

Quaternary environments, climate and man in the Netherlands

Field Guide

Edited by C. Kasse

Quaternary Research Association Annual Field Meeting, The
Netherlands, April 6-9th 1995, Vrije Universiteit, Amsterdam



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Pliocene and Lower-Pleistocene deposits in the Tegelen-Reuver area (SE-Netherlands)

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General introduction

The Tegelen-Reuver area is situated in the western part of the Lower Rhine embayment, a tectonical subsiding area which is bounded by the Palaeozoic rocks of the Eifel/Ardennes massive. The excursion area is located on the so-called Venloër Scholle, an area strongly dissected by SE-NW running faults and bounded by the major fault lines of the Peelrand Fault and the Viersen Fault (Fig. 1). The tectonic history of the area is complex. Each single block shows considerable differences in tectonic movement with respect to the adjacent ones. For example, just east of the Peelrand Fault Miocene marine deposits are found near to the surface where they are covered by a few meters of Mid-Pleistocene fluvial deposits. On the contrary, thick sequences of Pliocene and Lower-Pleistocene fluvial deposits occur on some tectonic blocks just north of Venlo.

Periods of high tectonic activity in the area are correlated with the major uplift phases of the Eifel/Ardennes massive during the Late-Tertiary and Mid-Pleistocene. Forced by tectonic movements the river Meuse shifted gradually eastward from a position in the Central Graben to its present course. This process started during the Cromerian when Rhine and Meuse followed their own separate courses. The eastward shift of the Meuse ended with the development of a terrace escarpment along the Dutch-German border (Figs. 2 and 3). This escarpment forms a well-pronounced morphological feature near the villages of Tegelen and Reuver. The escarpment is formed in fluvial Rhine/Meuse deposits of Early-Cromerian age (Fig. 3). The exact age of the escarpment is not known, but it seems likely that it is formed during the maximum inland ice stage during the Saalian. On the low side of the escarpment surface deposits date from the Upper-Pleistocene (Weichselian).

In Table I and figure 5 a comprehensive overview is given of the lithostratigraphical sequence underneath the terrace.

In the terrace escarpment but also higher up on the terrace, a series of active and abandoned excavations can be found which are mainly developed for clay digging. It is generally believed that the name Tegelen is derived from the latin word 'Tiglia'. From archeological studies it is known that already during Roman Times clay digging has taken place amongst other for the production of roofpans and bricks. Originally, the exploitation of the clay started in small excavations along the terrace escarpment (Fig. 3). Later on, one also started clay digging on the higher part of the terrace. Large-scale excavation began around the end of the last century and culminated during the fifties. After about 1960 many excavations were abandoned but since the mid-eighties an increase of clay- and sand-mining activities has taken place. Nowadays the exploitation is concentrated in a few large excavations.

Table I: Lithostratigraphic sequence in the High Terrace east of Tegelen.

Twente Formation or Nuenen Group	Aeolian and periglacial deposits. < 2 m	Upper-Pleistocene	coversand and löss-like deposits, locally reworked by wind during the Holocene.
Sterksel Formation	Coarse-grained fluviatile sand and gravel, locally with clay-layers. ± 20 m	early Middle-Pleistocene (Cromerian)	heavy minerals: Instable association dominated by epidote (Weert zone). Erosional lower boundary.
Kedichem Formation	Fluviatile and periglacial medium grained sandy deposits. Locally soil horizons, cryoturbations and frost wedges. 0 - 5 m	late Early-Pleistocene	heavy minerals: stable association. Only locally present as erosional remnants.
Tegelen Formation	Fluviatile sand and gravel with extended and often thick clay-layers. 10 - 25 m	Early-Pleistocene	heavy minerals: instable associations, often with considerable garnet content.
Kiezeloöliet Formation	Fluviatile sands with thick clay-layers and intercalated brown coal seams on top. 10 - 60 m	Pliocene	heavy minerals: stable associations. The finer top part is dominated by Zircon. The uppermost part shows a gradual change into instable mineral associations.
Breda Formation	fine-grained glauconite bearing marine sands and clay. > 100 m	Miocene	heavy minerals: dominated by stable associations but generally 20 - 30% instable minerals.

Geological and Palaeobotanical research

Since the beginning of the 20th century, geologists have shown interest in the clay-pits near Tegelen and Reuver (southeast of Venlo). The earliest investigations mainly concerned the study of the mammal faunas and the macro-remains of the flora. It was Eugene Dubois who carried out studies on the remains of (mainly) the larger mammals. The former director of the Geological Survey, Dr. P. Tesch, published a monograph on the Pliocene fluviatile sediments in this region (Tesch, 1908). In order to enlarge the knowledge about the Late-Tertiary and Early-Pleistocene he emphasized the importance of palaeobotanical research on the fossil

plant remains found in the several clay layers of this area. Therefore, he invited the British researchers E. & CL. Reid to describe these fossil floras. As a result of their studies Reid and Reid (1907, 1908, 1910, 1915) discerned two different flora assemblages, namely:

- 1 - Reuver Flora: with species like *Sequoia*, *Taxodium*, *Nyssa*, *Aesculus*, *Liquidambar* and *Fagus*.
- 2 - Tegelen Flora: with low percentages of the so-called Tertiary relicts and *Tsuga*, *Pterocarya*, *Carya*, *Eucommia*, *Phellodendron* and *Actinida*.

Already Reid and Reid thought that the decrease in number of thermophilous species, which form the differences between the two floras, were caused by a climatic change. They sought the explanation in a gradual cooling which took place from the Tertiary into the Pleistocene. The disappearance of several species could be caused by the existence of the W-E orientated alpine mountain belts in south and central Europe.

In the thirties Florschütz started pollenanalytical studies of the clay and brown coal layers in the same area. The results confirmed the assumed relationship between the palaeovegetations and climatic changes. Florschütz, however, suggested that the deterioration of the climate at the Plio/Pleistocene transition could be connected with a cold period (the prae-Tiglien). He rejected the idea of a gradual change and thought that a more sudden deterioration of the climate had taken place. In his thesis Zagwijn (1960) published a pollendiagram of the Meinweg section which shows a clear decrease in thermophilous plants and an increase of herbs. This phase was assigned by Zagwijn to represent the Pretigian which is generally seen as the onset of the Quaternary.

Contemporarily sediment-petrological studies were carried out in order to characterize the fluviatile sandy sediments according to their heavy mineral composition. This produced more information about the provenance of the sediments and the changing patterns of the rivers.

The first firm combination of the results from the palaeobotanical and sediment-petrological investigations was made by Zagwijn. At the end, this has resulted in the presently still used system for the Quaternary lithostratigraphy in The Netherlands (Zagwijn & Van Staaldin, 1975; Fig. 4).

During the excursion, Pliocene as well as Pleistocene deposits will be shown. Some information on the most important sites is given. A precise scheme of the pits that will be visited can only be given a few days before the excursion takes place, because of the varying conditions in the exposures.

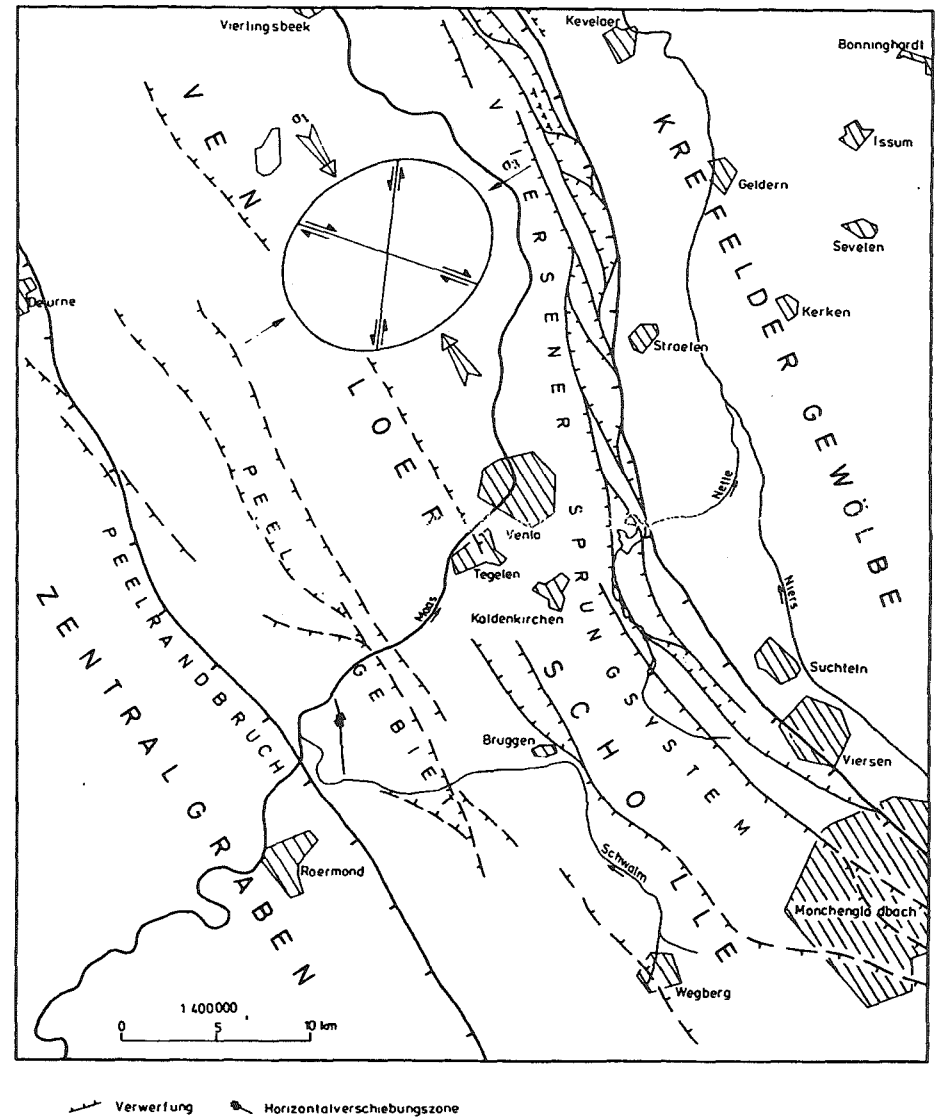
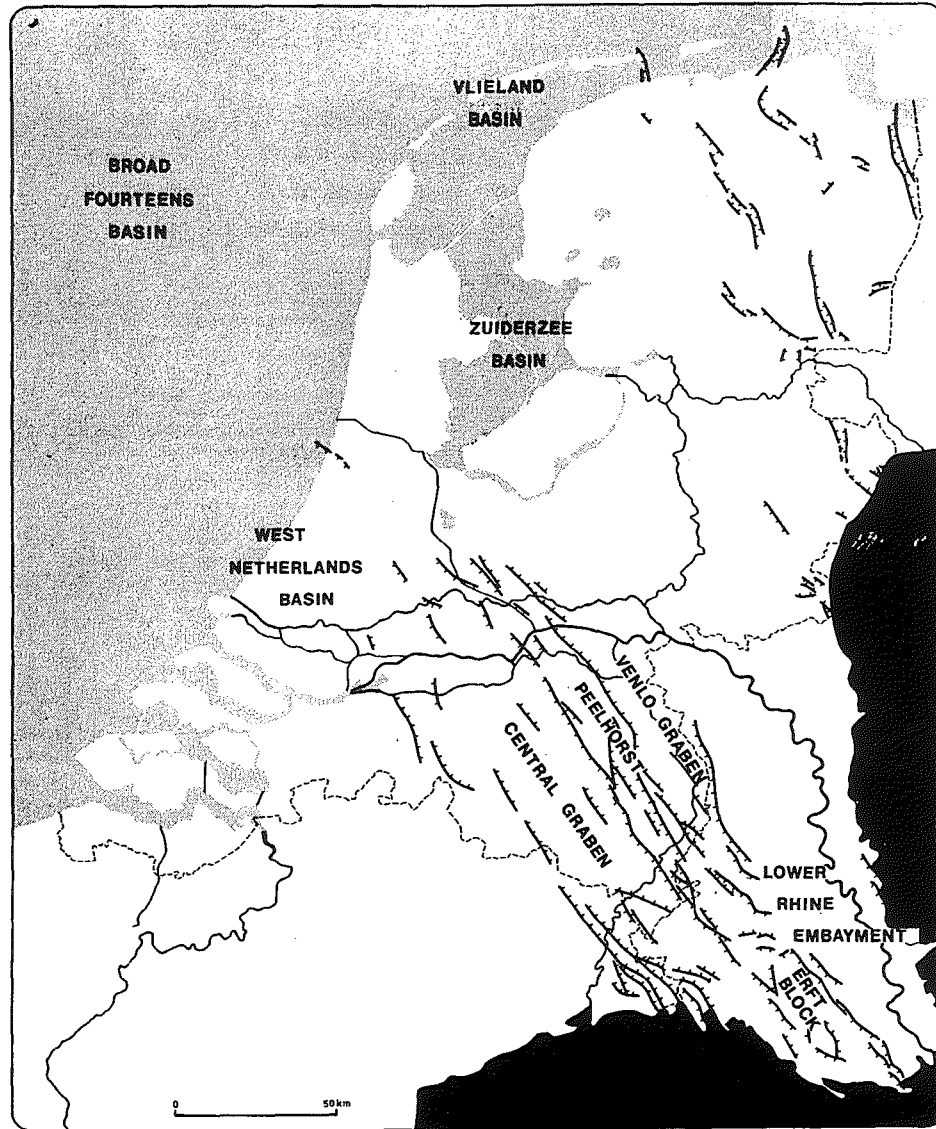


Figure 1: Tectonic framework Lower Rhine Embayment; A) General overview according to Zagwijn & Doppert, 1978; B) details of the Venlo area (Klostermann, 1983).

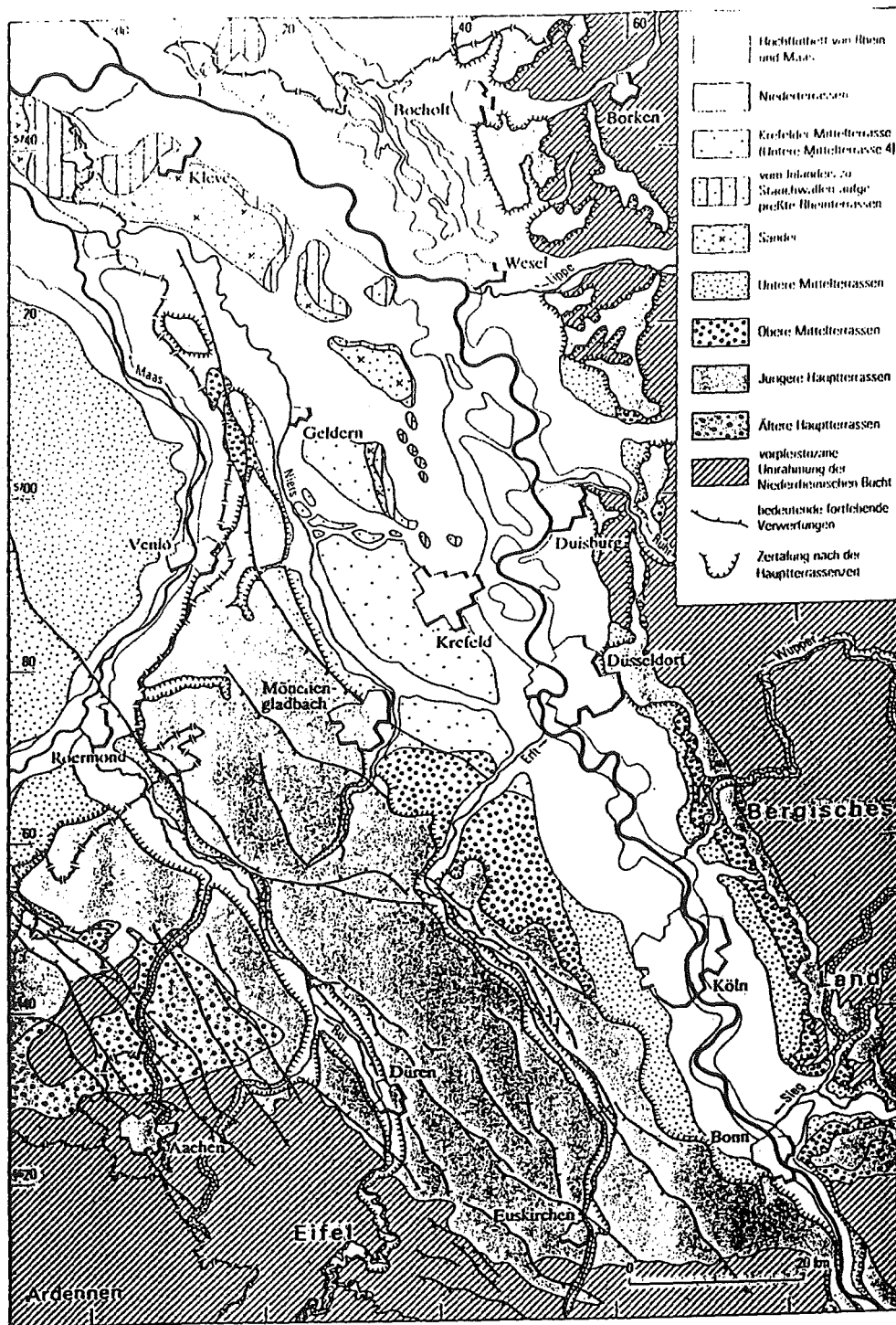
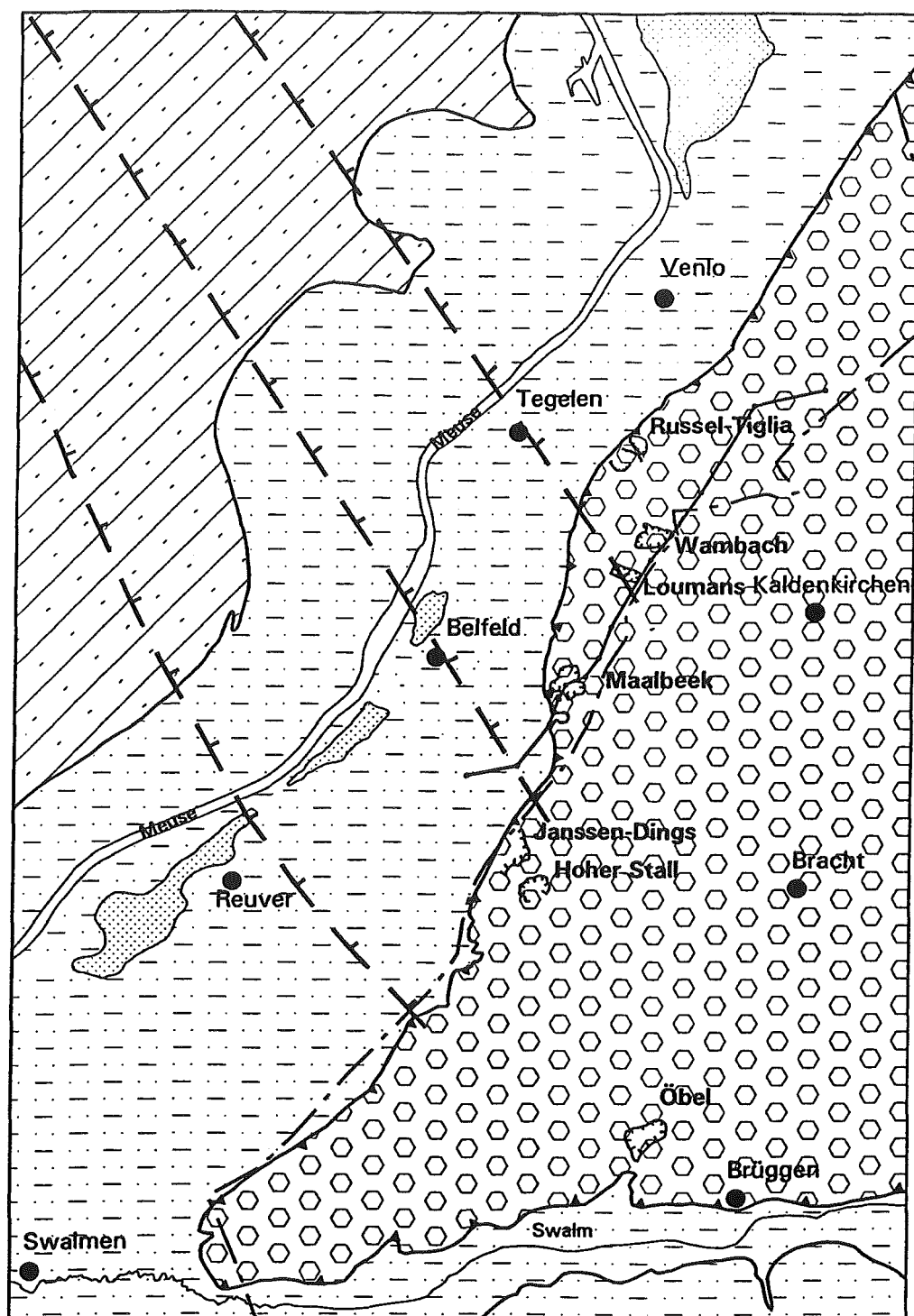





Figure 2: Terrace morphology of the Lower Rhine area (according Braun & Quitzow, 1961; published in Klostermann, 1988).



Fluviatile deposits of the Meuse

-  Upper - Pleistocene + Holocene
-  Middle-upper Pleistocene covered by Late Weichselian eolian deposits
-  riverdunes (Late Weichselian)

Rhine + Meuse deposits

-  Middle Pleistocene (Cromerian)






-  fault line
-  terrace escarpment
-  cross-section
-  excavation
-  state boundary

Figure 3: Simplified geological/geomorphological map of the Tegelen-Reuver area, with location of the most important excavations.

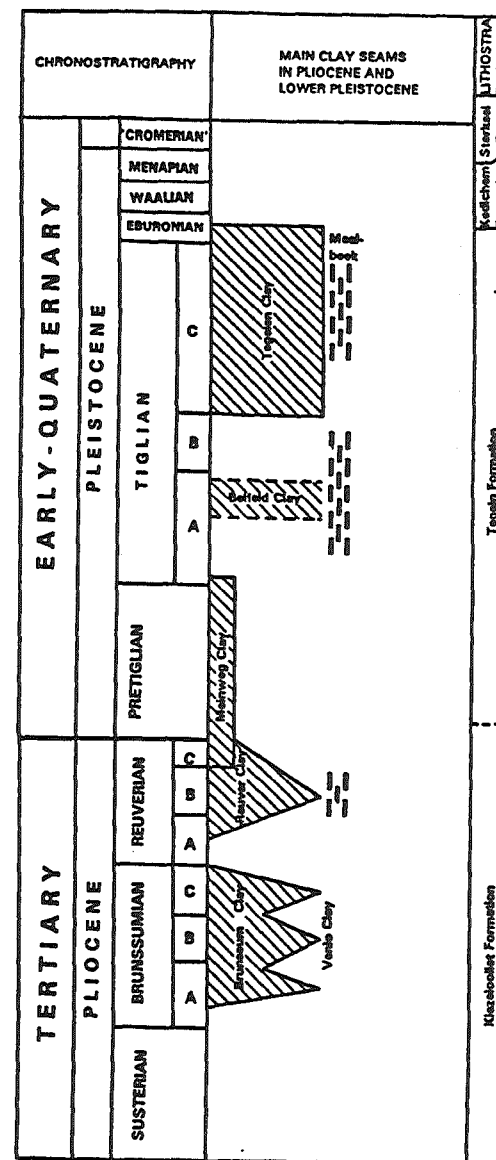
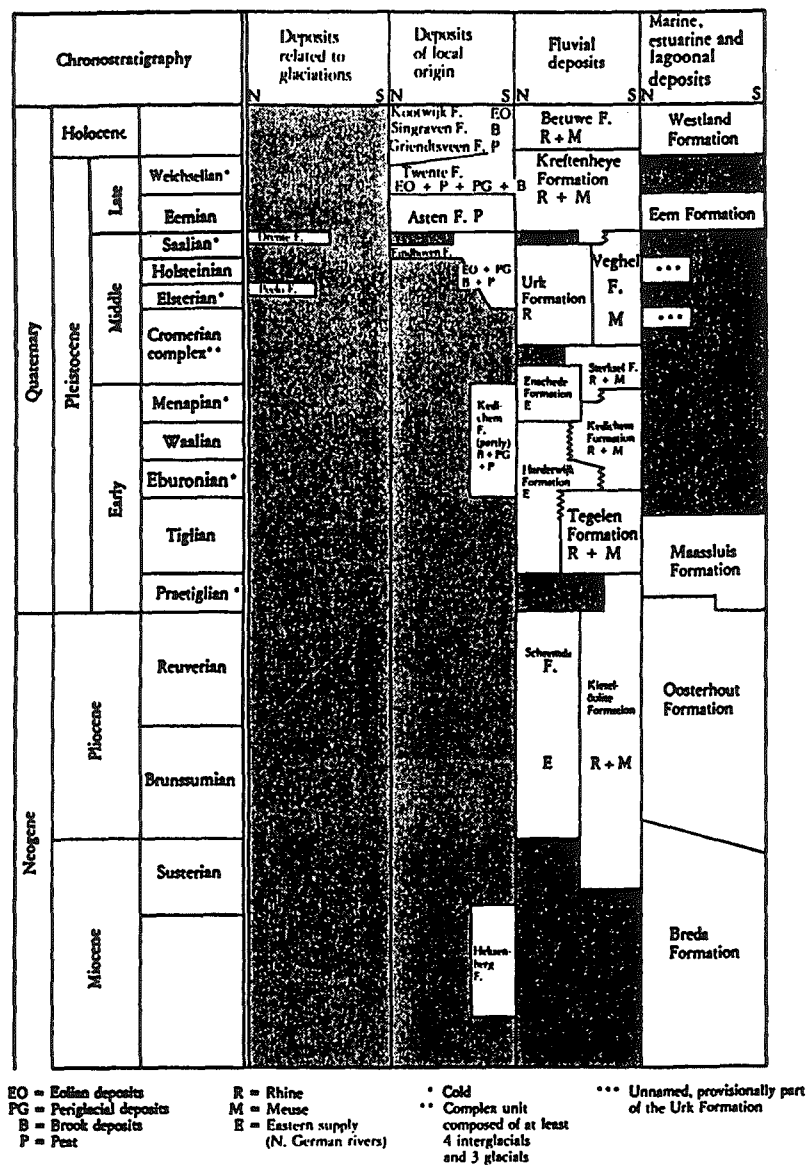


Figure 4: General Lithostratigraphy for Upper-Tertiary and Quaternary deposits in the Netherlands (Zagwijn & van Staaldunin, 1975)

Local Stratigraphy of the Maalbeek pit, thick hatches indicate the pollenanalytical dated clay seams in Maalbeek.

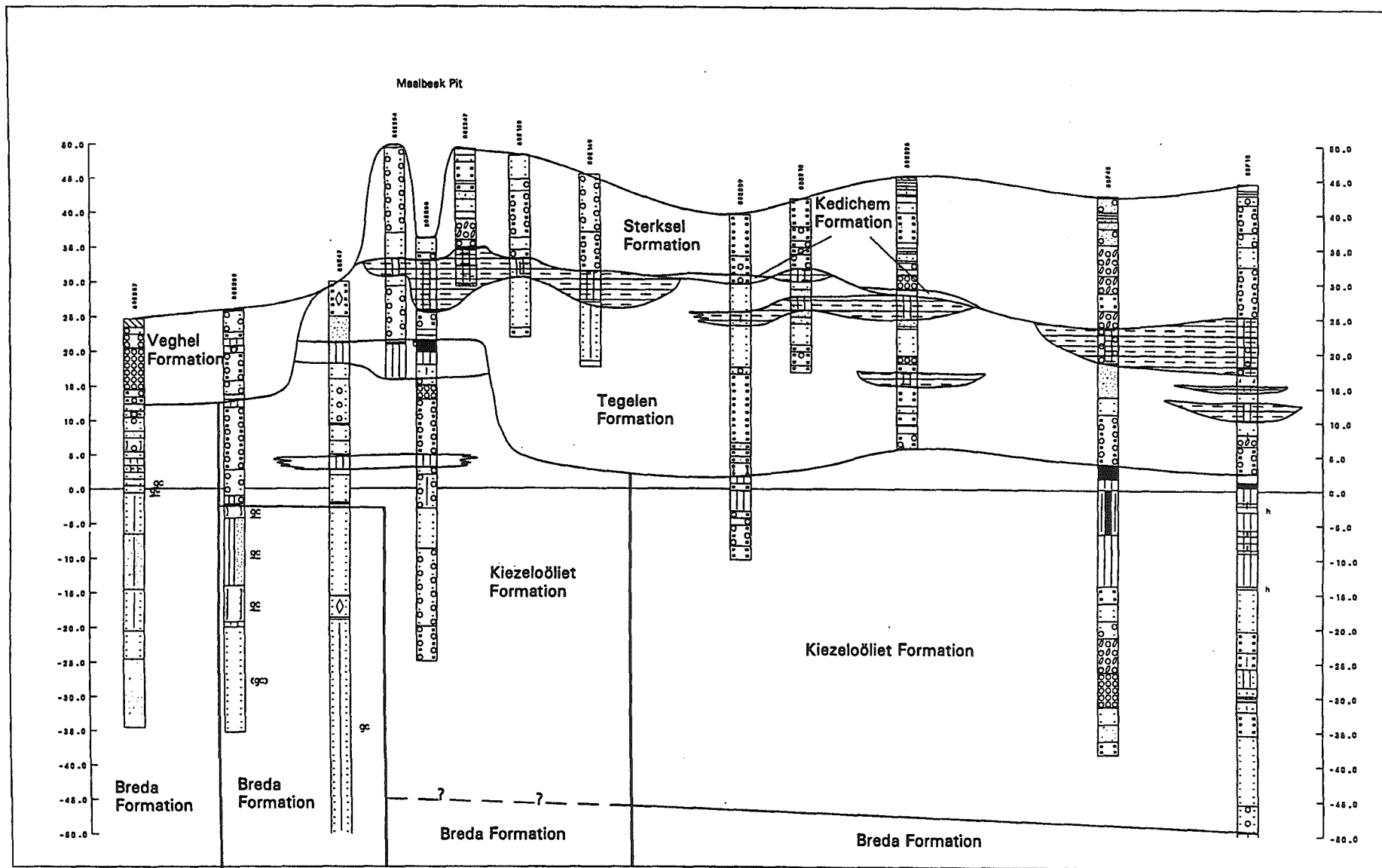


Figure 5: Geological cross-section along the state boundary near Tegelen (for location see fig.3)

The Öbel pit (Fig. 6)

The basal clay-layer in this pit is characteristic of the Reuver clay in the type area. It is a compact grey-blue coloured clay with two browncoal horizons in its top part. The clay contains many rootlets, leaves and other organic material. Concentrations and sometimes nodules of carbonate, siderite and sulphate bearing compounds are common features. In some place thin lines with pyrite are found. Pollenanalytical studies show that the clay is deposited during the Reuverian-B pollenzone. In the brown-coal seams conspicuous high peaks of *Ulmus* are noticed.

