben the Maas has a strongly meandering course with a broad floodplain (Van den Broek & Maarleveld, 1963). On the Peel Horst the Holocene floodplain is nearly absent along the straight, incised course. In the Venlo Graben the present river has a narrow floodplain and a low sinuosity meandering course.

Late Pleniglacial river evolution

During the glacial maximum of the Late Pleniglacial the Maas was a braided river system with large discharge fluctuations (fig. 3: levels 1 and 2; Excursion stop Grubbenvorst sand pit). Syngenetic ice-wedge casts point to permafrost conditions in the floodplain and a mean annual temperature below -8 °C (fig. 4). Aggradation prevailed because of the high sediment supply in the unvegetated landscape. During annual high water stages the braidplain was flooded and medium to coarse sand was deposited by transverse bars. In the last stage of each flooding event small-scale erosive channels were formed. Locally, stagnant pools developed in which silt was draped over the inactive bars. During a next high water stage the channels were filled laterally, with large-scale cross-bedded sand and climbing ripple cross-laminated sand, due to the migration of transverse bars over the floodplain.

Towards the end of the Late Pleniglacial, fluvial deposition became less important probably because of a stronger aridity in combination with a slight temperature increase. Parts of the braided plain became covered with aeolian sands (so-called coversands) and reworked aeolian sands (fluvio-aeolian facies of Grubbenvorst sand pit). The river maintained its braided character and flowed at more or less the same level as during the previous period (fig. 3: level 2).

Late Glacial river development and terrace morphology

Above the recent Holocene floodplain, four distinct Late Glacial to Early Holocene terrace levels have been distinguished along the Maas north of Venlo (fig. 3: levels 3, 4, 5, 6).

The transition of the Pleniglacial to the Late Glacial was accompanied by a strong temperature rise (fig. 4). The previously unvegetated Late Pleniglacial landscape was stabilized by vegetation. As a consequence the channel morphology of the Maas changed from braided into meandering and the river began to incise (Kasse et al., 1995; Vandenberghe et al., 1994; Huisink, 1997) (figs. 4 and 5). In the abandoned braided channels peat formation started during the Bølling, therefore dating the moment of change in the fluvial environment (Bohncke et al., 1993).

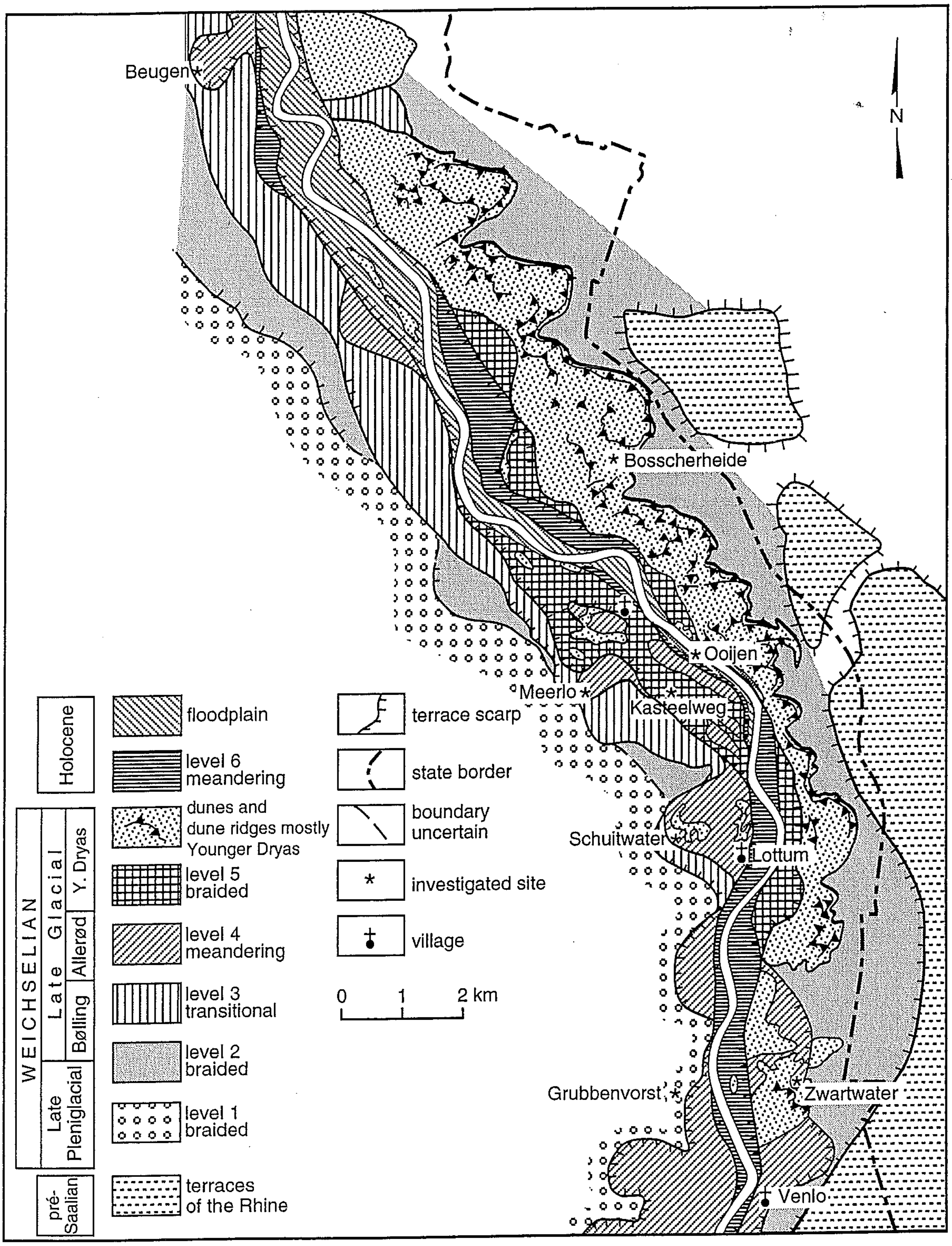

Fig. 3 Morphological map of the Late Pleniglacial, Late Glacial and Holocene terraces of the Maas north of Venlo (from Van Huissteden & Kasse, 2001; after Kasse et al., 1995 and Wolfert & De Lange, 1990).

This alteration from braided into meandering is characterized by a transitional phase with rather shallow (2-3 m), slightly incised, low-sinuosity channels (fig. 3: level 3). This transitional phase probably took place during or after the Bølling, but before the Allerød.

The next younger Late Glacial terrace is typified by large, high-sinuosity meander scars, especially at the outer terrace edges (fig. 3: level 4). This terrace level descends from 21 m above sea level at Venlo to 13 m at Beugen. The fine-grained scar fills are 3 to 5 m thick. The organic fills in the meander scars, dating from the Allerød and Younger Dryas, indicate that this high-sinuosity meandering phase occurred during the Allerød (figs. 4 and 5). Individual pointbars are poorly developed because of an aeolian cover or intense human occupation and cultivation.

The sedimentary sequence of the meander scar terrace is characterized by a thick (7.5 m) fining-upward sequence, formed by lateral migration of the channel and accretion on the meander pointbar (fig. 3: Lottum-Schuitwater). The fining-upward sequence consists of 2 m gravelly, poorly sorted, medium to coarse sand (300-850 μm) at the base overlain by a transitional bed of circa 1 m of moderately sorted, fine to medium sand (150-300 μm). This coarse-grained lower part was formed by strong tractional currents on the meander channel bottom and the lower part of the pointbar slope. The upper 4.5 m of the fining-up sequence are moderately or more often well sorted fine sands (105-210 μm) with thin sandy silt beds (1-13 cm), which increase in number and thickness towards the top. Some smaller fining-up sequences, separated by erosional boundaries, are present within this fine sand unit, probably reflecting reactivation surfaces of the meander inner bend during high discharges. The fine-grained, well sorted upper part was deposited by weaker tractional currents on the upper pointbar slope. The silt beds reflect slack water conditions following high water levels on the upper pointbar slope. These fines have been deposited from suspension by weak currents or in standing water.

The lowest Late Glacial terrace dates from the Younger Dryas (fig. 3: level 5). At the start of the Younger Dryas a strong temperature decline occurred (fig. 4), leading to local permafrost conditions. At Bosscherheide permafrost degradation was dated between 10,880 BP and 10,500 BP (Bohncke et al., 1993). Lower evapotranspiration and larger discharge fluctuations resulted in local flooding of the Late Pleniglacial braidplain at Bosscherheide. The high-sinuosity meandering channels of the Allerød period were abandoned by chute cut offs and a multi-channel to braided floodplain was formed. The higher discharges in combination with restricted sediment supply (because vegetation persisted) resulted in erosion of the floodplain and an up to 4 m high terrace scarp was formed. The Younger Dryas floodplain declines in altitude from c. 18 m near Venlo to c. 10 m near Beugen (fig. 3: level 5). It is characterized by its straight to low sinuous scars (fig. 3: Excursion stop Kasteelweg). Straight scars occur especially along the terrace edge to the higher meander scar terrace. Islands or bars in this floodplain are locally covered by river dune sand, which was blown from the multi-channel plain during periodic low water levels. The lithology of the Younger Dryas terrace level reveals laterally a larger heterogeneity than the higher meander terrace level 4. The palaeochannels are shallow and broad and locally contain up to 2 m of fine-grained infilling in isolated scour pits (Excursion stop Kasteelweg). They are underlain by coarse sand and gravel. The bars in between the channels consist of (gravelly) sand; fining-upward sequences being less pronounced and normally shorter than in the meander scar terrace.

