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EXCURSION GUIDE

Late Glacial and Early Holocene environmental evolutions, Dinkel and Maas valley, The Netherlands

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EXCURSION PROGRAMME ON 17-04-94

Late Glacial and Early Holocene evolution of the Maas,

Stop 1: BEUGEN

* Late Glacial meander neck cut off
* Allerød organic fill
* Younger Dryas reactivation

Stop 2: WANSSUM

* Morphology Late Pleniglacial - early Late Glacial transitional channels
* Low-sinuosity broad and shallow channels

Stop 3: MEERLO

* Early Younger Dryas shoot cut off
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* Morphology of terrace edge between Late Pleniglacial/Bølling floodplain and
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CLIMATIC CHANGE AND FLUVIAL EVOLUTION OF THE MAAS DURING THE LATE WEICHSELIAN AND EARLY HOLOCENE

Introduction

The Late Weichselian and Holocene evolution of the Maas 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 km². 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 m³/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 km²	185.000 km²
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.

Fig. 1 Catchment areas of the Rhine and the Maas.

Tectonics

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 and early Late Glacial evolution

During the glacial maximum of the Late Pleniglacial the Maas is a braided river system with large discharge fluctuations (fig. 4). Syngenetic ice-wedge casts point to permafrost conditions in the floodplain. Aggradation prevailed because of the high sediment supply in the unvegetated landscape. During annual high water stages the braidplain was flooded and coarse sand was deposited by transverse bars. In the last stage of the 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 because of the slight temperature increase and higher aridity. Parts of the braided plain became covered with eolian sands (so-called coversands). The river maintained its braided character and flowed at more or less the same level as during the previous period.
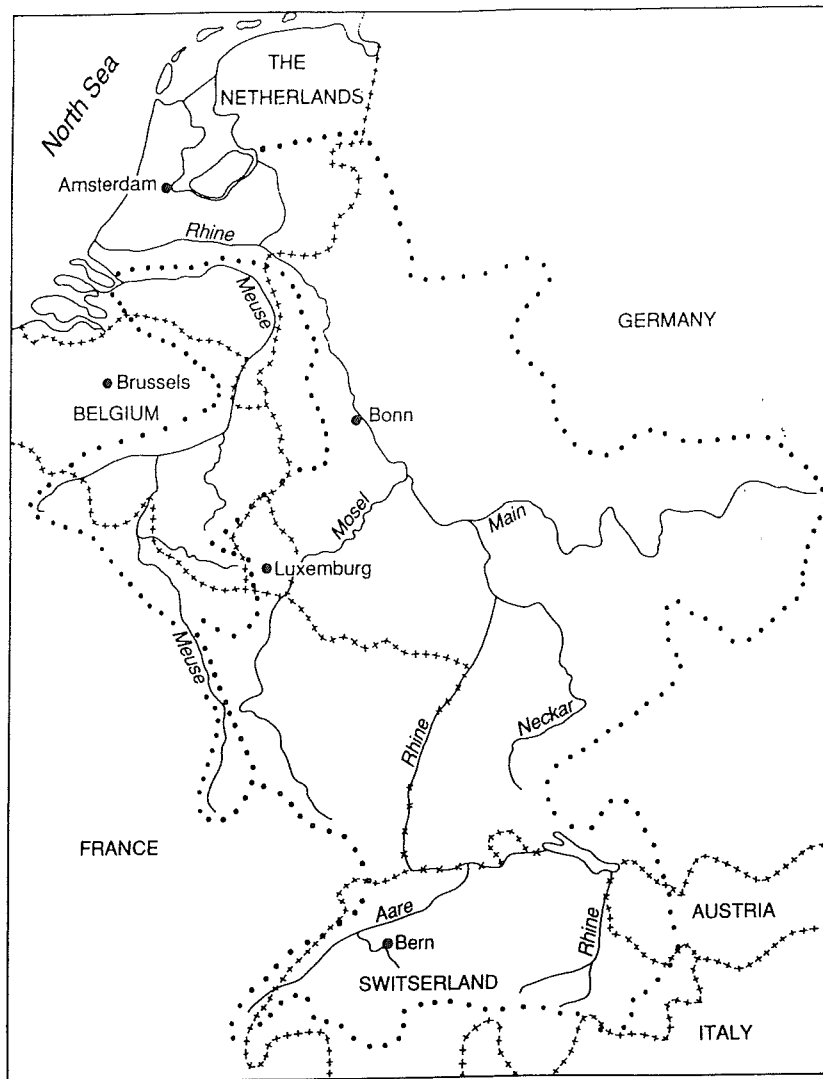
Fig. 1   Catchment areas of the Rhine and the Maas.
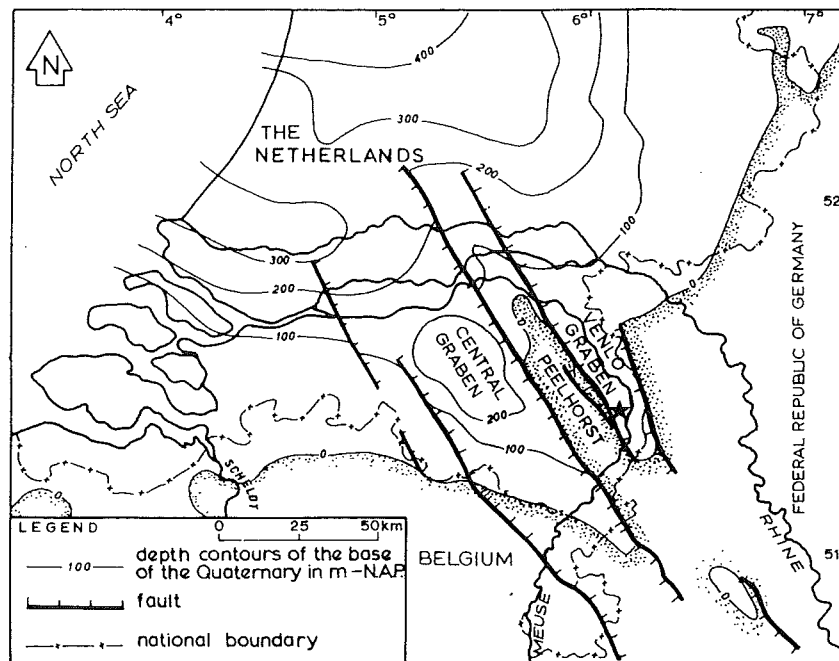

Fig. 2   Depth contours of the base of the Quaternary and major tectonic units of the southern North Sea basin (after Zagwijn & Doppert, 1978). Excursion area is indicated by an asteriks.

The transition of the Pleniglacial to the Late Glacial was accompanied by a temperature rise. 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. In the abandoned braided channels peat formation started during the Bølling, therefore dating the moment of change in the fluvial environment (excursion stop Bosscherheide). This alteration from braided into meandering is characterized by a transitional phase with rather shallow (2-3 m), slightly incised, low-sinuosity channels (excursion stop Wanssum). This transitional phase probably took place during or after the Bølling, but before the Allerød.

Late Glacial terrace morphology

Above the recent Holocene floodplain, two distinct Late Glacial to Early Holocene terrace levels have been distinguished along the Maas north of Venlo (Wolfert & De Lange, 1990)(fig. 3).

The highest Late Glacial terrace level descends from 21 m above sea level at Venlo (base of fig. 3) to 18 m at Holthees (top of fig. 3) and 13 m at Beugen. It is typified by large, high-sinuosity meander scars, especially at the outer terrace edges (excursion stop Beugen). The fine-grained scar fills are 3 to 4 m thick. The organic fills in the meander scars, dating from the Younger Dryas and Allerød, indicate that this high-sinuosity meandering phase occurred during the Allerød. Individual pointbars are poorly developed normally, because of an eolian cover and intense human occupation and cultivation.

The lithological sequence of the meander scar terrace is characterized locally (fig. 3: point A) by a thick (7.5 m) fining-upward, formed by lateral migration of the channel and accretion on the meander inner bend. 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 lower part of the inner channel slope. The upper 4.5 m of the fining-up 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 or it settled from suspension on the upper channel slope. The silt beds reflect high water levels on the upper pointbar slope with local weak currents and deposition of fines from suspension.

LEGEND

⊞ Late Glacial meander scar terrace
▨ Late Glacial - Early Holocene straight to low sinuosity scar terrace
☰ Holocene floodplain
⊡ Late Glacial dune field with dune ridges
◁◁ Terrace edge

Maashees

Bosscherheide

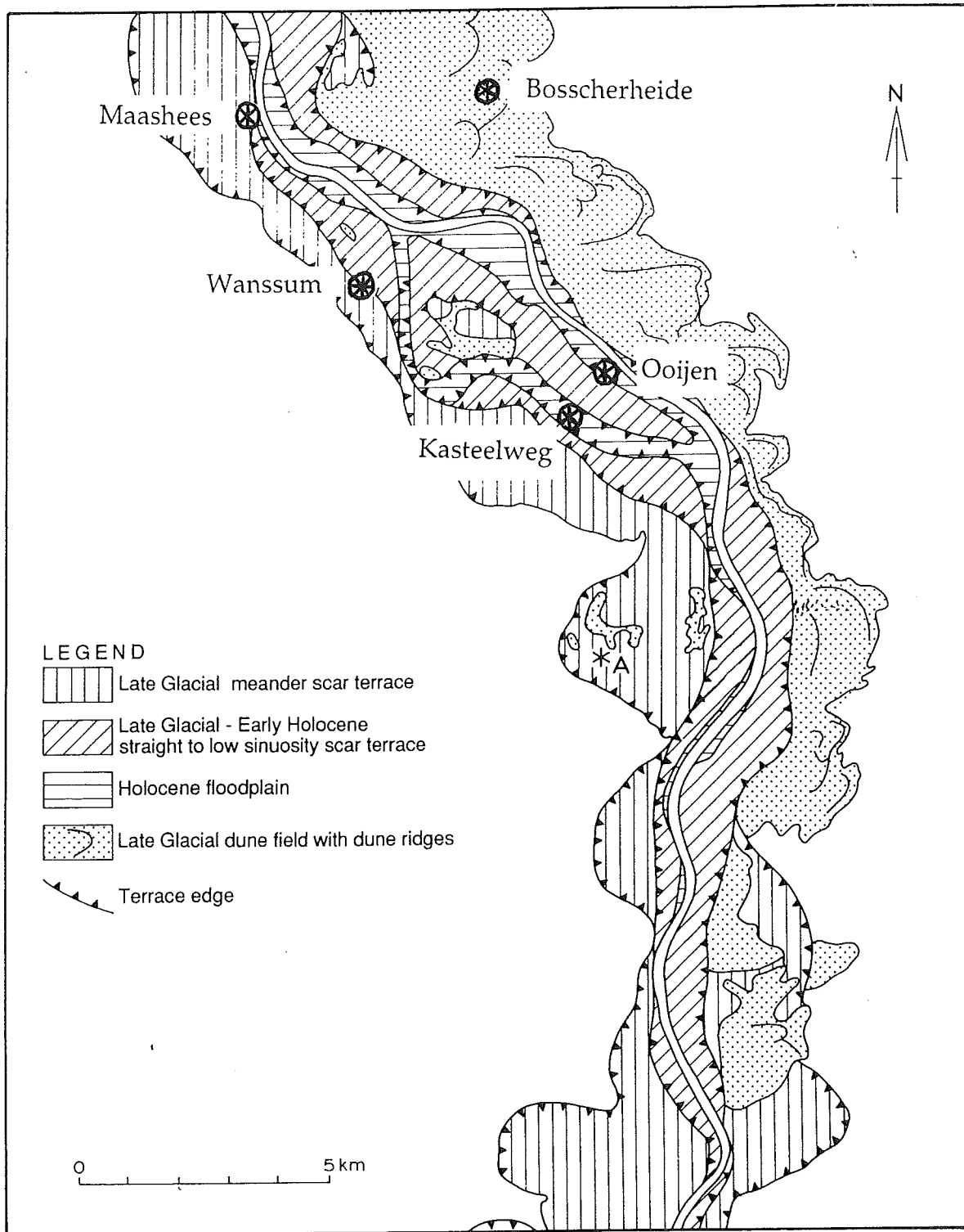Wanssum

Ooijen

Kasteelweg

*A

0   5 km

Fig. 3   Morphological map of the Late Glacial terraces in the Maas valley (after Wolfert & De Lange, 1990). Names refer to excursion sites.