With an erosional contact, the Reuver-clay is overlain by a fluviatile sequence dating from the Early-Pleistocene. These sediments belong to the Tegelen Formation. In Öbel the basal part of it consists of whitish, quarzrich gravels grading into sands which are covered by fine silts and clay. A small channel filled in with 3 - 5 m of clay can be found in the southwestern corner of the pit. A peat-layer, with a lot of wood fragments, is developed in the uppermost part of the channel fill. The pollendiagram of the channel fill points to a Tiglian age (pollenzone TC3 or TC5).

The Tegelen deposits are unconformably overlain by Middle-Pleistocene coarse-grained fluviatile sediments of the Sterksel Formation.

The heavy mineral composition of the exposed deposits shows many interesting aspects that are characteristic of the Upper-Pliocene and Lower-Pleistocene sediments in this area. In general the Pliocene sediments (Kiezeloöliet Formation) are characterized by a dominance of stable heavy mineral associations. On the contrary, the deposits of the Tegelen Formation are dominated by instable associations.

Formerly it was believed that this change marked the Plio/Pleistocene boundary. But according to the investigations of Boenigk (1970) this change already took place in the upper part of the Reuverian. As is shown in figure 6 the increase of a *garnet-epidote-alterite / saussurite-hornblende* association starts in the uppermost part of the Reuver-clay. Higher up in the sequence influxes of stable heavy minerals may occur. This has been observed on several locations in the area. An explanation for these phenomena will be discussed during the excursion.

ÖBEL

LEGEND

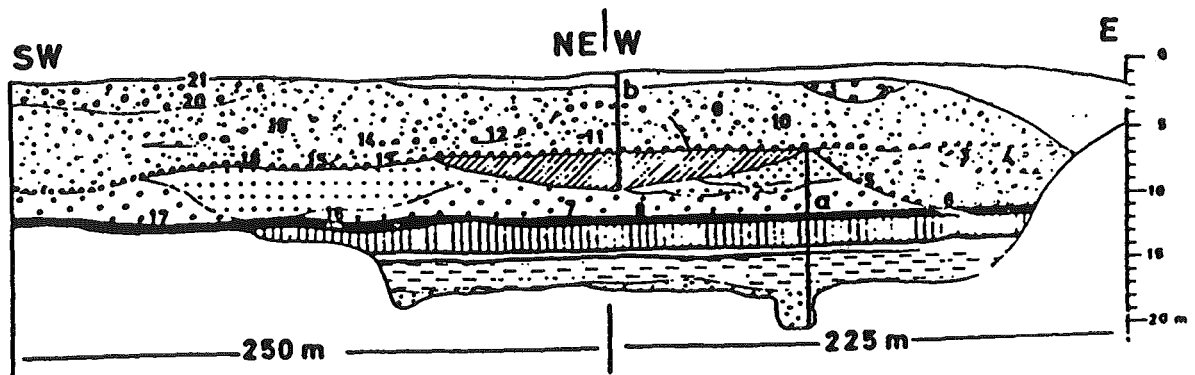
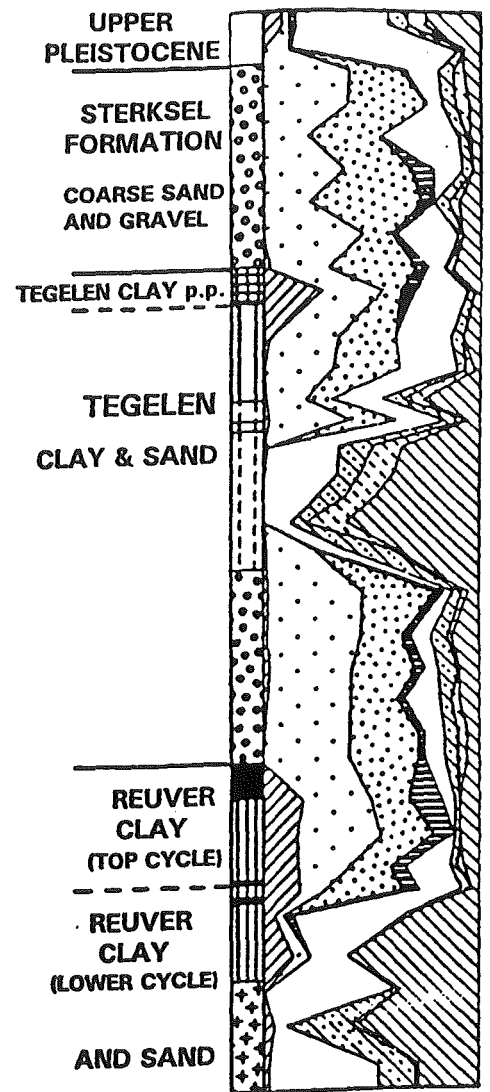
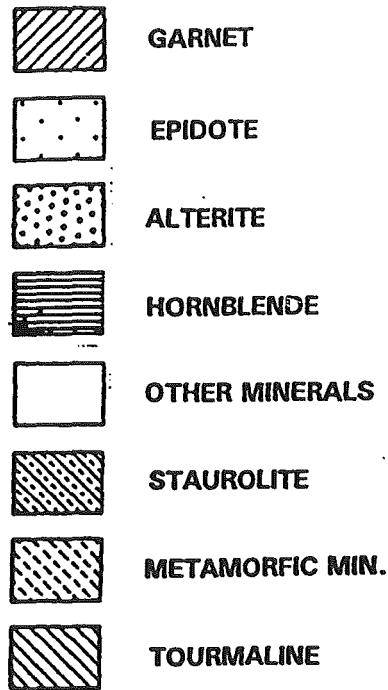


Figure 6: Section and heavy mineral diagram of the Öbel pit (Boenigk, 1970)

Hoher Stahl (pit van Eijck) (Fig. 7)

In recent years a number of smaller pits (Grube 104, 106, Boenigk, 1970) have been enlarged and connected to each other. In the basal part a typical example of the Reuver clay is exposed. Brown-coal seams with abundant wood remains are present. Pollenanalytical data point to the Reuver-B pollenzone.









The heavy mineral diagram of Boenigk (Fig. 7) indicates that instable heavy mineral associations are already important underneath the Reuver clay. The first clay layer in the Peistocene sediments shows an influx of stable heavy minerals. This clay is often correlated to the Belfeld Clay but so far no pollenanalytical sustainment for this has been found (the clay lacks any pollen).

A second clay layer in the Tegelen deposits of this excavation possesses an interglacial pollen assemblage. Probably, parts of the Tiglian-C pollenzone are reflected. Urban (1978) assigned this kinds of pollendiagrams as the so-called "Van Eyck-Interglazial".

BELFELD "Peter van Eyck"

15

LEGENDE

	GRANAT		ANDERE MINERALE
	EPIDOT		STAUROLIT
	ALTERIT		METAMORFE MIN.
	HORNBLLENDE		TOURMALIN

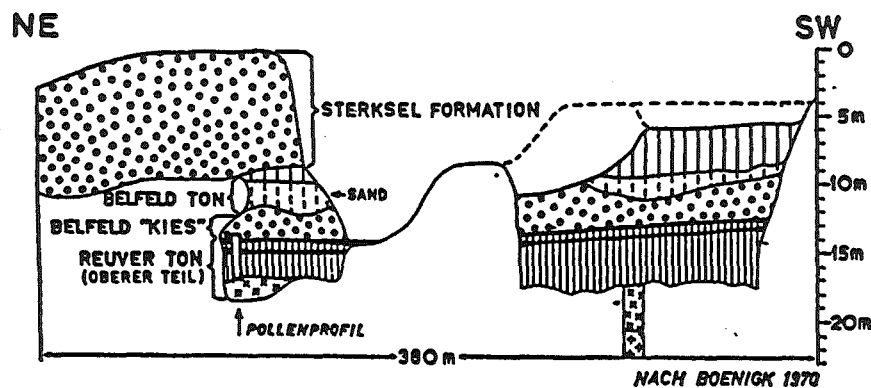
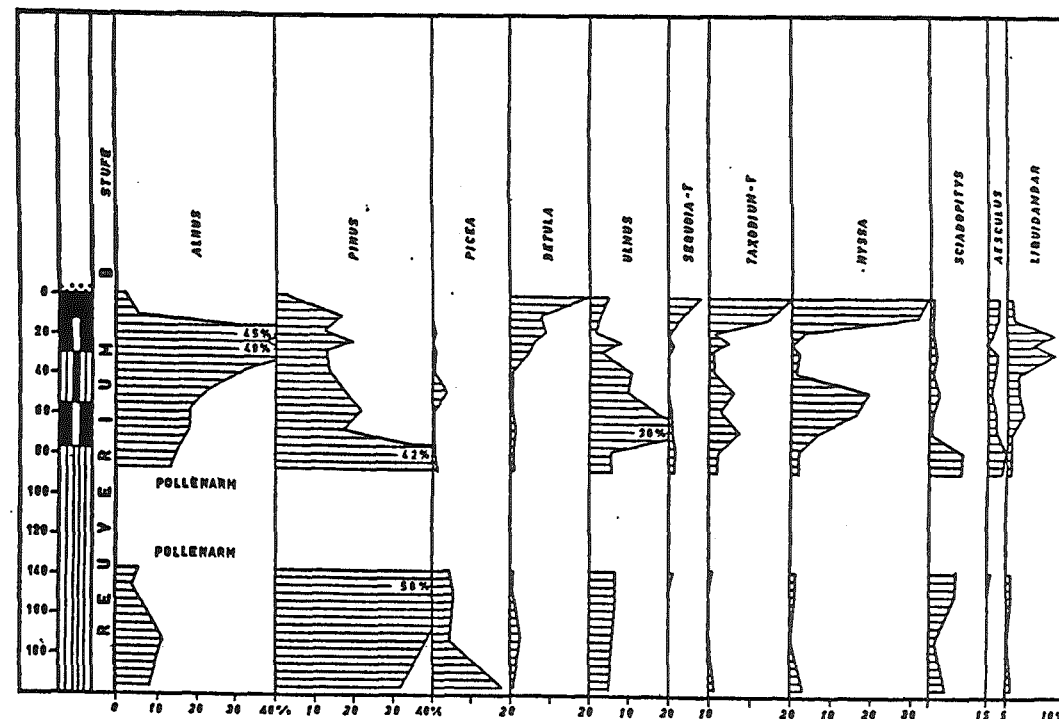
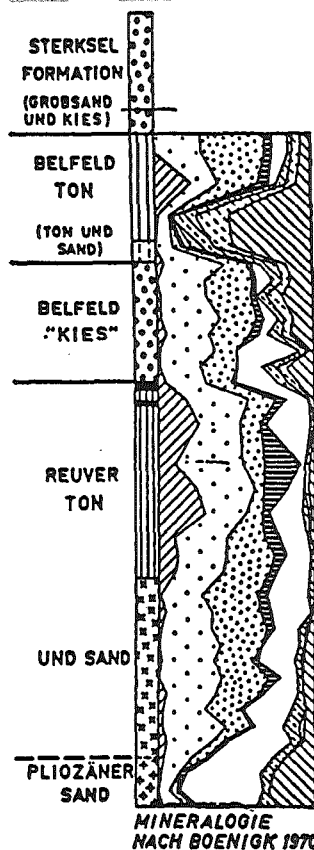


Figure 7: Section and heavy mineral diagram of the pit Peter van Eyck (Hoher Stall) according to Boenigk (1970) and pollendiagram of the Reuver Clay (Zagwijn, 1974).

The Maalbeek pit (Figs. 8, 9, 10, 11, 12)

Recent investigations in the Maalbeek pit, located near Belfeld, has yielded new information about the Lower-Pleistocene stratigraphy. One of the excavated clay layers in the pit has a pollen assemblage which clearly can be correlated with the so-called Belfeld Clay. It is the first find of this stratigraphical important layer on Dutch territory. Chronostratigraphically it is assigned with the Tiglian-A, the temperate period after the colder Pretiglian. The upper part of the investigated section in Maalbeek shows a cooler phase which is interpreted as the Tiglian-B period. The section as a whole is incised by a clayey channel-fill dating from the Tiglian-C period: indicating the complex geology of the study area (Fig. 8).

In the lowermost part of a section exposed in 1993 (Fig. 8) an up to 0.5 m thick brown-coal seam was found. Pollenanalytical investigations show a dominance of *Fagus* in the clay below the brown-coal seam (Figs. 9 and 10) and this clay is therefore dated as Tiglian-A and correlated with the Belfeld Clay (Fig. 11). This coal seam is overlain by a 3 - 5 m thick clay in which two soil horizon are developed. The interval directly overlying the brown-coal seam represents a cooler phase (Figure 10: high values for herbs, especially Ericaceae) and is interpreted as belonging to the Tiglian-B pollen-zone. The section is overlain by a channel infill which can be placed in the Tiglian-C pollen-zone.

In 1963 Zagwijn published a pollendiagram from the Maalbeek pit which is very similar to the Tiglian-B diagram of the recent sections (Fig. 12). He then interpreted it, on the basis of stratigraphical position and heavy mineral data, as part of the Eburonian glacial phase. But now it seems that a more complete sequence of the Tiglian has been found in the Maalbeek area.

Table II: Stratigraphy of Pliocene and Lower-Pleistocene in the Tegelen area.

chrono-stratigraphy	Pollen-zones	main climatic characteristics	local stratigraphy
TIGLIAN	T-C	Warm	Tegelen Clay and Tegelen Gravels
	T-B	Cool	
	T-A	Warm	Belfeld Clay and Belfeld Gravels
PRETIGLI	P-T	Cool	Meinweg Clay
REU-VERIAN	R-C/R-B/R-A	Warm	Reuver Clay

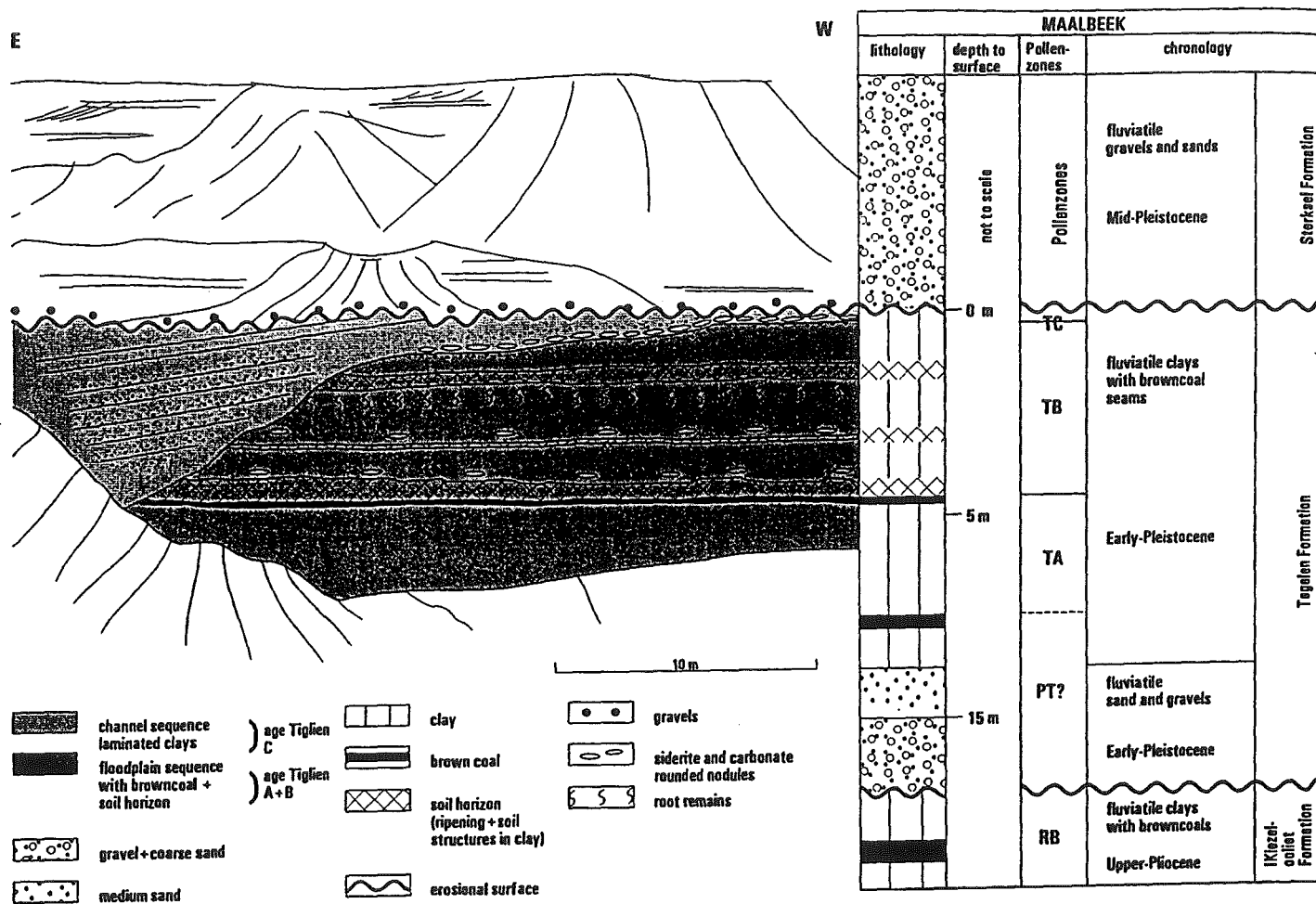


Figure 8: Section and stratigraphic position of the exposure Maalbeek in 1993.

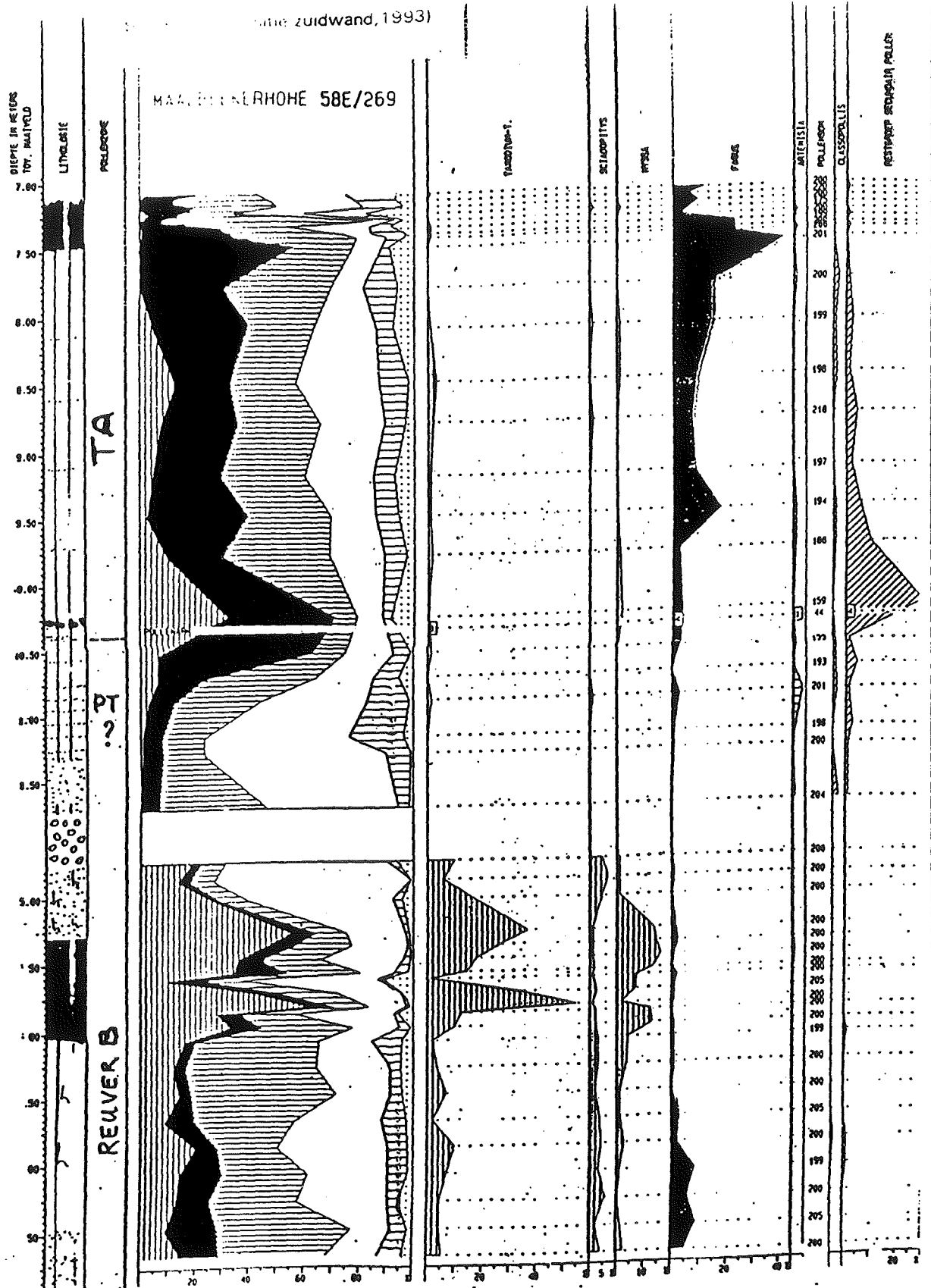


Figure 9: Pollendigram of a cored section underneath the exposure, shown in fig. 8, at Maalbeek.

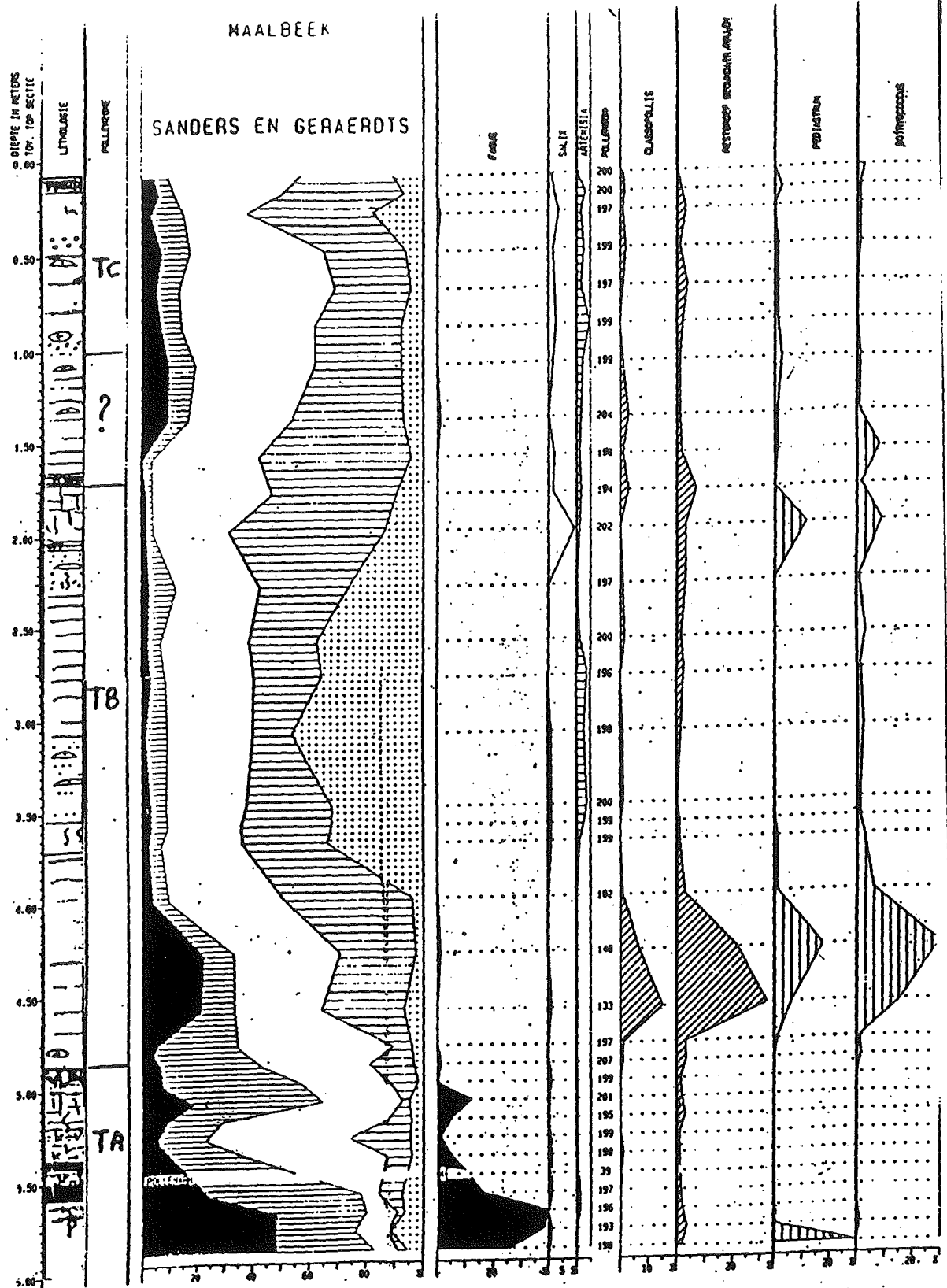


Figure 10: Pollendiagram of the exposure in Maalbeek (see fig. 8).

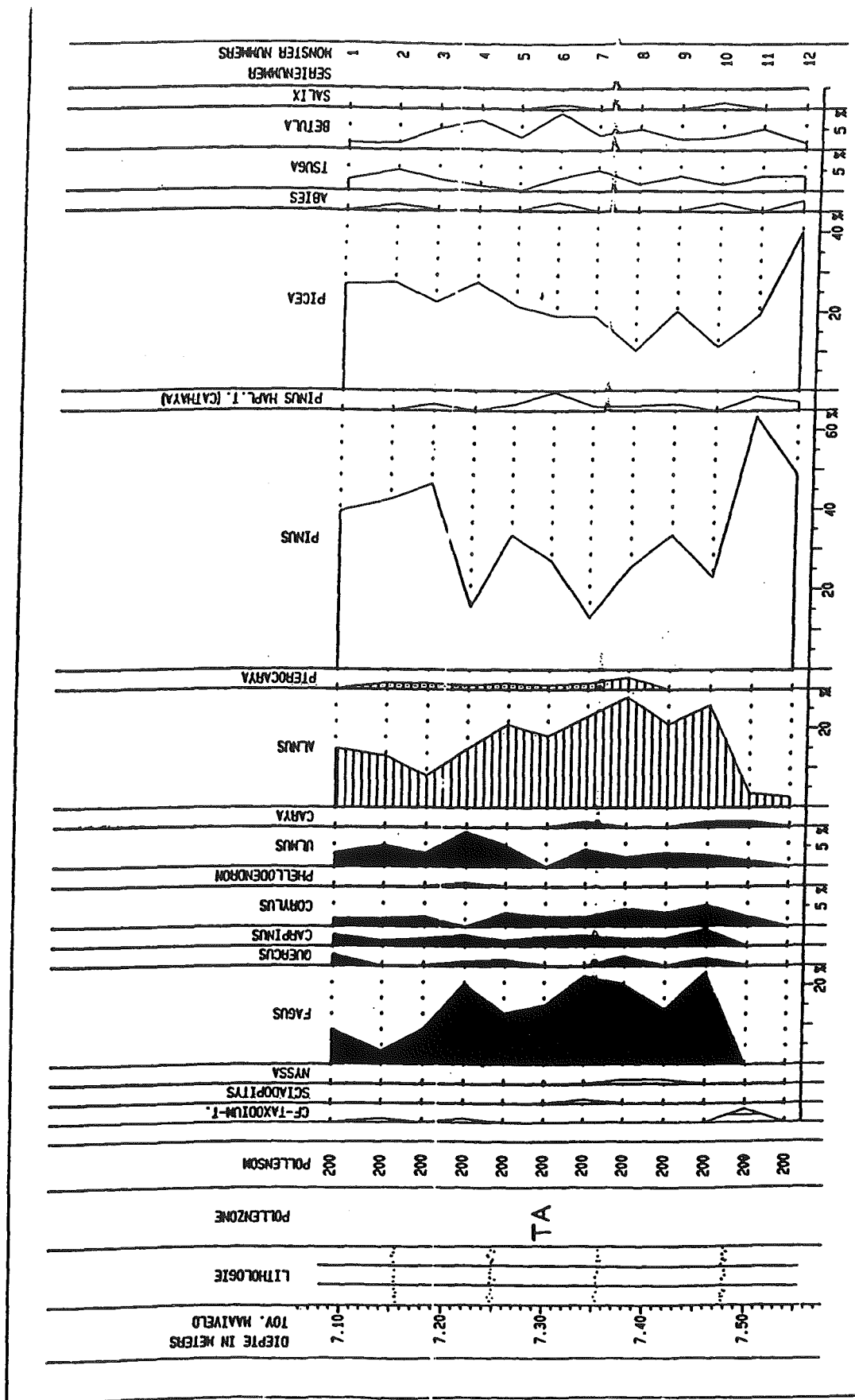


Figure 11: Pollendiagram of the Belfeld Clay (Tiglian-A, modified after van der Vlerk & Florschütz, 1953).

Figure 12: Pollendiagram of the Maalbeek pit published by Zagwijn (1963).

The Laumans pit (Fig. 13)

In this pit only the uppermost part of the Tegelen Clay is exposed. The sections, however, are of interest because it is one of the best sites in the area where remnants of the Kedichem Formation overlying the Tegelen deposits can be studied. During the summer of 1994 a clear polygonal pattern of frost wedges was observed in the top of the Tegelen Clay. This frost wedge level is overlain by another 30 cm of clay with the same habitus as the Tegelen Clay. Probably we found indications of a cold climate in the upper part of the Tiglian. However, pollenanalytical investigations of the section have still to be carried out.

Clear signs of cold conditions can be seen in the sediments of the Kedichem Formation. These generally fine-grained sands are locally named as "pap-zand" (literally this means porridge sand because of its susceptibility to liquefaction). In these deposits levels with small frost wedges and cryoturbations are found. It's believed that they are formed during the Eburonian glacial period. Soils in the top of the deposits from the Kedichem Formation (partly with peat from which pollen data have been obtained) point to a more temperate climate (probably the Waalian interglacial period). The soils are overlain by fine sand that again show cryoturbations and frost wedges (probably dating from the Menapian glacial period).

Because the deposits from the Kedichem Formation show an unconformity at the top we seldom see a complete section of it. The larger part of it has been eroded before the Middle-Pleistocene sediments of the Sterksel Formation were deposited.