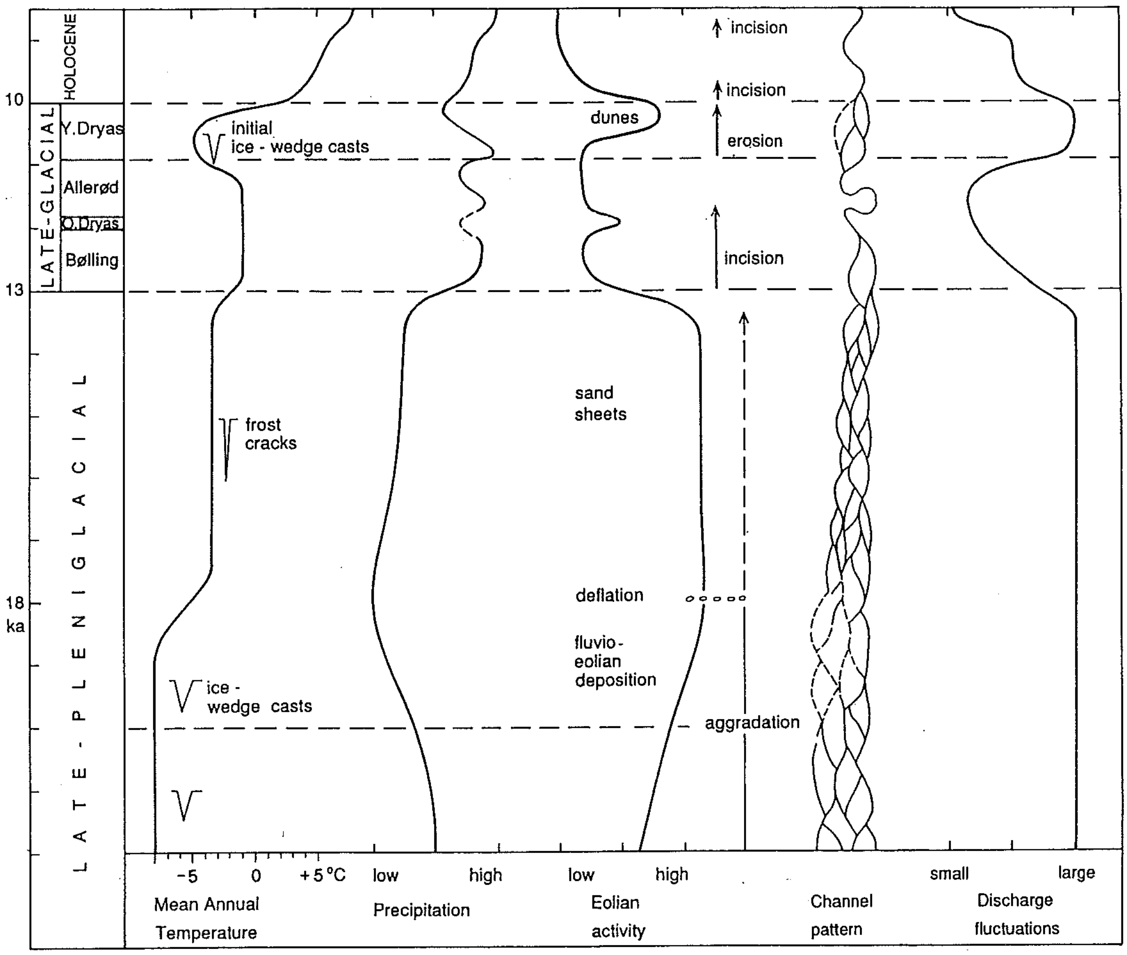

Fig. 4 Summary of the Late Pleniglacial and Late Glacial climatic changes and climate-related fluvial development of the Maas river (from Kasse et al., 1995).

The climatic amelioration and rapid vegetation development at the start of the Holocene led to a larger land surface stability and to a decrease in the discharge fluctuations, resulting in river incision (figs. 4 and 5). The braided channels of Younger Dryas age were abandoned and the river changed into a low-sinuosity meandering system during the Preboreal (fig. 3: level 6). The oldest fill in the incised channels dates from the Boreal (Excursion stop Kasteelweg-Ooijen), which indicates that the incision phase occurred between the Younger Dryas and the Boreal, i.e. the Preboreal. The Preboreal floodplain reveals straight and low-sinuosity scars, more or less conform to the actual low-sinuosity river course. Due to the lateral migration of the meander belt, channel side bars or large-scale pointbars developed. This clear fluvial morphology indicates that the aeolian activity, typical of the late Younger Dryas, had ceased.

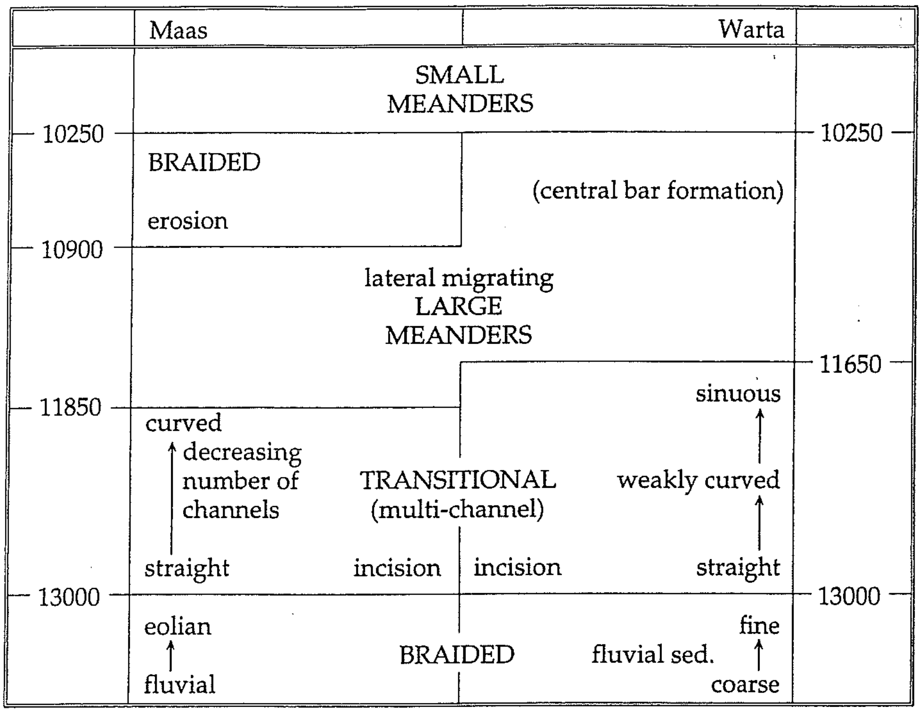

Fig. 5 Comparison of the Late Pleniglacial and Late Glacial climate-related fluvial developments of the Maas (the Netherlands) and Warta (Poland) rivers (from Vandenberghe et al., 1994).

Late Pleniglacial and Late Glacial aeolian activity

During the very end of the Late Pleniglacial aeolian deposition was the dominant process over large areas in the Netherlands. This widespread aeolian deposition can be attributed to aridity due to the absence of permafrost and/or lower precipitation values. In the Maas valley Late Pleniglacial fluvio-aeolian and aeolian deposits have been described by Mol et al. (1993) (see Excursion stop Grubbenvorst).

Late Glacial aeolian deposits are also very common in the Maas valley. Aeolian sediments dating from the Allerød and/or Older Dryas period are only locally present on the higher meander terrace level (fig. 3: level 4, Lottum-Schuitwater). At this site the meander terrace is covered by an up to 4 m thick unit of aeolian sediments. Since there is no evidence for an aeolian supply from outside the area, nor for a supply from the subsoil, we conclude that these aeolian sediments on top of the meander terrace were formed during the lateral migration of the meandering river channel. The comparable fine sandy grain-size of the aeolian deposits and the pointbar deposits support this hypothesis. The following mechanism has to be considered. During bankfull discharge, probably in spring due to melting of the snow cover, sediment was deposited on the meander upper pointbar. During the following low discharge this barren sediment on the upper slope could be deflated by the prevailing westerly winds (Maarleveld, 1960; Schwan, 1988) and was deposited on top of the point bar. The low dune morphology indicates aeolian deposition in a vegetated landscape. The start of the meander scar infill at Lottum-Schuitwater was dated palynologically in the Younger Dryas. This means that the lateral channel migration and the connected deflation and dune accumulation took place just before the Younger Dryas, i.e. the Allerød and/or Older Dryas period.

The younger, far more extensive, Late Glacial aeolian deposits date from the Younger Dryas. On the east side of the Maas valley widespread river dune complexes occur, lying on the Late Glacial meander scar terrace or on older fluvial deposits (fig. 3: Bosscherheide). The dune morphology is characterized by parabolic forms in the eastern (downwind) part of the dune field. Because of this morphology a westsouthwestern sand-transporting wind is inferred for the Younger Dryas period. It is stressed here that this wind direction is the dominant sand-transporting wind, probably generated by the passage of low-pressure systems with associated high wind velocities. The dominant annual wind direction may have been different. At the base of the dune sediments Late Glacial organic sediments and peats are present, which offer the opportunity to date the start of the aeolian deposition. From the top of the peaty layer, characterized by an alternation of moss-laminae and aeolian sand-laminae, a ^{14}C date of 10,500 ± 60 BP was obtained (Bohncke et al., 1993), which places the overlying dune body in the late Younger Dryas period.

The aeolian sediment of this phase is fine to medium grained. Because of the prevailing westsouthwesterly winds, the source area must have been the Younger Dryas floodplain west of the dune field. It is proposed that the aeolian sediment has been blown from the Younger Dryas floodplain during periodic low water levels. Only the sediments that accumulated outside the floodplain have been preserved. The aeolian sediment within the floodplain had a low preservation potential because it was easily eroded during subsequent high water floods.

The accumulation of large amounts of sand in the extensive dune field on the east bank of the Maas could only occur due to continuous sediment supply from upstream and/or fluvial reworking from the braided floodplain subsoil followed by deflation. These alternating processes of supply/reworking and deflation indicate that during the late Younger Dryas the Maas had a more intermittent character than during the previous Late Glacial periods. This is in agreement with the morphological evidence from the floodplain, which shows a braided river system during this period. The larger fluctuations in discharge are attributed to the climatic deterioration at the start of the Younger Dryas period. The larger volumes of snow melt water will have led to larger peak discharges (nival discharge regime). As a result the Late Glacial meandering river course changed into the braided Younger Dryas course.

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EXCURSION STOP 4: GRUBBENVORST

- Late Pleniglacial fluvial to aeolian facies
- Late Pleniglacial cryogenic stuctures (ice-wedge casts, deformations, vertical platy structures
- Late Pleniglacial and Late Glacial aeolian sediments

Stratigraphy

The Grubbenvorst sandpit is situated on a Maas terrace that is overlain by a sheet of aeolian coversand (Fig. 1). The stratigraphical position of the exposed section (see Figs. 2 and 3) can be deduced from the following data:

1. The occurrence, some four km northwest of the exposure, of a subsurface peat layer that has been dated at GrN-16949 : 26,130 ± 180 BP at its base and at GrN-16948 : 25,200 ± 180 BP in its top (Westerhoff & Broertjes, 1990). This bed is presumed to be time-correlatable with a level found at a shallow depth below the floor of the Grubbenvorst sandpit.

2. The presence, in the top of unit B3 of the Beuningen Complex. At the site under consideration, this important marker bed consists of multiple deflation levels which formed during a protracted phase of intense drought. Following Kolstrup (1980) the development of this feature took place, roughly, between 15.5 and 14 ky BP.

3. The presence of a strongly iron-stained level in the upper part of the exposure. This feature supposedly represents a palaeosol of Alleröd Interstadial age and as such should have formed in the timespan from 11.8 to 11 ky BP.

On the basis of the above evidence, a Late Weichselian Pleniglacial to Weichselian Late Glacial age is attributed to the section under consideration. Recently, OSL-dating has provided absolute ages for the fluvial to aeolian sequence ranging between 28 and 13 ka (Vandenberghe, 2004). The OSL-ages fully confirm the previous chronostratigraphic age estimates.

Fluvial to aeolian sedimentary facies

The Grubbenvorst section is a fine specimen of a fluvial to aeolian sequence, a type of succession that is wide spread in the Pleistocene lowlands of northwestern Europe (Schwan, 1987). In the exposure a gradual transition from a purely fluvial sediment at the base to a purely aeolian one in the top of the section is present (Figs. 2 and 3).

Unit A

In the fluvial sands of unit A, in the basal part of the section, large-size channels are the prevailing structural characteristic. The channels occur in both intersecting as well as solitary positions, mostly have a concordant filling and are incised in horizontally bedded sands. In upward direction, the channel fills give way to generally thinner sets which, on average, are also finer-grained, so that the whole of unit A represents a sequence with fining-upward and thinning-upward tendency.