The lower Late Glacial terrace dates from the Younger Dryas. At the start of the Younger Dryas a strong temperature decline occurred, leading to local permafrost conditions (between 10,880 BP and 10,500 BP at excursion stop Bosscherheide) (Bohncke et al, 1993). Lower evapotranspiration and larger discharge fluctuations resulted in local flooding of the Late Pleniglacial braidplain at Bosscherheide. The high sinuosity Allerød channels were abandoned by chute cut offs. The higher discharges resulted not only in a change of channel morphology from meandering into braided, but also in incision. The floodplain declines in altitude from 18 m in the south to 14 m in the north of fig. 3. to 11 m near Beugen. 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 lower terrace level reveals laterally a larger heterogeneity than the higher terrace level. The paleochannels are shallow and broad and contain up to 2m of fine-grained infilling (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.

The climatic amelioration at the start of the Holocene led to a higher land surface stability and to a decrease in the discharge fluctuations, resulting in river incision (fig. 4). The braided channels of Younger Dryas age were abandoned and the river changed into a low-sinuosity meandering system. The oldest fill in the incised channels dates from the Boreal (excursion stop Ooijen), which indicates that the incision phase occurred between the Younger Dryas and the Boreal, i.e. the Preboreal. The Holocene 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. The clear fluvial morphology indicates that the eolian activity, typical of the late Younger Dryas, had ceased.

Late Glacial eolian phases

Two eolian "phases" are recognised in the Late-Glacial river morphology (fig. 3). The oldest "phase" is less pronounced and only locally present on the higher meander terrace level (fig. 3: point A). At point A the terrace is covered by a 0 to 4 m thick unit of eolian sediments. Since there is no evidence for an eolian supply from outside the area, nor for a supply from the subsoil, we conclude that these eolian sediments on top of the meander terrace east of point A were formed during the lateral migration of the meandering river channel. The comparable grain-size of the eolian deposits and the upper channel slope deposits support this hypothesis. The following mechanism has to be considered.
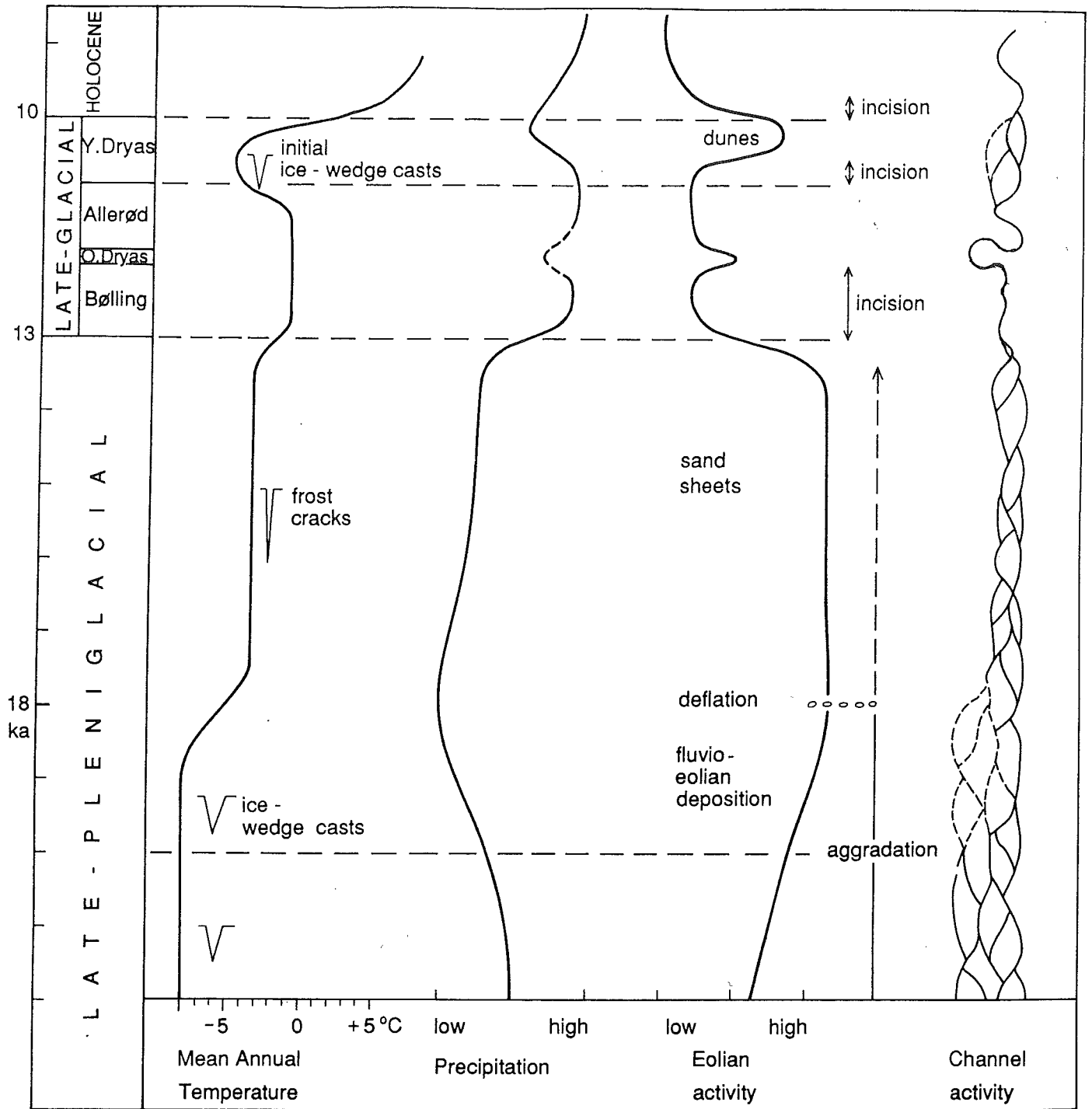
Fig. 4   Synthesis of the Late Pleniglacial and Late Glacial climatic changes and fluvial evolution of the Maas in The Netherlands.

During bankfull discharge, probably in spring due to melting of the snow cover and low evapotranspiration, sediment was deposited on the meander upper inner bank. During the following low discharge this barren sediment on the upper slope was deflated by the prevailing westerly winds (Maarleveld, 1960; Schwan, 1988) and deposited on top of the point bar. The low dune morphology indicates eolian deposition at the edge of a vegetated landscape.

The start of the meander scar infill at point A 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.

The younger Late Glacial eolian phase dates 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: excursion stop Bosscherheide).

The dune morphology is characterized by parabolic forms in the eastern (downwind) part of the dune field. Because of this morphology a southwestern wind is inferred for this eolian phase.

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 eolian deposition. From the top of the peaty layer, characterized by an alternation of moss-laminae and eolian sand-laminae, a C14 dating of 10,500 ± 60 BP was obtained, which places the overlying dune body in the late Younger Dryas period.

The eolian sediment of this phase is fine to medium grained. Because of the prevailing southwestern winds, the source area must have been the Younger Dryas paleofloodplain west of the dune field. It is likely that the eolian sediment has been blown from the Younger Dryas floodplain during periodic low discharges. The sediments that were deposited outside the floodplain have been preserved. The eolian sediment within the floodplain had a low preservation potential, since it occurred on top of the fluvial sediments and therefore was easily eroded during subsequent periods of high discharges.

The accumulation of large amounts of sand in the extensive dune field on the eastern bank of the Maas was possible due to the fluvial reworking of the braided floodplain followed by deflation. These alternating processes 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 indicates a braided river system during this period. The larger fluctuations in discharge are attributed to the climatic deterioration during the Younger Dryas. The lower evapotranspiration and the larger volumes of snow melt water will have led to larger peak discharges. As a result the Late Glacial meandering river course changed into the straight Younger Dryas course.

References

Bohncke, S., J. Vandenberghe, J. & A. Huijzer 1993 Periglacial paleoenvironments during the Late Glacial in the Maas valley (The Netherlands) - Geol. Mijnbouw 72: 193-210.

Jongman, R.H.G. 1987 The Rhine-Maas megalopolis and its waterresources. Workshop on interrelated bioclimatic and landuse changes, Noordwijkerhout: 1-14.

Maarleveld, G.C. 1960 Wind directions and cover sands in the Netherlands - Biul. Peryglacj. 8: 49-58.

Schwan, J. 1988 Sedimentology of coversands in northwestern Europe. Thesis, Free University Amsterdam, 137 pp.

Van den Broek, J.M.M. & G.C. Maarleveld 1963 The Late-Pleistocene terrace deposits of the Meuse - Med. Geol. Sticht. 16: 13-24.

Van den Toorn, J.C. 1967 Toelichting bij de geologische kaart van Nederland 1:50.000, Blad Venlo-West (52W). Geol. Sticht., Haarlem, 163 pp.

Wolfert, H.P. & G.W. De Lange 1990 Toelichting bij de geomorfologische kaart van Nederland 1:50.000, Blad 52 Venlo. Staring Centrum/Rijks Geol. Dienst, Haarlem, 27 pp.

Zagwijn, W.H. 1989 The Netherlands during the Tertiary and the Quaternary: A case history of coastal lowland evolution - Geol. Mijnbouw 68: 107-120.

Zagwijn, W.H. & J.W.Chr. Doppert 1978 Upper Cenozoic of the southern North Sea basin: palaeoclimatic and palaeogeographic evolution - Geol. Mijnbouw 57: 577-588.

Zagwijn, W.H. & C.J. Van Staalduinen 1975 Toelichting bij geologische overzichtskaarten van Nederland. Rijks Geol. Dienst, Haarlem, 134 pp.

LATE GLACIAL AND EARLY HOLOCENE VEGETATIONAL AND CLIMATIC HISTORY IN THE NETHERLANDS

The earliest accumulation of organic matter under favourable palaeohydrological conditions seems to have taken place shortly after 13 Ka BP. An AMS-date from the type-locality of Usselo (Van Geel et al., 1989) provides a date of 12,930 BP. At 12,885 BP organic sedimentation started in the former Dinkel Valley (Ran, 1990). Conventional C-14 dates, based on larger quatities of organic matter, indicate an onset of organic production in abandoned shallow river channels at around 12,700-12,600 BP (Bohncke et al., 1987; Teunissen, 1983).

Prior to 13 Ka BP a shortage in effective precipitation may well have been the limiting factor for the spread of shrubs and trees (e.g. Van Campo, 1984). In the absence of a reasonable vegetation cover the surface was prone to wind activity and a continuous transport of sand led to the formation of coversands (Older Coversands II, Van der Hammen, 1971). The Older Coversands II consist of horizontally bedded fine to medium sands. Sedimentary structures that indicate running water are rare and periglacial features are absent except for some faint micro drop-soil structures. Pollen analyses of the oldest infill of pingo remnants (e.g. Mekelermeer; Bohncke et al., 1988a) indicate the presence of a heliophilous herbaceous vegetation. *Helianthemum*, *Rumex acetosa/acetosella*, *Polemonium*, *Thalictrum*, *Artemisia* and some Chenopodiaceae and *Plantago major/media* form important constituants of this unique vegetation.