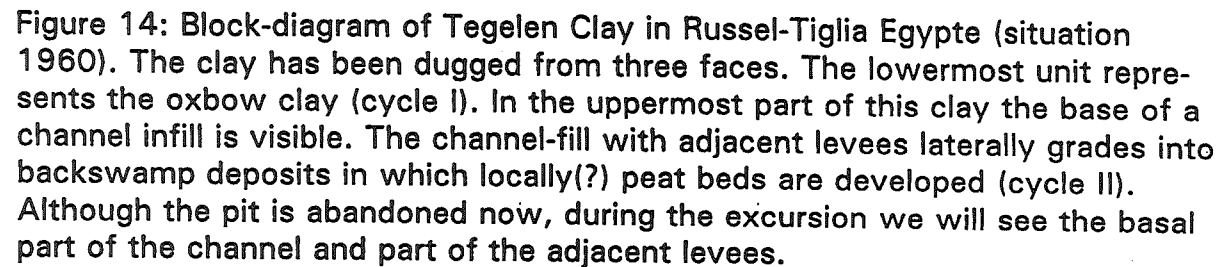
		FORMATION	climate	LITHOLOGY/SEDIMENTOLOGY	cryoturbations	MINERAL ZONE AND PROVENANCE	HEAVY MINERALS
UPPER PLEISTOCENE	WEICHSELIAN	TWENTE FORMATION	cold	COVERSAND (ABOUT 1m)			
MIDDLE PLEISTOCENE		STERKSEL FORMATION	cold	ABOUT 15 m OF COARSE SAND AND GRAVEL SOME LARGER BLOCKS MEANDERING RIVERS p.p.	frostwedge casts	"WEERT ZONE" (RHINE)	EPIDOTE HORNBLende SAUSSURITE
LOWER PLEISTOCENE	MENAPIAN	KEDICHEM FORMATION	cold	FINE SANDS	(large) frostwedges and involutions		
	WAALIAN (pp.)			SOILS			
	EBURONIAN		cold	FINE SANDS PARTLY FLUVIAL PARTLY PERIGLACIAL-EOLIAN (ABOUT 3m)	small frostwedge casts (in other pits)	"LOWER FINE ZONE" (LOCAL)	TOURMALINE STAUROLITE ZIRKONE METAMORFIC MIN
	TIGLIAN (top)	TEGELEN FORMATION	cool- temp.	CLAY FLUVIAL (MEANDERING RIVERS) (ABOUT 2-3m)	in other pits: involutions	"TEGELEN ZONE" (RHINE)	GARNET EPIDOTE HORNBLende SAUSSURITE

Figure 13: Stratigraphy of the Laumans pit (according to Zagwijn).

The Russel-Tiglia-Egypte pit (Figs. 14 and 15)

This pit is abandoned for more than 25 years now, after having been excavated for about 20 years. The pit is directly adjacent to the former pit of Canoy-Herfkens, from which Dubois and Cl. and E. Reid collected many fossils in the beginning of this century.

The Tegelen Clay exposed in the Russel-Tiglia-Egypte pit is used as the stratotype for the Tiglian (Zagwijn, 1963). Two sedimentary cycles have been described. Cycle I consists of a lake clay which is considered to be the infilling of an oxbow lake with a width of approximately 200 m. In my opinion this clay shows much resemblances to the Tiglian-C clay of the Maalbeek pit. The second sedimentary cycle consists of a narrow gully incised in the top beds of the oxbow lake clay. The gully is bordered with silty to fine sandy levees laterally grading into backswamp clays. A block diagram (constructed by Kortenbout van der Sluis) shows an impression of the excavation in 1960 (Fig. 14). From the base of the gully infilling of cycle II many fossil wood, fruits, seeds and mammal remains have been collected. The larger part of these fossil remains have been described by Zagwijn (1963), Freudenthal, Meijer & Van der Meulen (1976) and Van der Burg (1974).



Malacological research in the Tegelen area

Tom Meijer, Geological Survey of The Netherlands, Malacology Dept.

The Quaternary non-marine molluscs of the Tegelen area were studied by Tesch in the first half of the century, but never published separately. The first modest, and as yet only, attempt to review the molluscan data was made in Freudenthal et al. (1976) who published an annotated preliminary list of species which were collected in the Russel-Tiglia-Egypte pit. The list showed many undescribed taxa. Later, the Quaternary freshwater species of the Netherlands were reviewed (Meijer, 1990a, 1990b) and in these publications the freshwater species of the Tiglian of the Tegelen area were included. A review on the land species is in progress.

The most important exposures which have been studied for their molluscan content are the Russel-Tiglia-Egypte, the Maalbeek, and the Reuver pits. The latter are of Late Pliocene (Reuverian) age and produced numerous sideritic casts of freshwater mussels belonging to *Unio*, *Potomida*, *Anodonta*, *Sinanodonta*, and *Hyriopsis* (Van Regteren Altena, 1955). *Hyriopsis* at present is restricted to Japan, *Sinanodonta* to the Far East. These species testify the Asian imprint of the Late Pliocene molluscan assemblages.

The Tiglian assemblages of the Russel-Tiglia-Egypte and Maalbeek pits resemble each other very much, although obvious differences can be observed. From both pits the molluscs of the Tiglian-C deposits were studied. In Maalbeek also a small assemblage dating from the Tiglian-A pollen zone was collected. This assemblage does not show any species which is absent in the younger strata. In Russel-Tiglia-Egypte assemblages from sediments assigned (by definition) to pollen zones Tiglian-C3 and Tiglian-C5, were studied.

Essentially all assemblages are characteristic of slowly moving to stagnant fluvial environments, which are bordered by moisty, closed forests and reedlands. Both land and freshwater molluscs show a strong continental influence in the climate. Land species which point to such type of climate are: *Vertigo alpestris*, *Pupilla sterri*, *Discus ruderratus*, *Discus perspectivus*, *Perforatella dibothryon*, *Perforatella belgrandi*, *Monachoides* sp. nov., *Soosia* sp. nov., *Helicigona*, sp. nov., and *Cepaea* (s.l.) div. sp. nov. Freshwater species with the same affinities may be found among the mostly extinct Prosobranchs, such as *Viviparus glacialis*, *Valvata salebrosa*, *Valvata goldfussiana*, *Parafossarulus crassitesta*, *Lithoglyphus jahni* and *Tournouerina belnensis*. The still extant *Bithynia troscheli* certainly prefers a continentally influenced climate. Asian affinities are still present in the molluscan fauna of the Tiglian as is shown a.o. by the occurrence of *Parafossarulus*.

Several extinct freshwater species are only known from North European river systems. Amongst these, *Viviparus glacialis* is the most important. This species occurs in the Tegelen and Maassluis Formations and is further only known from the East Anglian Weybourne Crag which is supposed to be of Late Tiglian age. Tesch was aware of the stratigraphical significance of this species. He proposed two freshwater

molluscan biozones for the Quaternary of the Netherlands: "Horizon met *Viviparus glacialis*", correlating with the Middle and Late Tiglian and "Horizon met *Viviparus diluvianus*" which roughly correlates with the Holsteinian (Tesch, 1929).

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

QRA-excursion Saturday 8 April 1995

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

C. Kasse

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 terrace 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 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 ($Q_{\text{mean winter}}/Q_{\text{mean summer}}$) of the Maas in comparison with the Rhine. Furthermore, the fluctuations in discharge of the Maas during the winter ($Q_{\text{max winter}}/Q_{\text{mean winter}}$) and especially during the summer ($Q_{\text{max summer}}/Q_{\text{mean summer}}$) 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 m³/sec and 2870 m³/sec were measured during the two floods. The floods were caused by exceptionally high rainfall in the Ardennes and northern France.

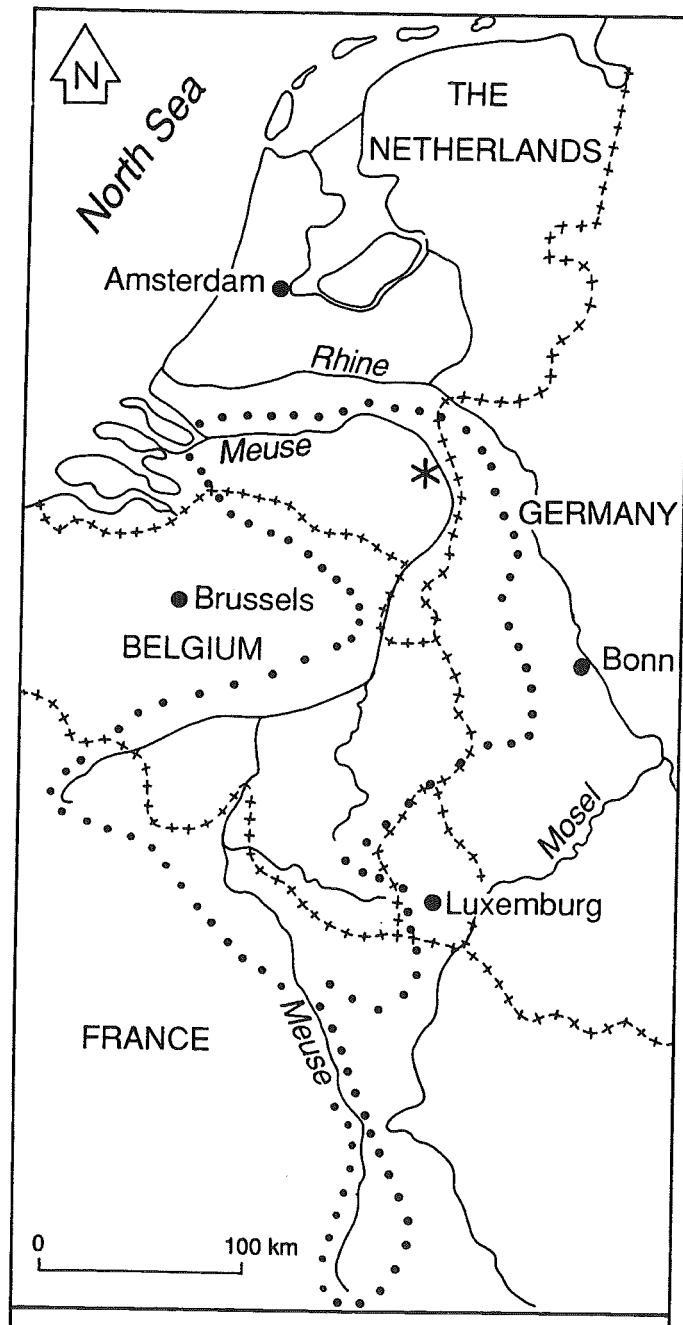


Fig. 1 Catchment area of the Maas (from Kasse et al., 1994)

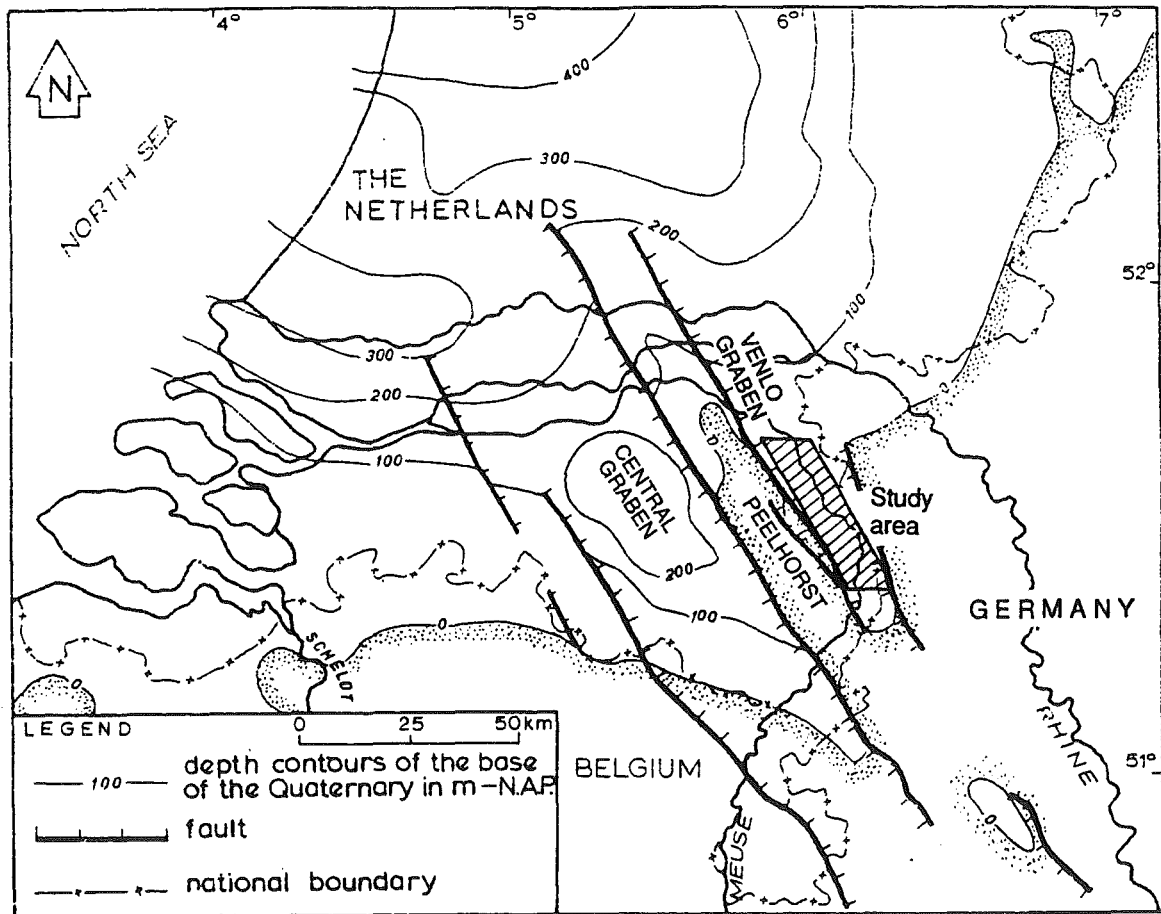


Fig. 2 Major tectonic units of the Southern North Sea basin and depth contours of the base of the Quaternary deposits (after Zagwijn & Doppert, 1978).

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 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., 1994; Vandenberghe et al., 1994) (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; Excursion stop Bosscherheide).

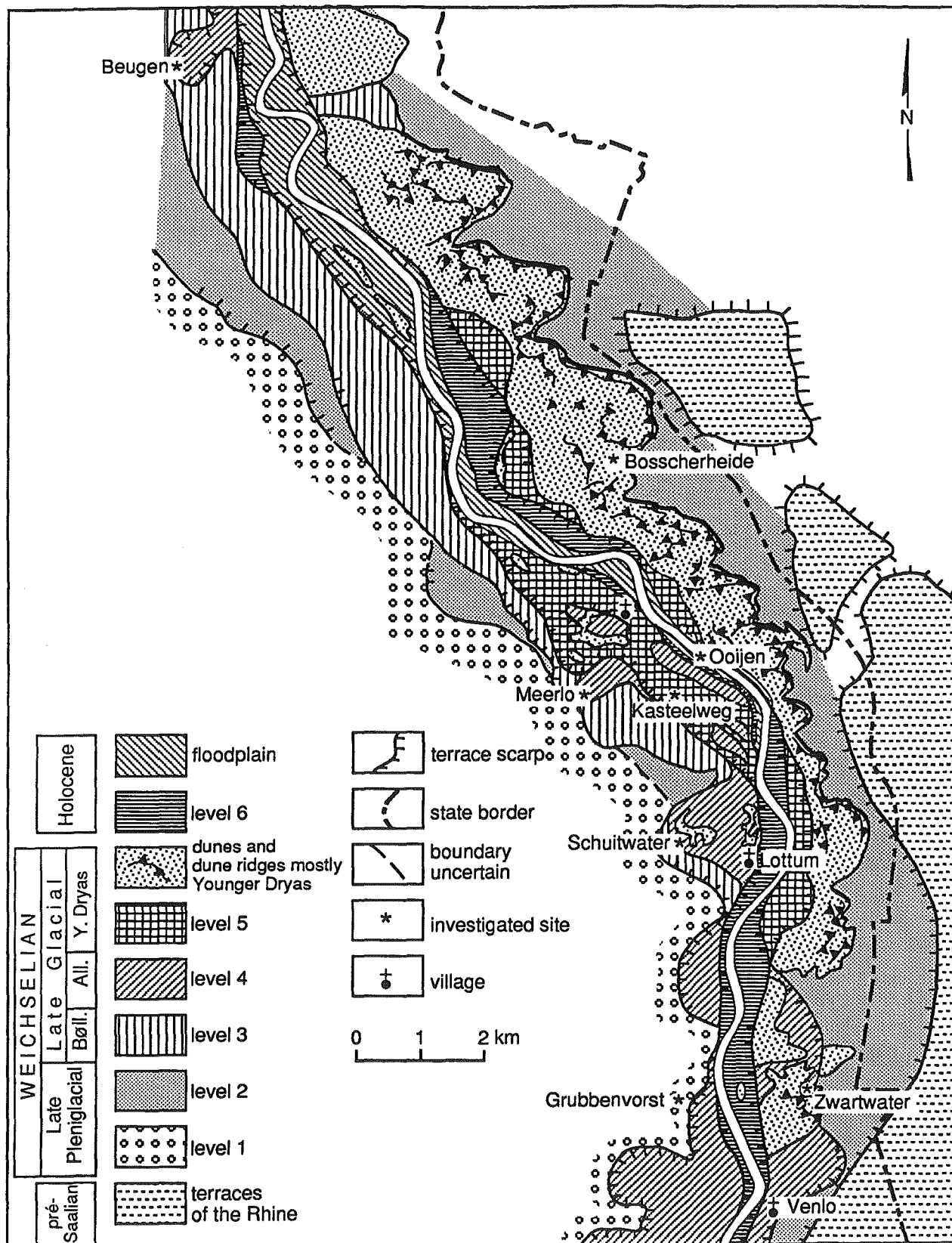


Fig. 3 Morphological map of the Late Pleniglacial, Late Glacial and Holocene terraces of the Maas north of Venlo (from Kasse et al., 1994, partly after Wolfert & De Lange, 1990).

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; Excursion stops Meerlo and Beugen). 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 Excursion stop 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 braided river floodplain was formed. The higher discharges in combination with a rather restricted sediment supply 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.

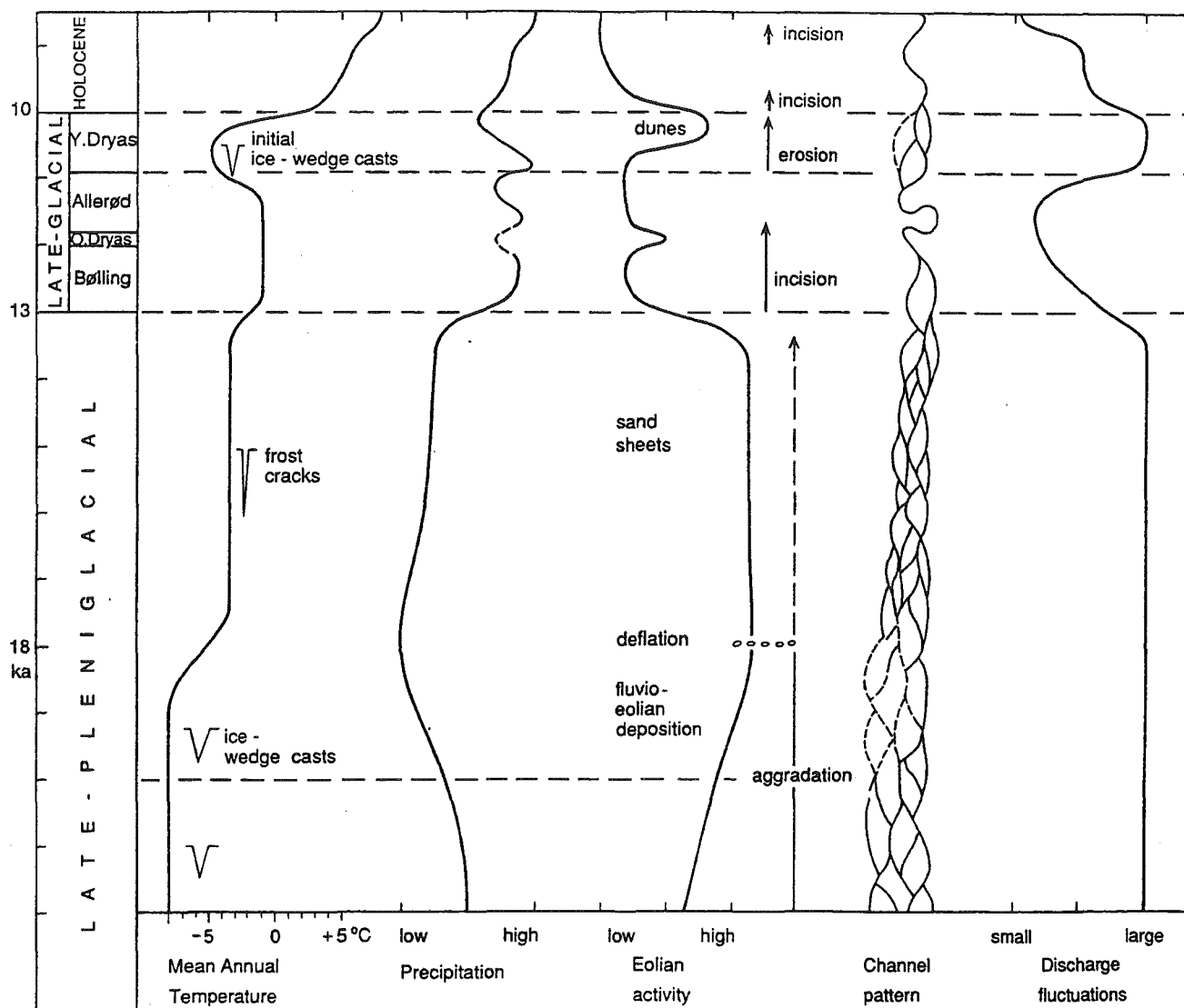


Fig. 4 Summary of the Late Pleniglacial and Late Glacial climatic changes and climate-related fluvial development of the Maas river (from Kasse et al., 1994).

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 (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 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.

	Maas	Warta	
	SMALL MEANDERS		
10250	BRAIDED	(central bar formation)	10250
	erosion		
10900	lateral migrating LARGE MEANDERS		
11850	curved ↑ decreasing number of channels	TRANSITIONAL (multi-channel)	sinuous ↑ weakly curved
	straight	incision	straight
13000	eolian ↑ fluvial	BRAIDED	fluvial sed. ↑ fine coarse

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: 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 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 completely 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 $10,500 \pm 60$ 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 the repeated fluvial reworking of the braided floodplain followed by deflation. These alternating processes of 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. As a result

the Late Glacial meandering river course changed into the braided Younger Dryas course.

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Late Glacial and early Holocene vegetational and climatic history in the Netherlands

S. Bohncke

The Late Glacial vegetation development of the Netherlands and North-West Europe has recently been summarized by Bohncke (1993) and Walker et al. (1994) (figs. 6 and 7).

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) provided a date of 12,930 BP. At 12,885 BP organic sedimentation started in the former Dinkel Valley (Ran, 1990). Conventional ^{14}C -dates, based on larger quantities 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 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 constituents 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* (c. 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.

Fig. 6
Summarized Late Glacial
vegetation succession
for the Netherlands
(from Bohncke et al., 1988).

HOLOCENE	P.A.Z.	CHRONOSTRATIGRAPHY	VEGETATION CHARACTERISTICS	SOIL CONDITION
	10,000			
	<i>Betula</i> , <i>Empetrum</i> , Cyperaceae	YOUNGER DRYAS	Expansion of a sedge-rich <i>Empetrum</i> heath, intermingled with <i>Betula</i> and <i>Salix</i> dwarf shrubs. Tree birches and pine trees are sparse, but more prominent in the S. Netherlands.	Soil conditions instable. Deposition of cover sands. Leaching and acidification prevails.
	11,000			
	<i>Betula</i> / Pinus Pinus / <i>Betula</i> <i>Betula</i> (Pinus)	ALLERÖD	Some <i>Empetrum</i> heath development with <i>Betula</i> forest expansion, with <i>Populus tremula</i> ; <i>Filipendula</i> increases. Development of a boreal birch forest intermingled with <i>Populus tremula</i> and preceded by a <i>Juniperus</i> belt.	Soil formation and start of podsolisation process.
	11,800			
	<i>Salix</i> , <i>Betula</i> , <i>Artemisia</i> .	OLDER DRYAS	Regional shrub vegetation replaced by some <i>Juniperus</i> and <i>Hippophae</i> shrubs. In favourable conditions tree birches maintain.	Soil conditions instable, soil erosion and deposition of coversands
	12,000			
	<i>Betula</i> , <i>Salix</i> , <i>Artemisia</i> .	BÖLLING	Shrub vegetation with <i>Betula nana</i> , <i>Salix</i> spp. and some <i>Juniperus</i> . Tree birches gradually spread.	Development of a vegetation cover and start of soilforming processes.
	12,400			
	<i>Betula nana</i> , Cyperaceae, Poaceae.	OLDEST DRYAS	Heliophilous herbaceous vegetation with steppe tundra elements, intermingled with dwarf willow and dwarf birches.	Instable soil conditions. Lime-rich soils.
	13,000			

¹⁴ ka BP	IRELAND	ENGLAND, WALES & S. SCOTLAND	SCOTLAND HIGHLANDS & ISLANDS	NORTH BELGIUM	NETHERLANDS	N.W. GERMANY
9.5	WOODLAND <i>Corylus</i> <i>Betula</i>	WOODLAND <i>Corylus</i> <i>Betula</i>	OPEN WOODLAND <i>Betula</i>	<i>Pinus-Corylus</i> woodland	<i>Betula-Pinus</i>	WOODLAND <i>Corylus</i> <i>Betula pendula</i> woods
10.0	OPEN WOODLAND <i>Betula-Juniperus</i> <i>Juniperus</i> OPEN SCRUB	TRANSITION TO WOODLAND <i>Juniperus</i>	TRANSITION TO OPEN WOODLAND <i>Juniperus</i> <i>Empetrum</i>	<i>Pinus-Betula</i> woodland TRANSITION TO CLOSED WOODLAND	BOREAL FOREST <i>Pinus-Betula-Populus</i> TRANSITION TO WOODLAND	with <i>Populus tremula</i> and <i>Pinus</i> <i>Betula pubescens</i> woods TRANSITION TO WOODLAND
10.5	<i>Artemisia</i> <i>Rumex</i> Gramineae	<i>Artemisia-Rumex</i>	<i>Artemisia-Rumex</i>	<i>Betula-Pinus</i> open woodland	<i>Juniperus-Betula nana</i> <i>Pinus</i> in S. <i>Empetrum</i> in N. <i>Betula-Salix</i>	<i>Juniperus</i> copses park Tundra with <i>Betula</i> , <i>Empetrum</i> & <i>Betula pubescens</i> in N. & W.; <i>Betula-Pinus</i> woodland in SE
11.0	TUNDRA & LOW ALPINE SCRUB <i>Cruceae</i>	TUNDRA & LOW ALPINE SCRUB <i>Rumex</i> Gramineae	TUNDRA & LOW ALPINE SCRUB Gramineae-Cyperaceae Caryophyllaceae Lycopodium	<i>Pinus-Betula</i> open woods with <i>Empetrum</i> TRANSITION TO BOREAL WOOD & HEATHLAND	<i>Empetrum-Artemisia</i> TRANSITION TO BOREAL WOOD & HEATHLAND	
11.5	Gramineae-Helianthemum	<i>Pinus</i> (SE England) <i>Betula</i> woods (Sporadic in N)	<i>Empetrum</i> & <i>Erica</i> Heathland	<i>Pinus-Betula</i> woodland <i>Betula</i> woods	<i>Pinus</i> Boreal woodland <i>Betula</i> Boreal woods	<i>Betula pubescens</i> woods with <i>Populus tremula</i> & <i>Pinus</i> in S.
12.0	OPEN GRASSLAND	WOODLAND	Scattered <i>Betula</i> wood & scrubland	TRANSITION TO WOODLAND OPEN GROUND COMMUNITIES	TRANSITION TO WOODLAND <i>Salix-Betula-Juniperus-Hippophae</i> OPEN SCRUBLAND	TRANSITION TO WOODLAND <i>Juniperus</i> copses HERBACEOUS COMMUNITIES
12.5	<i>Juniperus</i> or <i>Juniperus-Empetrum</i> (Depending on soils)	<i>Betula-Juniperus</i> wood & scrub	<i>Betula-Juniperus</i> scrub	Open <i>Betula</i> woods <i>Juniperus</i>	<i>Betula-Salix-Juniperus</i>	<i>Betula nana</i> , <i>Juniperus</i> , <i>Hippophae</i> , <i>Salix</i> (edaphically-controlled)
13.0	LOW SCRUB & HEATHLAND COMMUNITIES <i>Rumex-Salix</i> <i>Artemisia</i> -Gramineae OPEN GROUND COMMUNITIES	<i>Juniperus</i> <i>Salix</i> OPEN WOOD & SCRUB	<i>Juniperus</i> OPEN WOOD & SCRUB; HEATHS (N & W)	<i>Salix-Betula</i> TRANSITION TO WOOD & SCRUB OPEN GROUND COMMUNITIES	TRANSITION TO WOOD & SCRUB Gramineae-Artemisia-Cyperaceae Gramineae-Helianthemum Thalictrum Chenopodiaceae TUNDRA COMMUNITIES	TRANSITION TO SCRUB <i>Artemisia</i> STEPPE COMMUNITIES

Fig. 7 Summary of the vegetation history during the Late Glacial in northwest Europe (from Walker et al., 1994).

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. 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 hydrosereal succession 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 preceding 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 $11,990 \pm 70$ BP is available to pinpoint this event (e.g. site Mariahout; Bohncke, 1993).

11,900-10,900 BP: 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 hydrosereal succession (Bohncke et al., 1993). The pollenrecord demonstrates a regional spread of tree birches in the southern Netherlands while at more northern latitudes this *Betula* increase is preceded by a *Juniperus* maximum. Both taxa rely on increased amounts of precipitation.

At c. 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., 1993). The local hydrosereal succession is set back to open water. In the subsequent local hydrosereal succession the rhizomatose perennials like *Typha* spp. and *Phragmites* are lacking (e.g. Excursion site Bosscherheide, Bohncke et al., 1993) 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 *Isoetes* sp., *Elatine hydropiper*, *E. hexandra* (Bohncke & Wijmstra, 1988). Both species require pioneer conditions. *Isoetes* 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., 1993) 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 response to climatic changes at around 11,300 BP implying more severe winter temperatures and increase in freeze-thaw incidences. 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 responsible for this phenomenon.

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

At c. 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 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 palustre* and *Sparganium* spp. and pleurocarpe mosses. This period came to an end when

precipitation diminished and the previously established large shallow lakes dried out at c. 10,500 BP.

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

Soon after, c. 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., 1993).

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 harsh for plant-life. 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 c. 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, embracing 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 preceded 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 hydrosere 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 ($9,705 \pm 55$ BP, Schuitwater Lottum, 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 BP) *Betula* restores from its temporary low values during the Rammelbeek-phase. An increase in the lake-levels for this period (Bohncke, 1991) indicates a restoration 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 c. 8,400 BP *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 c. 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 January temperature higher than -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 successful 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 its rational limit (a continuous curve).