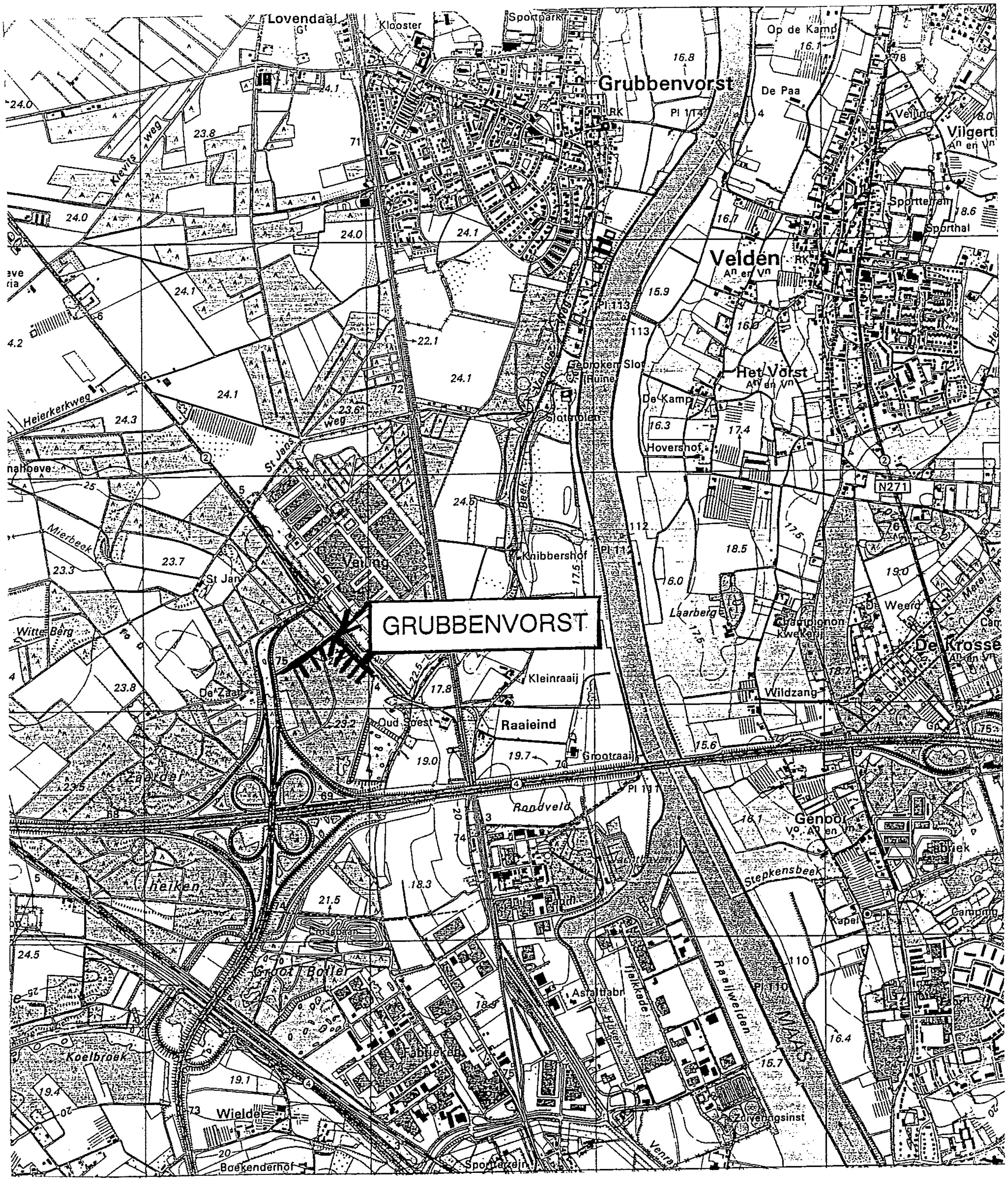
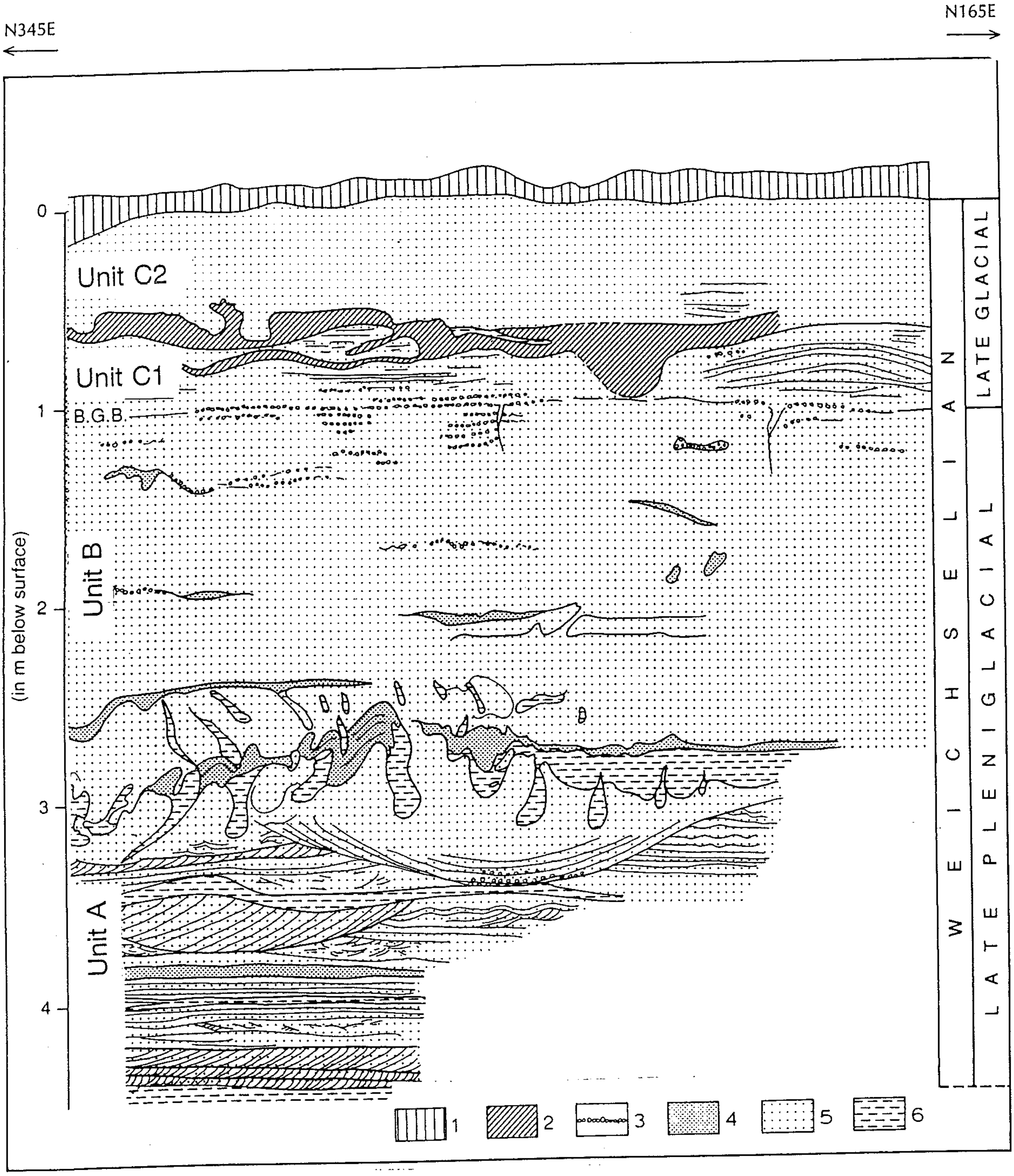


Fig. 1 Location map of the Grubbenvorst sand pit.



1, disturbed; 2, firm horizon; 3, gravel; 4, coarse sand; 5, very fine to medium sand; 6, silt; B.G.B., Beuningen Gravel Bed.

Fig. 2 Sedimentary succession in Grubbenvorst (from Mol et al., 1993).

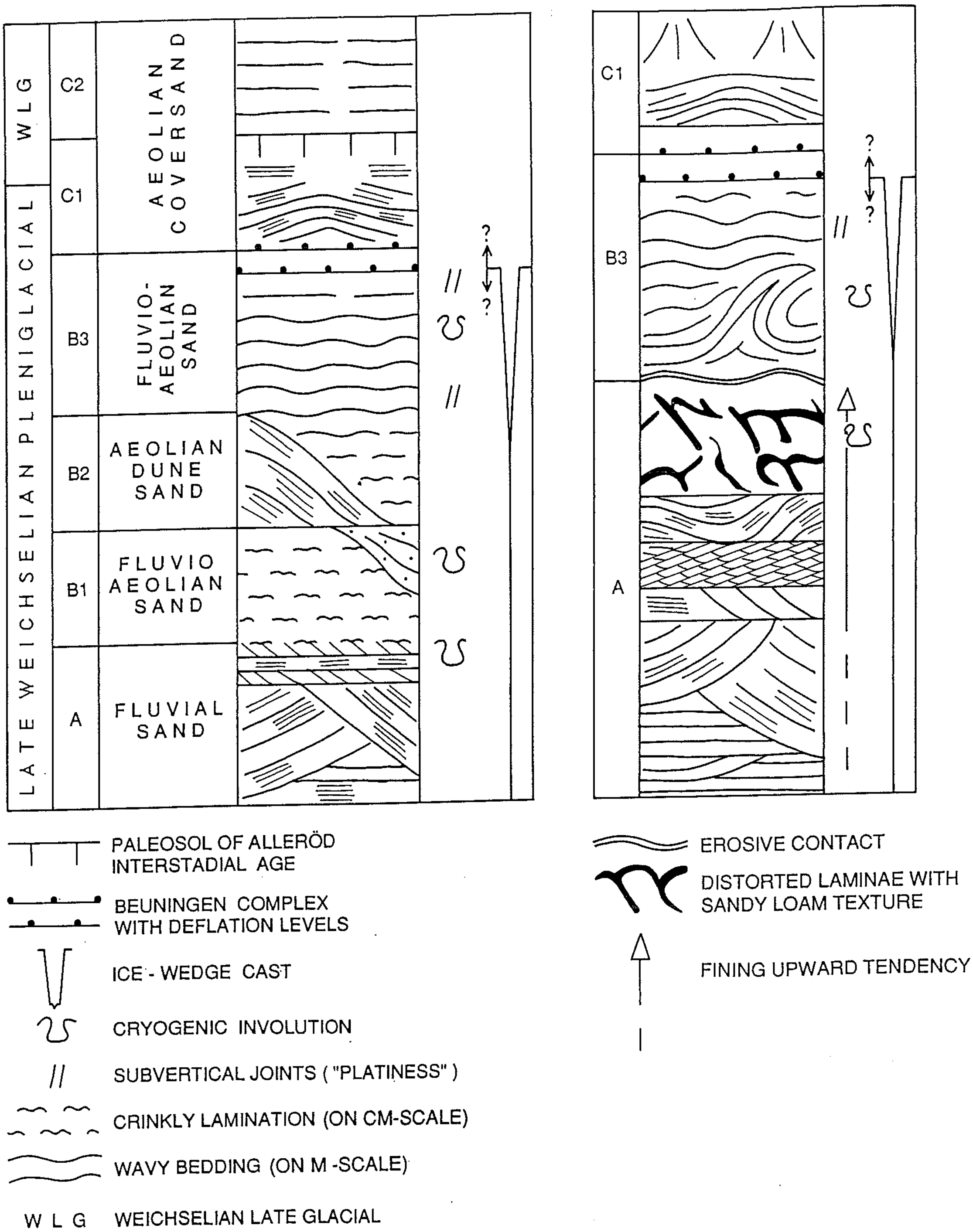


Fig. 3 Schematic sedimentary successions in sand pit Grubbenvorst (from Schwan & Vandenberghe, 1991).