12,600-12,000 BP: the Bølling biozone.

With an increase in effective precipitation shortly after 13 Ka BP a *Betula nana* (dwarf birch) and *Salix* spp. shrub tundra expands and *Artemisia* becomes more frequent. The substratum gradually became fixed and aeolian activity faded out. The presence of freely drained soils in combination with an increase in precipitation favoured the spread of *Juniperus* (ca 12,600 BP) which often forms a distinct vegetation belt before tree birches shade these out.

With the establishment of tree birches the AP/NAP ratio increases and a boreal birch forest seems to have been the dominant vegetation type. Biostratigraphically this pollen assemblage zone (*Betula-Salix-Artemisia* p.a.z.) is correlated with the Bølling.

12,100-11,800 BP: the Early Dryas biozone.

The Early Dryas biozone, which is defined as a temporary decline in the AP and a concomitant increase in the NAP and which is recognizable in Late Glacial lake sediments, does not show up in the pollen assemblages of terrestrial peat sequences. Two possibilities can account for this discrepancy 1) terrestrial organic deposits underwent a period of non registration and 2) site inherent factors causing the same time-stratigraphic event to be registered differently in the pollen assemblages. In both cases the palaeohydrological conditions of the sites involved play an important part. The palaeobotanical record suggests a dry continental climatic episode for this time interval.

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The lowering of the phreatic level during the Early Dryas is reflected in the pollenrecord by a regional decline in the *Betula* boreal woodland. A return to a shrub and heliophilous herb vegetation is registered (the *Salix-Juniperus-Artemisia*-Gramineae p.a.z.).

In terrestrial sites, where the hydroseral successsion is already more towards telmatic and semi-terrestrial, conditions will react to this relatively dry period either with a next stage in the vegetational succession, being the development of a local birch carr on top of the preceeding stage with sedges and pleurocarpe mosses or will respond with hiatus(es) to this period with pronounced decline in effective precipitation.

Soil instability and the inwash of minerogenic sediments, which is reflected in the geochemical record of lake sediments (Bohncke & Wijmstra, 1988), characterize this zone. Terrestrial sites show the influx of aeolian sediments trapped in the local vegetation. A date of 11,990 ± 70 BP is avaible to pinpoint this event (e.g. site Mariahout, Bohncke in prep.).

11,900-10,900: the **Allerød** biozone.

Palaeohydrological studies reveal the return to high lake-levels (Bohncke & Wijmstra, 1988) for the start of this period. The increase in the phreatic level determines the termination of a period with possible hiatuses. Locally, conditions became suitable for aquatic species to participate in the hydroseral succession Bohncke et al., in press). The pollenrecord demonstrates a regional spread of tree birches in the southern Netherlands while at more northern latitudes this *Betula* increase is preceeded by a *Juniperus* maximum. Both taxa rely on increased amounts of precipitation.

At ca 11,300 BP sites adjacent to river systems register a second phase with wetter conditions and occasionally the deposition of loamy lenses in their sequence is evident. An increase in (fluvial) erosion is inferred from this (Bohncke et al., 1987; Bohncke et al., in press). The local hydroseral succession is set back to open water. In the subsequent local hydroseral succession the rhizomatose perennials like *Typha* spp. and *Phragmites* are lacking (e.g. site Bosscherheide, Bohncke et al., in press) which leads to the supposition that winter temperatures must have dropped and the annual temperature range augmented. Consequently, a more intensive action of the freeze-thaw cycle started to operate and instable soil conditions re-established.

Moreover, the lacustrine environments at this moment in time show an increase in *Isoëtes* sp., *Elatine hydropiper, E. hexandra* (Bohncke & Wijmstra, 1988). Both species require pioneer conditions. *Isoëtes* as a lake bottom dweller and *Elatine* spp. on the borders of the lake. It is not unlikely that longterm pioneer conditions were created by ice action.

Remarkable in this respect is the almost simultaneous westward spread of *Pinus*. Both in Usselo (Van Geel et al., 1989) and in Bosscherheide (Bohncke et al., in press) this increase is dated to 11,300 BP. Since *Pinus* can be regarded as a more continental species, with its plant geographical centre of distribution at that period located in eastern Europe, it is concluded that the two features are somehow related. The spread of *Pinus* in this context must be interpreted as a respons to climatic changes at around 11,300 BP implying more severe winter temperatures and increase in freeze-thaw incidenses. The

latter process is responsible for the increase in instable soil conditions and possible the increase in sediment load in the fluvial systems as reflected in the deposition of overbank deposits.

The termination of the **Allerød,** both in lake and terrestrial peat sequences, shows relatively dry conditions. Both a decline in effective precipitation and an increased evapotranspiration may be responsable for this phenomenon.

10,900-10,500: the Late Dryas biozone.

At ca 10,850 BP both vegetation and coleoptera (Bohncke et al., 1987) indicate a distinct drop in the average July temperature from between 18 and 15 °C to 11 and 10 °C. Consequently, the mean annual temperatures declined (estimated to between -2 and -5 °C) and conditions approached those of permafrost environments. Simultaneously effective precipitation increased considerably resulting in a pronounced rise in the lake-levels (the lacustrine environment) and large-scale floodings and deposition of suspension load in large shallow lakes in the fluvial environment. Reconstruction of the vegetational history over the Allerød - Late Dryas transition quite often is obscured by erosion hiatuses and redeposition.

Evaluation of the available data in the Netherlands, taken into account the limitations set out above, provides a picture of a rapidly declining *Pinus* boreal forest. The pine forest was replaced by an open shrub vegetation with *Betula (nana)* and *Salix* spp. (e.g. *Salix reticulata*). Especially the presence of dwarf willows, generally known to be chinophilous, may indicate that at least a large part of the precipitation fell as snow. Heliophylous herbs became relatively more frequent (*Artemisia, Thalictrum,* Cyperaceae, Chenopodiaceae, *Helianthemum* and in some cases *Polemonium*). The local vegetation is characterized by the increase in aquatic taxa such as *Potamogeton* spp., *Batrachium* spp., *Hippuris vulgaris, Myriophyllum spicatum, Myriophyllum alterniflorum, Menyanthes trifoliata, Equisetum palustris* and *Sparganium* spp. and pleurocarpe mosses. This period came to an end when precipitation deminished and the previously established large shallow lakes dried out at ca 10,500 BP.

10,500-10,200 BP: the Late Dryas biozone.

Soon after, ca 10,500 BP, summer temperatures appear to have risen while effective precipitation declined considerably resulting in a decline in the lake-levels and periodically emerging river beds. Aeolian activity prevailed building up large parabolic river dunes (Bohncke et al., in press).

At around 10,500 BP permafrost had disappeared completely and mean annual temperatures increased to about -1 °C. With the disappearance of the discontinuous permafrost the phreatic level declined, not to be restored by sufficient precipitation. In the absence of a protecting snow cover conditions may have been even more hars for plant-live. This is emphasized in the pollen assemblages of this period by the presence, although sparse, of *Ephedra* spp. and *Polemonium*. The vegetational response to this dry upper part of the **Late Dryas** differs considerably, depending on the geographical position

of the study-site.

In the northern part of the Netherlands this episode is preeminently characterized by *Empetrum* heath intermingled with dwarf birches and dwarf willows. Locally some tree birches and pine trees may have survived. The NAP values are relatively low and dominated by sedges.

In the southern Netherlands *Pinus* spreads to reach dominance over *Betula* (tree-birch and dwarf-birch) by ca 10,400 BP (Vandenberghe & Bohncke, 1985).

In the SE Netherlands both *Betula* and *Pinus* persisted in equally high percentages intermingled with some *Juniperus* and *Betula nana* shrubs. The herb vegetation is dominated by grasses and *Artemisia*. Locally, *Helianthemum* and *Polemonium* may become important species in the sequence.

The prevailing dry conditions on the other hand frequently led to the hiatus(es) in the sequences, embrazing the later part of the Late Dryas and the early Holocene.

10,200-9,100 BP: the **Preboreal** biozone.

The Late Dryas period comes to an end when precipitation restores leading to an expansion of *Betula pubescens* s.l. In the northern and southeastern Netherlands this birch maximum is preceeded by a *Juniperus* maximum. In the southern Netherlands, where *Pinus* forms the major tree during the later part of the Late Dryas, *Betula* increases and sometimes an interval in which *Betula* and *Pinus* reach equally high values is registered. This first episode of climatic amelioration of the early Preboreal is called the **Fiesland-phase** (10,200-9,850 BP).

Subsequently more continental conditions involving a decline in effective precipitation and possibly also in winter temperatures, determines the vegetational development: the **Rammelbeek-phase** (9,850-9,750 BP). Again tree-birches (*Betula pubescens* s.l.) suffer from a decline in the phreatic level in the upland regions. An increase in grasses and a rapid hydroseral succession is registered at the site of the Borchert, northern Netherlands (Van Geel et al., 1981). Towards the end of this phase *Populus tremula* becomes a pioneer species on the emerging fringes of the mires. In the area within the pine-forest limits of that time, the more continental conditions result in a further expansion of *Pinus*, while the areas adjacent to these limits register their first Holocene *Pinus* expansion (9,705 ± 50 BP, Schuitwater Broekhuizen, Maas valley). At many localities the dry conditions again may result in a period of non-registration.

During the **Late-Preboreal** (9,750-9,150) *Betula* restores from its temporary low values during the Rammelbeek-phase. An increase in the lake-levels for this period (Bohncke, 1991) indicates a restorance of the effective precipitation, which may account for the registered *Betula* increase. This time *Betula* verrucosa is involved in the vegetational succession (Van Geel et al., 1981).

In the southern Netherlands, where *Pinus* dominates the picture, small-scale fluctuations in the *Pinus* pollen values are registered. The dips in the *Pinus* curve consistently seem to coincide with an increase in local wet conditions (aquatics, *Salix*) and a slight increase in the *Betula* values. At 9,400 BP *Pinus* spreads in the northern part of the Netherlands, some 1000 C-14 years later than the Pinus increase in the SE

Netherlands. *Pteridium aquilinum* and *Melampyrum* form part of the understory of the pine forests that dominated the late Preboreal vegetation cover.

The record demonstrates that the *Pinus* immigration and expansion is not simply a gradual northwards spread during the early Holocene but that superimposed on this process, periods of swift expansion designate the step-like expansion of the species.

9,150-8,000 BP: the **Boreal** biozone.

The start of this zone is characterized by a gradual increase in *Corylus* (Hazel). *Corylus* is a heliophilous species and its spread during the Boreal implies a rather open forest for this period. Almost simultaneously *Quercus* (Oak) and *Ulmus* (Elm) appear in the vegetational record in the S. Netherlands accompanied by low values for *Tilia* (Lime). Towards ca 8,400 *Pinus* generally shows a maximum before it started to decline during the later part of the Boreal. Although lake-levels are generally low during the whole of the Boreal, this *Pinus* maximum coincides with a minimum in the lake-levels.

From ca 8,400 the forest canopy becomes more dense. *Pinus* values decline, while *Corylus* and the Quercetum mixtum (*Quercus, Ulmus, Tilia* and some *Fraxinus* (Ash)) expands. Climbers like *Hedera helix* and *Viscum album* occur in low frequencies, indicating that winter temperatures from now on are less severe (mean T Jan. above -2 °C). Towards the end of the Boreal a last *Pinus* maximum is often evident in the pollen record.