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Saturday 8th April 1995

Excursion program

Stop 1: GRUBBENVORST SAND PIT

- * Late Pleniglacial fluvio-aeolian facies
- * Late Pleniglacial cryogenic structures (ice-wedge casts, deformations, vertical platy structures)
- * Late Pleniglacial and Late Glacial aeolian sediments

Stop 2: BOSSCHERHEIDE GRAVEL PIT

- * Late Pleniglacial braided floodplain
- * Bølling to Allerød peat in abandoned channel
- * Younger Dryas embryonal ice-wedge casts and cryoturbations
- * Younger Dryas flooding and moss peat growth
- * Younger Dryas sand sheet and river dune formation

Stop 3: MEERLO CHANNEL

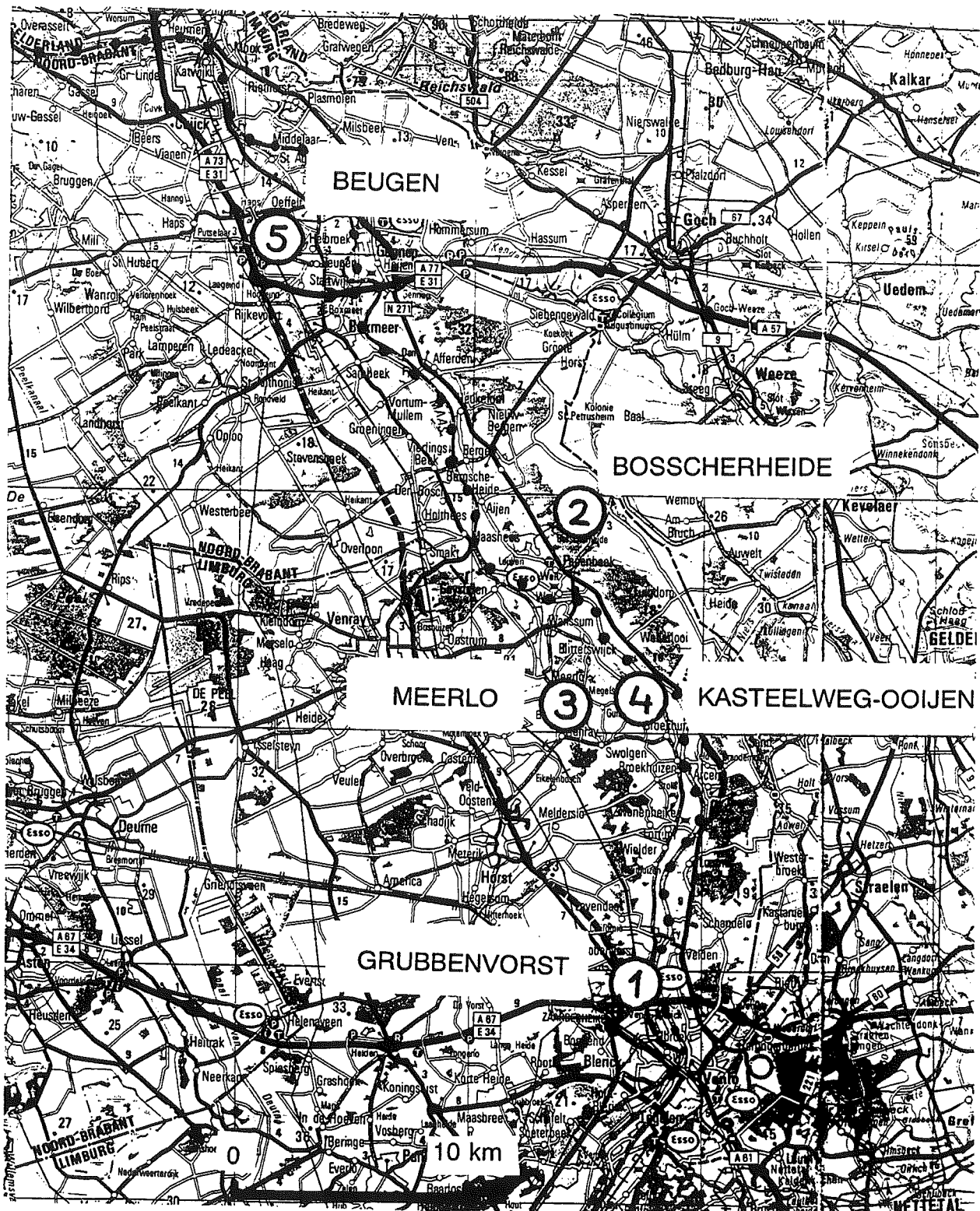
- * Allerød meander scar
- * Younger Dryas clastic infill

Stop 4: KASTEELWEG - OOIJEN CHANNELS

- * Younger Dryas braided floodplain morphology
- * Late Younger Dryas and Preboreal fill
- * Younger Dryas dune complex
- * Early Holocene incision and terrace edge
- * Boreal and Atlantic channel fill

Stop 5: BEUGEN

- * Late Glacial (Allerød) meander neck cut off
- * Allerød organic fill (gyttja)
- * Younger Dryas reactivation and clastic deposition



Location map with the excursion stops in the Maas valley.

EXCURSION STOP 1: GRUBBENVORST

Weichselian Late Pleniglacial fluvio-aeolian deposits and cryogenic structures

after J. Schwan & J. Vandenberghe (1991)

Stratigraphy

The Grubbenvorst sandpit is situated in a Maas river 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 : $26,130 \pm 180$ BP at its base and at GrN-16948 : $25,200 \pm 180$ 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 may be attributed to the section under consideration.

Fluvial to aeolian sedimentary facies

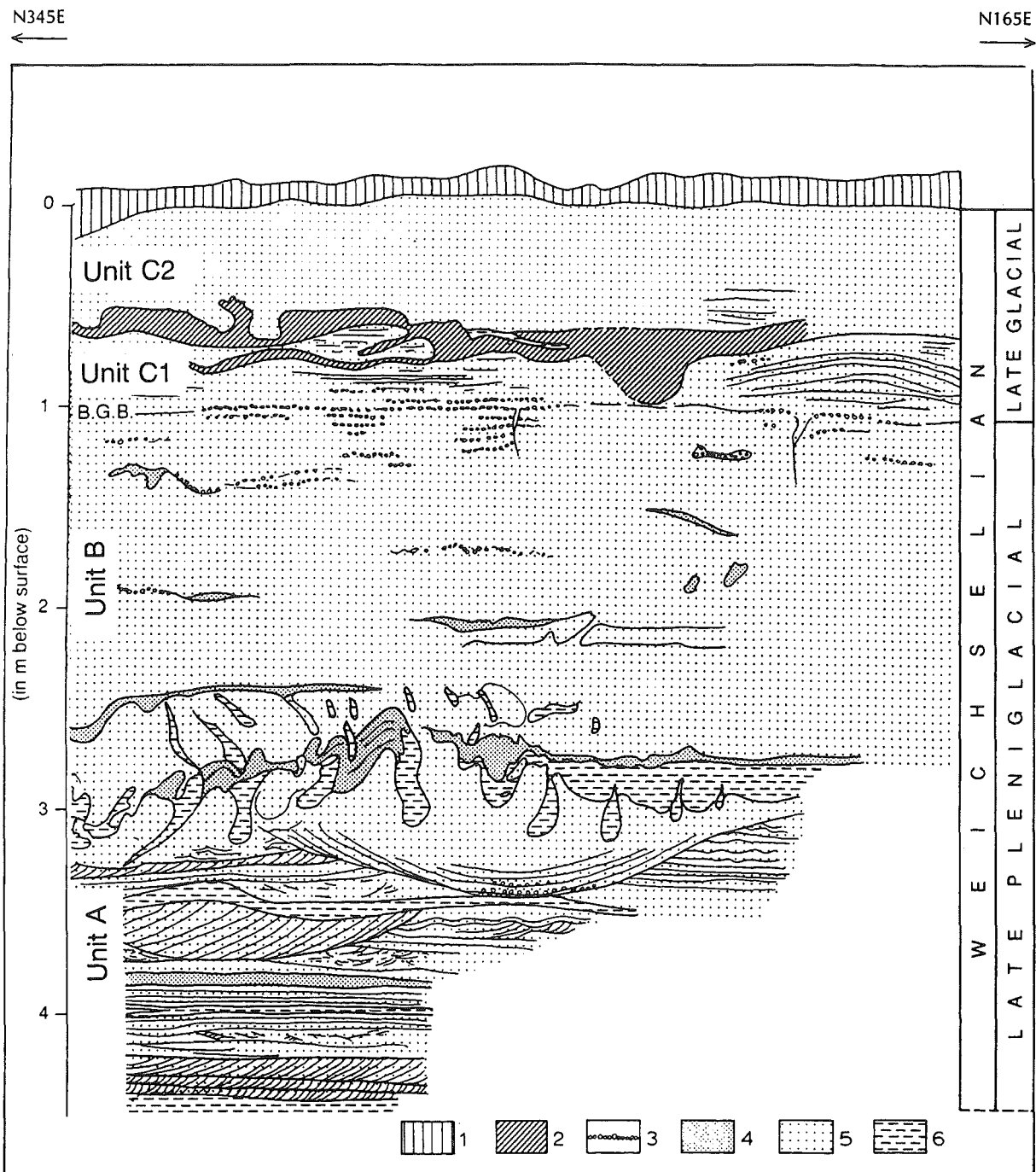
The Grubbenvorst section is a fine specimen of a fluvial to aeolian sequence, a type of succession that is wide spread in the Pleistocene lowlands of northwestern Europe (Schwan, 1987). In the exposure a gradual transition from a purely fluvial sediment at the base to a purely aeolian one in the top 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 infilled channels are the prevailing structural characteristic. The channels occur in both intersecting as well as solitary positions, mostly have a concordant infilling 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 sand pit Grubbenvorst in July 1989 (taken from Mol et al., 1993).

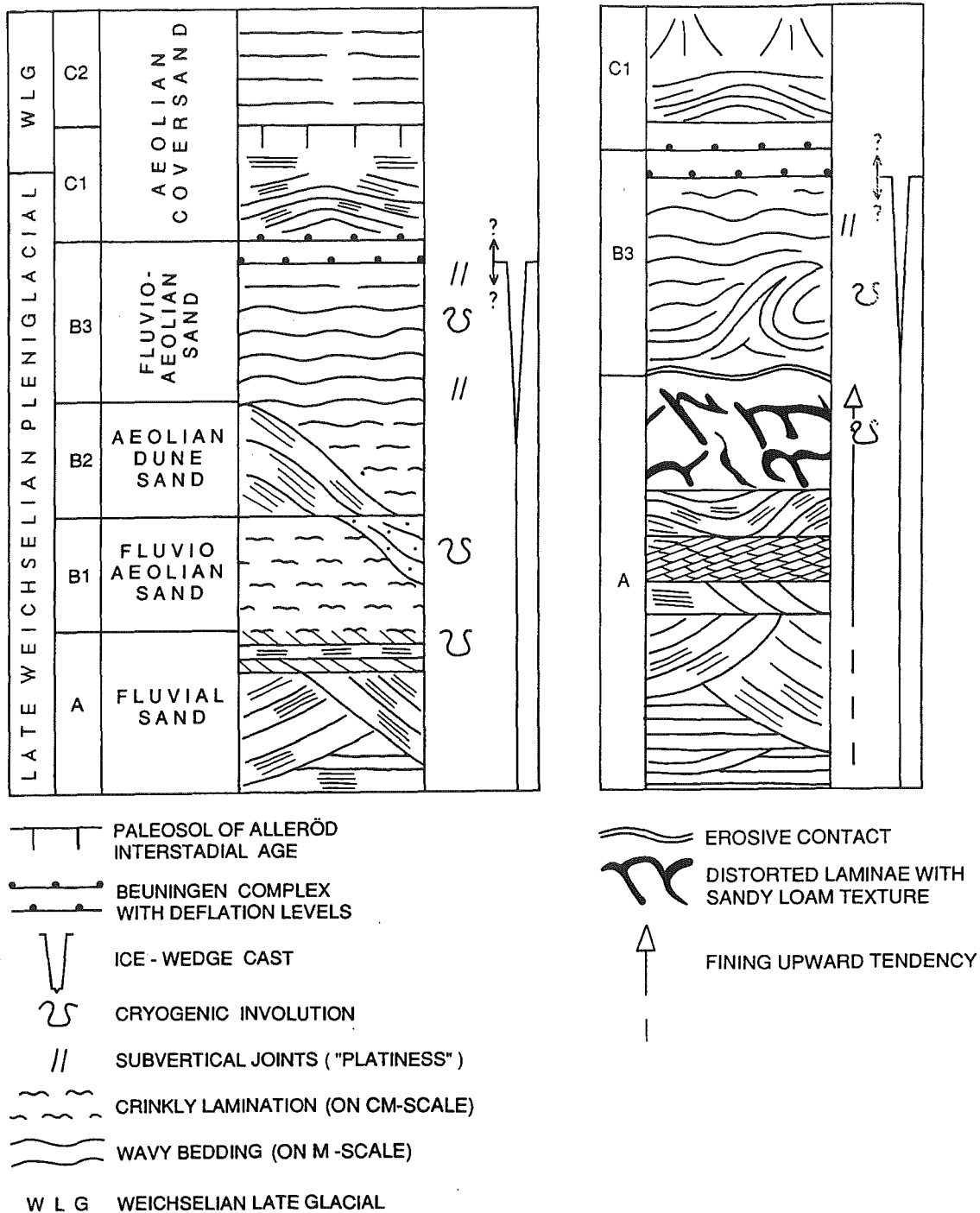


Fig. 3 Schematic sedimentary succession in sand pit Grubbenvorst in August 1988 (left side) and Novembre 1990 (right side) (taken from Schwan & Vandenberghe, 1991).

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 infilling, rather than being concordant, is lateral; the attitude of the foresets suggests that they were formed by a streamflow 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 within 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 streampower, - seasonal or otherwise, - followed by a phase of infilling 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 must have been involved in the buildup of the braidplain.

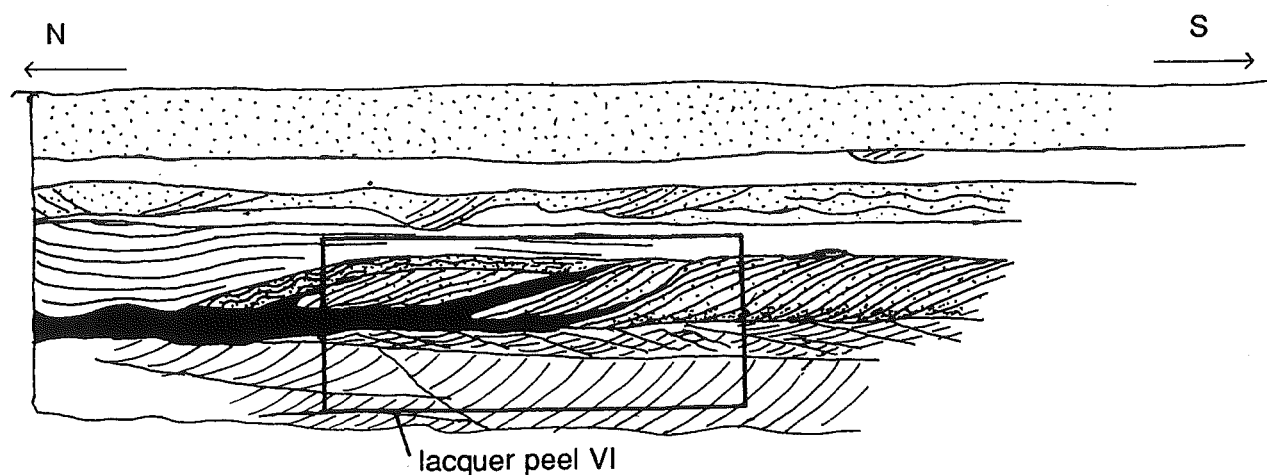


Fig. 4 Silt drapings on the foresets of large-scale transverse bars at the base of fluvial unit A. These lateral changes illustrate the recurrent alternation of sand deposition during peak discharges and silt deposition during standing water conditions on the braided floodplain.

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 sheetlike 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. They generally occur in association with ice-wedge cast and therefore are assumed to have been generated under the very severe cold conditions of the last glacial maximum.

With respect to the above it is noted that:

- 1). The degree of development of any one 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 the 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 timespan 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 2: BOSSCHERHEIDE

Introduction

Bosscherheide is located on the east bank of the Meuse in an extensive dune area, characterized by its general parabolic outline (Fig. 1). As a result of intensive exploitation of the area for its gravels and sands numerous profiles became available in which fluvial and aeolian environments, dated to the Weichselian Late Glacial, could be studied (see Fig. 2). Well-developed Younger Dryas cryoturbation structures have been found at the transition of both environments (Fig. 6 and 7). The organic infill of an abandoned gully and its lateral extension over the adjacent terrace (see Fig. 2) has been sampled for pollenanalyses and dating purposes (Figs. 3 and 4). The cryoturbation structures have been studied macroscopically (Fig. 7).

It is quite common in the Dutch and German lowland region that, buried under parabolic dunes of Younger Dryas age, a palaeosol of Allerød age, the so-called Usselo soil (Van der Hammen, 1951) occurs. This is also the case at excursion site Bosscherheide where, at the base of the dune sediments, Late Glacial organic sediments and peats are present.

Sandwiched between the base of the dune and the organic sediments the lithological sequence at Bosscherheide reveals the results of a series of processes during the Younger Dryas, that previously had been observed at a site called Notsel (Vandenbergh et al., 1987 and Bohncke et al., 1987). In Notsel, however, dating was difficult due to the absence of organic material in the Younger Dryas interval. Two important features were described at Notsel and placed in the early Younger Dryas being, 1) a reactivation of the fluvial regime visualized by the deposition of fluvial loams and sands and 2) the local establishment of deep seasonal frost or discontinuous permafrost as witnessed by the occurrences of a transitional form between an ice-wedge cast and a frost fissure (Fig. 5).

Dating of the terrace at Bosscherheide

In the area between Arcen and Bergen the terraces are partly of Rhine partly of Meuse origin (Zonneveld, 1956). At Bosscherheide the substratum is formed by the Well Sands, which have a Rhine heavy mineral assemblage. According to Zonneveld (1956, 1958) these gravels and coarse sands were deposited in the period in which the expanding Saalian inland ice forced the Rhine to divert its course in westerly directions; the depressions between the terraces of Walbeck and Twisteden would have functioned as the southern most escape route of the Rhine waters. During the Weichselian Middle and Late Pleniglacial these Rhine deposits have been eroded and reworked by a braided system to form the Kreftenheye Formation (Pons, 1957; Teunissen, 1983). The Well Sands of Saale age are still present in the eastern part of the exposure, while the Kreftenheye Formation is found in a somewhat lower position in the western part of the pit.

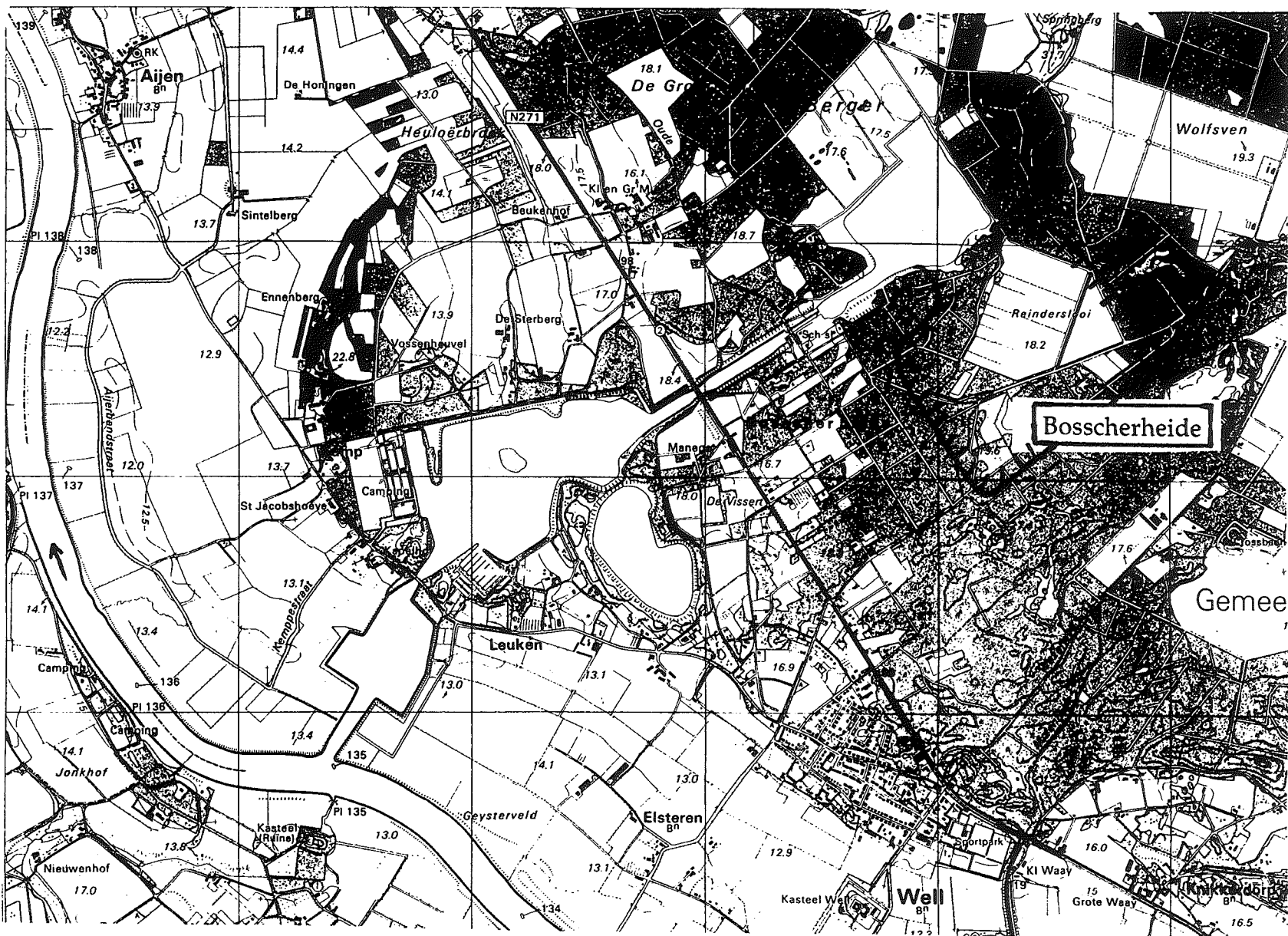


Fig. 1 Location map of excursion stop Boscherheide on the east bank of the Maas.

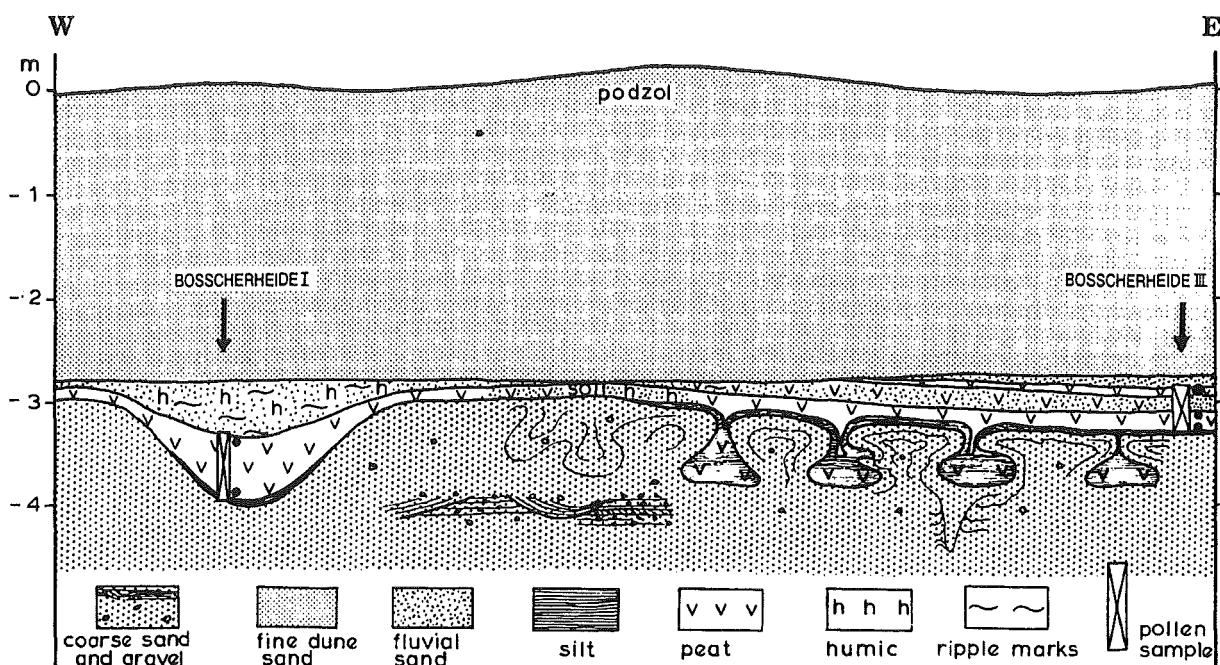


Fig. 2 Schematic profile of exposure Bosscherheide showing the location of the pollen samples. The black dots refer to radiocarbon-dated levels (taken from Bohncke et al., 1993).

Bosscherheide I from top to bottom:

- 10,940 ± 60 BP
- 11,300 ± 60 BP
- 12,100 ± 70 BP
- 12,110 ± 70 BP

Bosscherheide III from top to bottom:

- 10,500 ± 60 BP
- 10,880 ± 50 BP
- 11,500 ± 50 BP

A dominant feature of this Late Pleniglacial fluvial unit is the quick alternation of gullies. Many gullies contain horizontal, sharply bounded sets of sand and gravel. These relatively thin sets often show an internal oblique lamination and may be interpreted as channel bars. Other gullies are filled completely with obliquely laminated sands, caused by lateral accretion. The described sedimentary structures point to a high accumulation rate in a multi-channel river system.

This fluvial unit is capped by a sandy grey loam at the time the fluvial system merged into an incising single-bedded channel, transforming in this way the earlier fluvial deposits into a terrace. This terrace forms part of the Meuse lower terrace (see also Teunissen, 1983). In the abandoned gullies organic sediments started to accumulate. Virtually no time seems to have elapsed between the abandonment of the gully and the first accumulation of organic matter, since aquatic species are abundant in the bottom most samples of the infill (see Fig. 3, the pollen record).

The pollenrecord Bosscherheide I and BH III; chronology and paleoenvironment of the Late Glacial sediments (Figs. 3 and 4)

Biostratigraphically registration in the abandoned gully of Bosscherheide starts with the Bølling (Fig. 3: zone 1b and 1c) as demonstrated by the high *Betula* values in these zones. 7,5 km southeast of Bosscherheide, Teunissen (1983) obtained the oldest radiocarbon date of $12,760 \pm 150$ BP (GrN-4478) for the transition between the underlying thin sandy loam and the organic channel infill. A similar date ($12,600 \pm 60$ BP) has been obtained from the organic infill of an abandoned braided channel at Notsel (Bohncke et al., 1987).

The relatively high *Betula* values show a sudden drop at the transition to the Older Dryas at the base of zone 2 (Fig. 3). The top of the Bølling has been dated here at $12,100 \pm 70$ BP. The Older Dryas is characterized by a herbaceous rich pollen assemblage in which *Artemisia* is frequently encountered. Subsequently, *Betula* gradually restores (Fig. 3: zone 2 *Betula*-phase) after which *Pinus* becomes the dominant species in the vegetation cover (zone 2 *Pinus*-phase). Zone 2 reflects the *Betula*- and the *Pinus*-phase of the Allerød.

The lateral extension of the peat over the adjacent terrace has been studied in a site similar to Bosscherheide III (see Fig. 2) and it reveals the *Pinus*-phase of the Allerød (Fig. 4: zone 2) followed by a herbaceous rich vegetation (Fig. 4: zone 3), where the lithology changes into a fluvial loam. More distal from the alluvial plain of the Maas this Allerød peat changes into a sandy, charcoal rich soil, but the overlying fluvial loam of Younger Dryas age remains a consistent feature.

The top of the Allerød peat has been dated in the gully and on the adjacent terrace providing dates of $10,940 \pm 60$ BP and $10,880 \pm 50$ BP respectively. At Notsel the termination of the Allerød was dated to $10,970 \pm 50$ BP and south of Bosscherheide, Teunissen (1983) obtained a date of $10,870 \pm 100$ BP. These radiocarbon dates unambiguously put the overlying fluvial loam in the early Younger Dryas period.

At the transition to zone 3 (Figs. 3 and 4) the *Pinus* forest suddenly drowns as appears from the return of aquatic species and the deposition of fluvial loams over the interstadial peat. The reactivation of the fluvial system is clearly demonstrated by the occurrence of *Classopollis*, an element of the Tertiary brown coal layers, in the pollen record. *Azolla tegeliensis* was encountered in the fluvial loams that cap the interstadial Bølling/Allerød peat at Notsel (Bohncke et al., 1987) emphasizing the regional occurrence of this process of reactivation of the fluvial regime during the early Younger Dryas. The lateral extension of the Younger Dryas loam reflects large-scale floodings and deposition of suspension load in large shallow lakes (Pons & Schelling, 1951).