From borehole-observations in the vicinity of the exposure, it is known that a succession of three such sequences is present in the deeper subsoil and extends to a depth of approximately 7.5 m below the floor of the sandpit.

Occasionally, within unit A, channels are found whose filling, rather than being concordant, is lateral; the attitude of the foresets suggests that they were formed by a stream flow which was roughly perpendicular to the axis of the host channel.

Besides channel sediments, unit A contains horizontally bedded units with large-scale tabular cross-bedding, which are attributed to deposition by straight-crested bars on the floodplain (fig. 4). The dip of the foresets indicates a stream direction towards the northwest or north. The silt drapings on the bar foresets reflect periods of low current velocities or standing water conditions following high current velocities during flood.

The unit-A deposits are believed to have formed by an aggrading braidplain. The large-scale channel fills are suggestive of episodic peak discharges with erosive stream power, - seasonal or otherwise, - followed by a phase of filling when the flow strength had diminished. The wide lateral spread of the channel occurrences is most likely to be a result of shifting behaviour of the ancient Maas. Because of the threefold succession alluded to above, some kind of cyclicity may have been involved in the buildup of the braidplain.

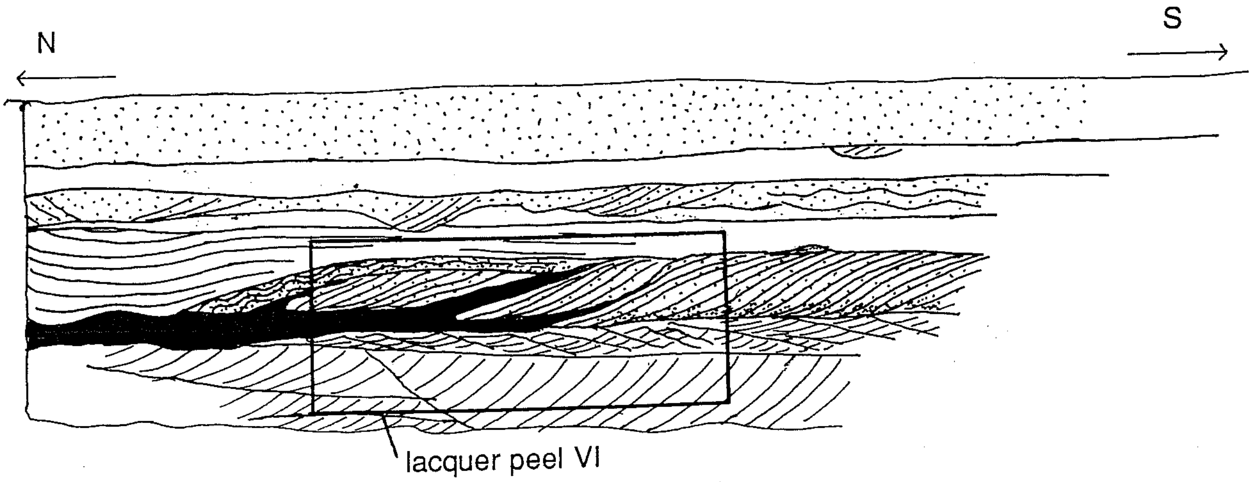


Fig. 4 Silt drapings on the foresets of large-scale transverse bars at the base of fluvial unit A. They illustrate the recurrent alternation of sand deposition during peak discharges and silt deposition during standing water conditions on the braided floodplain.

Units B1 to B3

Whereas units B1 and B3 correspond to the fluvio-aeolian type, unit B2 is a deposit of small dunes having a purely aeolian origin (Fig. 3). As far as known, the occurrence of unit B2 is restricted to a single stretch less than 10 m in length which, by now, has fallen prey to excavation. There, the unit had a thickness of 1.5 m only and, by attitude-readings on the dune-foresets, deposition from a west-northwestern direction could be deduced. Wherever unit B2 is absent, the upward transition from unit B1 to B3 is gradational.

Unit B1, the lower fluvio-aeolian unit, has an overall granular composition of fine sand with a few thin intercalations of either silty or coarser-than-average sandy texture; layers of the latter type are lense-shaped with a concave lower boundary and a flat upper one. Whereas finely-laminated parallel bedding is the prevailing stratification type, low-angle cross-lamination, ripple-foreset cross-lamination and small scour-fill structures are subordinate characteristics that occur regularly. Normal grading from (fine) sand to a more silty texture is frequently found in one to two cm thick sets. Almost certainly as a result of cryogenic stress, all these features have a generally crinkled appearance when viewed on a cm-scale. Moreover, one to three cm thick micro-loadcasted levels add to the impression of small-scale distortion.

Channel fills laterally merging with the parallel-laminated sands occur here and there in unit B1; their scarcity testifies to the dwindling importance of fluvial activity.

Unit B3, the upper fluvio-aeolian unit, has a make-up quite similar to that of B1, except for the absence altogether of fluvial channel-fills. In unit B3, laminae are organised in rather thick sets separated by indistinct bedding planes. Owing to this, the waviness of the beds on a m-scale stands out clearly. Like the crinkliness discussed before, this feature must be attributed to cryogenic stress with the difference between the two being merely a matter of size.

Units B1 and B3 are thought to be a result of fluvial aeolian interaction as, in the subject sediment, planebedded layers of generally accepted aeolian origin (e.g. Schwan, 1988) occur side by side with features that strongly suggest aquatic conditions with or without current flow. In elaboration of the above it is assumed that (i). Buildup of the fluvio-aeolian sediment takes place on an abandoned braidplain which is flooded, now and then, by the river and (ii). Deposition of the blown-sand component proceeds, mainly, in sheet-like fashion, so that the former braidplain will retain its overall horizontality whilst it aggrades. When these requirements are met, the following processes may become operant:

1. Aeolian deposition of a thin sheet of sand on a dry or damp land surface,
2. Sand grains driven by wind or fines falling out from the air which come to rest in a temporary and shallow pool of still water,
3. Reworking of previously deposited material by high-energy sheet flows during a flooding event.

Clay-rich mud drapes that often characterise sandy overbanks subject to flooding (e.g. Langford, 1989) are rare in the sediment under consideration. Possibly, this is due to a primary lack in very fine particles in the suspension load of the river.

Locally, units B1 and B3 are separated by an aeolian dune facies. As already told, the dunes are low and insignificant in areal extent. This data, therefore, does not contradict the assumption of aeolian deposition mainly in the form of low-angle sand sheets.

Unit C

Unit B3 is capped by, successively, the Beuningen Complex and a packet of aeolian coversands (unit C) (Figs. 2 and 3). Whereas the former type results, mainly, from deflation, the second one is a characteristic product of aeolian deposition. Unit C1 is characterized by horizontal lamination and low-angle cross-bedding and is assumed to be deposited in a sand sheet of low dune environment. Unit C2 is separated from Unit C1 by a reddish brown sandy silt, which on stratigraphical grounds may be equivalent with the Usselo soil of Allerød age. The soil represents a phase of local surface stability. Locally, the transition between the sand units C1 and C2 is not a single bed of sandy silt, but rather an interval with alternating bedding of fine sand and silt. Rather than a period of complete absence of aeolian deposition and associated soil formation, this alternating bedding suggests a period of decreased aeolian deposition.

Cryogenic deformation structures

In the Grubbenvorst exposure, four different types of cryogenic deformation features are observable. These are:
* Casts of epigenetic and syngenetic ice wedges (see Fig. 5)
* Cryoturbation structures (= periglacial involutions) (see Fig. 2)
* Warping and crinkling of originally evenly bedded strata.
In the foregoing, the corresponding deformations were termed "waviness" and "crinkliness" respectively. Though the latter category is attributed to cryogenic stress in the first place, it cannot be excluded that microloading of water-saturated sediment is another agent capable of producing crinkly structures.
* Sets of subvertical joints developed in an en-echelon pattern (see Fig. 6).
In literature, the sets of subvertical joints are sometimes referred to as "platiness". Mol et al. (1993) attributed the development of these vertical platy structures to intense cooling and cracking of the surface in periglacial environments. Stratigraphically, the vertical platy structures are most common in Weichselian Upper Pleniglacial beds below the Beuningen Gravel Bed. They generally occur in association with ice-wedge casts and therefore are assumed to have been generated under the severe cold conditions of the Last Glacial Maximum.

With respect to the above it is noted that:
1). The degree of development of anyone of the cryogenic features is laterally variable. This is to say that, within one and the same unit, a phenomenon like cryoturbation may be well expressed in one place and lack in another one. Granular composition and height above the groundwater table are two often quoted parameters to account for this variability.
2). Periglacial deformation structures occur at several different levels in the exposure. Their morphology and stratigraphic position is as follows:
(i). Three levels of involutions with amplitudes of respectively 40, 50 and 80 cm in the top of unit A,
(ii). Strongly cryoturbated channel fills with silty texture in unit B1,
(iii). Waviness locally changing into involuted structures in unit B3,
(iv). Ice-wedge casts departing from both the top of unit A as well as from unit B; in the lower part of the exposure, some of the casts have been truncated by fluvial channels.
3). The strata directly overlying the youngest ice-wedge casts are entirely free of large-scale distortion, except for their wavy and crinkly appearance. Normally, melting of ice wedges is accompanied by involution of the waterlogged active layer. Absence of the resultant structures suggests either removal by erosion of the affected levels or inactivity of the periglacial-loadcasting process when the permafrost degraded. The second alternation is thought to apply to the

Grubbenvorst site.

4). The strong development of the cryogenic features in units A and B of the Grubbenvorst section is in full agreement with the chronostratigraphic age attributed to them (see section Stratigraphy). Following various sources (e.g. Vandenberghe, 1985) the early part of the Late Weichselian Pleniglacial in northwestern Europe was a period of continuous permafrost corresponding to the second cold maximum of the Last Ice Age.

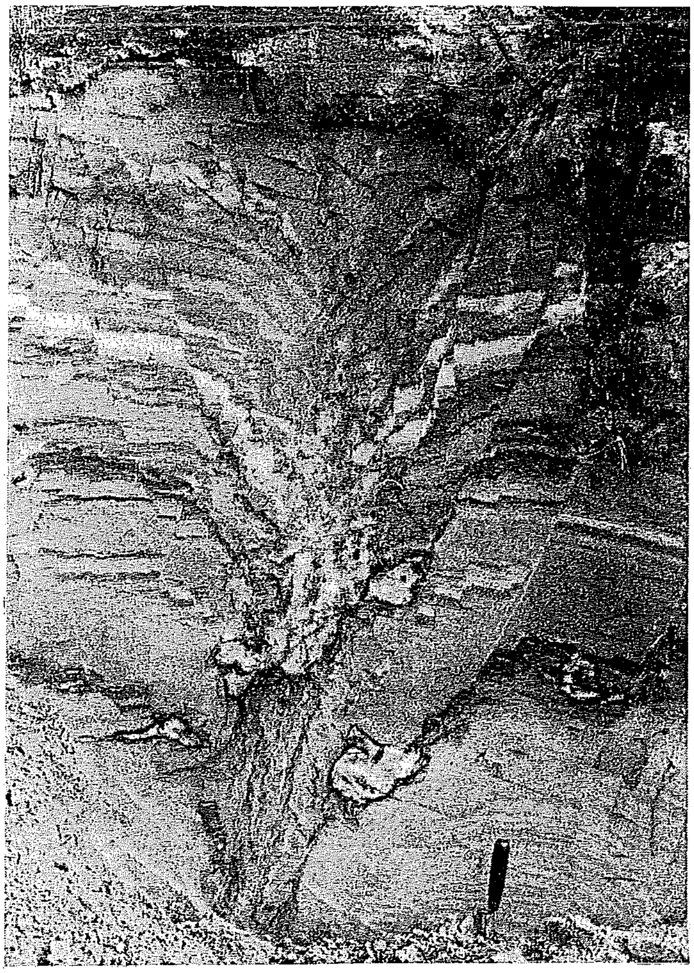


Fig. 5 Ice-wedge cast in fluvial unit A at Grubbenvorst demonstrating the presence of continuous permafrost during the Weichselian Late Pleniglacial (trowel is 25 cm) (from Mol et al., 1993).