8,000-5,000 BP: the **Atlantic** biozone.

The Atlantic period is characterized by a fast expansion of *Alnus*, which quickly outranges the *Pinus* values of the Late Boreal. The rapid increase in *Alnus*, a species that requires rather wet conditions, implies major environmental changes. It is assumed that air circulation patterns have changed dramatically and led to an increase in (effective) precipitation. This process will have initiated a rise in the phreatic level and a spread of water logged sites, the habitat that *Alnus* needs for its succesfull spread in the early Atlantic period. This is in agreement with the distinct rise in lake-levels found for the early Atlantic period (Bohncke, 1991). Between 7,000 and 6,700 BP *Fraxinus* reaches it rational limit (a continuous curve).

References

Bohncke, S. 1991 Palaeohydrological changes in The Netherlands during the last 13.000 years. Thesis, Vrije Universiteit, Amsterdam, 187pp.

Bohncke, S., J. Vandenberghe, R.G. Coope & R. Reiling 1987 Geomorphology and Palaeoecology of the Mark Valley (Southern Netherlands): palaeoecology, palaeohydrology and climate during the Weichselian Late Glacial - Boreas 16: 69-85.

Bohncke, S., T.A. Wijmstra, J. Van der Woude & H. Sohl 1988a The Late Glacial infill of three lake successions in The Netherlands; regional vegetational history in relation to NW European vegetational developments - Boreas 17: 385-402.

Bohncke, S. & T.A. Wijmstra 1988b Reconstruction of Late Glacial lake-level fluctuations in The Netherlands based on palaobotanical analyses, geochemical results and pollen density data - Boreas 17: 403-425.

Bohncke, S., J. Vandenberghe & A.S. Huijzer 1993 Periglacial palaeoenvironments during the Lateglacial in the Maas valley, The Netherlands - Geol. Mijnbouw 72: 193-210.

Ran, E. 1990 Dynamics of vegetation and environment during the Middle Pleniglacial in the Dinkel Valley (The Netherlands) - Meded. Rijks Geol. Dienst 44-3: 141-220.

Teunissen, D. 1983 The development of the landscape of nature reserve De Hamert and its environments in the northern part of the province of Limburg, The Netherlands - Geol. Mijnbouw 62: 569-576.

Van Campo, M. 1984 Relations entre la vegetation de l'Europe et les temperatures de surface oceaniques apres le dernier maximum glaciere - Pollen et Spores 26: 497-518.

Vandenberghe, J. & S. Bohncke 1985 The Weichselian Late Glacial in a small lowland valley (Mark river, Belgium and The Netherlands) - Bull. de l'AFEQ 2-3: 167-175.

Van der Hammen, T. 1971 The Upper Quaternary stratigraphy of the Dinkel valley. In: Van der Hammen, T. & T.A. Wijmstra (eds): The Upper Quaternary of the Dinkel valley - Meded. Rijks Geol. Dienst, N.S. 22: 81-85.

Van Geel, B., R.G. Coope & T. Van der Hammen 1989 Palaeoecology and stratigraphy of the Lateglacial type section at Usselo (The Netherlands) - Rev. Paleob. Palynol. 60: 25-129.

Van Geel, B., S.J.P. Bohncke & H. Dee 1981 A Palaeoecological study of an upper Late Glacial and Holocene sequence from "De Borchert", The Netherlands - Rev. Palaeob. Palynol. 31: 359-448.

EXCURSION SITE 1

THE ALLERØD MEANDER SCAR AT BEUGEN

Beugen is located on the left bank of the Maas, 20 km southeast of Nijmegen (fig. 1 and 2). At this excursion point a well developed Late Glacial meander scar is present, previously investigated by Teunissen (1990: pollen diagram Helbroek). This scar is situated on the Late Glacial terrace, 11-13 m above sea level (fig. 2 and 3). To the south this terrace with large meander scars is separated from the Pleniglacial floodplain by a terrace edge. At the excursion point Beugen, the Late Glacial meander occurs at nearly the same topographic level as the Pleniglacial floodplain in the west (see fig. 3). However, erosion did occur during the transition from the braided Late Pleniglacial/Early Late Glacial phase into the Late Glacial meandering phase. In figure 4 it is clear that coarse grained channel sediments in the inner bend (pointbar) of the meander occur at a higher level than the base of the channel at the moment of cut off. This means that during lateral migration of the channel also vertical erosion took place.

The meander scar gives a good impression of the Late Glacial fluvial system. The scar is approximately 200 m wide and the infilling is up to 3 m thick. The scar is a fine example of a neck cut off. Such neck cut offs are caused by the meandering process itself and they are not related to climatic changes, in contrast to chute cut offs which can be caused by climatic changes as well. Therefore, the start of the infilling of the cut off channel gives a good date of the period in which the floodplain was characterized by large-scale meanders. The fill was studied by several detailed cross sections (fig. 5). The fill consists of fine laminated gyttja at the base (fig. 5: unit 1), abruptly lying on the channel sediment deposited before the cut off (coarse sand and gravel). Palynological analyses (see fig. 6) point to an Allerød age (*Betula* phase) for the start of the infilling. This date shows that the Maas was a large scale meandering river at least during the Allerød (moment of neck cut off).

The gyttja is erosively overlain by a gray, fine sandy clay, often calcareous at the base (fig. 5: unit 2). This unit represents a phase a renewed fluvial activity in the meander scar. Current velocity in the channel was low (deposition of sandy clay) and the Allerød channel morphology was not modified. The erosional transition from gyttja to clay correlates pollen analytically with the Allerød - Younger Dryas boundary. This means that fluvial inundations with clay deposition reached the channel again after a period of organic deposition (unit 1). The Younger Dryas cooling is held responsible for this change in fluvial dynamics. The decreased evapotranspiration and probably also the increased snowfall in the higher regions of the Maas catchment (Ardennes) resulted in higher discharges and higher peak discharges. These higher peak discharges are the reason for the renewed inundation of the Allerød meander scar at Beugen and the Late Pleniglacial floodplain at Bosscherheide (stop 2). Unit 3 (fig. 5) represents a sand layer, which was probably deposited by a brook flowing through the scar during the Holocene. Unit 4 is a peat bed of Holocene age. Locally, it is thicker and the base dates from the Late Boreal or Early Atlantic (Teunissen, 1990). The peat overlies the Younger Dryas clay, which indicates a hiatus due to non-deposition during the Preboreal and Boreal.
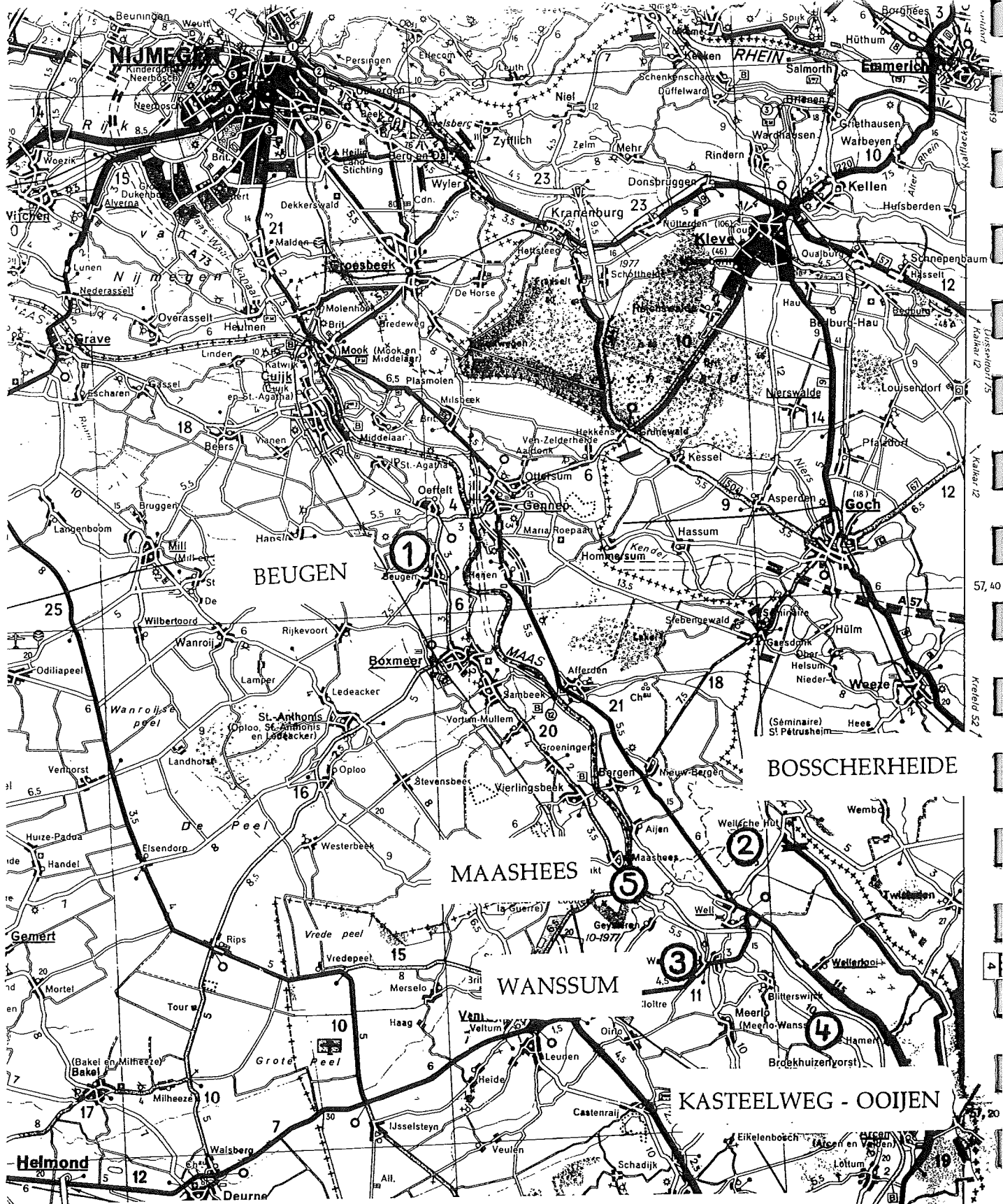
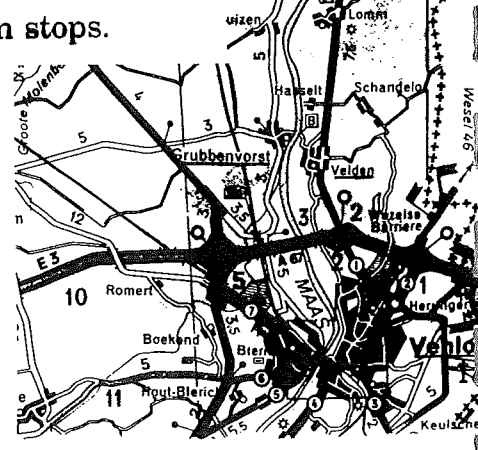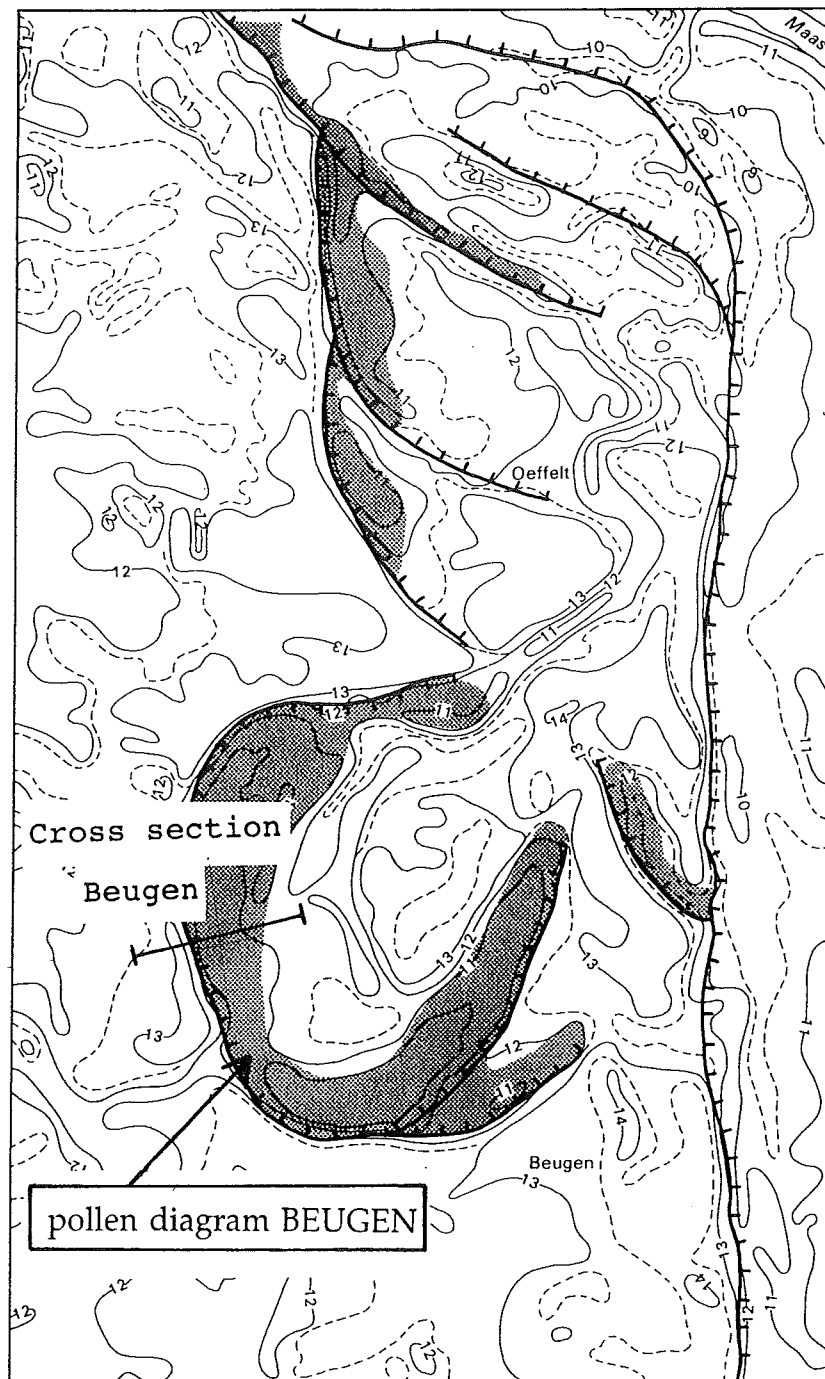Fig. 1   General location map with the excursion stops.