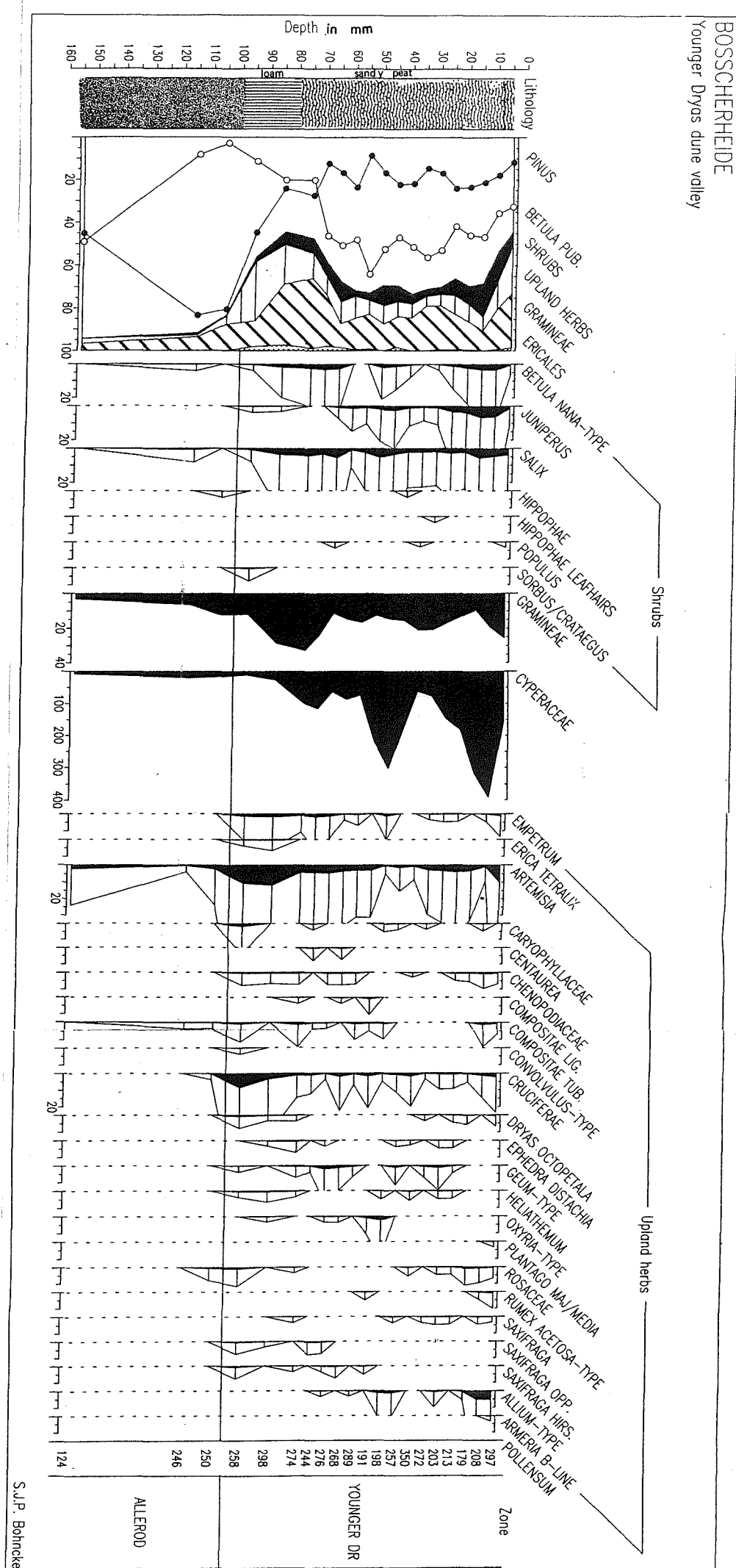
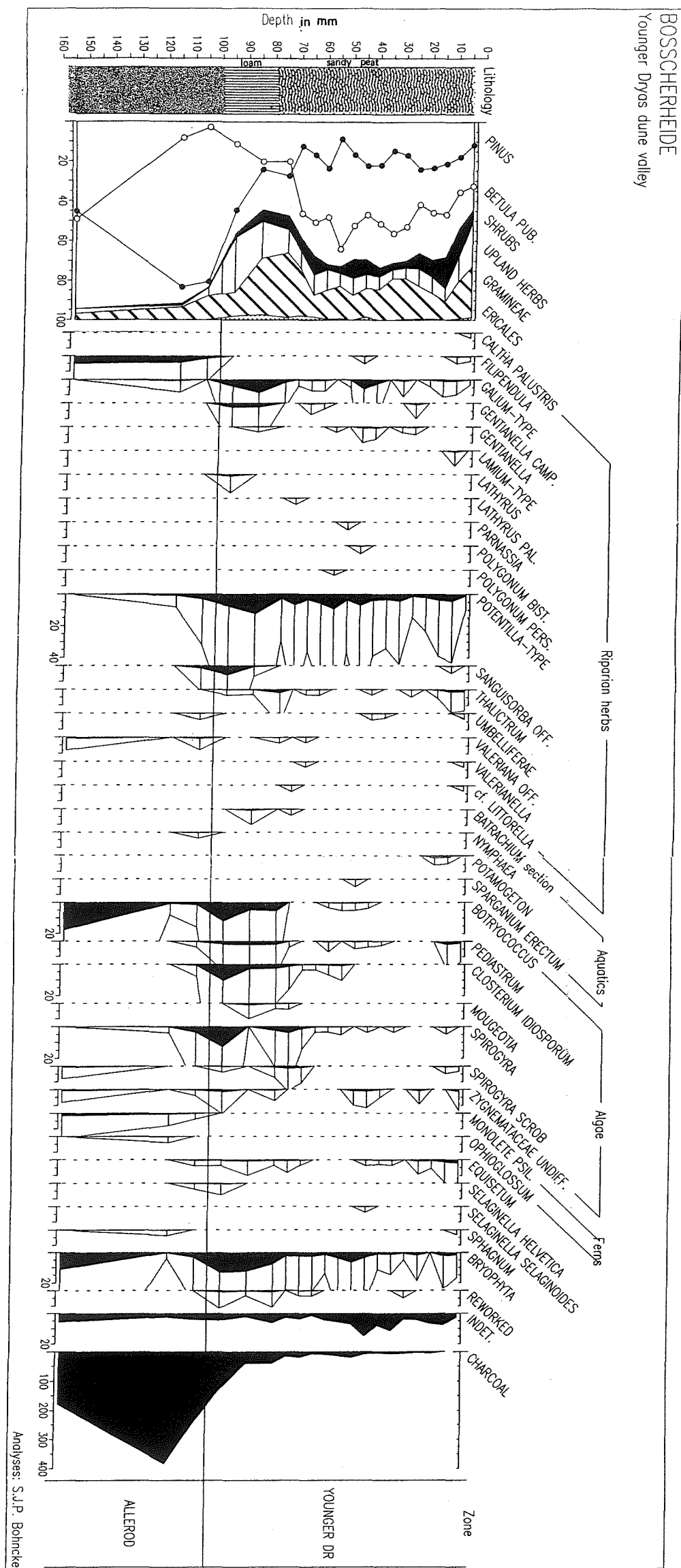


Fig. 4 Pollen diagram of Allerød soil and Younger Dryas loam and peat adjacent to the abandoned channel in a position similar to Bosscherheide III (see Fig. 2).

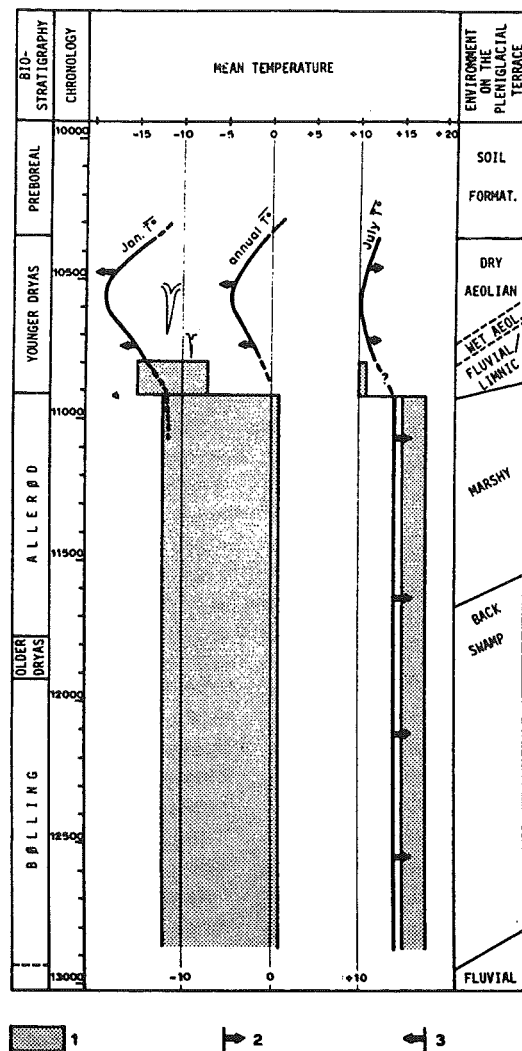


Fig. 5 Tentative reconstruction of temperature fluctuations during the Late Glacial based on sequence Notsel (Bohncke et al., 1987). The arrows correspond to the levels with specific temperature indicator species within the peat section at Notsel, based on Coleoptera (1), palaeobotanical evidence (2) and periglacial phenomena (3).

The profiles at Bosscherheide show that, subsequent to the deposition of organic matter during the Bølling and Allerød and preceding the deposition of the overlying fluvial loam, a period with deep seasonal frost or incipient permafrost established (Figs. 6 and 7). According to the ^{14}C -dates provided by profile Bosscherheide III, where the top of the Allerød peat had been dated, this local permafrost established directly following 10,880 BP (see below). One may assume that the local permafrost could have installed already in due course of the Allerød, but this is strongly contradicted by the temperature curve from the peat sequence at Notsel (see Fig. 5).

With the deposition of fluvial loams on top of the Allerød layer the subsoil thawed, leaving nicely developed loading structures (see Figs. 6 and 7). The loam, although partly filling the morphology that established following the permafrost degradation, is horizontally laminated and hence postdates the fase with permafrost. Another argument for the fact that permafrost degraded during or preceding the deposition of the Younger Dryas loam is the absence of segregation ice in the loam, whereas this feature is present in the underlying organic deposit and Late Pleniglacial fluvial loam as appears from the study of thin sections of the deposits (see below: Periglacial features).

More distal these Younger Dryas fluvial deposits are intercalated by thin moss layers and clayey gyttjas and peat. This sediment series represents the end of the fluvial activity at this location. On top of this sequence an aeolian layer of varying thickness and consisting of horizontally bedded and low-angle cross-bedded sands has been deposited. The start of the aeolian activity at this location could be dated by the moss layer directly underlying the aeolian deposits (Figs. 4 and 6). A date of $10,500 \pm 60$ BP has been obtained, and places the aeolian activity at this location in the late Younger Dryas.

Periglacial features

The upper part of the terrace sands and gravels and the lower part of the overlying fine-grained, organic layer are deformed by a regular pattern of well-pronounced involutions (Figs. 6 and 7). The mean amplitude is ca. 60 cm. The updoming parts have a diapir-like form and the sinking parts reach a remarkably constant depth (Fig. 6). Comparison with similar involution structures allows the interpretation as load casts (Vandenberghe & Van de Broek, 1982). The horizontal boundary between the flat-bottomed involutions and the underlying, actually very permeable sediments may be interpreted as a former permafrost table. Although the amplitude of the involutions is relatively small in comparison to cryoturbations from other sites in the same region (Vandenberghe & Krook, 1981 and 1985), the very regular development of the involutions testifies of their single development and excludes a multiple formation, for instance as a recurrent process in yearly thawing soils (French, 1986).

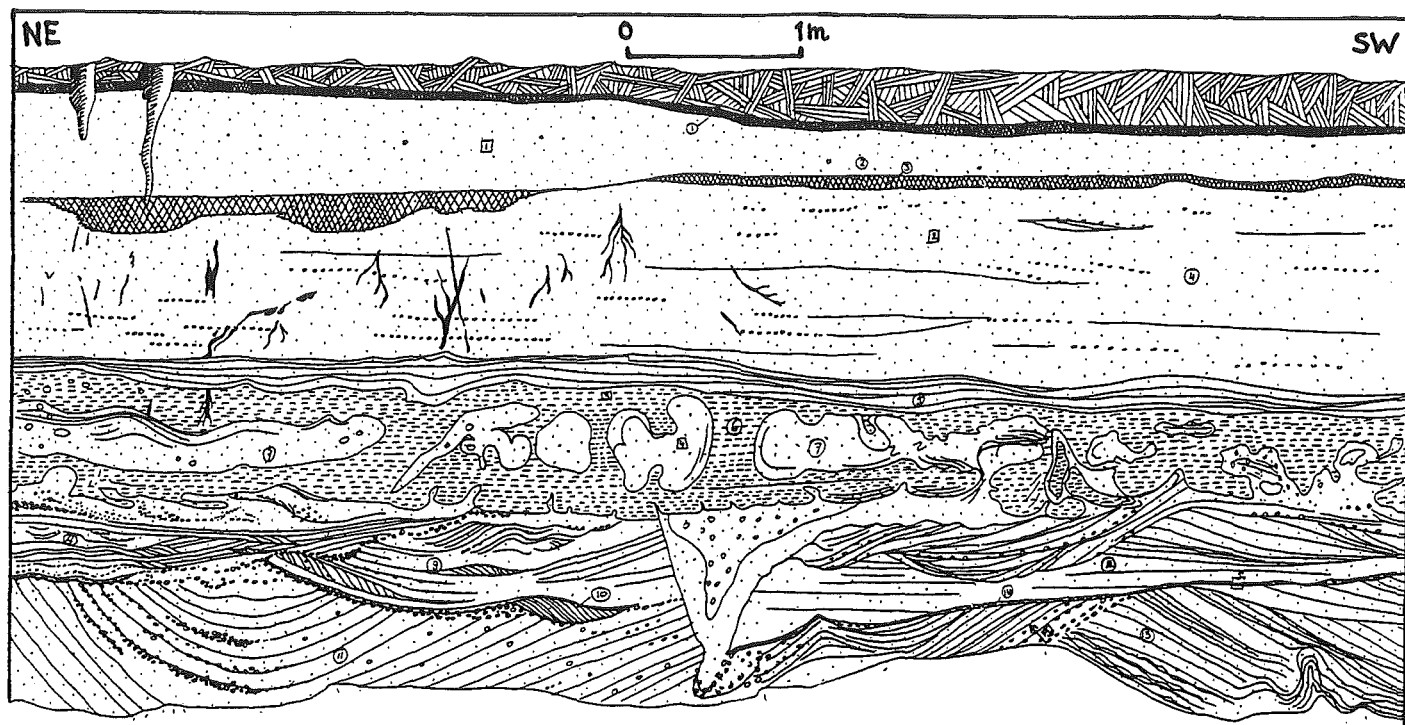


Fig. 6 Drawing of a profile face at Bosscherheide showing a) Late Pleniglacial to early Late Glacial braided sediments, b) Younger Dryas involutions and initial ice-wedge cast and c) late Younger Dryas dune sands.

The generation of the described involutions on top of a degrading frozen subsoil may further be argued by the wedge forms just below the involution zone (Fig. 6). They reach a depth of 45 cm and a maximum width of 15 cm. Very distinct upturning has been observed, but downturning occurs as well. The infilling with completely homogenized sand sharply contrasts with the surrounding distinctly bedded sediments. The limited depth of these wedges (c. 1 m from the former surface) and the absence of subsidence features (faults and slump structures) questions their interpretation as ice-wedge casts. The absence of vertical lamination and the only weakly upturned edges reject an interpretation as sand wedge casts. A similar wedge form, but the upper part not considerably disturbed by cryoturbation, has been described by Vandenberghe et al. (1987) at Notsel in the Mark valley c. 100 km to the west. It could be interpreted either as an intensely developed frost fissure, formed during numerous successive winters in a deep (min. 1 m) seasonal frozen ground or as an initial form of ice-wedge cast. Given the considerations about the overlying cryoturbation zone, the latter interpretation now seems more probable. It is concluded that in this region local permafrost was present for a very short time.

Shortly after the intense cryoturbation, the soil underwent again an involution process, but the resulting deformations - although regularly developed - have a very

small amplitude (max. 30 cm) and a large wavelength (c. 120 cm). Small slump structures occur on the steeper parts of the undulations. The described forms resemble "earth hummocks" (f.i. Tarnocai & Zoltai, 1978) and "thufurs" (f.i. Schunke, 1977). Besides, the undulations at Bosscherheide consist of a peat and a peaty loam layer at their top as do thufurs and earth hummocks. It is, however, at present not clear if the same processes are involved.

Thanks to radiocarbon analyses of surmounting and underlying peat layers the periglacial phenomena can be dated in detail. The rate of peat growth of the main peat layer could be calculated from the radiocarbon dates of $10,880 \pm 50$ BP and $11,500 \pm 50$ BP in Bosscherheide III (see Figs. 2 and 7). It follows that the peat formation ended at c. $10,750 \pm 50$ BP. The intense cryoturbation took place just after this peat formation. An overlying peat layer, not affected by cryoturbation, is dated at $10,500 \pm 60$ BP. Assuming the same growth rate the latter peat formation started at c. $10,550 \pm 50$ BP. Thus the degradation of the frozen layer, manifested by the intense cryoturbation, took place between 10,550 and 10,750 BP. Also the later formation of weaker undulations took place before $10,550 \pm 50$ BP. The age of the (local) permafrost has to be situated shortly before that cryoturbation phase. It has to be stressed that all these periglacial phenomena may be formed in a short time. The described wedge form may, according to its maximum width of 15 cm and an assumed mean growth rate of 0,5 à 1 mm/yr be formed in 150 to 300 years. Involutions on top of a frozen soil are due to loading which is an instantaneous process during the melting of the frozen subsoil (Vandenberghe, 1983).

Finally, the degradation of the (local) frozen subsoil is not necessarily the expression of a climatic amelioration. Especially in valleys the talik underneath the river may move laterally causing the melting of the neighbouring soil (Bryant, 1983). This seems likely at Bosscherheide since the Maas river plain was nearby. According to the micromorphological analyses, however, permafrost did not return afterwards. In the Mark valley at Notsel, the above mentioned wedge has developed and degraded also shortly before the dune formation which started probably around 10,450 BP. As a conclusion it may be derived that the coldest phase of the Younger Dryas has to be situated between c. 10,800 and 10,500 BP.

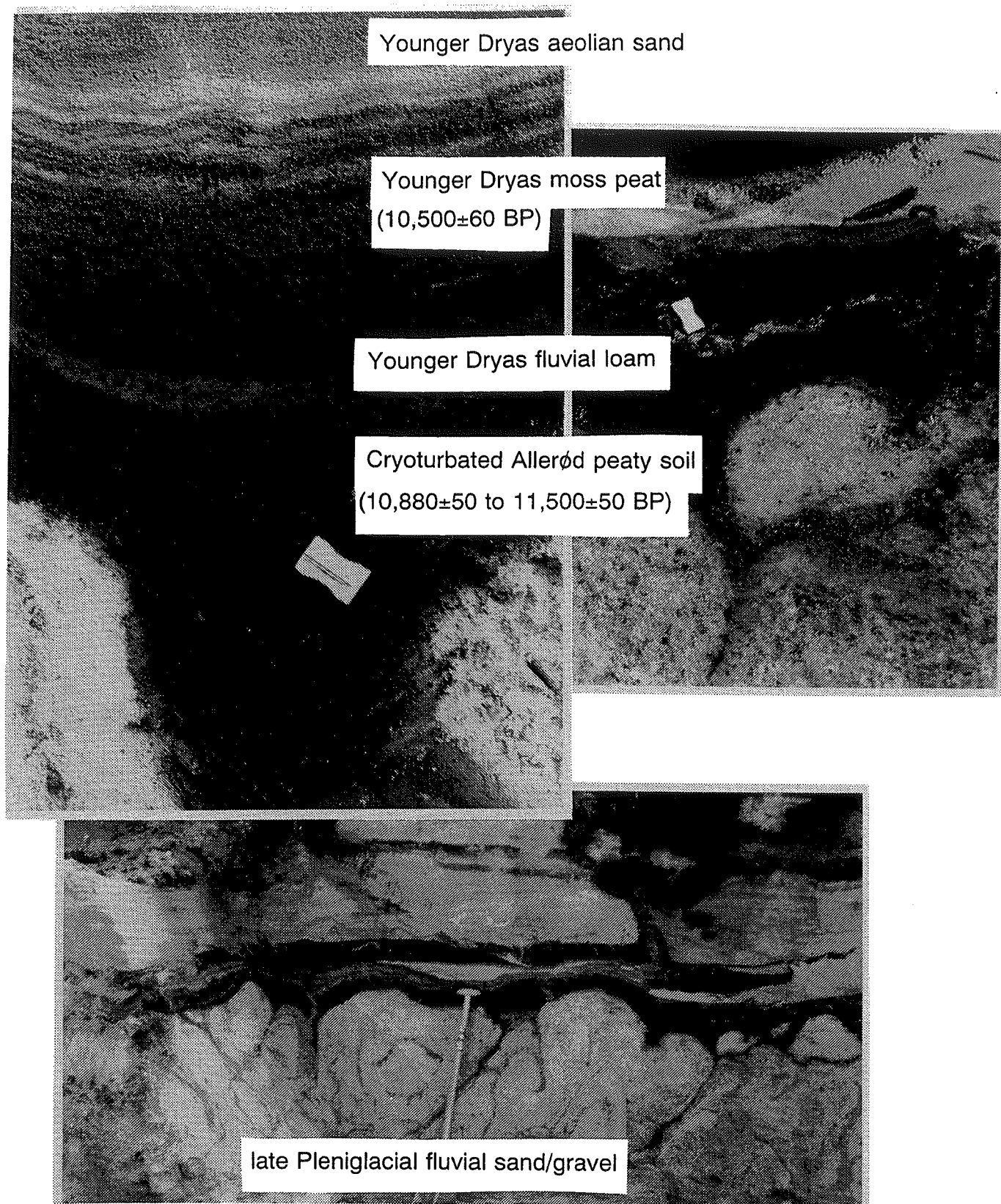


Fig. 7 Younger Dryas involutions at Bosscherheide dated between c. 10,880 and 10,500 BP

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EXCURSION STOP 3: LATE ALLERØD MEANDER SCAR AT MEERLO

The meander scar at Meerlo is situated at the same Late Glacial terrace as the one at Beugen (Excursion stop 5). The scar is up to 175 m wide and the fine-grained infill is up to 5 m thick. 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 the process of meander chute cut-off. The Younger Dryas floodplain is located less than 1 km north of the meander scar (see Fig. 1 and Excursion stop Kasteelweg-Ooijen, Fig. 3).

The meander has been abandoned due to the rapid changes in the fluvial regime at the Allerød-Younger 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 (see Excursion stop 4: Kasteelweg-Ooijen). A more straight and wide (multi-) channel floodplain was required for the larger peak discharges that characterized the climate at the time of the chute cut-off.

During the higher Younger Dryas peak discharges the abandoned Allerød channel of Meerlo was still flooded and here deposition of fines from suspension load took place (see lithology of the pollen record in Fig. 2). The fine-sandy clay at the base of the meander infill reflects this episode of peak discharges straight after the cut-off. The scar infill reveals a fining-upward sequence. The sandy clay at the base gradually changes into humic clay and gyttja. The clay is calcareous and locally contains ostracods, fresh water snails and burrowing structures.

The pollenassemblage (Fig. 2) shows relatively high values for the Non Arboreal taxa, among which the grasses and the sedges are dominant. Furthermore, the relatively high values for the *Artemisia* and the continuous presence of *Empetrum* are indicative for the **Younger Dryas** chronozone (Fig. 2: 350-520 cm). The presence of *Classopollis*, a reworked taxa from the Lower Cretaceous beds in the upper course of the Maas, characterizes the clastic 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 swift morphodynamic adaptation of the fluvial regime, which in turn was forced by a dramatic climate cooling. This change in climate involved an increase in precipitation, predominantly in the form of snow.

At the Younger Dryas-Preboreal transition a change in the lithology from humic silty clay to gyttja is present. *Classopollis*, derived from Cretaceous beds upstream in the Maas catchment, and other reworked species (e.g. *Corylus*, *Quercus*) disappear at this level. These changes in lithology and pollen assemblage indicate a decrease in the flooding frequency of this site by the Maas river.

The reasons for these changes at the Younger Dryas-Preboreal transition are a series of processes involving climate improvement, the related change in fluvial regime from braided back to meandering and incision of the river. The nearby western branch of the Younger Dryas braided system became abandoned, as has been established at Excursion stop 4 (Kasteelweg) (see Fig. 1 for locations). The eastern branch of the Younger Dryas system near Ooijen (Fig. 1) became the principal river course and developed into the early Holocene floodplain (see Excursion stop 4: Ooijen). Because of these hydrological changes the Meerlo meander scar became situated more than 3 km from the Preboreal floodplain, in contrast to only 1 km during the Younger Dryas. The distal position of the meander scar in combination with the early Holocene incision evidently led to the changes in the sediment and pollen record at this level (Fig. 2).

The **Preboreal** (Fig. 2: 268-350 cm) zone starts with a distinct lacustrine environment with high values for the aquatic taxa. The pollen record 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 $9,750 \pm 50$ BP. The local hydroseral succession at this time reached the semi-terrestrial phase, leading to the formation of peat. At the top of the pollen record (Fig. 2: 240-268 cm) the increase in autochthonous *Corylus* pollen determines the transition to the **Boreal** (c. 9100 BP).

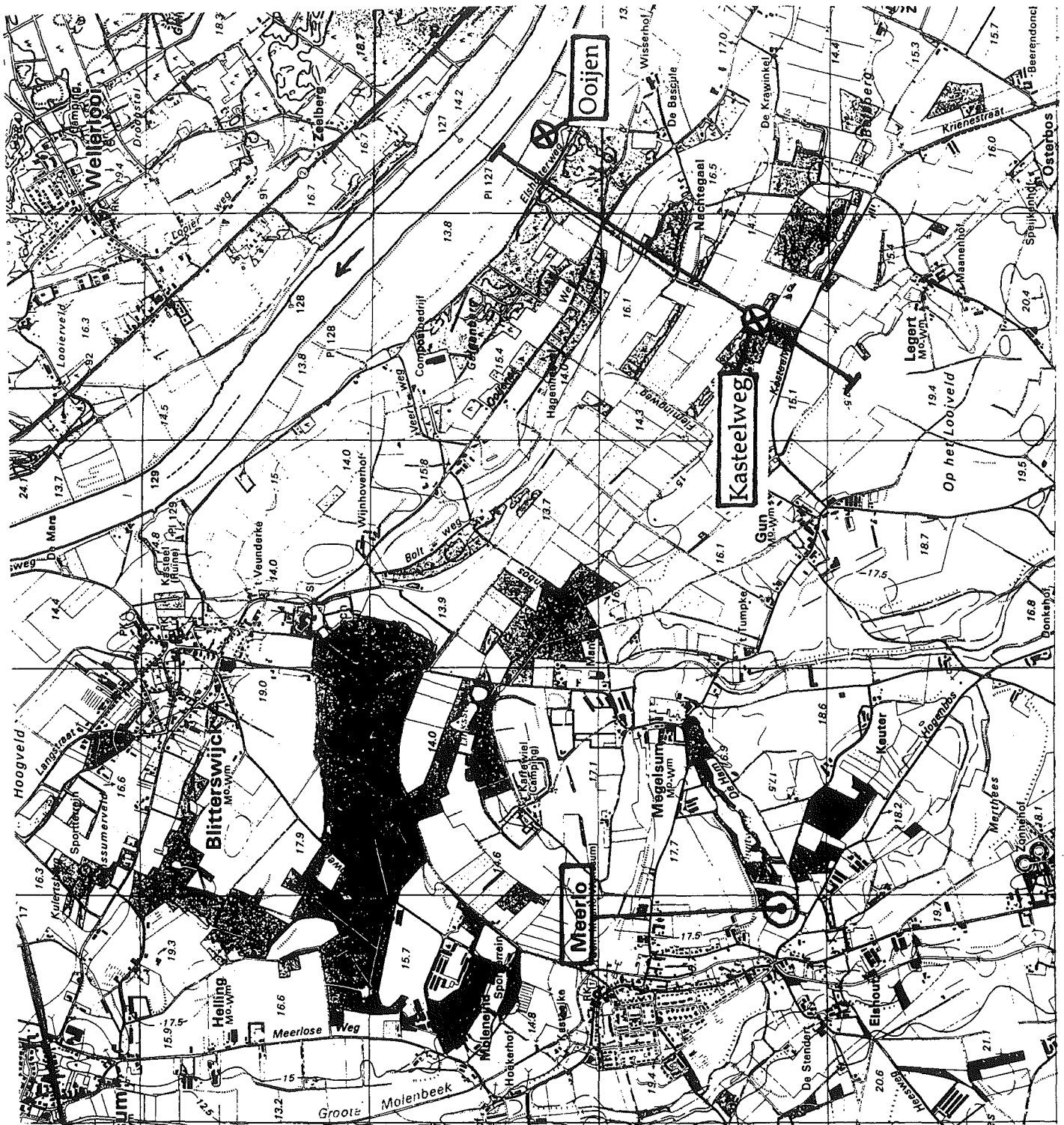


Fig. 1 Location map of the meander scar at Meerlo.

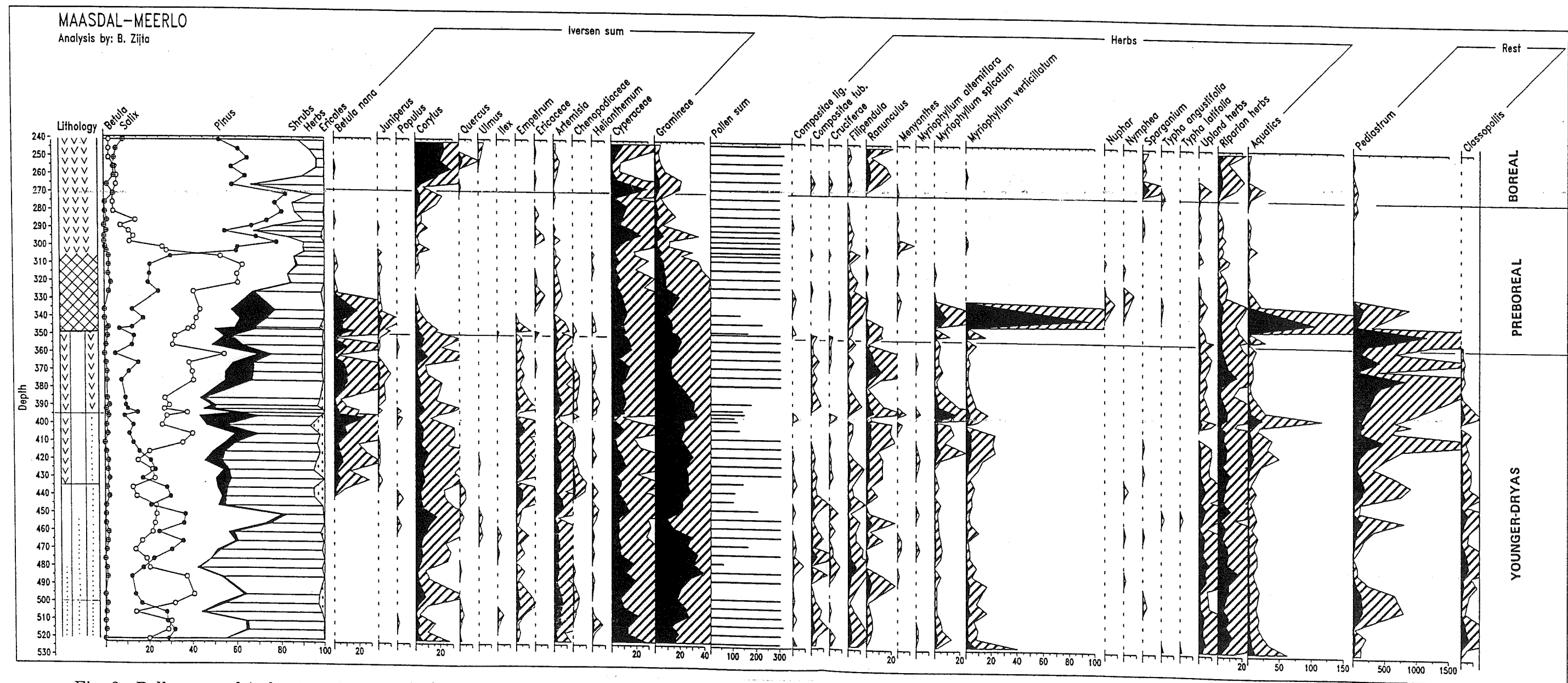


Fig. 2 Pollen record (selection of taxa) from the infill of the meander scar at Meerlo.

EXCURSION STOP 4: YOUNGER DRYAS AND HOLOCENE FLOODPLAIN AT KASTEELWEG-OOIJEN

Introduction

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. 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, in press):

1. Change in channel pattern. The Younger Dryas floodplain is generally 1 km wide and characterized by a 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 led to lower evapotranspiration and higher peak discharges. The morphological change was accompanied by 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 sandy clay during the Younger Dryas.

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 (Excursion stop 2) a fluvial loam was deposited over the Allerød soil/peat and the Younger Dryas involutions.

