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EXCURSION GUIDE

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EXCURSION SITE 6

LATE-GLACIAL AND HOLOCENE EVOLUTION OF THE MEUSE VALLEY

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Introduction

The Meuse 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. 6.1).

The Meuse 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 Meuse is a rain-fed river. The mean annual precipitation amounts to 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 Meuse at Borgharen (Netherlands-Belgian border) and of the meltwater-fed Rhine at Lobith (Netherlands-German border) (Jongman, 1987).

	Meuse	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 Meuse in comparison with the Rhine. Furthermore, the fluctuations in discharge of the Meuse 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 Meuse.

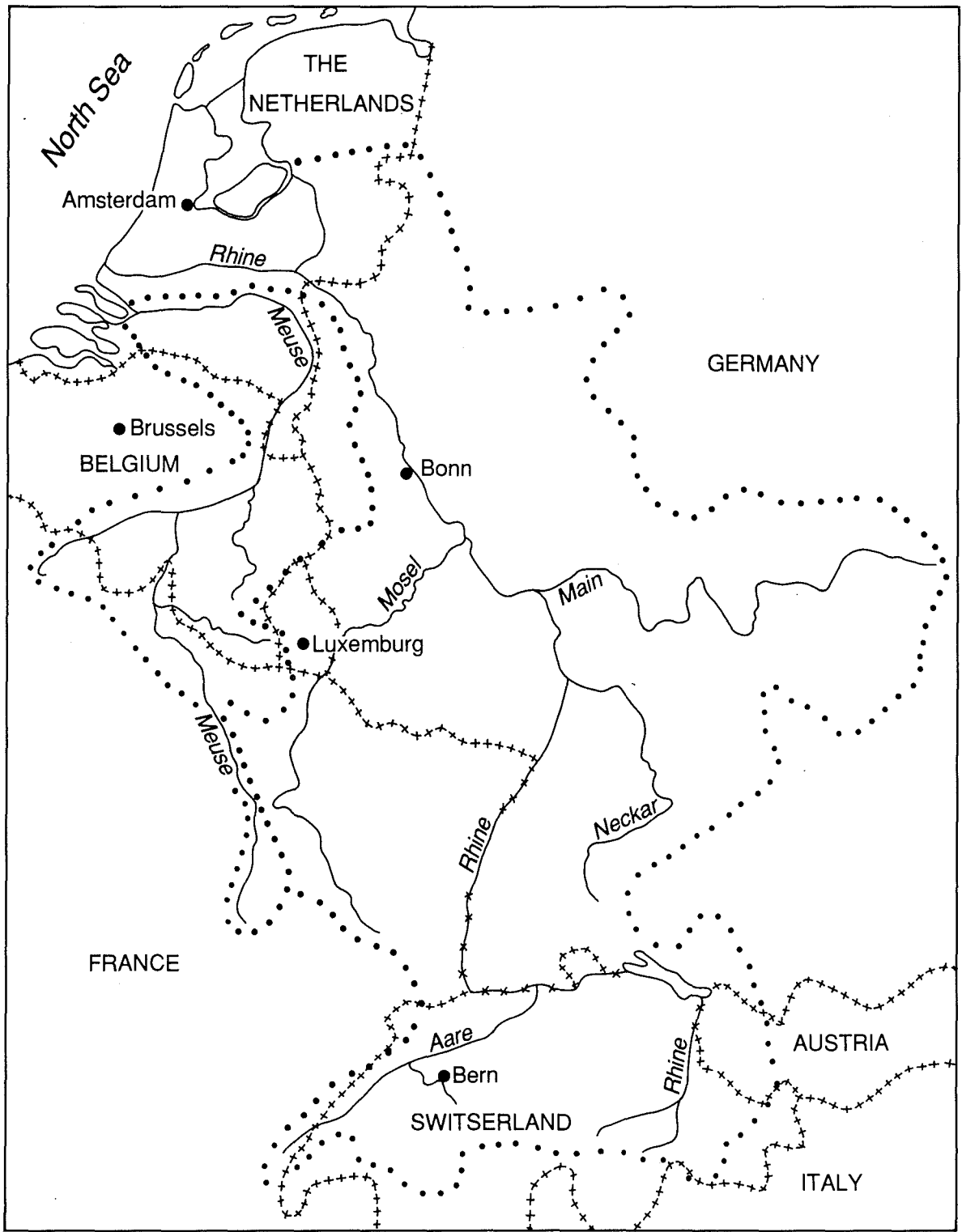


Fig. 6.1. Catchment areas of the Rhine and the Meuse.

Tectonics

North of Maastricht the Meuse enters the southeastern part of the North Sea Basin, which is characterized by the occurrence of southeast-northwest oriented faults (fig. 6.2). 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 Meuse 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 Meuse, which deposited the Veghel Formation. During the Elsterian, Holsteinian and Saalian the Meuse gradually shifted eastwards over the Peel Horst area into the Venlo Graben (Van den Toorn, 1967; Zagwijn and Van Staalduinen, 1975).

At present the Meuse 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 Meuse has a strongly meandering course with a broad floodplain (Van den Broek and 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.