Fig. 2   Topographical map of excursion stop Beugen (scale 1:25.000).

—10— contourlines in meters above NAP (sea level)
----- contourline in half meters above NAP
▓▓▓ meander scars
└─┴─┘ erosional edge

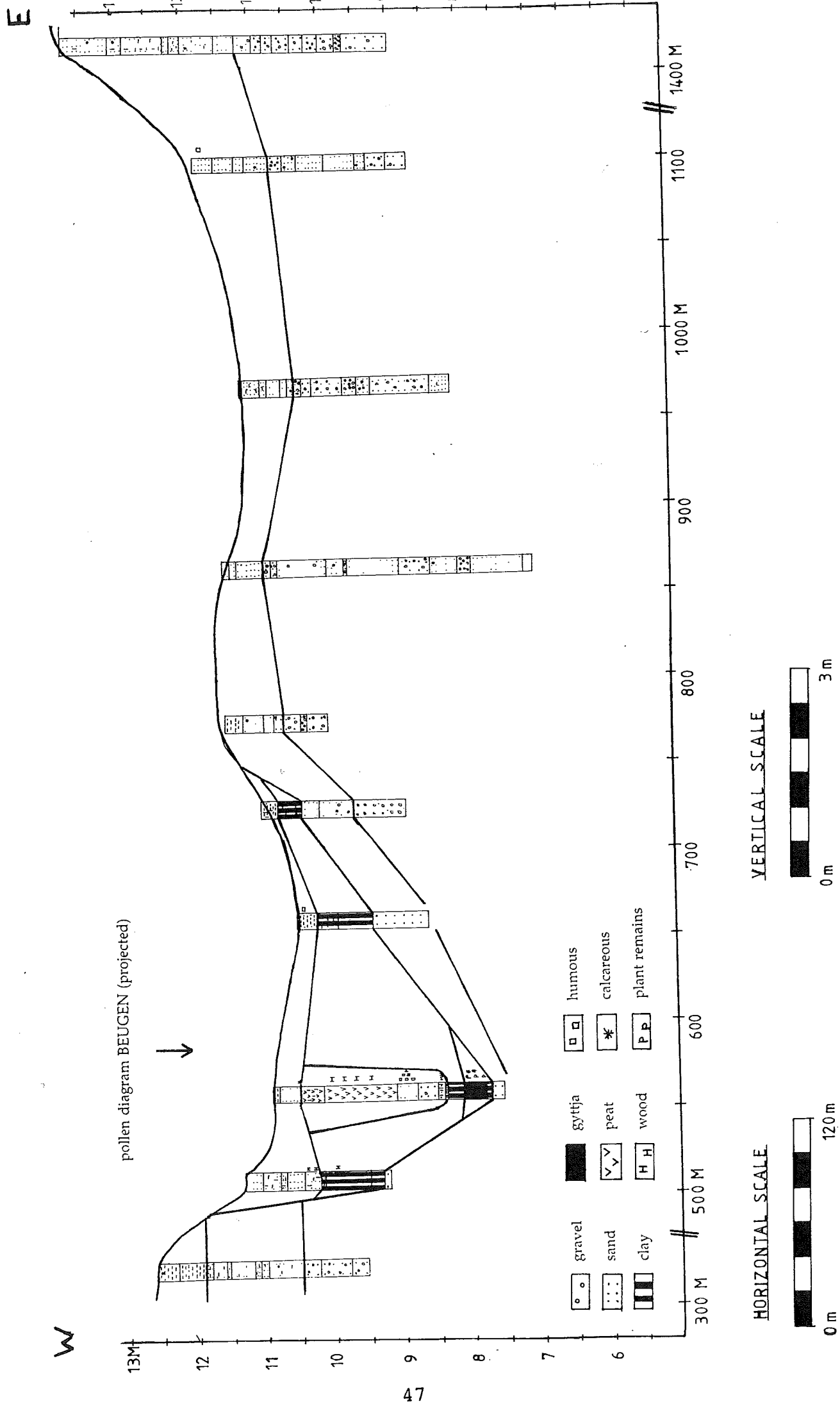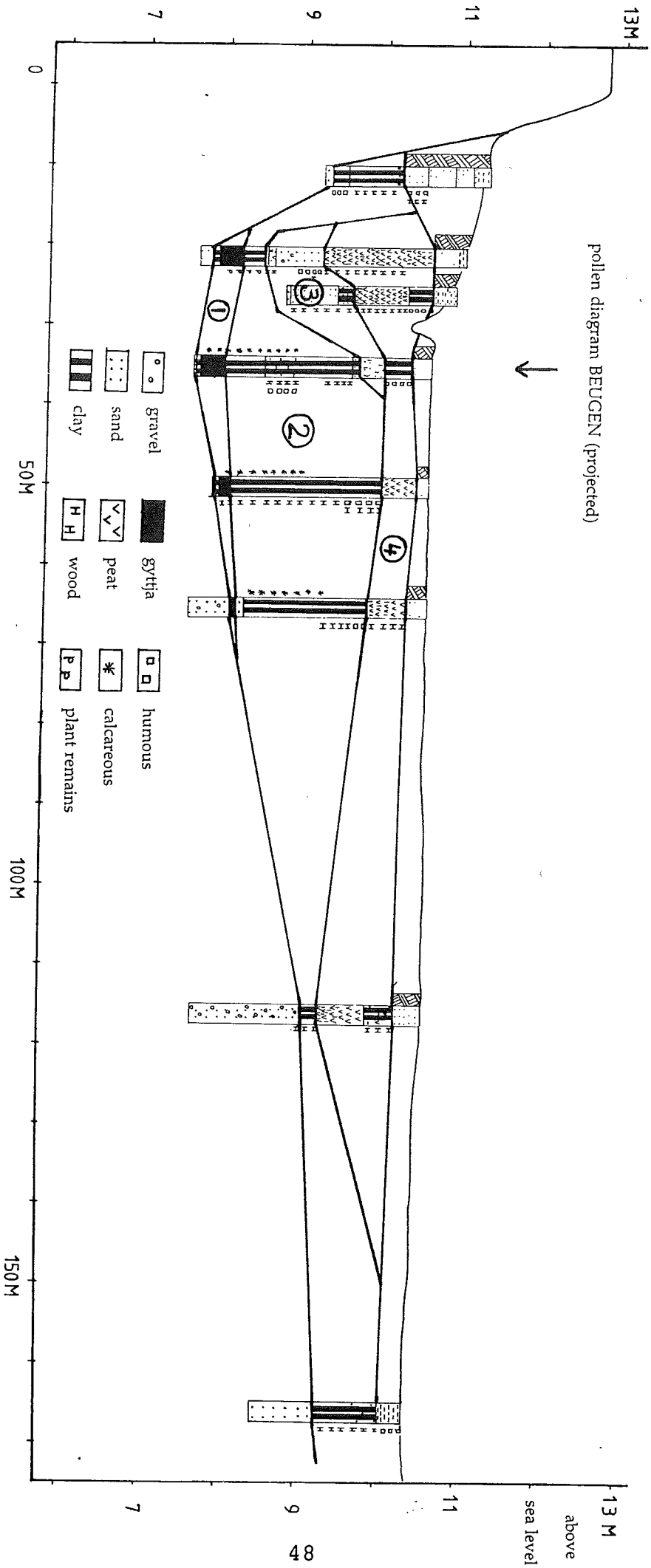Fig. 3   Morphological map of the Allerød meander cut off at Beugen (after Buitenhuis & Wolfert, 1988).

46

Fig. 4   Cross section over the incised Allerød meander at Beugen.
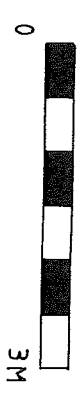
47

Fig. 5   Detailed cross section over the Late Glacial and Holocene channel fill at Beugen.

The pollenrecord

Preliminary results of the analyses are presented in fig. 6.

Local pollenzone BGN-1 (333-325 cm):

The basal spectra of the infill show part of the *Betula*-phase of the Allerød.

Local pollenzone BGN-2 (325-300 cm):

At 321 cm *Pinus* starts to rise. A date for this event at Bosscherheide (excursion stop 2) comes to 11,300 BP. For reference purposes the pollen record of Bosscherheide (a selection of taxa only) and the relevant C-14 dates have been given in fig. 7.

The *Pinus*-phase of the Allerød is relatively short and it cannot be excluded that during the overlying Late Dryas, part of the Allerød has been truncated. Lithologically a sharp transition from laminated gyttja to slightly sandy clay has been observed at 295 cm depth.