4. Lithological change in abandoned channels. Because of the higher peak discharges, the Maas was able to inundate not only older terraces but also abandoned meander channels of the Allerød period. At Beugen (Excursion stop 5) a gyttja of Allerød age at the base of the infill is erosively overlain by a gray fine sandy clay. 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 parabole dune morphology indicates a westsouthwesterly wind during their formation in the late Younger Dryas. At Bosscherheide (Excursion stop 2) the aeolian sediments overlie a peat layer dated at $10,500 \pm 60$ 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 multi-channel braided system. During low discharge, sediment could be deflated from this wide south-north oriented Maas floodplain by the westerly wind.

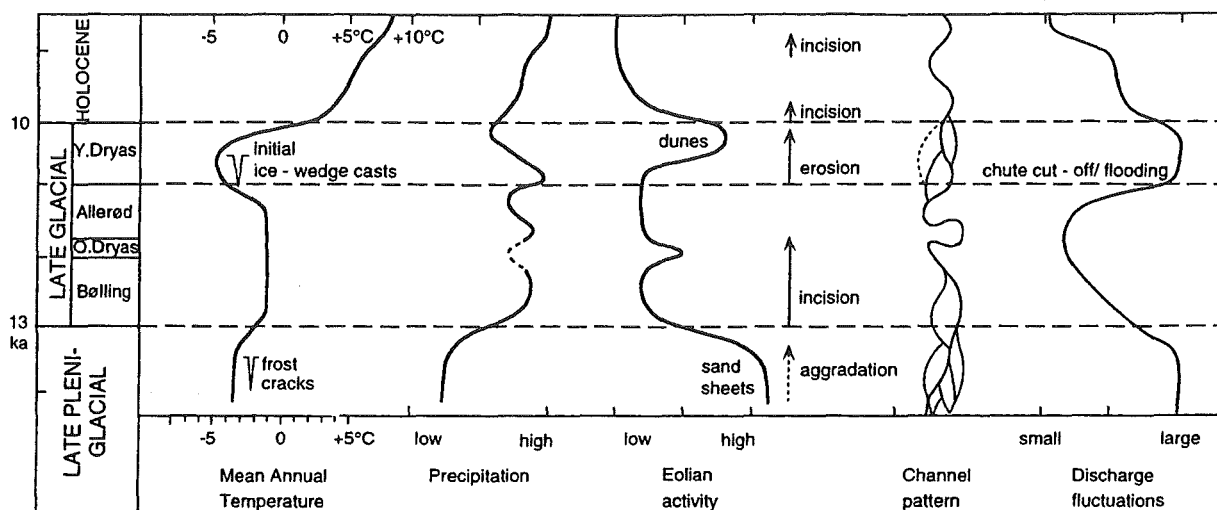


Fig. 2 Summary of the Late Glacial climatic and environmental changes in the Maas valley (from Kasse, in press).

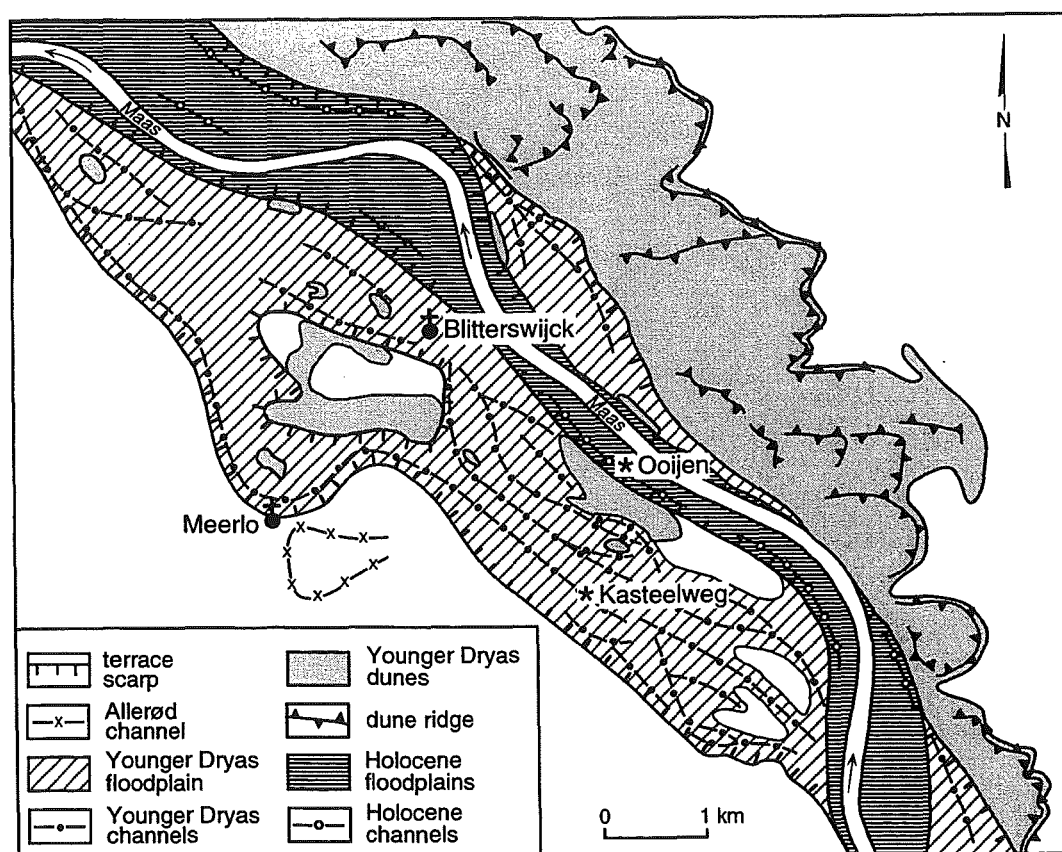


Fig. 3 Morphological map of the Younger Dryas braided floodplain and Younger Dryas dune field at Blitterswijk (from Kasse, in press).

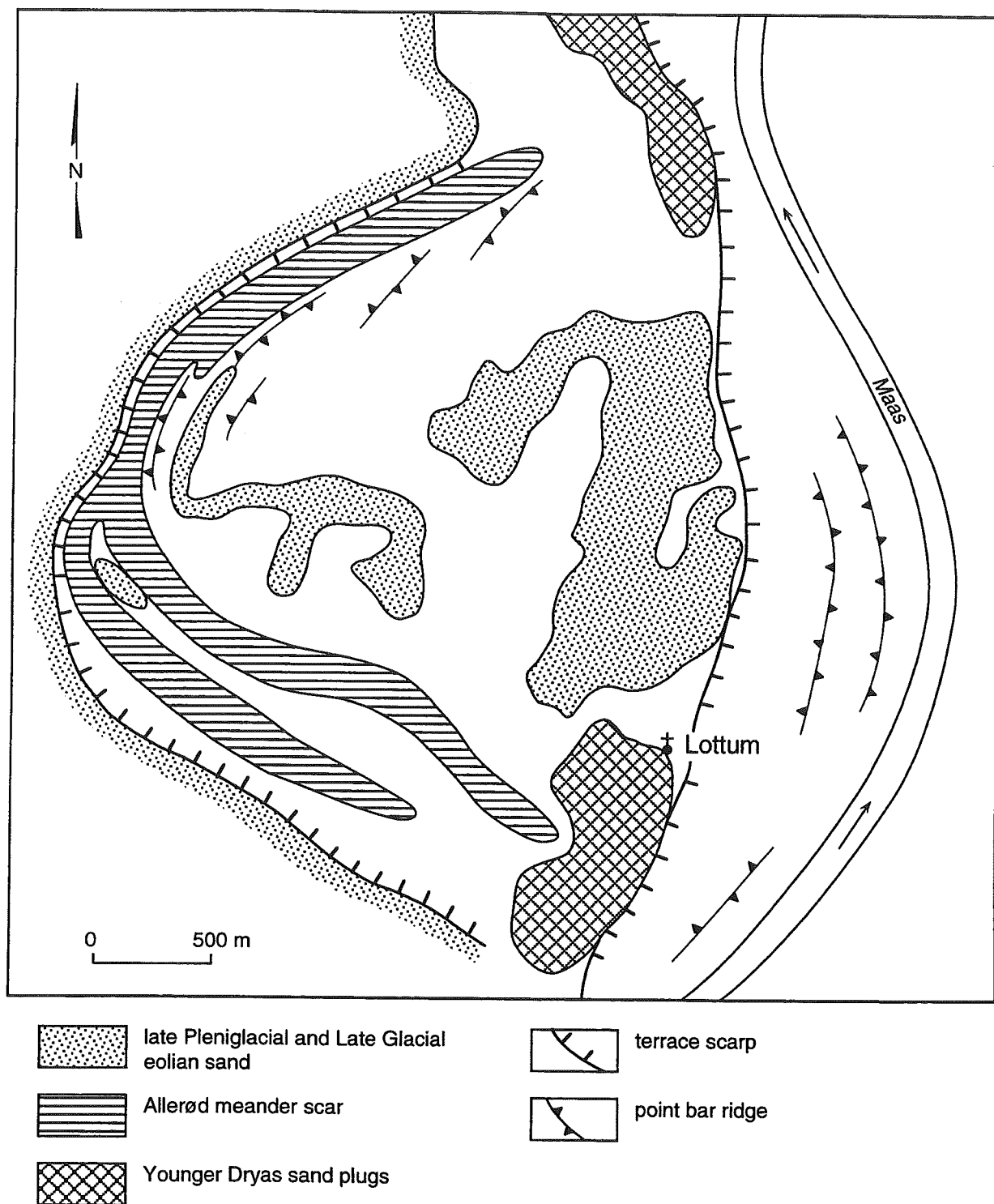


Fig. 4 Younger Dryas chute cut-off of the Allerød meander at Lottum (from Kasse, in press).

Geomorphology and fluvial development

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

1. The high-sinuosity meander scar terrace in the west. It consists of gravelly sands fining upward into sand and it dates from the early Late Glacial to the Allerød period.
2. The Younger Dryas channel system, underlain by gravelly sands. Local organic channel fill dates from the late Younger Dryas (Westerhoff & Broertjes, 1990: pollen diagram Blitterswijk 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 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 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. Corings in area 3 revealed the existence of older Late Glacial meander sediments in the subsurface.

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 dune field. 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.

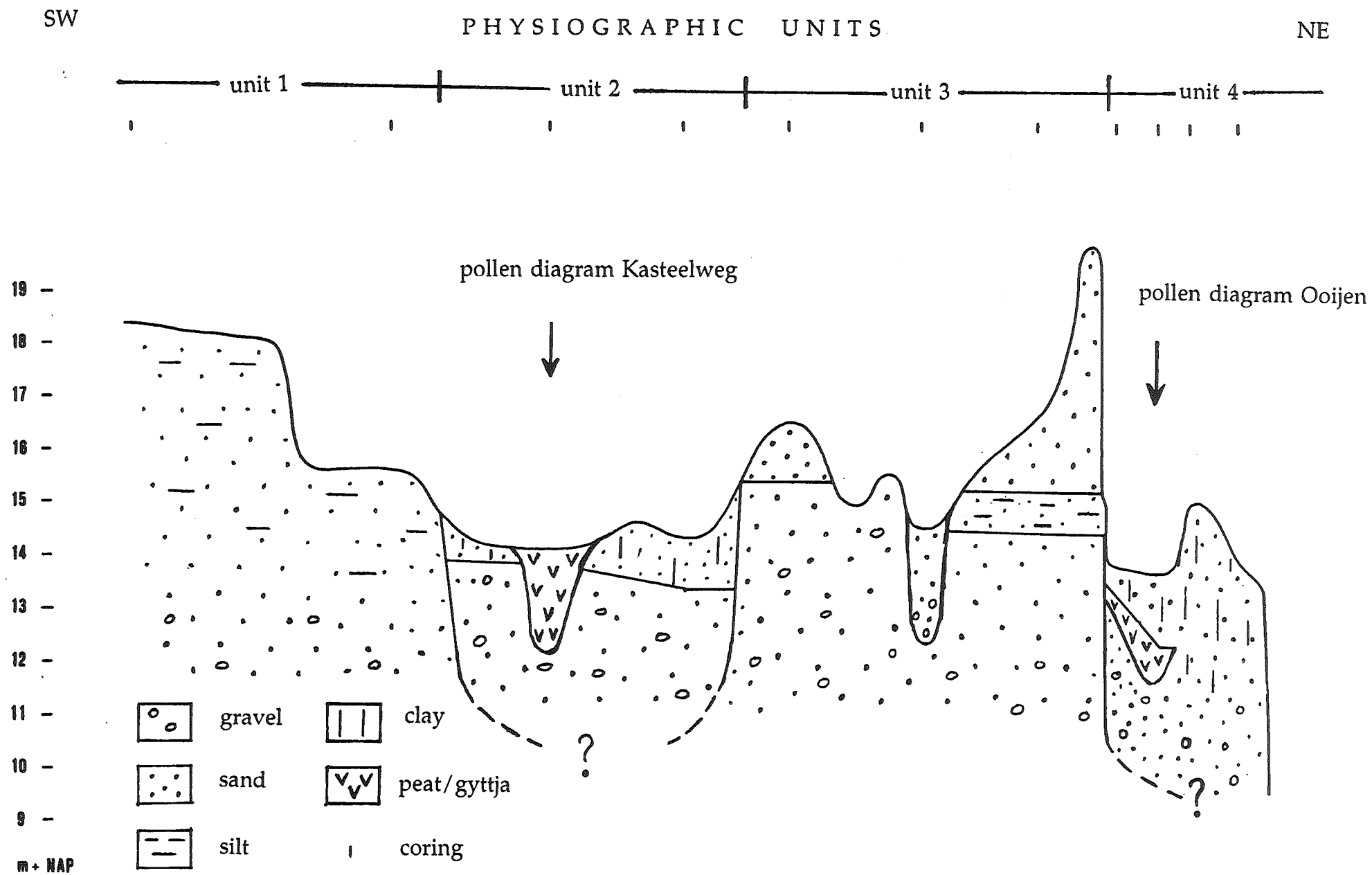


Fig. 5 Schematic cross section and physiographic units of the Younger Dryas and Holocene floodplains at Kasteelweg-Ooijen.

0 300 m

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 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 transition towards the Preboreal 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 within the limits of the former wide stream-bed in some depressions organics 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 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. 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.

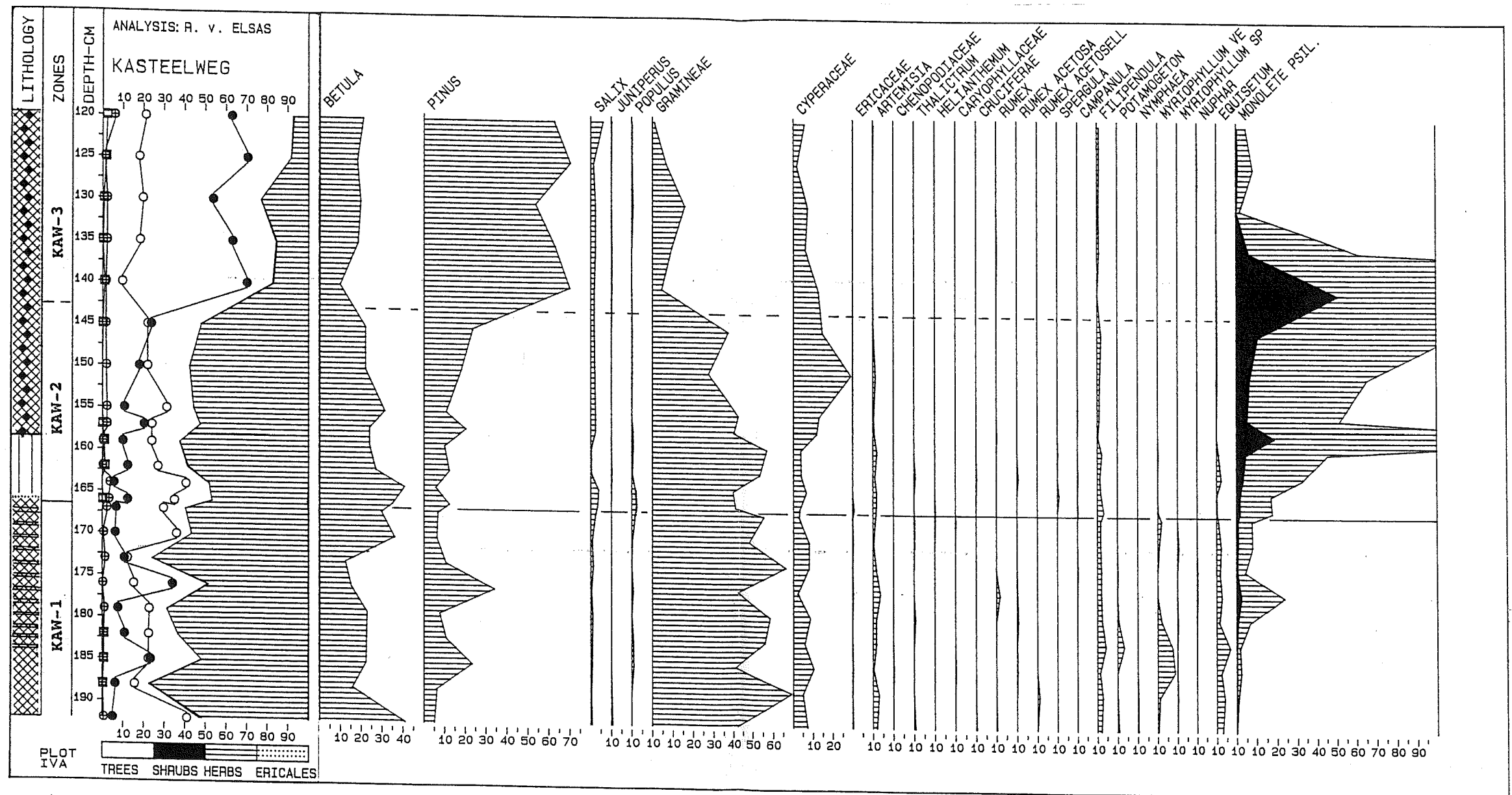


Fig. 6 Pollen record (selection of curves only) from the infill of the Younger Dryas braided channels at Kasteelweg.

The presence of *Populus* at this level in the sequence 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 hydrosereal 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 (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**, a hiatus towards the early Holocene and the later part of the **Preboreal**. At the start of the **Preboreal** (c. 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 then active early Holocene river system.

The Holocene flood plain at Ooijen

Two cores were taken in the early Holocene river plain at Ooijen (Figs. 1, 3 and 5) in order to establish the period during which the channel became abandoned and started to fill in with organogenic sediments (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 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 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. Further research may elucidate this problem.

Resuming, the biostratigraphical indications as provided by the pollen records from core 11 and 12 demonstrate that the infilling 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.

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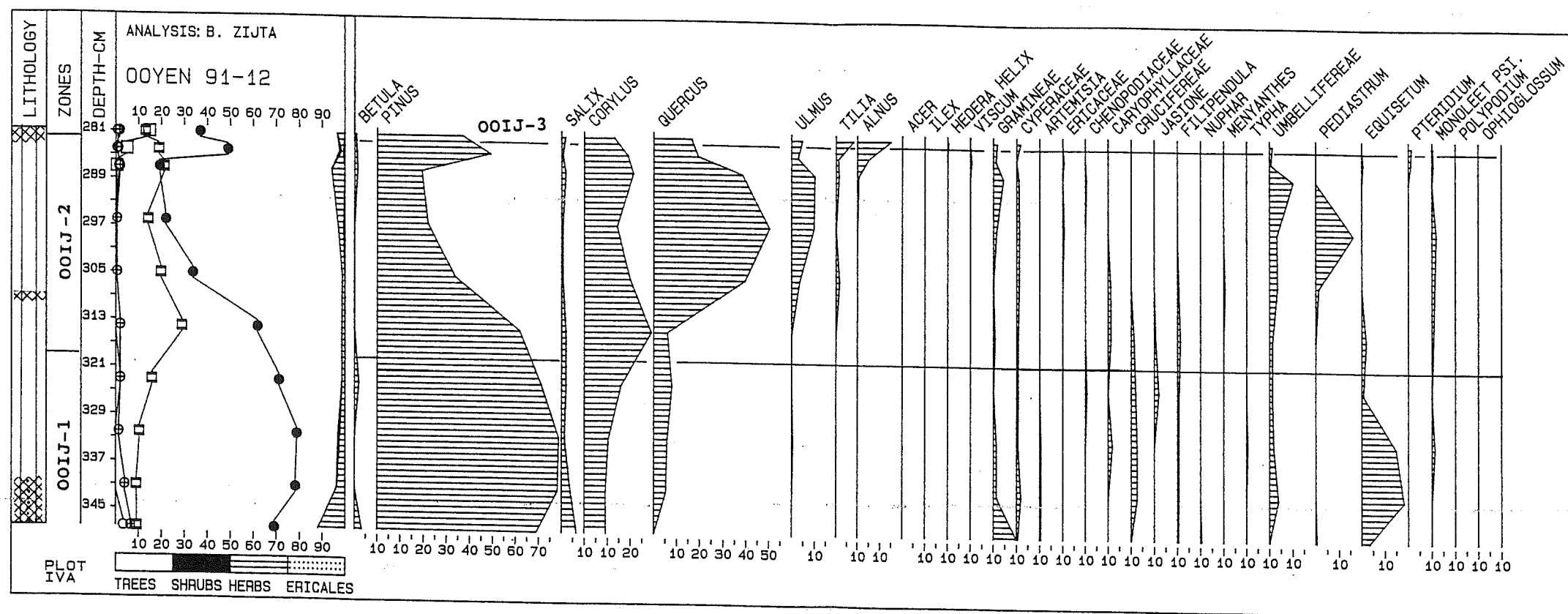


Fig. 7 Pollen record (selection of curves only) from core 91-12 in the Holocene river plain at Ooijen.

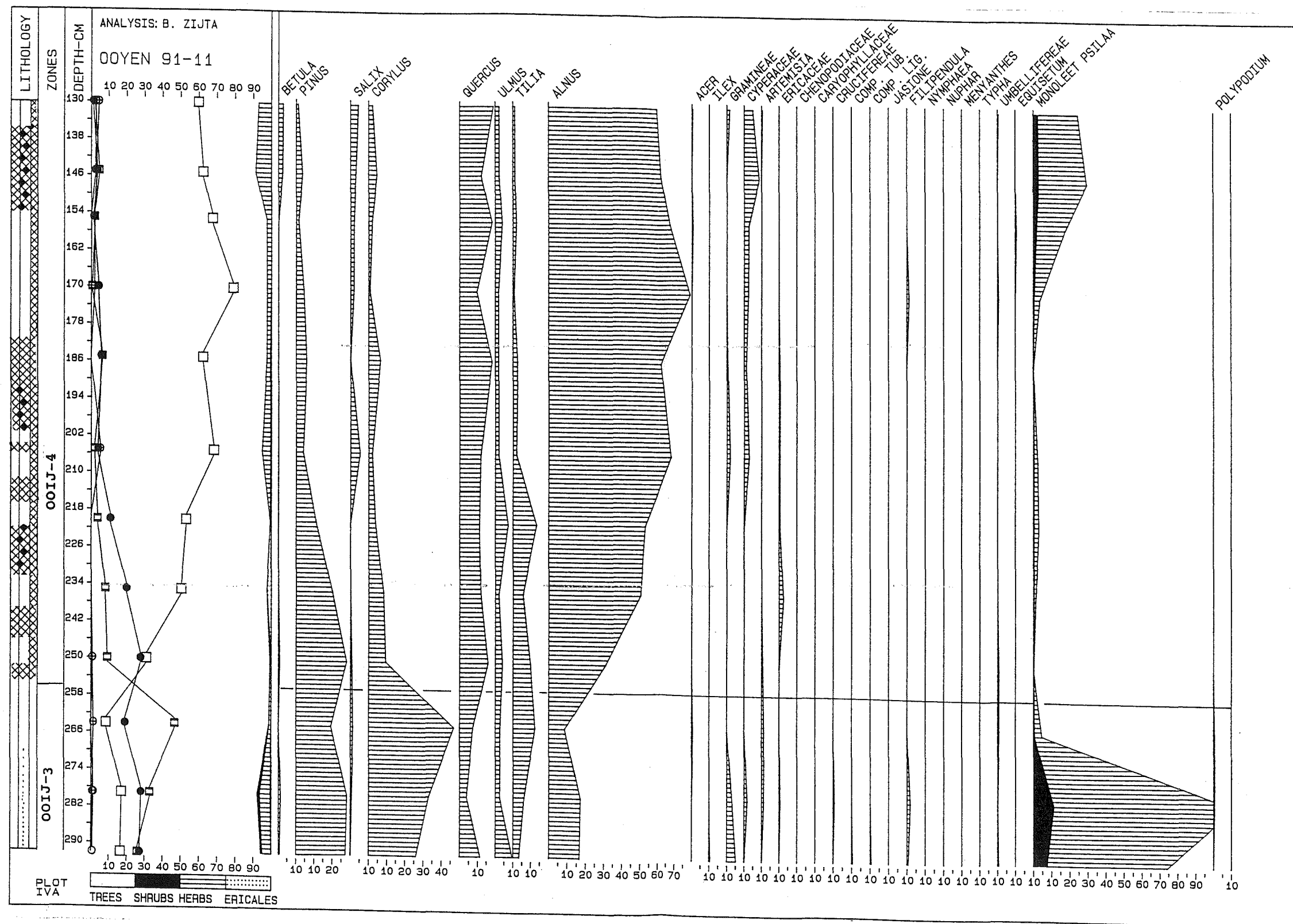


Fig. 8 Pollen record (selection of curves only) from core 91-11 in the Holocene river plain at Ooijen.

EXCURSION STOP 5: ALLERØD MEANDER SCAR AT BEUGEN

Introduction

Beugen is located on the left bank of the Maas, 20 km southeast of Nijmegen (Fig. 1). 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). 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. 2). However, erosion did occur during the transition from the braided Late Pleniglacial/Early Late Glacial phase into the Allerød meandering phase (Kasse et al., 1994). Corings in the meander pointbar showed that coarse-grained channel sediments occur at a higher level in the pointbars than in 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.

Infill of the meander scar

The meander scar gives a good impression of the Late Glacial Maas 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 Maas floodplain was characterized by large-scale meanders. The fill was studied by several detailed cross sections (Fig. 3). The fill consists of laminated, fine-detrital gyttja at the base (Fig. 3), abruptly lying on the coarse-grained channel sediment deposited before the moment of cut off (coarse sand and gravel). Palynological analyses (see Fig. 4) 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. 3). This unit represents a phase a renewed fluvial activity in the meander scar. Current velocity in the channel was low, however, (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. 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 (excursion stop 2).

The upper two units of the scar infill (Fig. 3) are probably of Holocene age. The sand layer was probably deposited by a local brook (Oeffeltsche Raam), coming from the southwest and flowing through the meander scar. The peat bed (Fig. 3), commonly a wood peat, is rather thin here. 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.

Pollenrecord of the infill

The results of the pollen analyses are presented in Fig. 4. Four local pollen zones have been distinguished.

Local pollenzone BGN-1 (333-322 cm):

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

Local pollenzone BGN-2 (322-297 cm):

At 321 cm *Pinus* starts to rise. A date for this event at Bosscherheide (Excursion stop 2) comes to 11,300 BP. 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 (297-279 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 repetitive feature in many diagrams of the Maas. At Bosscherheide remnants of birch trees have been revealed from the Late Dryas organic deposits in the backswamp, that were formed shortly after 10,880 BP.

Local pollenzone BGN-4 (279-170 cm):

From 279 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 *Classopollis*, 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 Holocene brook incision in the ox-bow infill.



Fig. 1 Location map of the Allerød meander scar at Beugen.

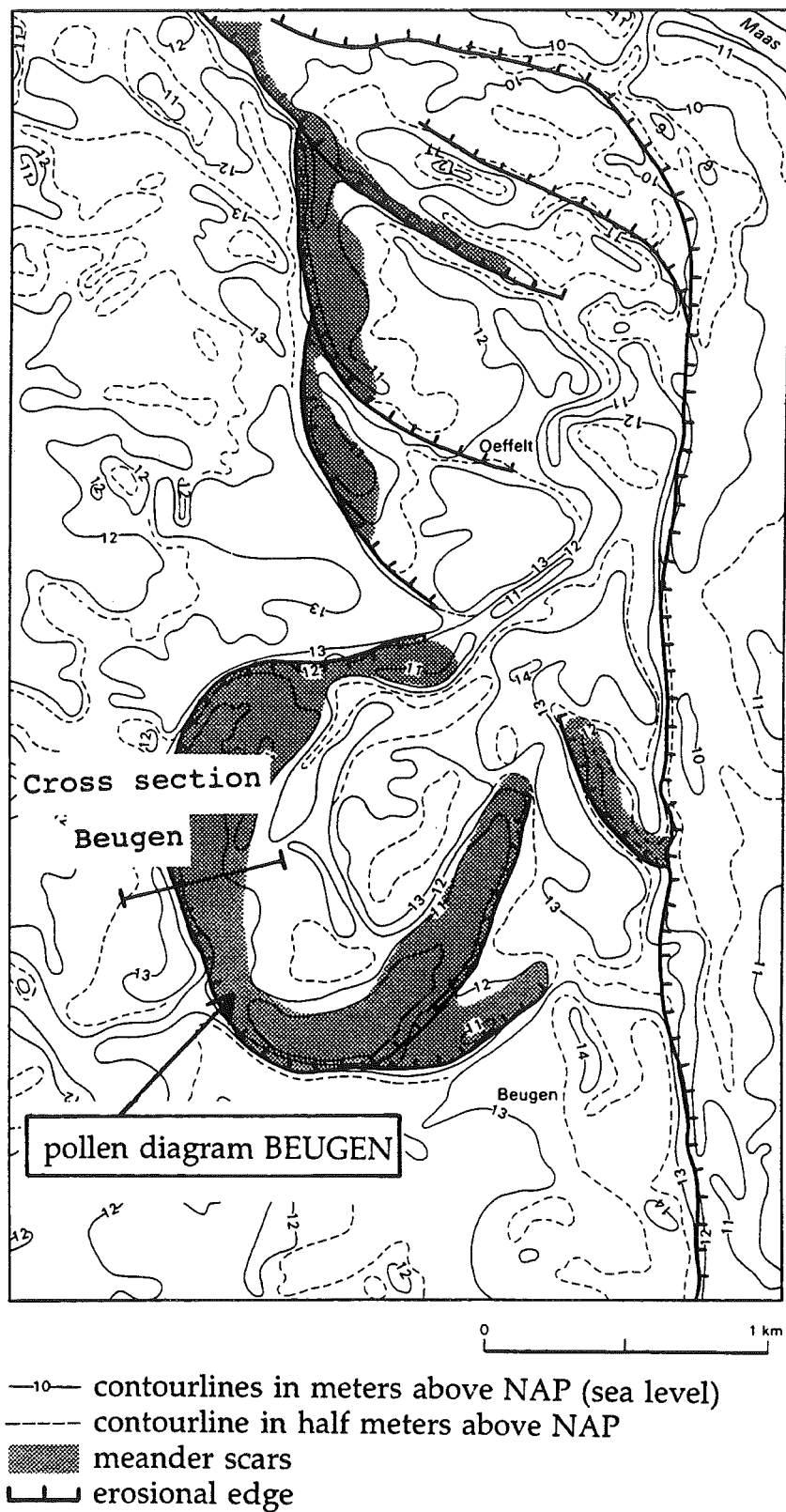


Fig. 2 Morphological map of the Allerød meander neck cut off at Beugen (after Buitenhuis & Wolfert, 1988).