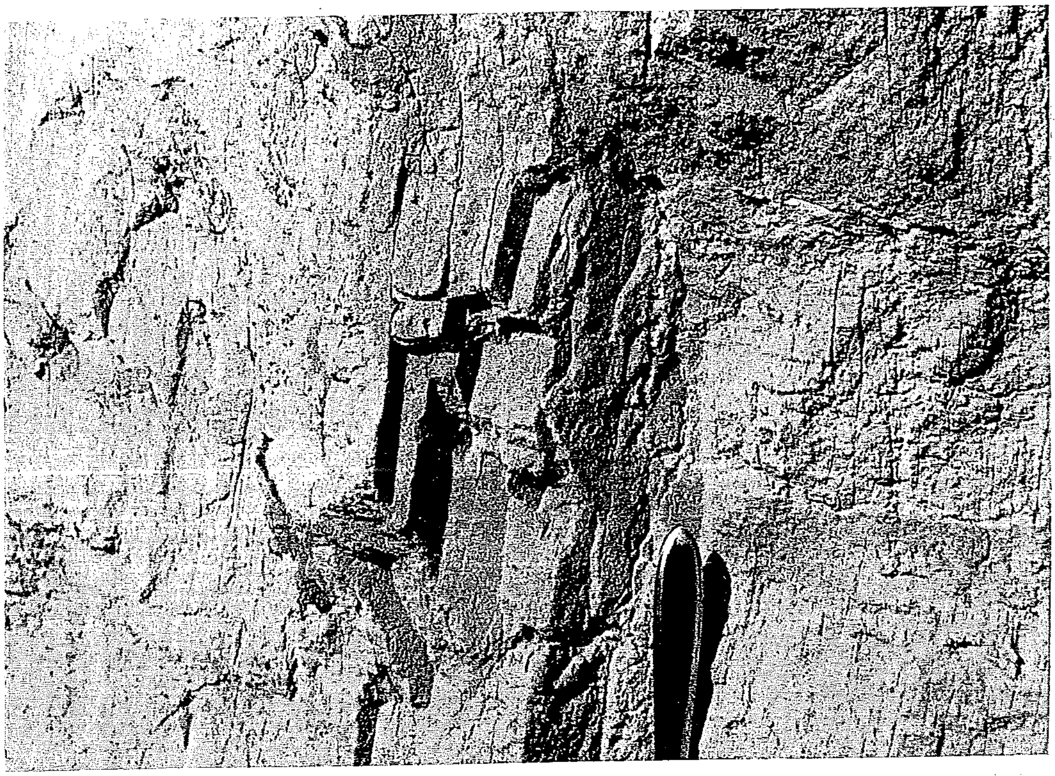


Fig. 6 Vertical platy structures in the fluvio-aeolian unit B at Grubbenvorst showing different directions of the microjoints (knife is c. 25 cm) (from Mol et al., 1993).

Palaeo-environment

In the period from 26 ky BP to approximately 18 ky BP, conditions of continuous permafrost prevailed at the exposure site and its wide surroundings. Concurrently, the depositional regime changed from a fluvial one to a fluvio-aeolian one, possibly caused by stronger aridity towards the end of this period.

In the time span from 18 ky to c. 14 ky BP, degradation of the permafrost was followed by a hyperarid phase with predominance of wind activity; in that interval, the desert pavement of the Beuningen Complex came into being. A subsequent amelioration of climate resulted in the deposition of windblown sand (Kolstrup, 1980). Possibly during the Allerød Interstadial of the Weichselian Late Glacial, the aeolian accumulation process slowed down or was interrupted by pedogenesis.

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EXCURSION STOP 5: YOUNGER DRYAS AND HOLOCENE FLOODPLAIN AT KASTEELWEG-OOIJEN

- Younger Dryas braided floodplain morphology
- Late Younger Dryas and Preboreal fill of channels
- Younger Dryas dune complex
- Early Holocene incision and terrace scarp

Introduction

At this excursion point a 1.5 km walk will be made, in which we cross the Younger Dryas and Holocene floodplains. Kasteelweg-Ooijen is located on the western bank of the Maas. The altitude is between 14 (channels) and 20 meters (river dunes) above sea level (Fig. 1).

In this region the Younger Dryas cooling of the climate has been registered in the following manners (Fig. 2) (Kasse, 1995a, 1995b):

1. Change in channel pattern. The Younger Dryas floodplain is generally 1 km wide and characterized by a multi-channel (braided) floodplain with straight to low-sinuosity channels (Fig. 3). The change in channel morphology from high sinuous to braided probably occurred during the start of the Younger Dryas, because the temperature decrease (especially in winter) led to a longer snow accumulation period and therefore higher peak discharges in spring or early summer (nival discharge regime). The morphological change was accompanied by floodplain-wide erosion of the high-sinuosity meander terrace (see Fig. 4).

2. Chute cut-off of meanders. The rather straight Younger Dryas floodplain was formed by several chute cut-offs of high-sinuosity Allerød meanders (Fig. 4). This is concluded for instance from a nearby meander (Meerlo, 2.5 km west of this excursion point) which was abandoned and filled with loam during the Younger Dryas. The formation of chutes over previous pointbar surfaces is also related to increased nival peak discharges.

3. Flooding of older terraces. Because of the higher peak discharges the Maas was able to inundate areas that were previously above the flood limit, for instance higher lying Late Glacial or Late Pleniglacial terraces. At Bosscherheide a fluvial loam was deposited over the Allerød soil/peat and the Younger Dryas involutions (Kasse, 1995a).

4. Lithological change in abandoned channels. Because of the higher peak discharges, the Maas was able to inundate abandoned meander channels of the Allerød period. At Beugen a gyttja of Allerød age at the base of the infill is erosively overlain by a gray loam. This lithological break correlates pollenanalytically with the start of the Younger Dryas.

5. Widespread aeolian deposition. On the east bank of the Maas valley an up to 4 km wide sand-sheet and dune belt is present, overlying older Late Glacial or late Pleniglacial deposits. The parabolic dunes indicate a westsouthwesterly wind during their formation in the late Younger Dryas (Kasse, 1995b). At Bosscherheide the aeolian sediments overlie a peat layer dated at 10,500 ± 60 BP. The reason for this large-scale aeolian deposition is related with the change in channel morphology at the Allerød-Younger Dryas boundary. The single-channel meandering system was at that moment replaced by a much broader braided system. During low discharge, sediment was deflated by the westerly wind from the wide south-north oriented Maas floodplain.

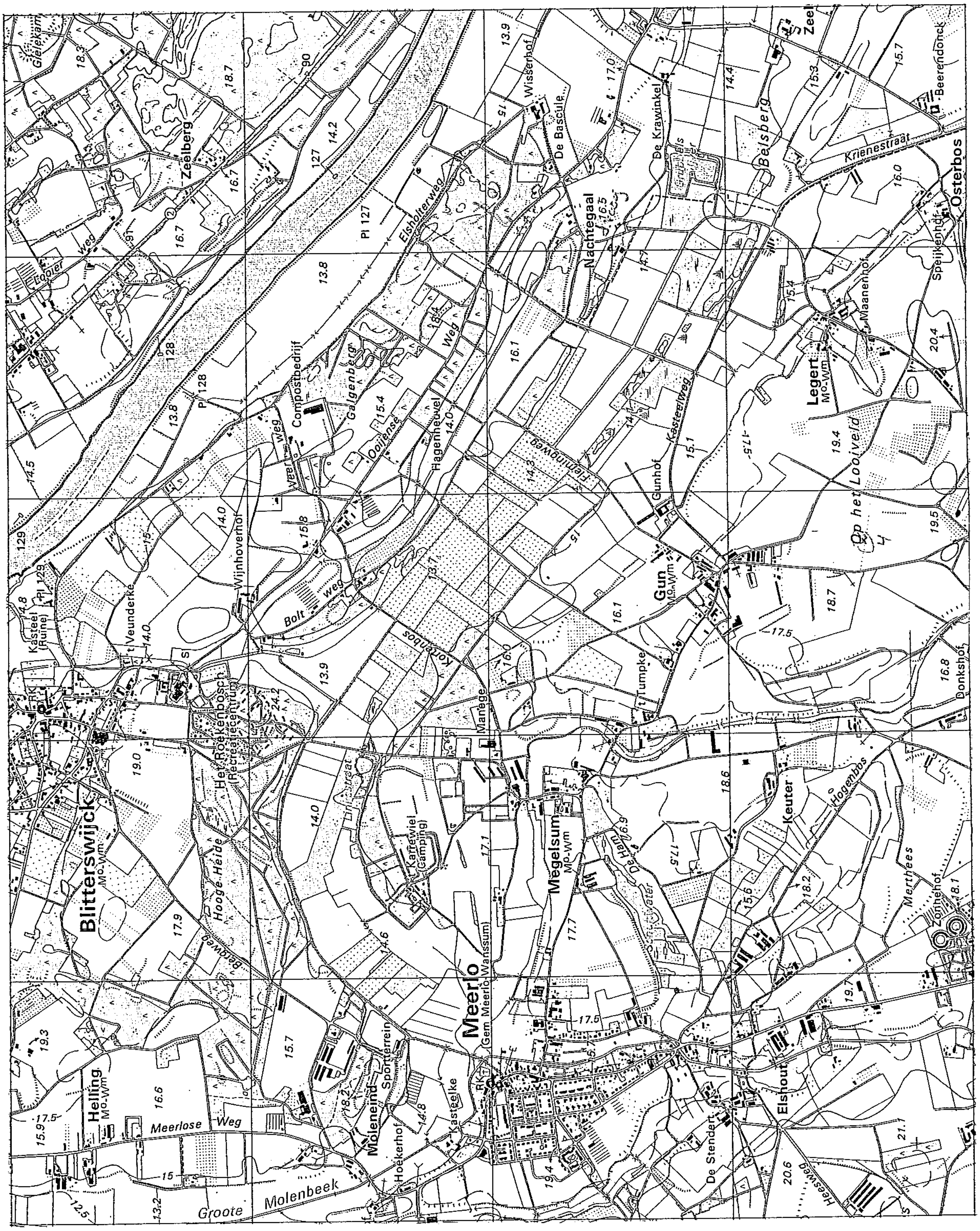

Fig. 1 Location of the Kasteelweg - Ooijen excursion stop.