Local pollenzone BGN-3 (300-275 cm):

At 296 cm the *Pinus* value drops, the NAP pollen rises, while the pollen of tree birches show an increase. This short lasting birch phase at the beginning of the Late Dryas is a repetative feature in many diagrams of the Maas. At Bosscherheide remnants of birch trees have been revealed from the Late Dryas organic deposits in the back-swamp, that were formed shortly after 10,880 BP.

Local pollenzone BGN-4 (275-185 cm):

From 275 cm depth the landscape gradually became more open and grasses, sedges and *Artemisia* together with dwarf birches and dwarf willows and some *Dryas, Helianthemum* and *Thalictrum* dominate the picture. The relatively high values for *Classo-pollis*, a pollen-type derived from the Lower Cretaceous Beds in the upper course of the Maas, clearly demonstrates that the ox-bow lake was again reached by the then active Maas river. Probably during periods of peak discharge these old streambeds were carrying water and deposition of suspension load took place.

A transition to the early Holocene has not been reached in the analyses and is probably absent due to late Holocene brook incision in the ox-bow lake.

References

Buitenhuis, A. & H.P. Wolfert 1988 Toelichting op kaartblad 46 Gennep. Stichting voor Bodemkartering, Wageningen; Rijks Geol. Dienst, Haarlem, 27pp.

Teunissen, D. 1990 Palynologisch onderzoek in het oostelijk rivierengebied; een overzicht. Meded. Afd. Biogeologie, No. 16, Katholieke Univ., Nijmegen, 162pp.

Fig. 6 Pollenrecord (selection of curves) from the infill of the ox-bow·lake at Beugen.
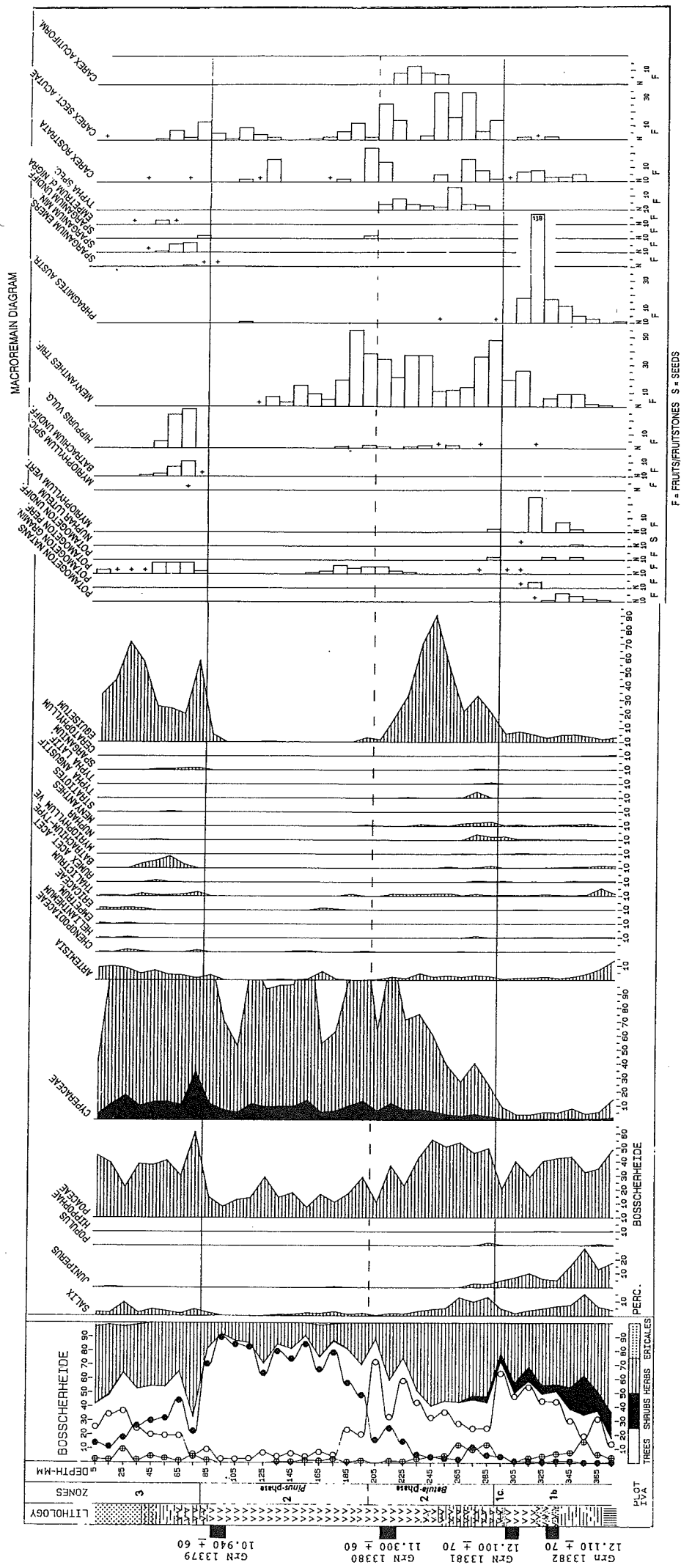
50

Fig. 7   Pollen and macrobotanical record (selection of Taxa) from the infill of an abandoned gully at Bosscherheide. Bosscherheide is situated on the Pleniglacial terrace on the east bank of the Maas (see fig 3. p. 33).

EXCURSION SITE 2

THE LOW-SINUOSITY CHANNEL AT WANSSUM

This excursion point is situated on the left bank of the Maas at 18 m above sea level (fig. 1). The morphology is characterized by low-sinuosity channels, with a northwestern stream direction. This channel zone separates the Late Pleniglacial terrace in the west from the Late Glacial terraces in the east.

This low-sinuosity system represents a transitional morphological stage between the Late Pleniglacial multi-channel braided system (see excursion stop 2 Bosscherheide) and the Late Glacial (Allerød) high-sinuosity system (see excursion stop 1 Beugen). Since the braided system at Bosscherheide was abandoned already during or before the Bølling (start of peat formation according to Teunissen, 1983: 12,700 ± 150 BP) and the high-sinuosity system was functioning during the Allerød, it is deduced that this transitional phase was active during the Bølling and/or Older Dryas (i.e. between 12,700 and 11,800). Direct dating of the system itself (by dating of the base of the infilling) is problematic. Because the channel is rather shallow peat formation did probably not begin immediately after channel abandonment (hiatus).

Teunissen, D. 1983 The development of the landscape of nature reserve De Hamert and its environs in the northern part of the province of Limburg, The Netherlands Geol. Mijnbouw 62: 569-576.
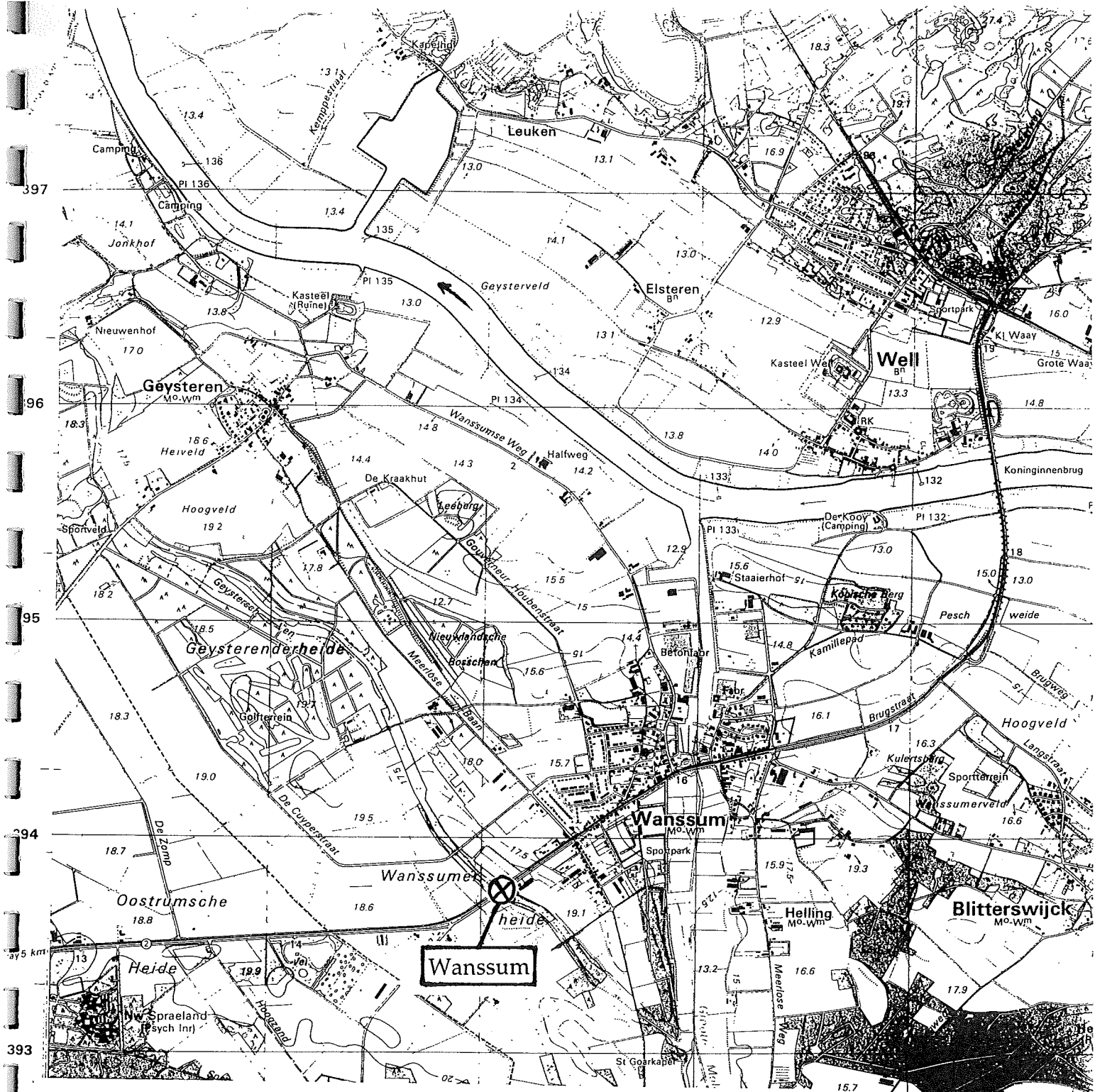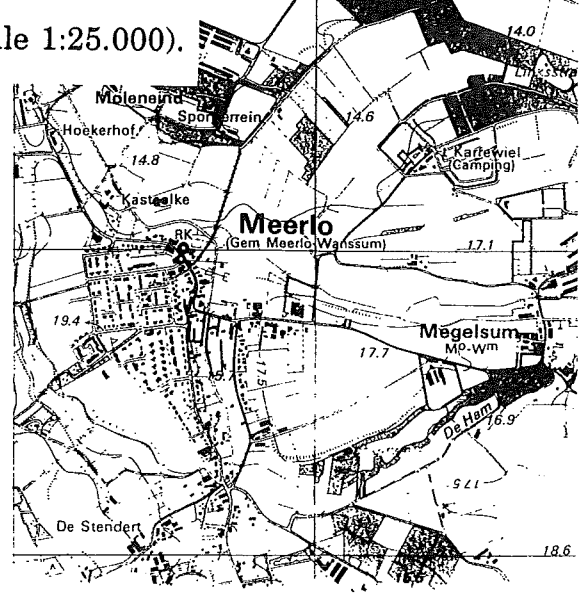
Fig. 1   Location map of excursion stop Wanssum (scale 1:25.000).