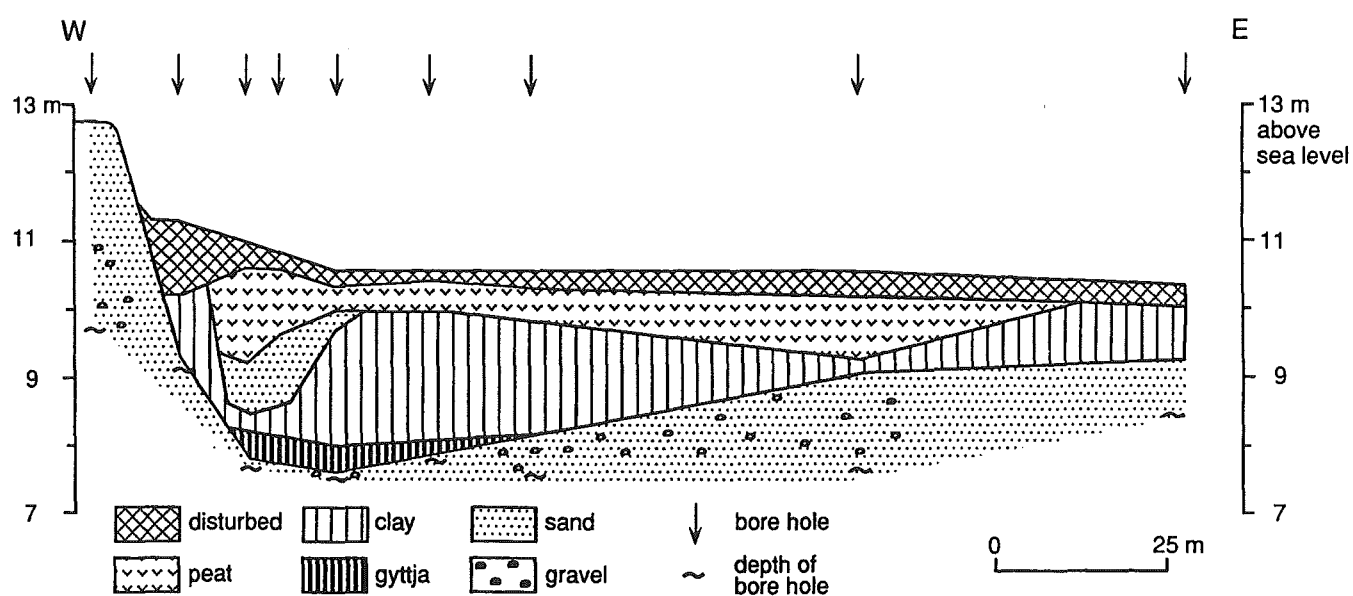


Fig. 3 Cross section over the Late Glacial and Holocene infill of the meander neck cut off at Beugen (from Kasse et al., 1994).

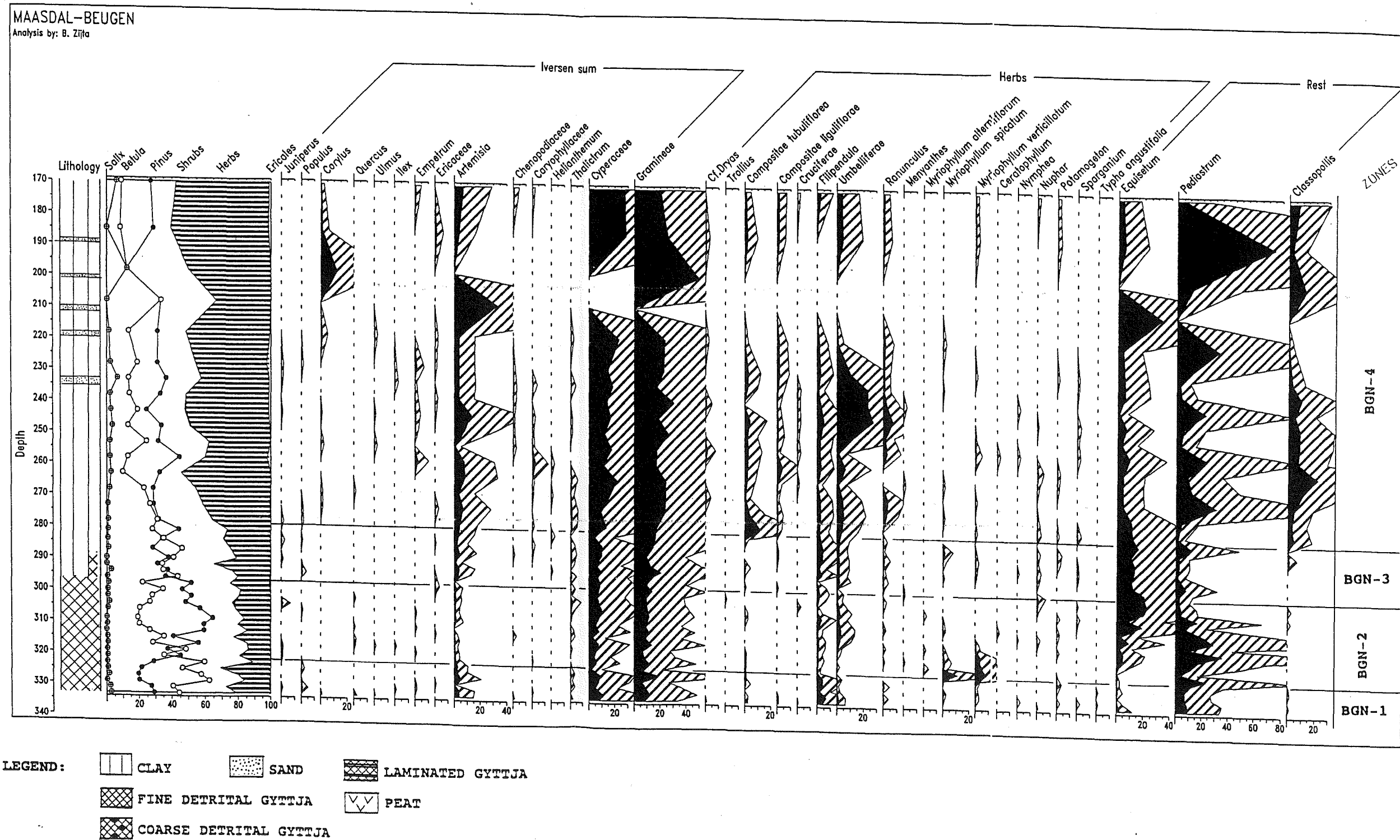


Fig. 4 Pollen diagram of the Allerød and Younger Dryas infill in the meander scar at Beugen.

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Holocene fluvial development and human occupation in the Rhine-Maas delta

QRA-Excursion Sunday 9 April 1995

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THE PEAT DISTRICT

Geology

East of the estuarine creek systems and direct marine influences and west of the river clay district quiet conditions prevailed throughout the Holocene. Extensive peat bogs came into existence in that region, that separated the coastal inhabitable area from the Pleistocene sands. The southern part of this intracoastal peat district stands out as a region with characteristics of its own, that favoured prehistoric occupation and explain the archaeological interest. First, tidal flats and salt marshes did not extent as far to the east as they did to the north of it during the earlier (I-III) Calais phases. Second, the peat bogs are crossed here by the lower courses of rivers that brought eutrophic water and mineral sediment (clay) in the area. High water clays, streambed and levee deposits form marked landscape units that were important for prehistoric settlement. The flow of eutrophic river water prevented the development of oligotrophic peat in this part of the peat district. Third, the pre-Atlantic river dunes are bound in their origin to the hidden Late Glacial Rhine/Meuse valley.

A greater part of this "riverclay/woodpeat area" has been preserved in the post-Roman erosion period. The "Grote Waard" south of the Merwede river, was lost in the storm surge of 1421, but the Alblasserwaard, to the north of it, just escaped this fate.

The "donken" are the outcropping tops of river dunes of Late Glacial - Early Holocene age. They have their bases on the Late Glacial surface that slopes from - 6 m in the east of the peat district to - 12 m in the west of the Alblasserwaard and attain there heights up till + 4 m NAP. The donken are arranged in WNW-ESE orientated rows or groups and are in fact only the highest tops of extensive dune systems of which the lower elevations are completely covered by younger sediments and hidden from the eye. Their steep slopes (up till 20%) are covered with peat and clay under the influence of the rising ground-water table. This process started in the beginning of the Atlantic, when the sea-level did reach this level. The increasing wetness of the river valley impeded further sand blowing and the river dunes became fossilized from then on.

Donken are only found in the former 20 - 30 kms wide valley of the Late Glacial Rhine and Meuse and lack in the coversand districts north and south of it. They are blown up from the river channel sands and rest on the top of a tough and sticky clay that marks the last activity phase of this river system. There are good (palynological) arguments to date this clay to the Late Weichselian (probably the very end) and the Preboreal and (Early) Boreal periods.

In the riverclay area to the east and the marine and estuarine areas to the west only a few donken are known. Many of them will have been eroded away, while in the west others will have been completely covered.

The post-Boreal sedimentation sequence is characterized by a succession of peat formation and alder carr and of clastic sedimentation in lakes and along numerous creeks.

The river sediments in the peat area show the same sedimentation pattern as those in the riverclay district but there are a number of important differences in texture and relative importance of the various sedimentation units. Of major importance are the sandy channel deposits of the main streams, that are the less coarse-textured counterparts of the wide "meanderbelts" in the Betuwe. The levee deposits are of a (very) restricted extent and consist of a clayey sand to a sandy clay. The back-swamp deposits have their counterparts in extensive high water clays, intercalated in the peat. The correlation of these clays with the channel deposits is not always easy, because of the compaction effects in the peat. There are numerous minor creeks that can only be made visible by very detailed mapping like that of J.D. van der Woude (1982), connected up with the main-streams.

The activity phases of the streams can be dated in various ways: with the oldest archaeological remains on the channel fills, C14-dates of rest gully fills or of the peat layers above and below the corresponding clay beds. Intersections of channels and stratification of the clay layers give relative ages.

Van der Woude identified four major phases of clastic deposition during the Atlantic/Subboreal period, in his 4 km² study area around the Hazendonk, in the centre of the Alblasserwaard. The sedimentation took place in a lagoonal, fluvial and/or lacustrine environment and he stresses that open water must have dominated the landscape in these phases. The lakes and swamps communicated with the main stream by means of numerous small creeks and gullies. The clastic depositional phases are separated by periods of extensive reed swamps and swamp forest, peat formation and gyttja sedimentation.

A major question is, to what extent these changes are controlled by the sea, that is in how far these can be connected with the transgression/regression cycles. Only a restricted correlation of both processes can be made out. Clastic-1 is synchronous with the central part of Calais-I, clastic-3 is correlated with C III and clastic-4 with C IV b. But cl-2 is dated in the time of the C I/II interval, C IV a is missing (but not further east) and Dunkirk-0 is not present as a separate clay bed, although there are indications that flooding and clay sedimentation continued on a modest scale. It is true that in this area sedimentation level was directly or indirectly determined by the level of the sea - so we can call it "perimarine" - but the type of sedimentation certainly was the result of the regime of the major rivers, of the higher and lower discharges in them and the influx of river water and sediment from the east.

An interesting landscape element is the Schoonrewoerd stream ridge, a c. 100 m wide sandbody, deposited by a relatively strong current in this quiet environment. Its major activity phase is dated 2100-1800 B.C. and linked with the first stage of clastic bed 4. It appeared that already in the foregoing phase parts of this stream were present in a perhaps more modest form. It was suggested earlier (Louwe Kooijmans 1974) that this stream in fact acted as a high water spillway during a restricted period of water surplus in the river clay district and that it could have been formed by the connection of existing drainage creeks. Deposition of clastic-4b after the stream was sanded-up (around 1800 B.C.) must originate from other streams

that still persisted in the environs and that might have been parts of the same stream that were not yet silted-up or even rejuvenated. These inundations caused channels that cross the ridge, called "break-through channels" in contrast to crevasses, that are synchronous with the activity of the river and are spillways from the river through one of its levees into a backswamp.

After the formation of clastic bed 4 a swamp forest (mainly consisting of alder) spread over the whole region, with the exception of the Schoonrewoerd stream ridge and the donken. Dunkirk-0 deposits are very restricted and Dunkirk-I is absent. Estuarine creek systems reached the western part of the Alblasserwaard and river sediments were laid down not farther west than Vianen and Leerdam.

An interesting new contribution to the possible correlation of marine and fluvial sedimentation is given by Berendsen (1982) in his study on the river area in the southern part of Utrecht, where the transition can be made out between the riverclay district proper and the district of peat with stream ridges. He could conclude that there are two correlations. First between the beginning of the peat formation in residual gullies of river channels and regression phases and second between periods of avulsion (i.e. the natural diversion of river sections) and transgression phases. It appears that back-swamp stratigraphy is of local value only and that it is very hard or even impossible to give reliable estimations on sedimentation intensity in this area. Both peat formation and sedimentation seem to have continued in an ever varying pattern. But it might be possible to reach an estimation on a subjective way as done until now. The frequency of fluvial sedimentation in the peat district at some distance of the river clay area (like around the Hazendonk) might, moreover, be interpreted as a measure of river water and sediment surplus farther east.

Archaeology

The western part of the river area could form a contact zone between the coast and the Pleistocene hinterland by the occurrence of water courses, fossil stream deposits and donken, which made human settlement and traffic possible. We can expect traces of occupation on the donken from Late Boreal times onward and so from times from which the stream ridges are beyond reach. But no mesolithic remains are identified until now. The hunter-fishers will not have settled on the dune tops that are the only parts open for prospection, but more probable at their bases. But in later prehistory they formed dry refuges - be it of restricted space - in an otherwise wet environment, apt for permanent settlement as far as dimensions and natural resources made that possible. They are, however, in general too small for a fully agrarian subsistence economy. Pre-Vlaardingen remains are - with one exception of a few sherds - exclusively known from the Hazendonk. But this site might be considered as representative since no prospection programs on other potential sites have been carried out. Vlaardingen-culture remains are found on eight dune tops between Ridderkerk and Zaltbommel, but no sites on fluvial deposits are known in this region. It might be that these were lacking (it is in fact a period between two phases of deposition) but it is also true that sediments of this age are covered and below the reach of normal prospection.

From late Bell Beaker times onward occupation has been attested on the Schoonrewoerd stream ridge. The fragmentary condition of the remains makes it difficult to prove that occupation was continuous, but in view of the agrarian character of the settlements one can assume a permanent occupation of the ridge. There are no reasons to assume occupation breaks before the Late Bronze Age.

The wide-spread peat deposits gradually covered all inhabitable locations, with the exception of some donken. The district as a whole became unsuitable for a fully agricultural way of life, like the peat district farther north. Roman settlements lie on rows on the fillings of the main channels of Dunkirk-I estuarine creek systems in the western part of the Alblasserwaard. An unoccupied peat marsh separated these sites from the westernmost sites in the dense inhabited riverclay district. The contact between both occupation areas was formed by two major roads, the "north" and "south" road of the Tabula Peutingeriana. The former along the river Rhine to Leiden, the latter south of the Alblasserwaard, probably along the Meuse. Most of it is, however, lost. Some Roman sites along the Merwede river reflect the occupation in a zone along this road.

An even wider district was unoccupied in the Early Middle Ages. Occupation starts not earlier than the 11th century, when colonists settled along the main rivers and along the water courses that drained the peat. It are well-organized reclamations of the "wilderness", supported by the authorities, namely the Count of Holland and the Bishop of Utrecht. Rectangular "blocks" were drained, embanked and parcelled in long strips. Farmyards were built on artificially raised places as a protection against floods. The reclamation was completed in the 13th century and a collective Water Control Board (Hoogheemraadschap) was founded by the Count of Holland in 1277. This and similar Boards are still in function as autonomous bodies (with taxes of their own !) that control drainage, maintenance of embankments, waterpollution, etc.

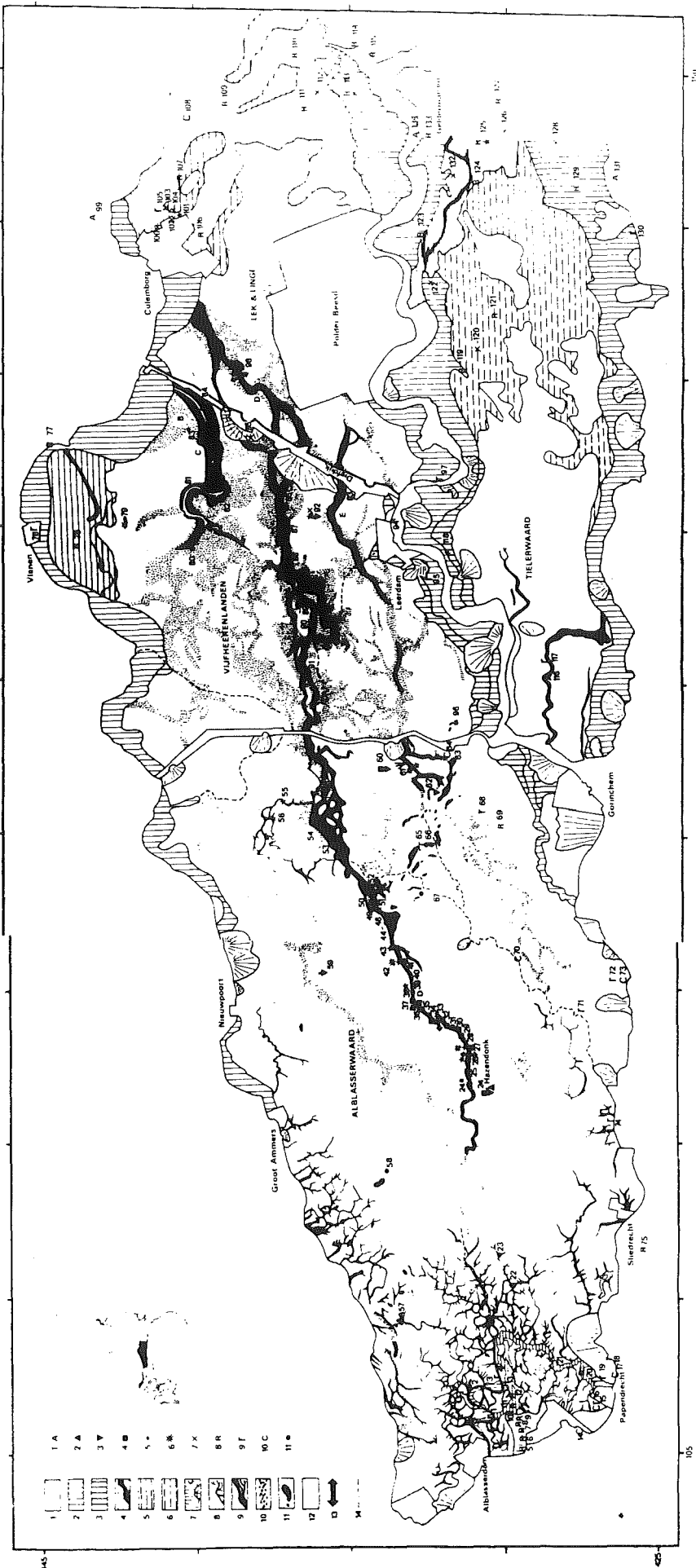


fig. 55. Map of the prehistoric and Roman find spots in the river clay/wood peat area, in relation to its geology. After Louwe Kooijmans 1974.

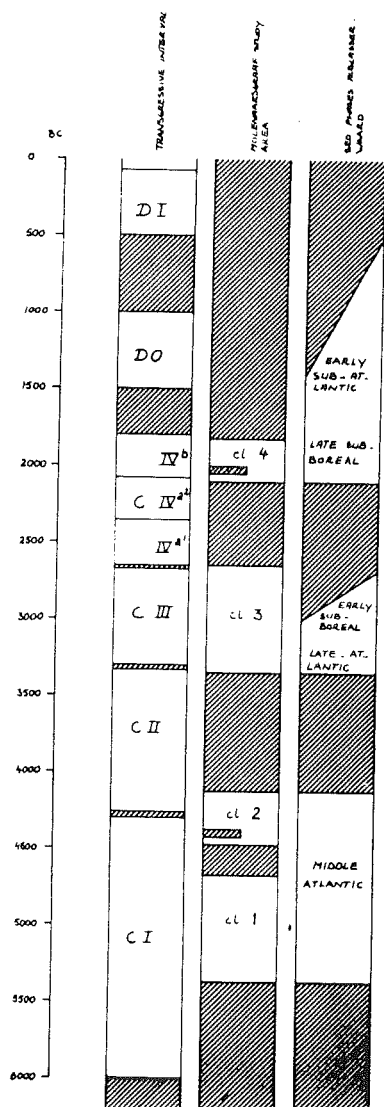


fig. 57. Chronostratigraphy of the Molenaarsgraaf study area compared with the marine transgressive phases (left) and a preliminary sequence of sedimentation phases in the river clay/wood peat area. Redrawn after Van der Woude 1981.

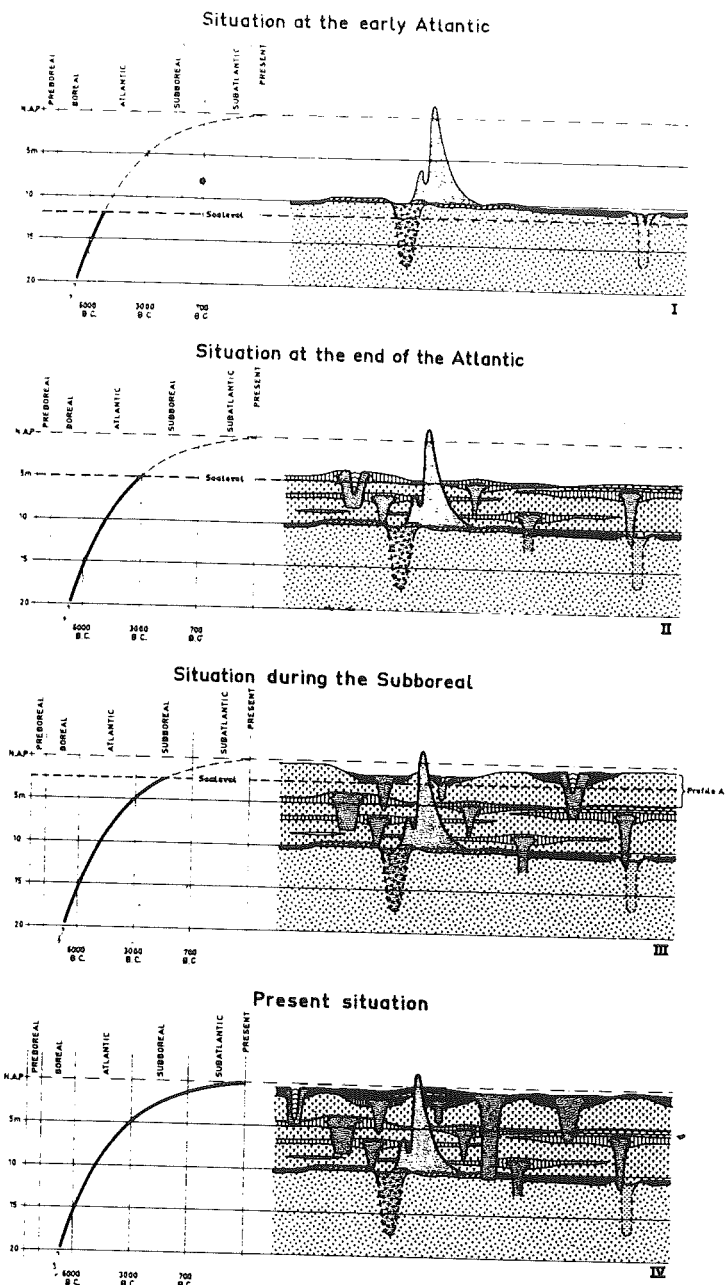
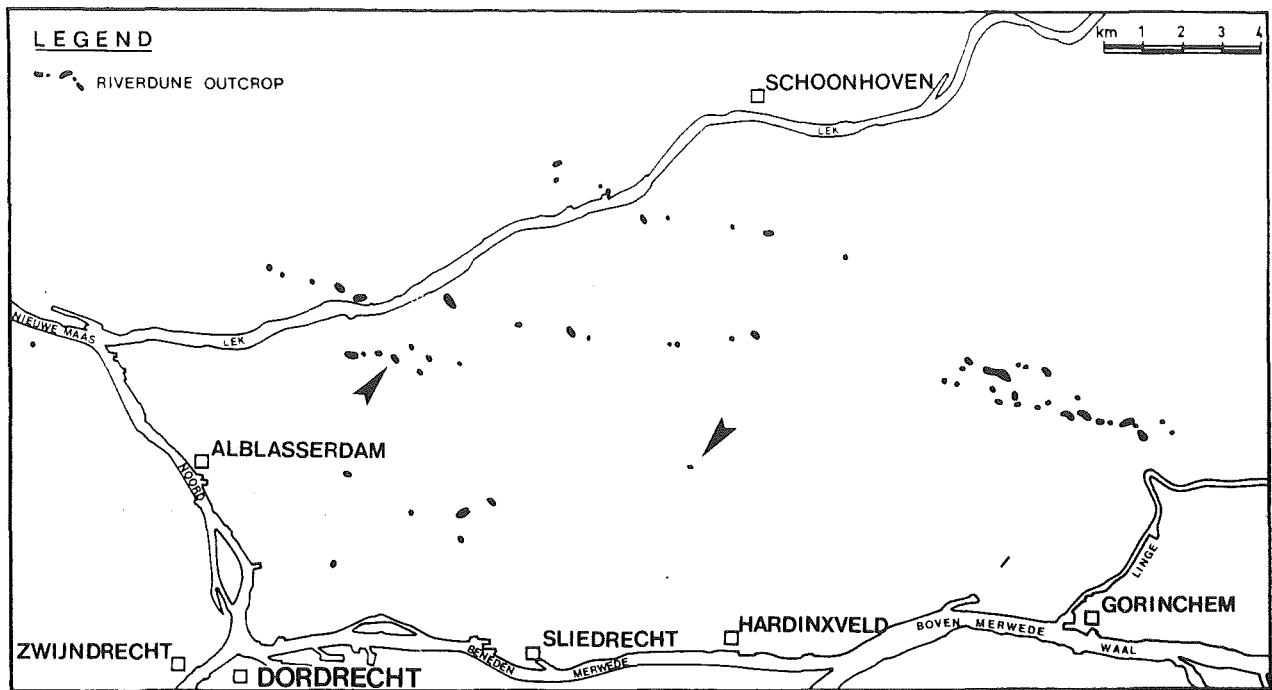
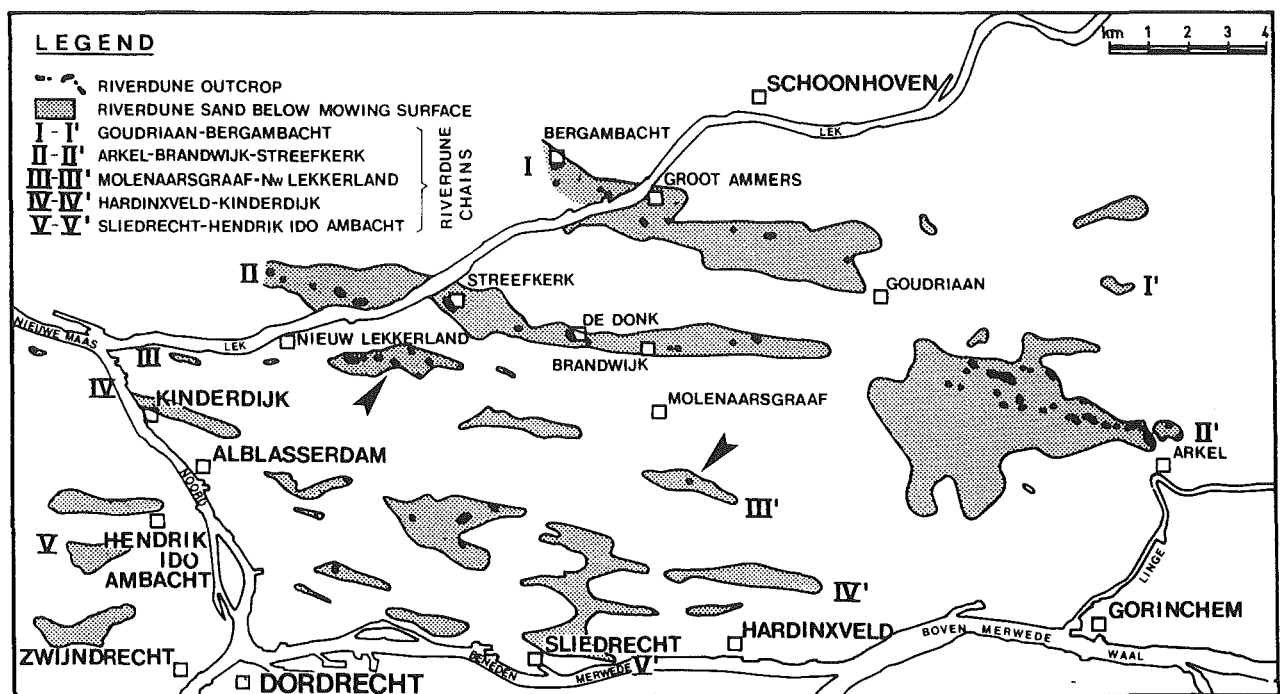


fig. 56. Stages of development of the river clay/wood peat area in relation to the curve of the relative sea-level rise. After Hageman 1969.



Outline map of the riverdune occurrence in the Alblasserwaard



Outline map of the riverdune chains in the Alblasserwaard

fig. 58. Distribution of the river dunes in the Alblasserwaard.
After Verbraeck 1974.