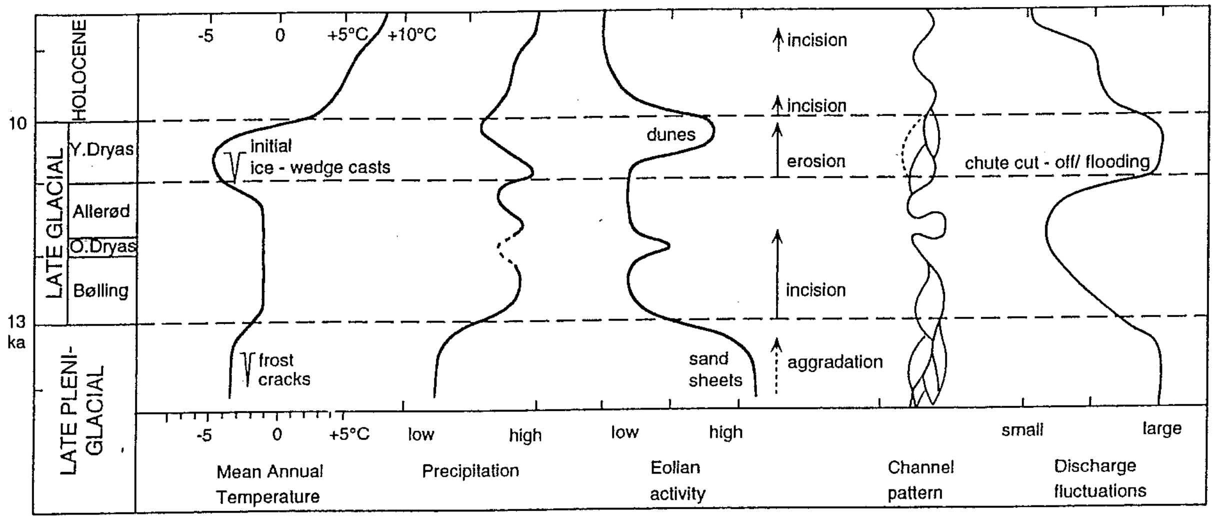

Fig. 2 Summary of the Late Glacial climatic and environmental changes in the Maas valley (from Kasse, 1995a).

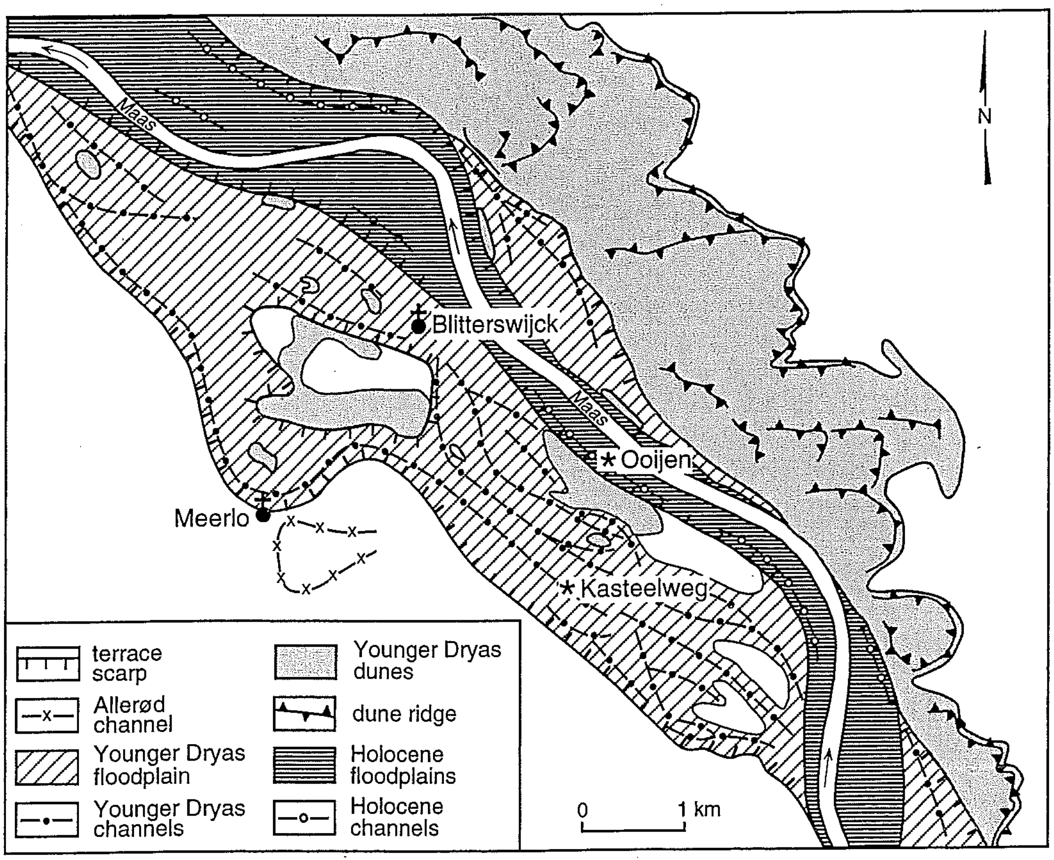

Fig. 3 Morphological map of the Younger Dryas braided floodplain near Blitterswijck and Younger Dryas dune field east of the Maas (from Kasse, 1995a).

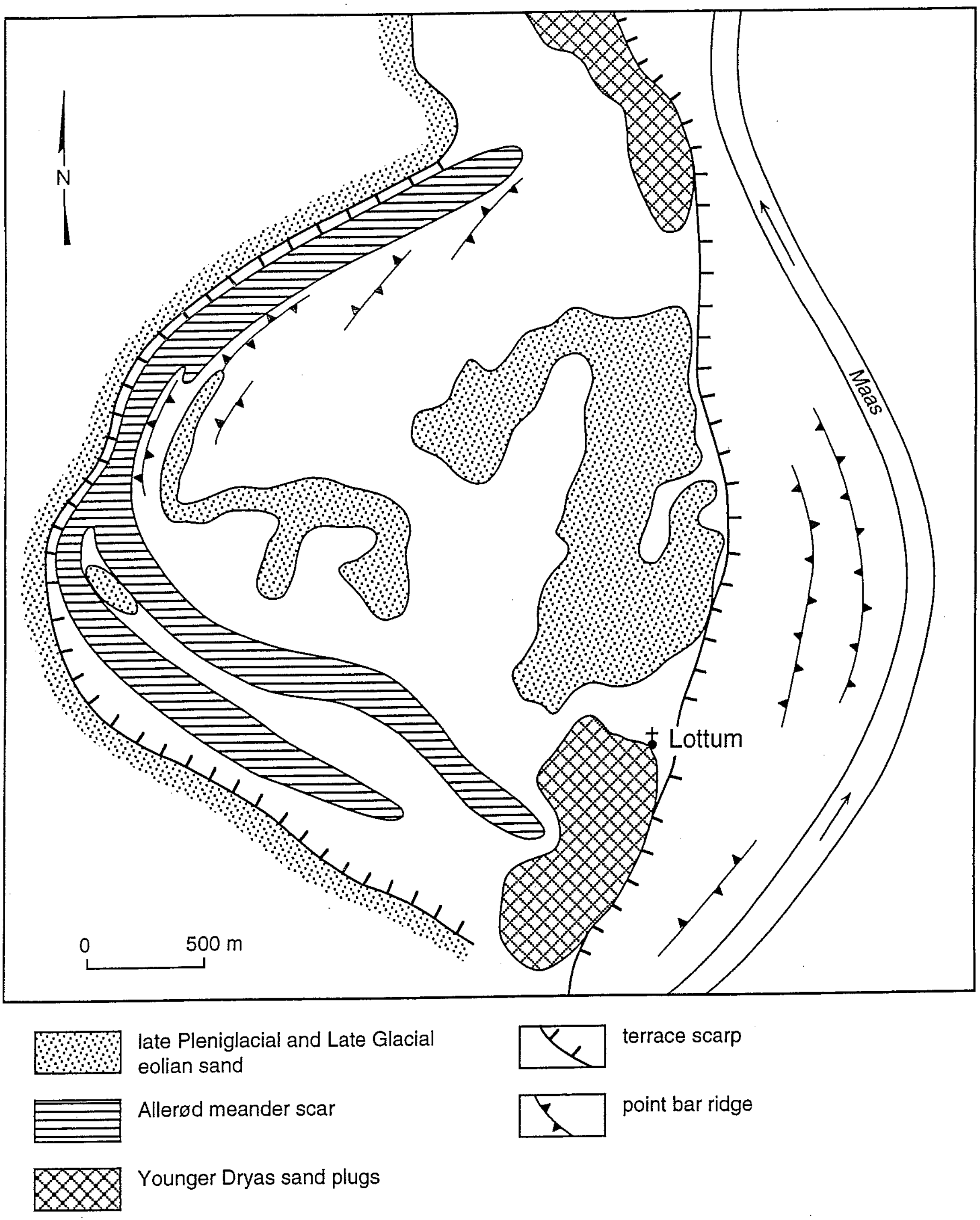

Fig. 4 Younger Dryas chute cut-off of the Allerød meander near Lottum (from Kasse, 1995a).

Geomorphology and fluvial development

A generalised cross section of the Kasteelweg-Ooijen area is given in figure 5. Four physiographic areas/units are distinguised:

1. The high-sinuosity meander scar terrace in the west. It consists of gravelly sands fining upward into sand. It dates from the early Late Glacial and Allerød period.
2. The Younger Dryas channel system, underlain by gravelly sands. Local organic channel fills dates from the late Younger Dryas (Westerhoff & Broertjes, 1990: pollen diagram Blitterswijck Linkstraat; Fig. 6: pollen diagram Kasteelweg).
3. The Younger Dryas channel and bar system, partly covered by river dunes, and resting on erosional remnants of the meander scar terrace.
4. The Early Holocene floodplain with channels filled with coarse-detrital gyttja, dating from the late Boreal-early Atlantic (Figs. 7 and 8: pollen diagrams Ooijen).

The first stop is in a 500 m broad, but shallow, Younger Dryas channel (Fig. 5: Kasteelweg). Locally, a 2 m thick organic fill was found in scours within the channel. Palynological research (see discussion on pollen diagram Kasteelweg, Fig. 6) indicates that the base of the filling (just after abandonment of the channel) dates from the end of the Younger Dryas. Two ^{14}C-dates of 10,150 ± 80 BP and 10,140 ± 60 BP at the base of the channel fill support the biostratigraphical interpretation. This age proves that the channel was active during the Younger Dryas itself.

Walking towards the northeast, physiographic unit 3 is crossed. Corings in area 3 revealed the existence of older Late Glacial meander sediments (fining-upward sequence capped by a loam layer) in the subsurface. Area 3 is a higher area of straight gravelly channels and interchannel bars, covered by aeolian sand. There is no organic material in these higher lying channels, but on morphological grounds, this channel-bar system is regarded as part of the Younger Dryas floodplain. The dunes besides the channels are interpreted as Younger Dryas river dunes, which were formed by deflation from the neighbouring channels during low water discharges. A well-developed podzol soil is present in the top of the dunes which is in support with the Younger Dryas age of the aeolian sands. Young Holocene aeolian drift sands do not show a podzol soil profile.