EXCURSION SITE 3

THE MEANDER SCAR AT MEERLO

The meander scar at Meerlo is situated at the same Late Glacial terrace as the one at Beugen. The meander-scar at Meerlo differs from the one at Beugen in that it has been abandoned at the start of the Younger Dryas, possibly by a so called chute cut off. The meander has been abandoned due to a rather rapid change in the fluvial regime at the Allerød/Youger Dryas transition. Peak discharges, possibly due to snow melt in late spring, led to a strong modification of the river channels and the establishment of a new balance between channel morphology and climate. A more straight and wide (multi-) channel floodplain was required for the large peak discharges that characterized the climate at the time of the chute cut-off.

During the higher YD peak discharges the abandoned Allerød channel was still reached and here deposition of fines from suspension load took place (see lithology, pollenrecord fig.1).

The silty/sandy clay at the bottom of the meander infill reflects this episode of peak discharges straight after the cut-off.

The pollenassemblage shows relatively high values for the Non Arboreal taxa, among which the grasses and the sedges are dominant. Furthermore the relatively high values for *Artemisia* and the continous presence of *Empetrum* are indicative for the Younger Dryas chronozone. The presence of *Classopollis*, a reworked taxa from the Lower Cretaceous in the upper course of the Maas, characterizes the part of the infill during which the straight and wide (multi-) channel floodplain of the Younger Dryas was active.

Resuming, the Allerød large meandering system was abruptly abandoned as a result of a swift morphodynamic adaptation of the fluvial regime, which in turn was forced by a dramatic climate change. This change in climate involved an increase in precipitation and/or a more seasonally determined precipitation in the form of snow.

At the end of the Younger Dryas a change in lithology (humic silty clay - gyttja) coincides with a distinct lacustrine depositional environment with high values for aquatic taxa in the Preboreal. The pollenrecord shows a nice succession with *Juniperus* and *Betula nana*, followed by tree birch and pine. The increase in *Pinus* in the Maas area has been dated to 9,705 ± 50 BP. The local hydroseral succession at this time reached the semi-terrestrial phase, leading to the deposition of peat. At the top of the pollenrecord the increase in autochthonous *Corylus* pollen determines the transition to the Boreal (ca. 9100 BP).

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Fig. 1. Pollenrecord (selection of Taxa) from the infill of the ox-bow lake at Meerlo.

EXCURSION SITE 4

THE YOUNGER DRYAS FLUVIAL SYSTEM AT KASTEELWEG-OOIJEN

At this excursion point a 1,5 km walk will be made, in which we cross the Younger Dryas and Holocene floodplain. Kasteelweg-Ooijen is located on the western bank of the Maas, 5 km southeast of Wanssum (stop 3). The topography is between 14 (channels) and 20 meters (river dunes) above NAP (fig. 1).

A generalised cross section is given in figure 2. Four physiographic units are distinguised:

1. The high sinuosity meander scar terrace in the west. It consists of gravelly sands fining upward into sand and it dates from the Allerød or early Late Glacial.
2. The Younger Dryas channel system, underlain by gravelly sands. Local organic channel fill dates from the late Younger Dryas (fig. 3: pollen diagram Kasteelweg).
3. The Younger Dryas channel and bar system, partly covered by river dunes, and resting on 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 (fig. 4: pollen diagram Ooijen).

In this region the Younger Dryas terrace is very well developed; it is approximately 1 km wide and characterized by a braided floodplain with straight to low sinuosity channels (fig. 1 and 2). The rather straight floodplain was formed by chute cut off of the high-sinuosity Allerød meanders. This is concluded from a nearby meander (Meerlo, 2.5 km west of this excursion point) which was abandoned and filled with sandy clay during the Younger Dryas. The change in channel morphology from high sinuous to braided probably occurred during the start of the Younger Dryas, because the temperature decrease led to lower evapotranspiration and higher peak discharges. The morphological change was accompanied by erosion of the high-sinuosity meander terrace (see fig. 2).

The first stop is in a 500 m broad, but shallow, Younger Dryas channel (fig. 2: Kasteelweg). Locally, a 2 m thick organic fill was found in scours within the channel. Palynological research (see discussion on pollen diagram Kasteelweg, fig.3) indicates that the base of the infilling (= moment of abandonment of the channel) dates from the end of the Younger Dryas. This age proves that the channel was in use during the Younger Dryas itself.

Walking towards the northeast, physiographic unit 3 is crossed. It is a higher area of straight gravelly channels and inter channel bars, covered by eolian sand. There is no organic material in these higher lying channels, but on morphological grounds, this channel-bar system is correlated with 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. Corings in unit 3 revealed the existence of older Late Glacial meander sediments in the subsurface.

Fig. 1   Location map of the Kasteelweg-Ooijen region (scale 1:25.000).

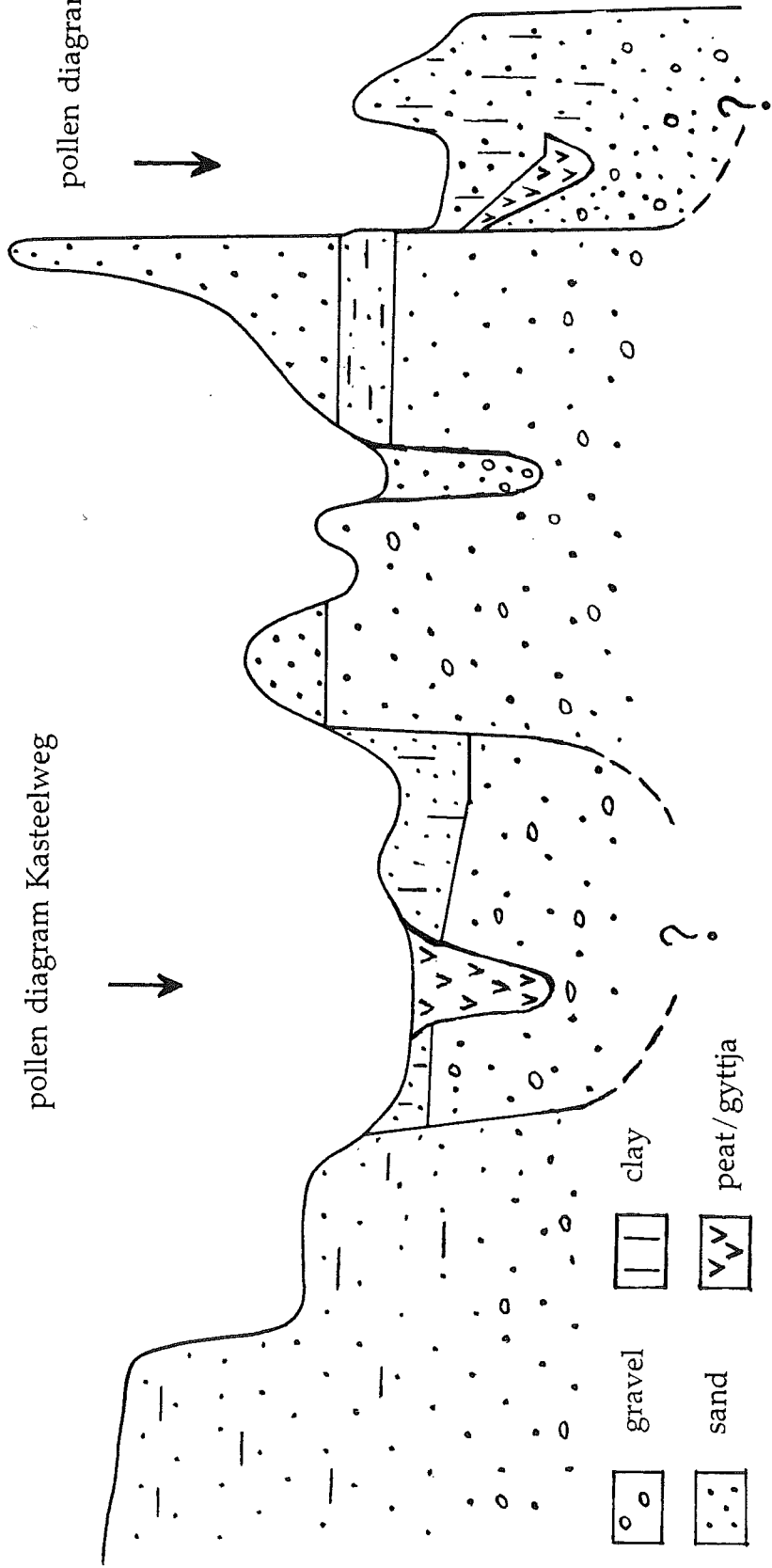Fig. 2 Schematic cross section and physiographic units of the Younger Dryas and Holocene floodplain.

The high topographic position of unit 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 unit 3, gradually incised towards their position in unit 2 and perhaps also unit 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 in front of the Younger Dryas dune field. After channel abandonment it was filled with up to 3.5 m of clay and coarse detrital gyttja (fig. 4: pollen diagram Ooijen). Several detailed cross-sections were made over the channel. Generally, the fill is 3 m thick and registration started in the Atlantic. Only very locally the fill is thicker (3.5 m) and registration started already during the late Boreal, which stresses the importance of a detailed survey. The palynology of the fill is discussed below (fig. 4). Since the base of the fill dates from the late Boreal it is clear that the erosion of the Younger Dryas dunes 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. 2: Ooijen) is situated at almost the same level as, or slightly lower than, the base of the Younger Dryas channel (fig. 2: Kasteelweg). This possibly indicates that at the end of the Younger Dryas two channel systems existed, west and east of physiographic unit 3. The climatic transition towards the Preboreal 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 coarse at the start of the Holocene.

The pollenrecord at Kasteelweg (fig. 3)

Organogenic deposits in the straight braided channels of Younger Dryas age are rare, but within the limits of the former wide stream-bed some depressions have been preserved. At Kasteelweg a black gyttja going into a greenish gyttja was found underlying the organic sediments that were more frequently encountered. At the transition between these two lithological units a 1 mm thin sand band and a thin grey clay layer are present (see lithological column fig. 3).

Fig. 3   Pollenrecord (selection of curves only) from the infill of the Younger
Dryas straight, braided channels at Kasteelweg.

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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. Since we are on the Younger Dryas terrace and the overall picture of this zone is one that is dominated by the NAP, the pollenrecord is thought to demonstrate the termination of the Late Dryas biozone. 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. The presence of *Populus* at this level in the sequence may possibly indicate that the vegetational record can be placed in the **Rammelbeek-phase** of the **Preboreal.**

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, indicative for the spread of *Pinus* in the region of the Maas valley, is dated to 9.700 BP in a meander scar due south of this location (Broekhuizen). *Corylus* only shows a slight increase in the uppermost samples of the analyzed core segment and indicates the end of the **Preboreal** period (ca 9150 BP).

Resuming, the pollenrecord shows the termination of the **Late Dryas**, a hiatus towards the early Holocene and the later part of the **Preboreal**. At the start of the **Preboreal** (ca 10.200 BP) or shortly after and before 9.700 BP, river erosion and subsequent deposition of clay from suspension load took place originating from the than active early Holocene river system.

The Holocene flood plain at Ooijen

Two cores were taken in the early Holocene river-plain at Ooijen in order to establish the period during which the channel became abandoned and started to fill in with organogenic sediments (fig. 4 and 5).

The pollenrecord from core 91-12 at Ooijen:

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 coarse detrital clay 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** period from ca 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, ca 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. Further research may elucidate this problem.