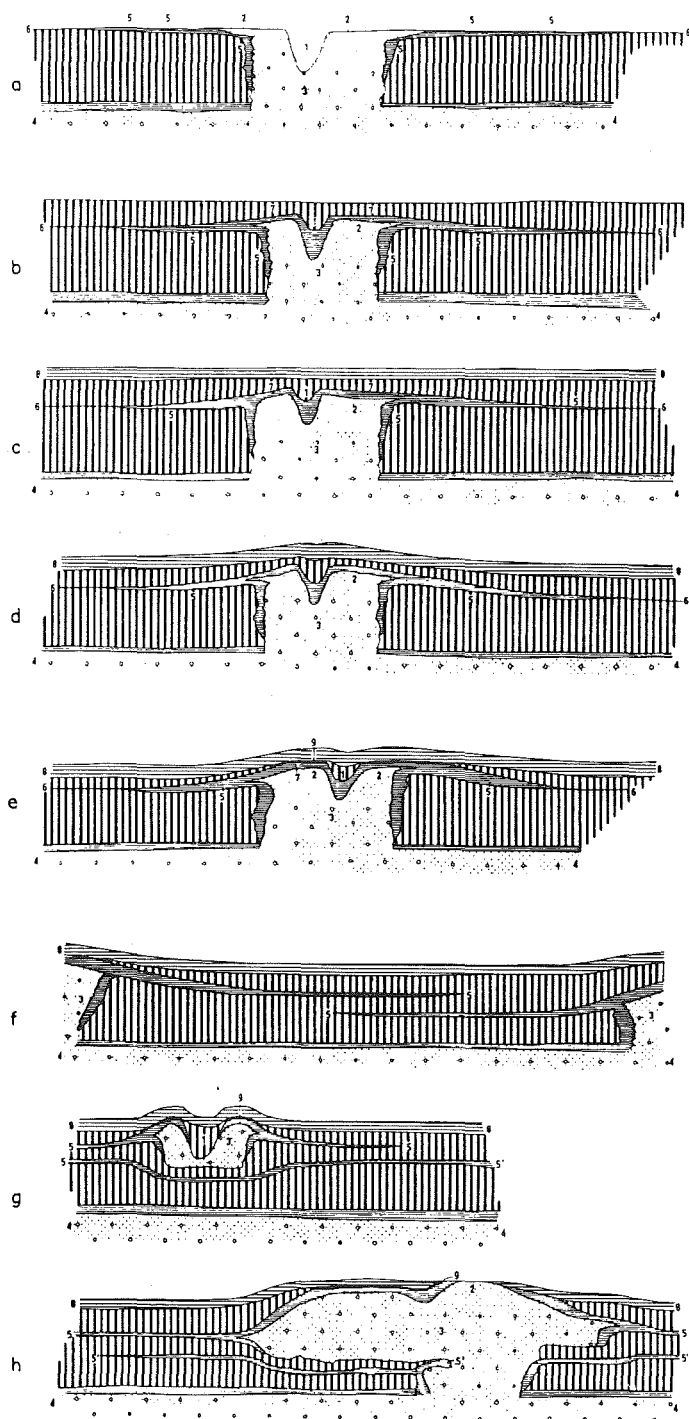


fig. 59. Stages of development of a stream ridge (a - e) and schematic cross-sections of various possible stratigraphies. After De Boer/Pons 1960.

Kinderdijk

The excursion route follows the narrow embankment of the Noord. River foreland is lacking completely and the dike has a little concrete wall at the water side against waves and water in the case of extreme high waters and storms. Two round ponds ("wielen") are the scars of former dike breaches.

Kinderdijk ("childrens dike") got its name by the story that a cradle with a baby was washed ashore in 1421 after the severe St. Elisabeth Flood, when great parts of Holland were destroyed and the Biesbos came into existence in a polder called Grote Waard, that could not be recovered. The cradle was kept in balance by a cat, that jumped from side to side ...

Kinderdijk is the main drainage point for the Alblasserwaard, that lies upstream of this place. This "waard" is divided in an upper (Overwaard) and a lower part (Nederwaard). Both have a lower storage basin for surplus water ("boezem") that leads to Kinderdijk where 19 large windmills of different types, built in 1738-1740, pumped the water into two storage basins with a higher water level. From these could the water be drained into the river at low tide. Today two electric pumping stations do the work, but the mills are kept in condition as stand-by. It is a remarkable situation that the Overwaard leads its surplus water to Kinderdijk via a long canal that crosses the total of the Nederwaard.

Lek Embankment

The Lek is a young lower course of the Rhine-system. The river came into existence after the Roman period but before 850 A.D., most probably as a peat drainage river that was used more and more by surplus river water in the Utrecht riverclay district. So far down-stream the river has lost most of its sediments. Hardly any forelands are formed and the river is closely locked up between the embankments. These were raised again and again to match up with rise of the river water level and are now vulnerable for erosion. We will pass a "dijkval" (dike fall), where a part of the embankment has sunken away into the river.

The increased drainage of the peat of the peat district during seven centuries has resulted here in a subsidence of the surface of circa 2 m. The result of this and of the raised sedimentation level along the embanked rivers is alarmingly visible along the lower Lek.

We have a good view from the high dike over the long strip-parcels, characteristic of the mediaeval reclamations of the Holland peat district.

Zijdedijk and Tiendweg

In contrast to the high embankment along the Lek stands the old Zijdedijk. This is the original western bank of the Alblasserwaard, dating from the time when Count Floris V of Holland founded the still existing Drainage Board for that district in 1277. The western part of the present Alblasserwaard was added in 1281. The Zijdedijk lost its function and never was raised since then. There are two embanked pothole ponds of 13th century dike breaches preserved, one where the bank crosses a "donk" and seepage easily could cause such a breach.

A narrow road, bordered by two ditches, crosses all narrow parcels. It is a so-called Tiendweg or a road for the 10% tax on hay and crops, that was removed directly from the fields along these roads. They saved the long ride over every parcel and the necessity to cross the yard of unfriendly farmers.

Schoonenburgse Heuvel

One of the highest and best preserved of the donken, named after the castle Schoonenburg that stood nearby. This donk has its top at + 4 m NAP and measures 100 x 250 m. It is part of a large group of high dunes of which more are well visible. In 1952 the farmer's son found a point-butted flint axe near the top. Professor Modderman documented the find and visited the site. Later the axe was bought by the National Museum of Antiquities for the price of a calf. Further prospections did not reveal traces of occupation of any importance.

Dunkirk-I creek ridges

Between the "lower storage canal" of the Overwaard and that of the Nederwaard the road crosses a number of small estuarine creek fillings that are visible as low ridges along both sides of the road. They are the easternmost extensions of the Dunkirk-I (and -II ?) systems of Alblasserdam and Oud Alblas. On the sandy fillings of the main channels a number of Roman settlement sites are known.

Alblas and Graafstroom

The winding river Alblas is one of the natural drainage rivers of the peat district from pre-embankment times. No Roman occupation has been found on its levees, so it must be dated in the Dunkirk-II transgressive phase. The first mediaeval colonists settled themselves along this stream in the 11th century. Up-stream, where the river became too narrow, it was widened and extended with a more or less straight canal called Graafstroom (The "Dug Stream").

Farms and Floods

Many farms were built on small artificial mounds. From some of these Pingsdorf but no Badorf pottery has been recovered. This proves the colonisation in the 11th century.

Originally four long parcels of generally 1.5 km in length belonged to one farm. So the old farms are separated by three parcels. These gaps were partly filled in the last century with labourers' houses and more recently by those of commuters, thus destroying the original village appearance.

Many leaning farms demonstrate that the subsoil of unsettled clay did not offer a good foundation, and so does the Molenaarsgraaf church.

Three stones in a Molenaarsgraaf farm are testimonies of the severe floods by which this district suffered. Between 1320 and 1820 no less than 36 dike bursts are documented, the most severe in up-stream locations, caused by ice dams in the rivers during severe winters. Especially when the Diefdijk - the eastern defence on the boundary with Guelders - gave way the consequences were catastrophic. Farmers often lost goods and chattles and were reduced to beggary. Sometimes a freedom of taxes was allowed for a couple of years to overcome the disaster.

The old farms were constructed with a heavy "flood attic", where the cattle could be stalled during floods and with a roof construction supported by the posts only. The free-standing walls could be washed away and replaced when the inundation was over. Some farms had "flood doors" halfway the frontal façade.

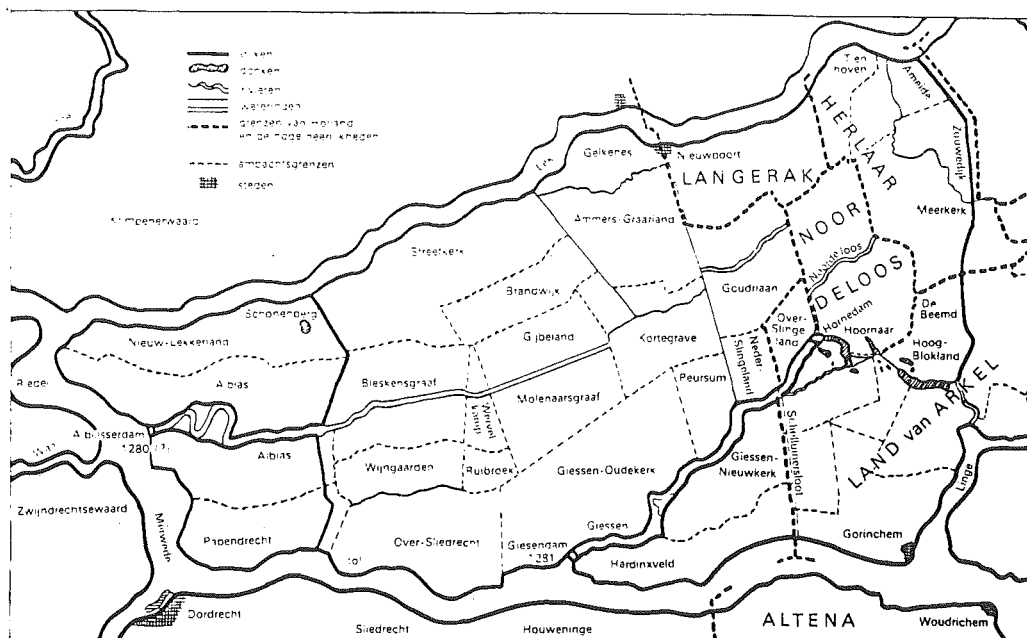


fig. 61. The Alblasserwaard and its reclamations. After Henderikx 1977.



fig. 62. Drainage system of Overwaard and Nederwaard.

COASTAL AND INTRA-COASTAL SEA-LEVEL RISE IN THE WESTERN NETHERLANDS DURING THE MIDDLE HOLOCENE

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Introduction

Probably the most 'dramatic' phase in the evolution of the coastal plain of the western Netherlands was the formation of the coastal barrier system, beginning around 5000 yrs BP or a little earlier. Among the factors that controlled this evolution, characterised by morphodynamic feedback mechanisms, are: pre-Holocene topography, input of freshwater via rivers and groundwater seepage, sediment supply (both fluvial and marine), tidal and wave regime, and relative sea-level rise. Here, I highlight what is presently known of the rise of relative sea level and what influence it may have had on the evolution of the western Netherlands, particularly during the mid-Holocene.

Tidal damping in the intra-coastal zone

The area landward of (insipient) coastal barrier islands is referred to as the intra-coastal area or zone. The tidal wave, upon entering an intra-coastal zone, can be damped to various degrees and become reduced to an inconspicuous oscillation. This decrease in intra-coastal tidal range is called the floodbasin effect (Fig. 1). It develops when the tidal wave enters a tidal (or 'flood') storage basin behind the coastline of greater capacity than the tidal prism of the coastal inlet. Frictional loss of energy is another important factor in tidal wave attenuation. It is important to know the evolution of this floodbasin effect as it determines the intra-coastal mean high water (MHW) level, accomodation space, tidal energy, sediment transport, and salinity, hence the lithological, morphological and ecological characteristics of the evolving depositional environment. The floodbasin effect is defined by the distance of local MHW below coastal MHW. The present-day mean tidal range along the Holland coast is 2.2 m in the south (Haringvlietsluizen) and 1.4 m in the north (Den Helder) (for locations see Fig. 3).

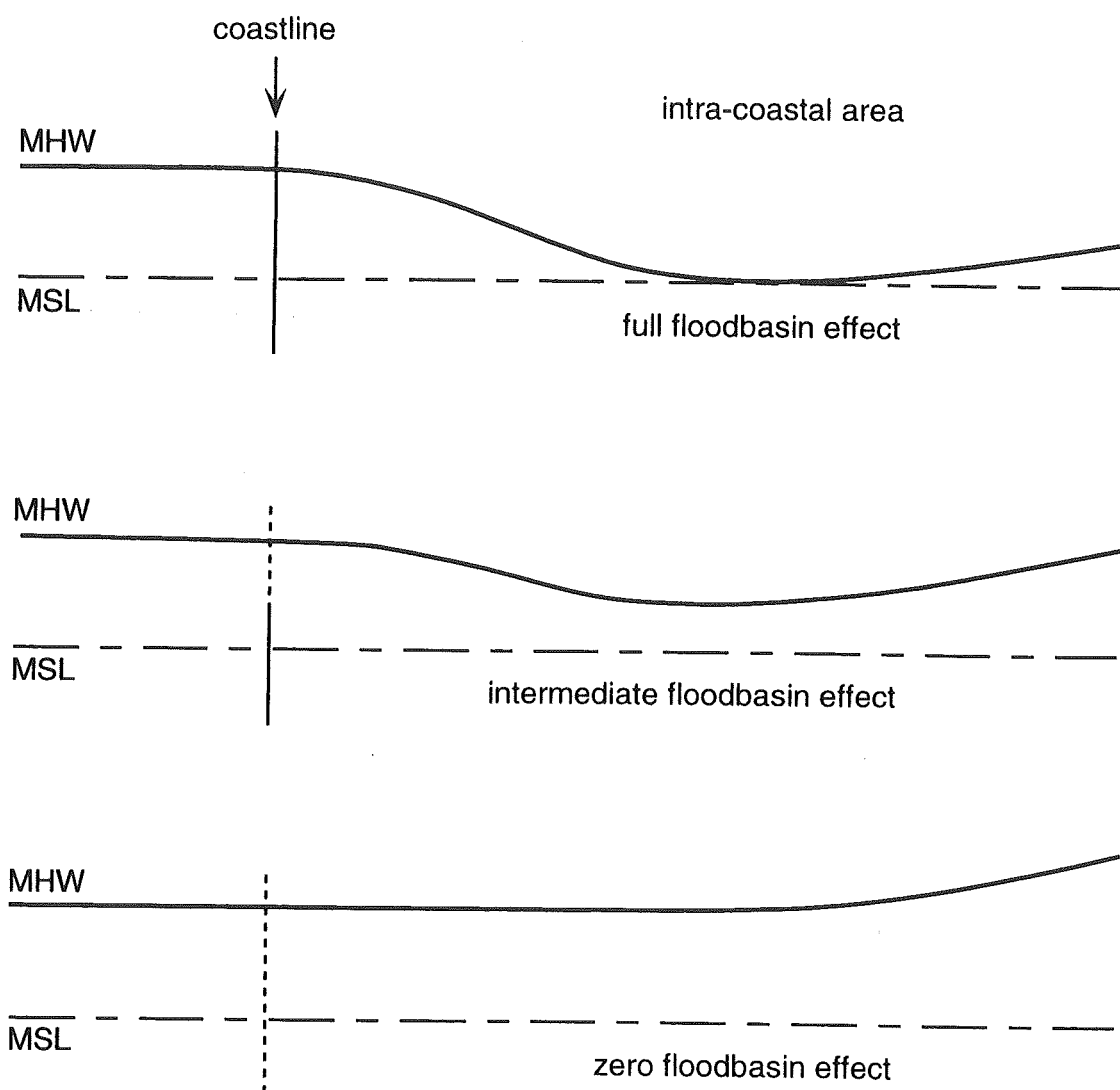


Figure 1. Schematic illustration of zero, intermediate and full tidal damping or floodbasin effect in a river-influenced intra-coastal area.

As will be shown below, tidal damping was a widespread phenomenon in the western Netherlands, beginning as early as ca. 5250 cal BC (ca. 6275 BP). In the Rhine-Meuse delta, this reduction in the local tidal range enhanced the effect of the slowing down of MSL rise and resulted in a decrease in the rate of local MHW rise from about 0.8 m/100 yr around 5300 cal BC (6375 BP) to less than 0.2 m/100 yr around 4800 cal BC (5800 BP).

Mid-Holocene sea-level curves for the W. Netherlands

A summary of relevant mid-Holocene sea-level curves for the western part of the Dutch coastal area is presented in Figure 2, which features:

one coastal MHW curve: F-E,

two local MHW curves: F (post-5250 cal BC) and G, and

three MSL curves: A, A-H, and I-H.

Coastal MHW rise

The curve for the relative rise of coastal MHW, curve F-E, is obtained by connecting curve E and the pre-5250 cal BC part of curve F. Curve E is based on an analysis of sedimentary characteristics of coastal barrier deposits along the western coastline, and can be considered a reliable approximation of the coastal MHW rise (Roep and Beets, 1988). Curve F is based on basal peat dates obtained from the slopes of two river dunes in the lower Rhine-Meuse delta (Hillegersberg and Rijksweg river dunes; Fig. 3) and represents the rise of local MHW in the Rotterdam area. It is assumed that the pre-5250 BP part of curve F approximates the relative rise of coastal MHW because: (1) it supports the position of local MHW levels in the lower Rhine-Meuse delta based on a reconstruction of river-gradient lines (Van Dijk et al., 1991), and (2) does not call into question a) the interpretation of curve A as representing the relative rise of mean sea level (MSL), b) the assumption of a mean coastal tidal range of 2 m, or c) the argument that little floodbasin effect is expected to occur in the Rotterdam area prior to 5300 cal BC (6375 BP). Strong tidal damping in the Rotterdam area as early as 5300 cal BC is not expected for two reasons: firstly, the rate of relative MSL rise in the foregoing centuries was still considerable (1.0-0.9 m/100 yr) prohibiting extensive shallowing of the intracoastal area, and secondly, although barrier islands or tidal flat expanses surrounding the tidal inlets of the delta front probably existed (Beets

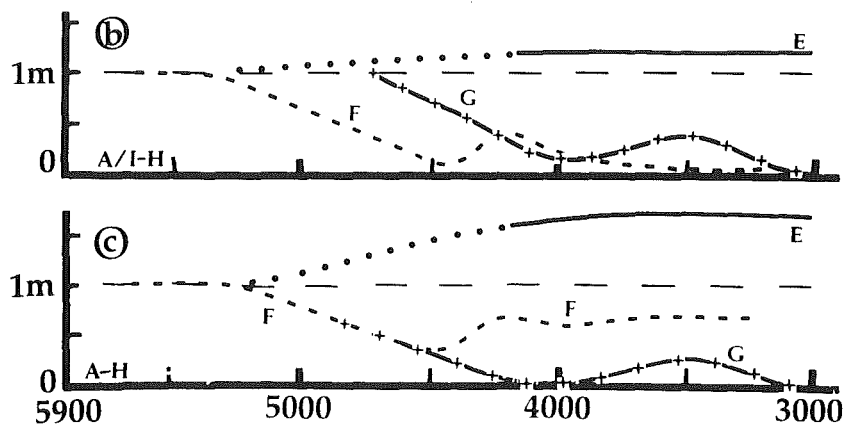
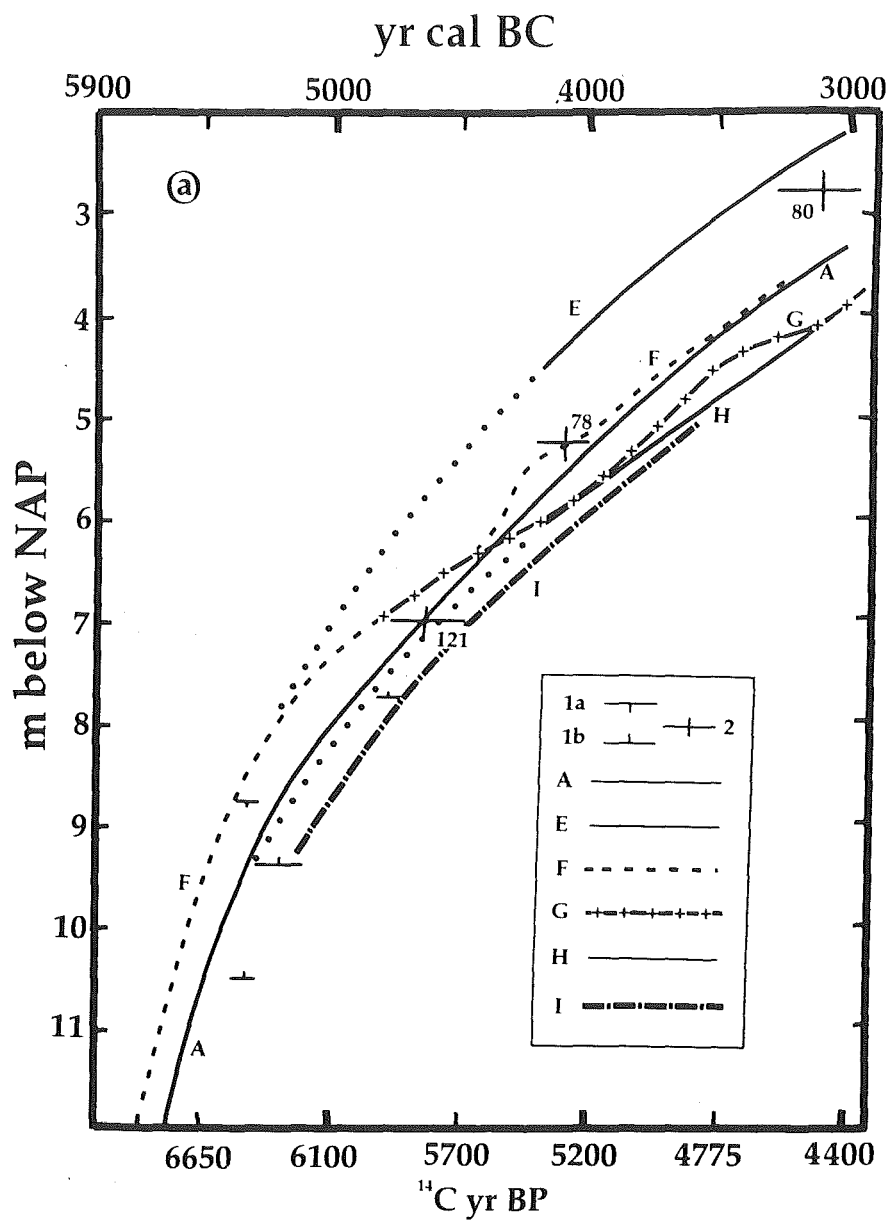


Figure 2. (a) Age/depth diagram with mid-Holocene sea-level curves for the western Netherlands; (b) Plot of curves F-E and F relative to curve A, and of curve G relative to curve I-H (case of greater crustal subsidence of the Schokland area); (c) Plot of curves F-E, F, and G relative to curve A-H (case of no differential crustal movements). Explanation in the text.

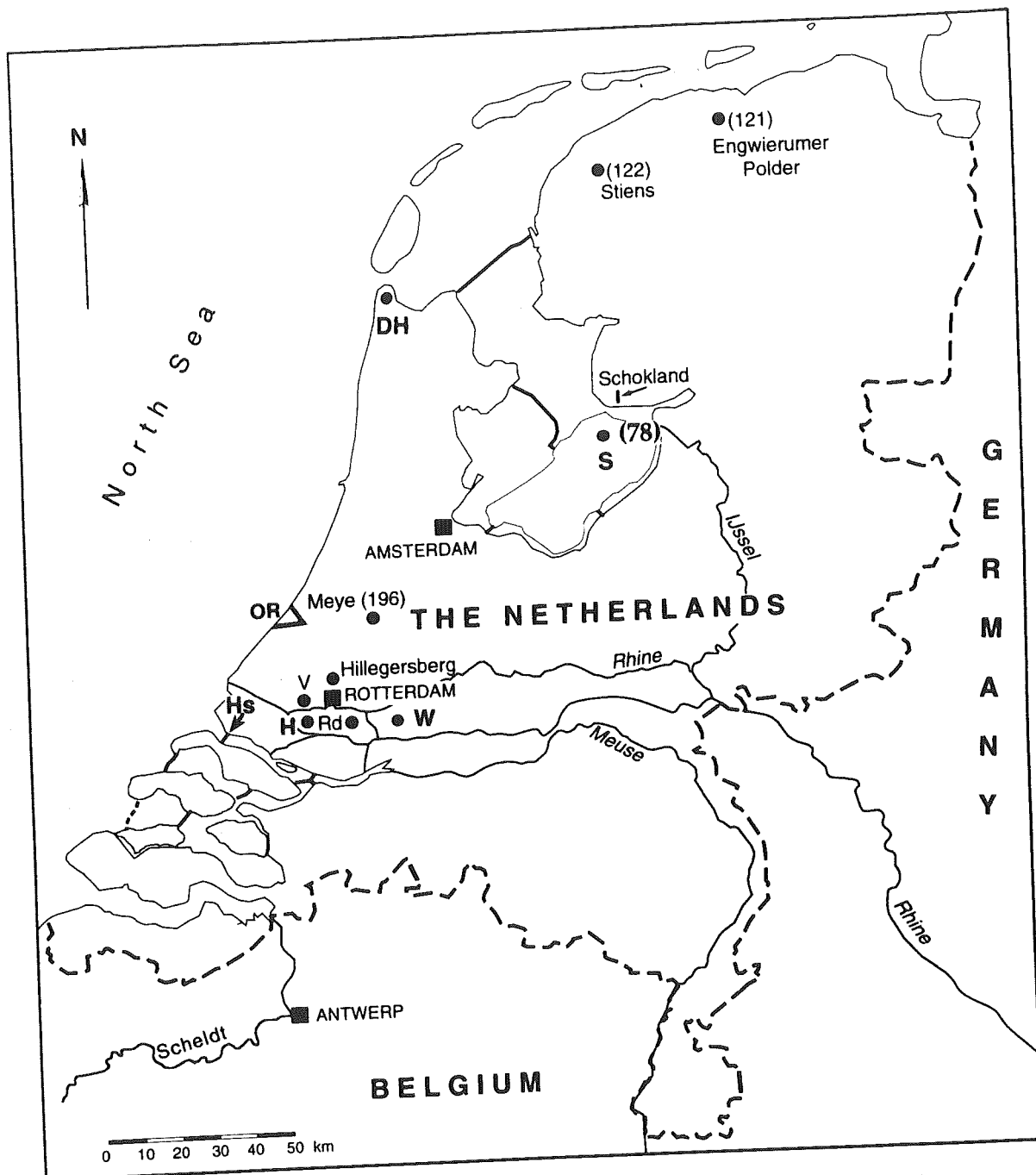


Figure 3. Location map for sites mentioned in the text. Key: DH: Den Helder; Hs: Haringvliet sluizen; Rd: Rijksweg river dune, S: Swifterbant; V: Vlaardingen; 78: number of sea-level index point.

Key (inset Fig. 2 a): 1a, 1b: respectively, upper and lower limit for MSL; 2: sea-level index point with error bars for depth and age (1 sigma); age/depth graphs: A: 1982-MSL-limit curve (Van de Plassche, 1982); E: coastal MHW curve (Roep and Beets, 1988); F: coastal MHW curve (pre-5250 cal BC) and local MHW curve (Rotterdam area; post-5250 cal BC) (Van de Plassche, 1995); G: (ground)water level / local MHW curve for the Schokland area (Roeleveld and Gotjé, 1993); H: MSL-limit curve for the Schokland area (Van de Plassche, 1995); and I: hypothetical MSL-limit curve for the Schokland area (Van de Plassche, 1995), assuming this area subsided at a rate of 0.1 m/1000 yr relative to the Rhine-Meuse delta and the northern Netherlands (curve A).

et al., 1992; Van der Spek, 1994), the tidal prism is likely to have been still too large to permit narrow inlets causing significant restriction of tidal inflow.

Local MHW and MSL rise

Mean sea level curve A is based on a critical evaluation of all age-depth data from the western and northern Netherlands available by 1982 (Van de Plassche, 1982). Curve F (Van de Plassche, 1982, 1995) represents the rise of local MHW in the Rotterdam area, while curve G (Roeleveld and Gotjé, 1993) approximates the local MHW rise near the former island of Schokland (Fig. 3). Both curves are based on basal peat dates obtained from the slopes of river dunes.

The oldest part of curve G coincides with curve F, but then dives below the 1982-MSL curve A, indicating either that the younger part of curve A does not represent MSL or that the Schokland area has subsided relative to the Rhine-Meuse delta. In the case of no differential crustal subsidence, curve A needs to be replaced by curve A-H. In the other case, curve I-H represents the relative rise of MSL in the Schokland area based on the assumption that this area subsided relative to the Rhine-Meuse delta at a constant rate of 0.1 m/1000 yr during the last 7000 years. All MSL curves are in fact upper-limit-for-MSL curves.

Conclusions

The current sea-level database for the western, east-central and northern Netherlands is inconclusive with respect to differential crustal movements during the last 7000 calendar years. Available relative MSL data may indicate that the east-central coastal plain (the Schokland area) subsided relative to the Rhine-Meuse delta and the coastal barrier area at a mean rate of 0.1 m/1000 years during the last 5000 years.

Case of greater subsidence of the Schokland area

For the case of possible differential crustal subsidence, with curve A approximating the relative MSL rise for the Rhine-Meuse delta and the northern Netherlands, and curve I-H being an inferred local MSL curve for the Schokland area (Fig. 2b), it is concluded that the coastal tidal amplitude has been close to 1 m throughout the period 5500-3000 cal BC (6625-4400 BP).

As for intra-coastal tidal ranges, a clear floodbasin effect began to develop in the Schokland area around 4750 cal BC (5850 BP) (curve G in Fig. 2b), about

500 years later than in the Rotterdam area (curve F, considered relative to curve A, Fig. 2b). A minimum estimate for local MHW near Swifterbant (index point 78, Figs. 2a and 3) suggests that the local tidal range can vary by as much as 1 m over a distance of 10-15 km. This latter conclusion remains valid for the case of no differential crustal movement.

Case of no differential crustal movements

For the case of no differential crustal movements, with curve A-H representing the relative MSL rise for the Rhine-Meuse delta and the Dutch coastal plain further to the north, it is concluded that the coastal tidal range along the Holland coast increased from about 2 m prior to 5300 cal BC (6375 BP) to close to 3 m by 4200 cal BC (5350 BP), maintaining this value until at least 3000 cal BC (4400 BP). This conclusion contrasts with results of palaeo-tidal modelling, which suggest that the tidal range along the Holland coast was smaller when sea level was lower.

As for floodbasin effects, MHW levels in the Rotterdam and Schokland areas began to fall rapidly relative to MSL from about 5250 cal BC (6275 BP) until ca. 4500 cal BC (5675 BP). In the firstmentioned area, the local tidal range temporarily increased again from about 0.7 m around 4400 cal BC (5650 BP) to ca. 1.3 m by 4300 cal BC (5400 BP). This increase corresponds to the time that a clay bed was deposited near Vlaardingen (Fig. 3). In marked contrast to curve A, curve A-H (Fig. 2c) implies that tidal activity did not disappear from the Rotterdam area after 4000 cal BC (5225 BP); the local tidal range remained close to 1.3 m.

The main general conclusion is, that an important part of the foundation for the evolution of the western Netherlands coastal plain during the later part of the middle Holocene was put into place in the period 5300-4300 cal BC (6375-5400 BP) when (a) the relative MSL rise slowed down from about 0.8 m/100 yr to 0.23 m/100 yr, (b) the coastal tidal range increased, possibly by as much as 1 m, causing increasing amounts of sediment to be transported into the intra-coastal area, while (c) intra-coastal tidal damping reduced rates of local MHW rise (even further), tidal current velocities and accommodation space, thus hampering landward transportation of sediment and promoting closure of the Holland coast.

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