The high topographic position of area 3 (the Younger Dryas channel and bar system) gives the impression of an island, surrounded by lower-lying channels of physiographic units 2 and 4. This might indicate that erosion occurred during the Younger Dryas period: the Younger Dryas channels, which first flowed on the topographical level of area 3, gradually incised towards their position in area 2 and probably also area 4.

At the east side of physiographic unit 3 a clear morphological boundary is present between the Younger Dryas dunes and physiographic unit 4: the Holocene floodplain. This boundary is an erosional one, made by a Holocene channel which flowed east of the Younger Dryas dunes. After channel abandonment it was filled with up to 3.5 m of clay and coarse-detrital gyttja (Figs. 7 and 8: pollen diagrams Ooijen). Several detailed cross-sections were made over the channel. Generally, the fill is 3 m thick and registration started in the Atlantic (Fig. 8). Only very locally the fill is thicker (3.5 m) and registration started already during the late Boreal (Fig. 7), which stresses the importance of a detailed survey.

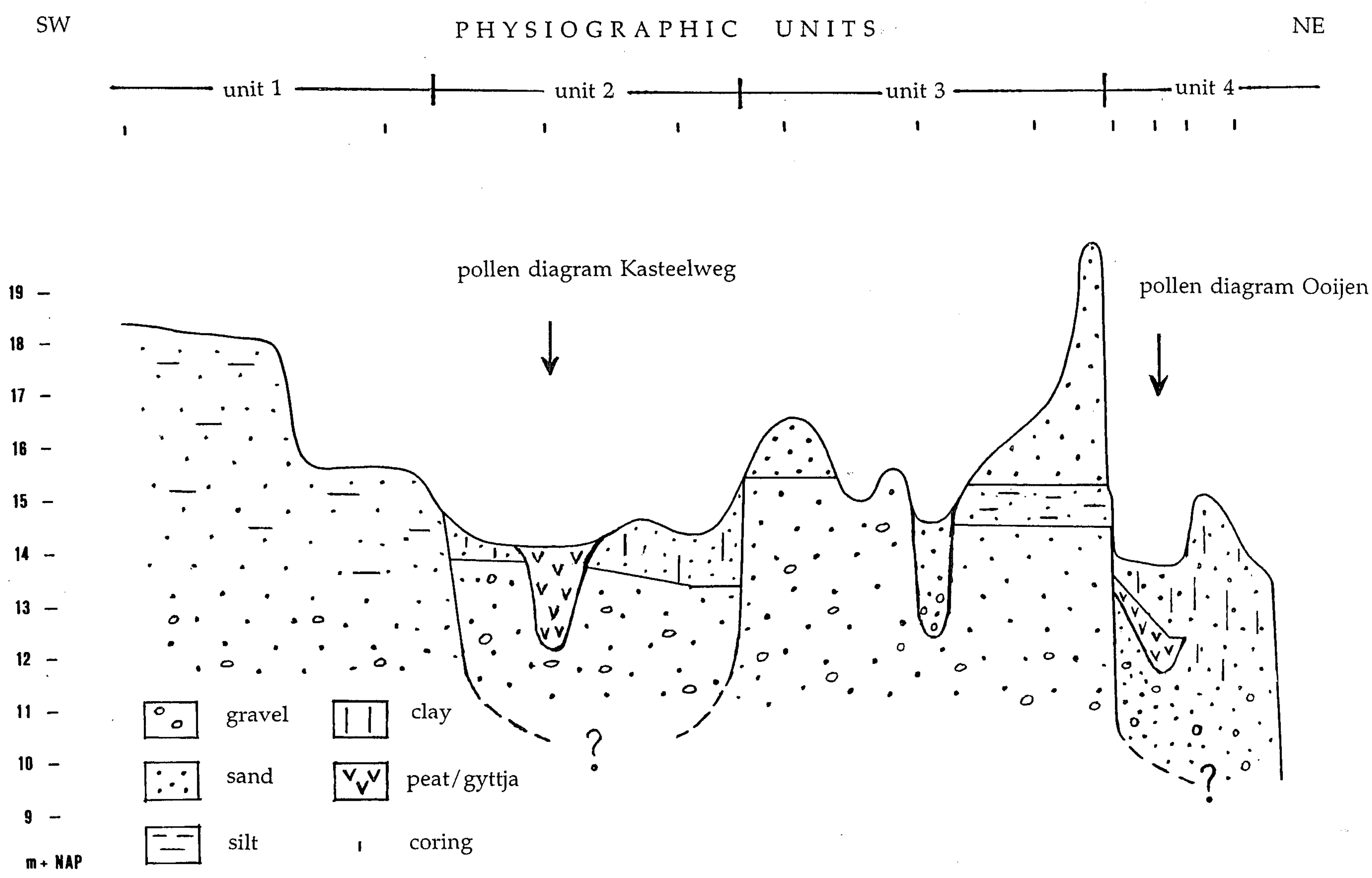

Fig. 5 Schematic cross section and physiographic units of the Younger Dryas and Holocene floodplains at Kasteelweg-Ooijen.

The palynology of the fill is discussed below. Since the base of the fill dates from the late Boreal it is clear that the erosion of the Younger Dryas dunes (boundary between physiographic units 3 and 4, fig. 5) occurred during the Preboreal or early Boreal. Some small finds of Mesolithic flint artifacts on top of the Younger Dryas dunes point to occupation along the erosional boundary and the abandoned channel.

The base of the abandoned early Holocene channel (Fig. 5: Ooijen) is situated at almost the same level as, or slightly lower than, the base of the Younger Dryas channel (Fig. 5: Kasteelweg). This possibly indicates that at the end of the Younger Dryas two channel systems existed, west and east of physiographic area 3. The climatic warming at the Younger Dryas to Preboreal transition probably resulted in a more regular and/or lower discharge, which led to the abandonment of the western branch (Kasteelweg) of the Younger Dryas system. By this change from a multi-channel system into a single-channel system the Ooijen channel became the principal river course at the start of the Holocene.

The pollenrecord at Kasteelweg (fig. 6)

Organogenic deposits in the straight braided channels of Younger Dryas age are rare, but in local scour pools of the largest braid channels organic material has been preserved. At Kasteelweg a black to greenish gyttja was found underlying the peaty sediments that were more frequently encountered. At the transition between these two lithological units a 1 mm thin sand band and a thin gray clay layer are present (see lithological column Fig. 6).

Local pollenzone KAW-1 (193-167 cm):
This zone is characterized by the presence of Gramineae, *Artemisia, Thalictrum* and *Betula*. *Betula* shows an increase towards the top of this zone (35%). *Juniperus* is only rarely present. The overall picture of this zone is one that is dominated by the NAP and the pollenrecord is thought to demonstrate the termination of the **Late Dryas biozone**. Two ^{14}C-dates of 10,150 ± 80 BP and 10,140 ± 60 BP from the base of the channel fill support the biostratigraphical interpretation of a Late Dryas to early Holocene age. The increase in *Betula* towards the top of this zone is interpreted as the Early Holocene *Betula*-rise. The absence of any reaction at this level in the *Juniperus* curve remains peculiar. It can not be excluded that the thin sand band, forming the transition to the overlying zone, reflects a truncation of the underlying zone and that the early Holocene is represented by a hiatus in the sequence.

Local pollenzone KAW-2 (167-143 cm):
After a possible hiatus at the level of the thin sand band registration resumes, while clastic sedimentation takes place. Reworked taxa occur at this level (*Alnus, Corylus, Quercus, Acer*). *Betula*, Gramineae and *Artemisia* maintain in the lower part of this zone, but subsequently the curves of these taxa show a declining tendency and *Thalictrum* is absent from the pollenrecord. Remarkable is the appearance of *Populus* directly after the supposed erosional hiatus.

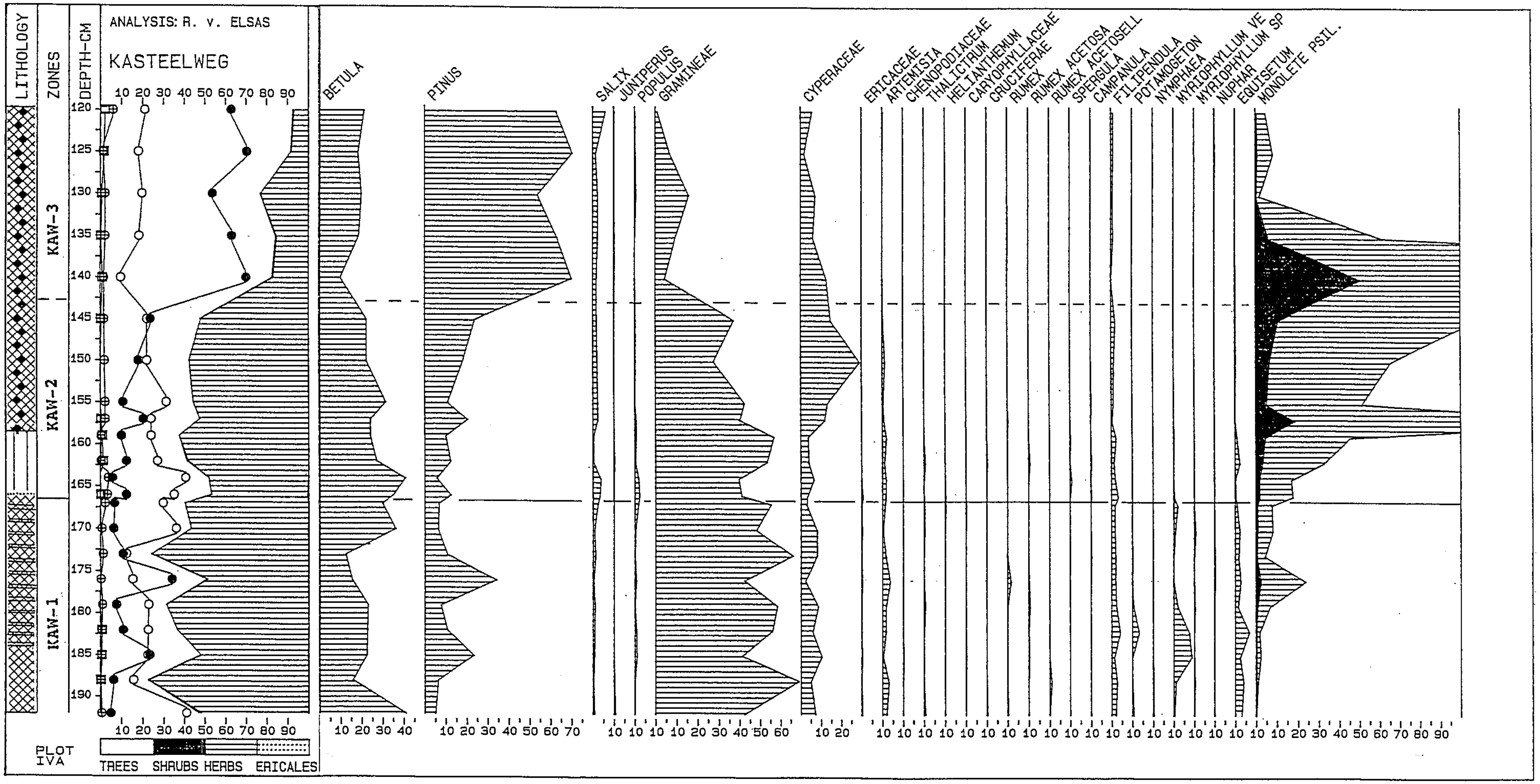

Fig. 6 Pollen record (selection of curves only) from the infill of the Younger Dryas braided channels at Kasteelweg.