Resuming, the biostratigraphycal indications as provided by the pollen records from core 11 and 12 demonstrates that the infilling of the channel at Ooijen started shortly after the **Preboreal/Boreal** transition. Accumulation of clayey detrital gyttja continued over the **Boreal** and **early Atlantic** period and probably even longer, but there are faint indication for an erosional hiatus in the sequence.
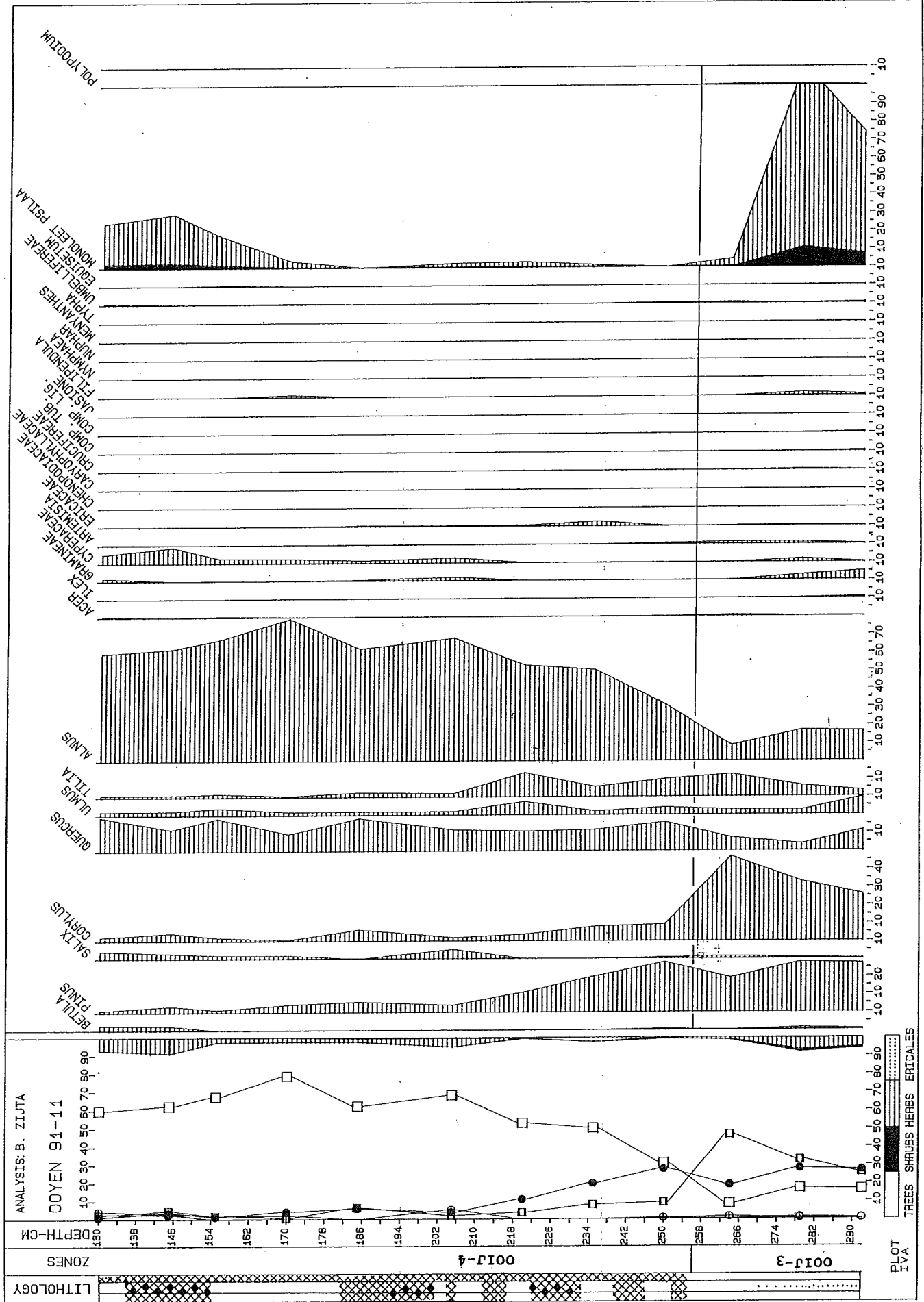
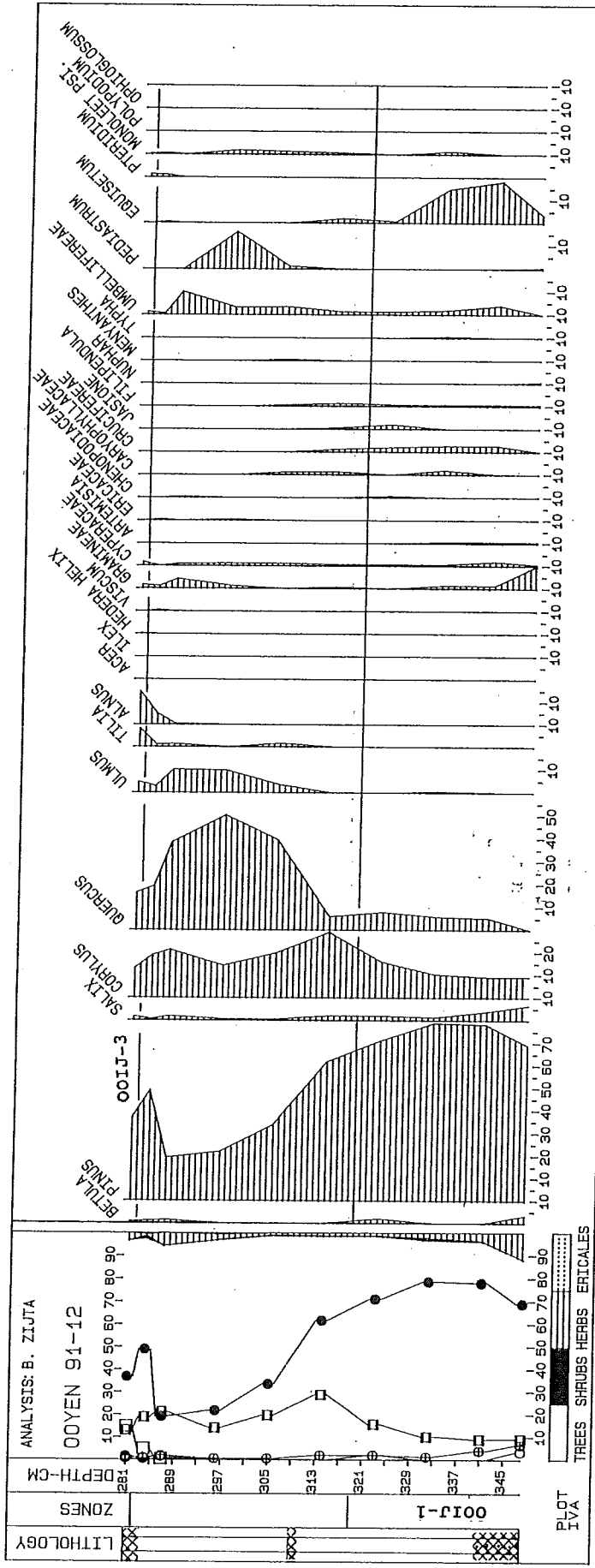Fig. 5   Pollenrecord (selection of curves only) from core 91-11 taken in the Holocene river plain at Ooijen.

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Fig. 4   Pollenrecord (selection of curves only) from core 91-12 taken in the Holocene river plain at Ooijen.

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EXCURSION SITE 5

VIEWPOINT MAASHEES

At this site the morphology of a terrace edge is to be seen, separating the early Late Glacial terrace from the Younger Dryas terrace (fig. 1). The edge is fairly high (4 m) and well developed because of the absence of the intermediate Allerød meander scar terrace. The Younger Dryas paleofloodplain is narrow here, since it is situated at the erosional outer bend of the present-day Maas.
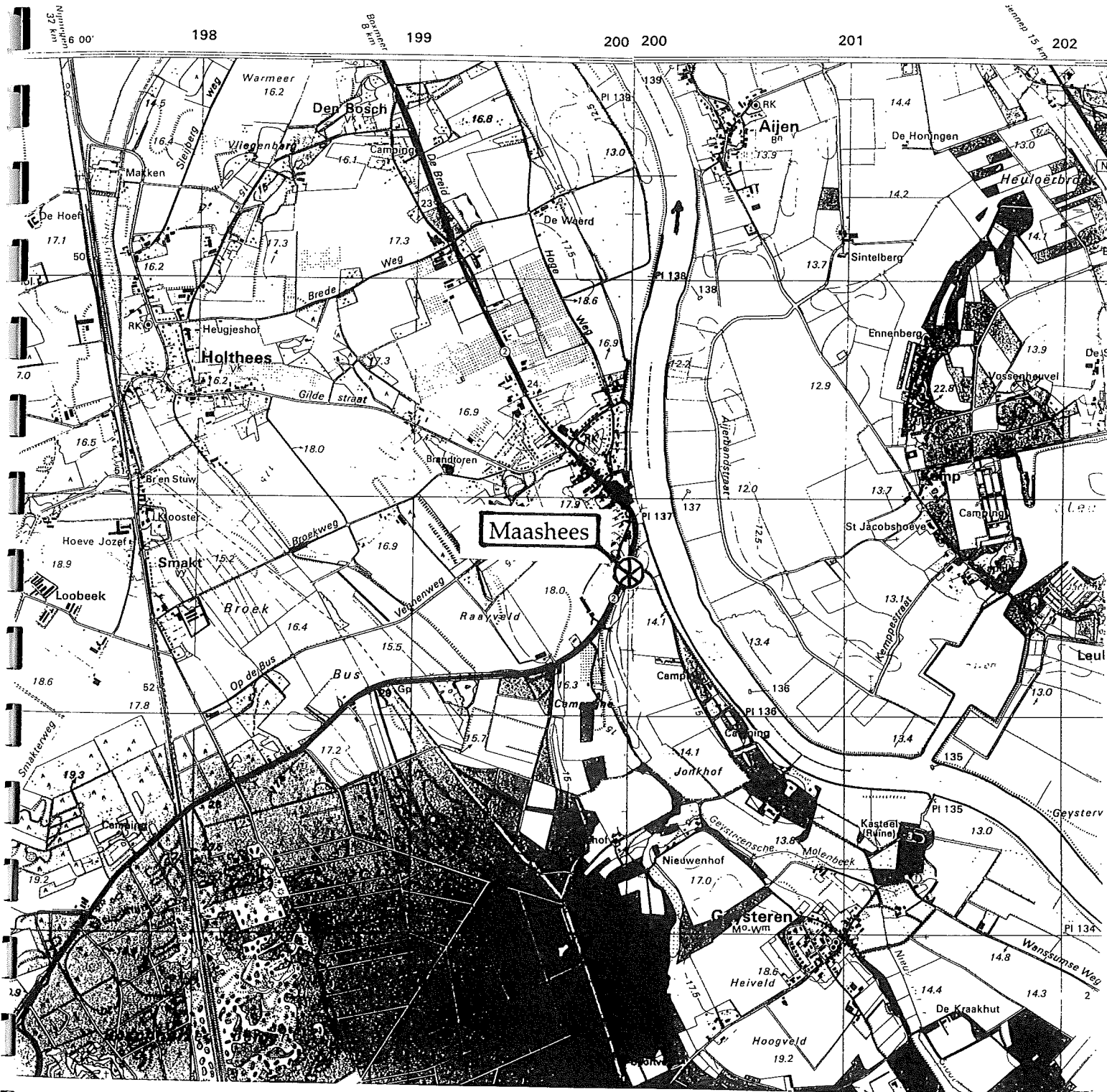

Fig. 1   Location map of excursion stop 5 (scale 1:25.000).