The presence of *Populus* at this level in the sequence indicates that the vegetational record can be placed in the **Rammelbeek-phase** of the **Preboreal**. A ^{14}C-date of 9750 ± 70 BP supports the biostratigraphical interpretation of the Rammelbeek phase, which is placed between 9,950 and 9,750 BP (Hoek, 1997).

The upper half of this zone is mainly characterized by a dominance in the Cyperaceae. The Cyperaceae peak is thought to be a local effect of an hydroseral succession following the deposition of the clay.

Local pollenzone KAW-3 (143-120 cm):
The sharp pine increase is indicative for the spread of *Pinus* in the Maas valley during the early Holocene. A ^{14}C-date of 9610 ± 80 BP supports the biostratigraphical interpretation and is in agreement with a date of 9,700 BP for the pine rise in a meander scar due south of this location (Lottum-Schuitwater). *Corylus* only shows a slight increase in the uppermost samples of the analyzed core segment and indicates the end of the **Preboreal** period (c. 9150 BP).

Resuming, the pollenrecord shows the termination of the **Late Dryas**, possibly a hiatus in the early Holocene and infilling during the **Preboreal**. Shortly after the start of the **Preboreal** but before 9,700 BP, possibly during the Rammelbeek phase, deposition of clay from suspension took place possibly reflecting a phase of higher discharge originating from the early Holocene river system in the east.

The Holocene flood plain at Ooijen (figs. 7 and 8)

Two cores were taken in the early Holocene river plain at Ooijen (Figs. 1, 3 and 5). The aim was to establish the moment of channel abandonment and to date the terrace scarp (i.e. incision phase) between the Younger Dryas braidplain and early Holocene meandering plain (Figs. 7 and 8).

Core 91-12 reached a greater depth than core 91-11 and also on lithological grounds appeared to contain a different type of sediment in its basal part. Here a clayey coarse-detrital gyttja was present below the grey sandy clay that was encountered everywhere in the channel fill.

Local pollenzone OOIJ-1 (348-320 cm, core 12):
This zone represents a *Pinus* dominated phase with pine values up to 80%. *Corylus* remains relatively low and *Ulmus* and *Quercus* occur in low frequencies. The pollen assemblage agrees well with the **early Boreal** vegetational history, where within a mature boreal pine forest *Corylus* gradually spread.

Local pollen zone OOIJ-2 (320-283 cm, core 12):
During this zone *Corylus* shows a gradual rise followed by a spread of the Quercetum mixtum (*Quercus, Ulmus, Tilia, Fraxinus*). This succession is typical for the upper half of the **Boreal**

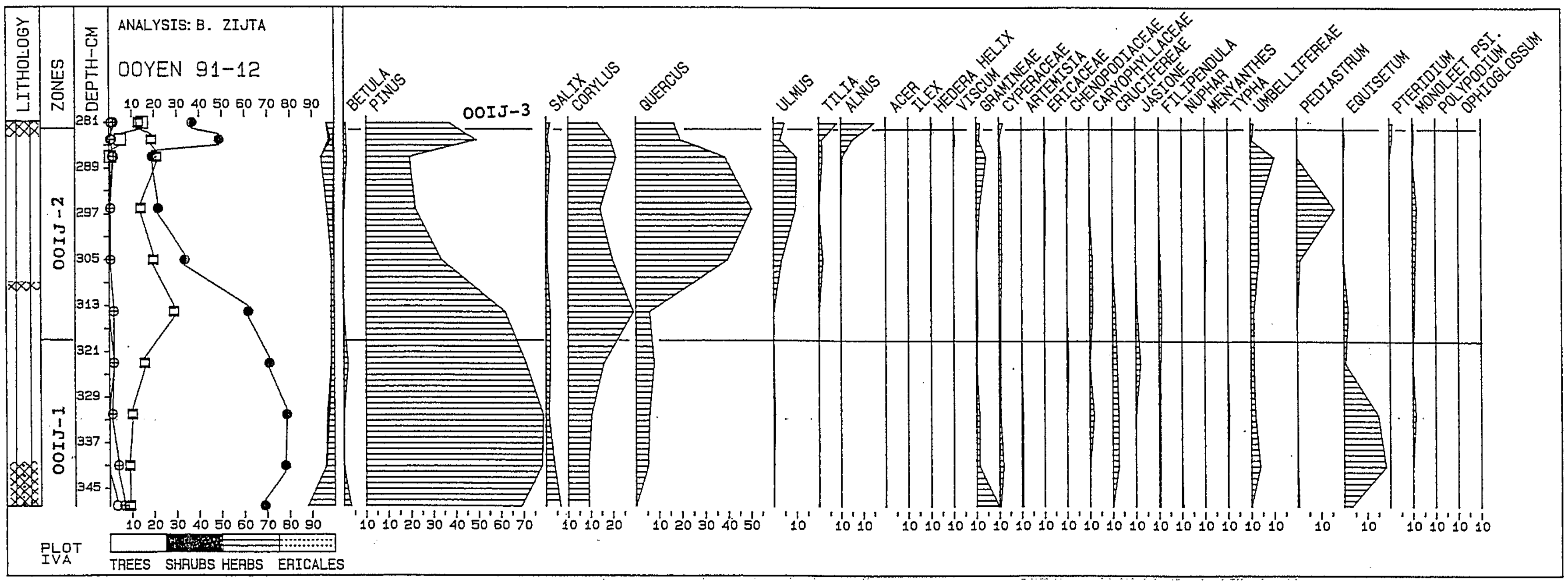

Fig. 7 Pollen record (selection of curves only) from core 91-12 in the Holocene river plain at Ooijen.

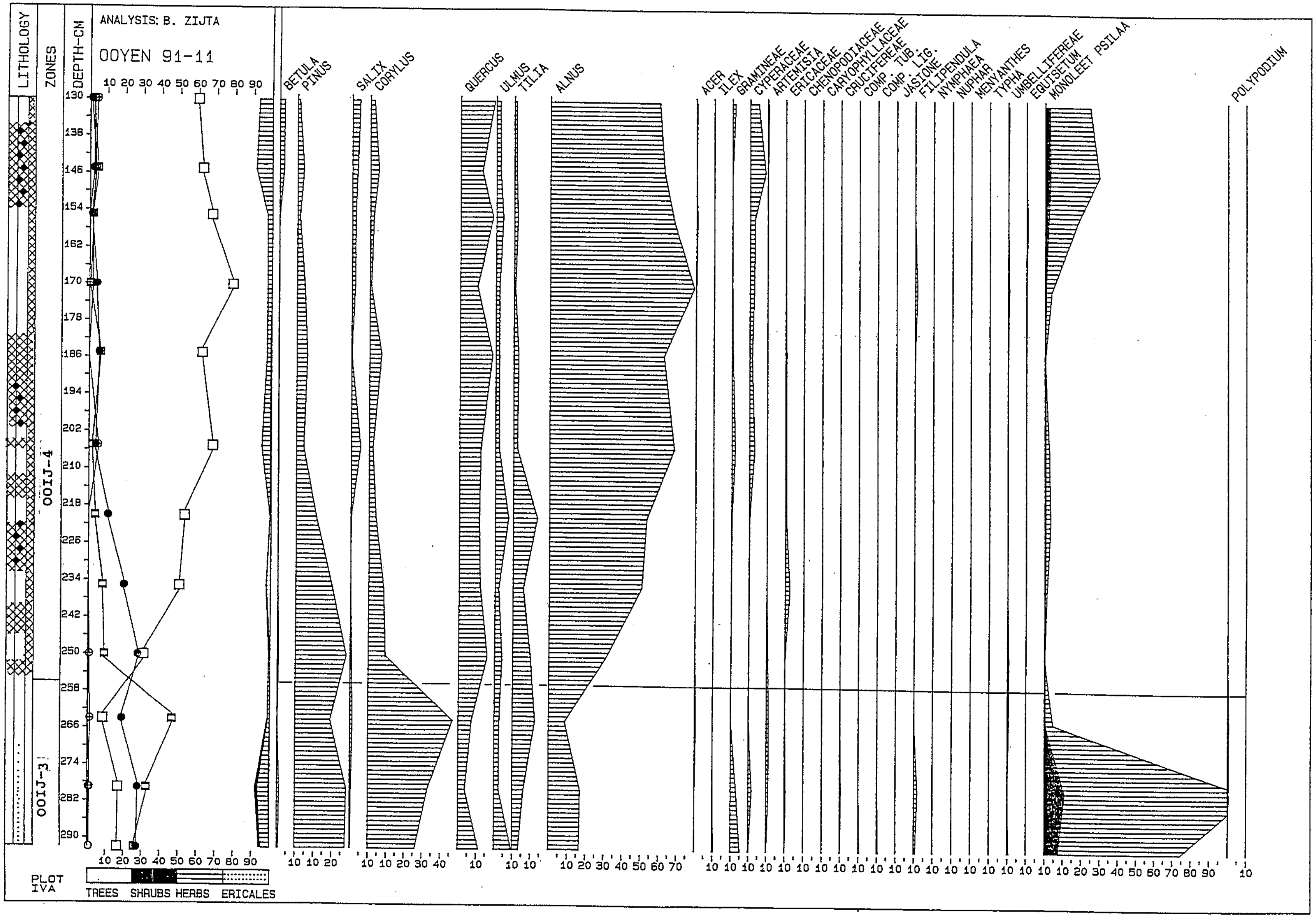

Fig. 8 Pollen record (selection of curves only) from core 91-11 in the Holocene river plain at Ooijen.

period from c. 8400 BP. The spike in the *Pinus* curve at 284 cm does fit the picture for the late **Boreal** where a last *Pinus* maximum is generally present in the pollen records. An estimated date for this dry interval comes to 8250 - 8050 BP (Bohncke, 1991).

Local pollenzone OOIJ-3 (283-281 cm, core 12; 292-256 cm, core 11):
The *Alnus* increase in the top most sample of core 12 indicates that the vegetational record approaches the transition to the **Atlantic** period, c. 8000 BP.

Local pollenzone OOIJ-4 (256-130 cm, core 11):
Core 11 demonstrates that after a gradual increase in *Alnus* a rapid expansion of the species takes place resulting in a relative decline in *Corylus* and *Pinus*. The *Pinus-Alnus* crossing in the pollenrecord is taken to represent the start of the **Atlantic** period. It can not be excluded that at 217 cm depth an erosion hiatus is present in the sequence. Both the lithology (the occurrence of clay pebbles) and the sudden decline in both *Tilia* and *Ulmus* may be taken as indicative for such a feature.

Resuming, the biostratigraphical indications as provided by the pollen records from core 11 and 12 demonstrate that the filling of the channel at Ooijen started shortly after the **Preboreal/Boreal** transition. Accumulation of clayey detrital gyttja and clay continued over the **Boreal** and **early Atlantic** period and probably even longer. The change in channel pattern from the Younger Dryas braidplain to the Holocene meandering system and the associated incision and formation of a terrace scarp can be bracketed between the end of the Younger Dryas (see Kasteelweg) and the start of the Boreal (see Ooijen).

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