Terrace morphology

Above the recent Holocene floodplain, two distinct Late-Glacial to Early-Holocene terrace levels are distinguished along the Meuse north of Venlo (Wolfert & De Lange, 1990)(fig. 6.3).

The highest Late-Glacial terrace level descends from 21 m above sea level at Venlo (base of fig. 6.3) to 18 m at Holthees (top of fig. 6.3). It is typified by large meander scars, especially at the outer terrace edges. Individual pointbars are poorly developed normally, because of an eolian cover and intense human occupation and recultivation.

At excursion stop A (Schuitwater) the top of the infill of the paleomeander is found at 17.5 m. This is approximately 5 m below the older Pleniglacial surface in the west, which consists of fluvial and eolian sediments (see excursion site 4: Grubbenvorst). The inner bend of the meander rises generally to 19.5 m, but higher locations up to 24 m are found locally, where eolian accumulations occur on the terrace surface.

The morphology of the meander inner bend northeast of point A shows an alternation of higher and lower areas, roughly conform the paleomeander scar (fig. 6.4). These ridges are too large and small in number to be individual pointbars. It is suggested that these large ridges are complexes of several high pointbars, while the depressions in between the large ridges might be complexes of low pointbars.

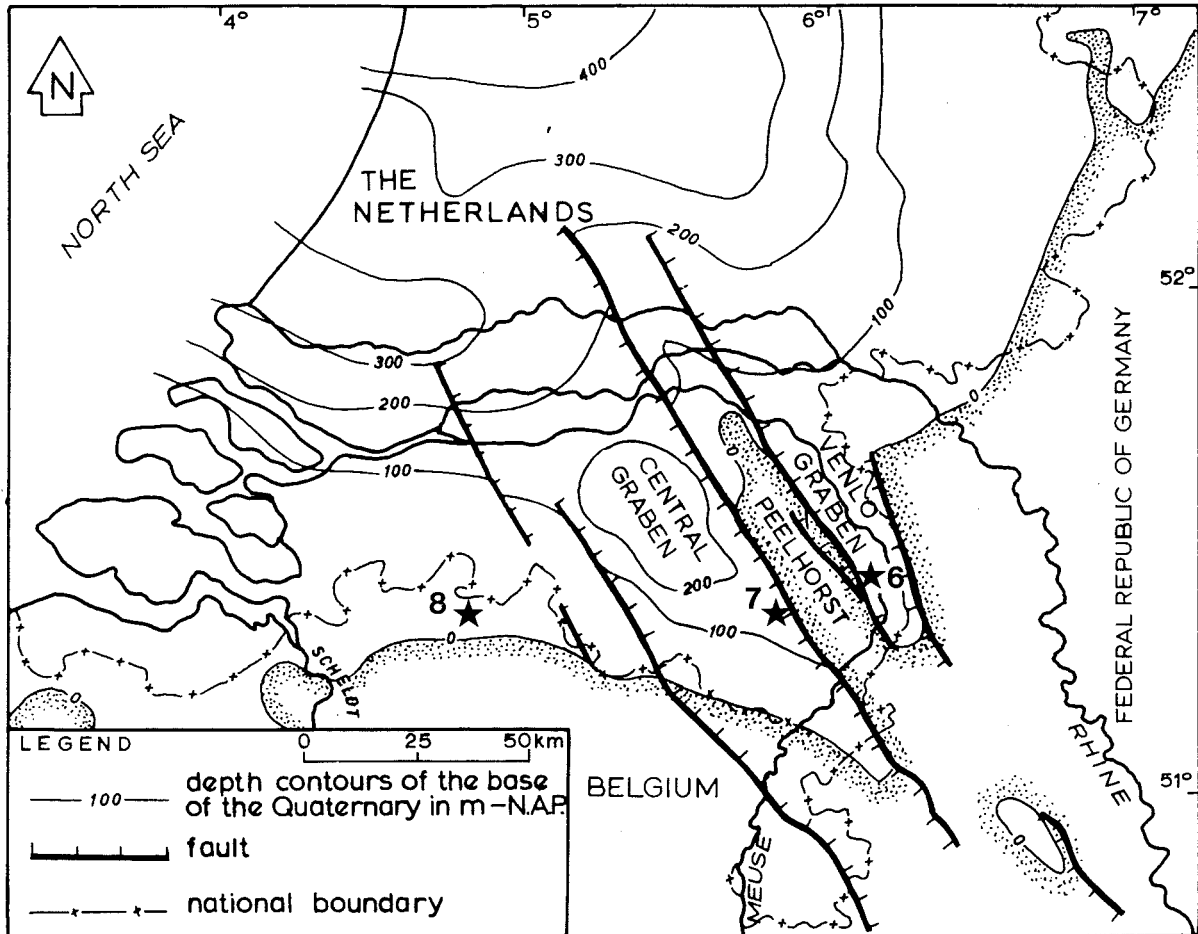


Fig. 6.2. Depth contours of the base of the Quaternary and major tectonic units of the southern North Sea basin (after Zagwijn & Doppert, 1978). Excursion sites 6, 7 and 8 are indicated by an asterisk.

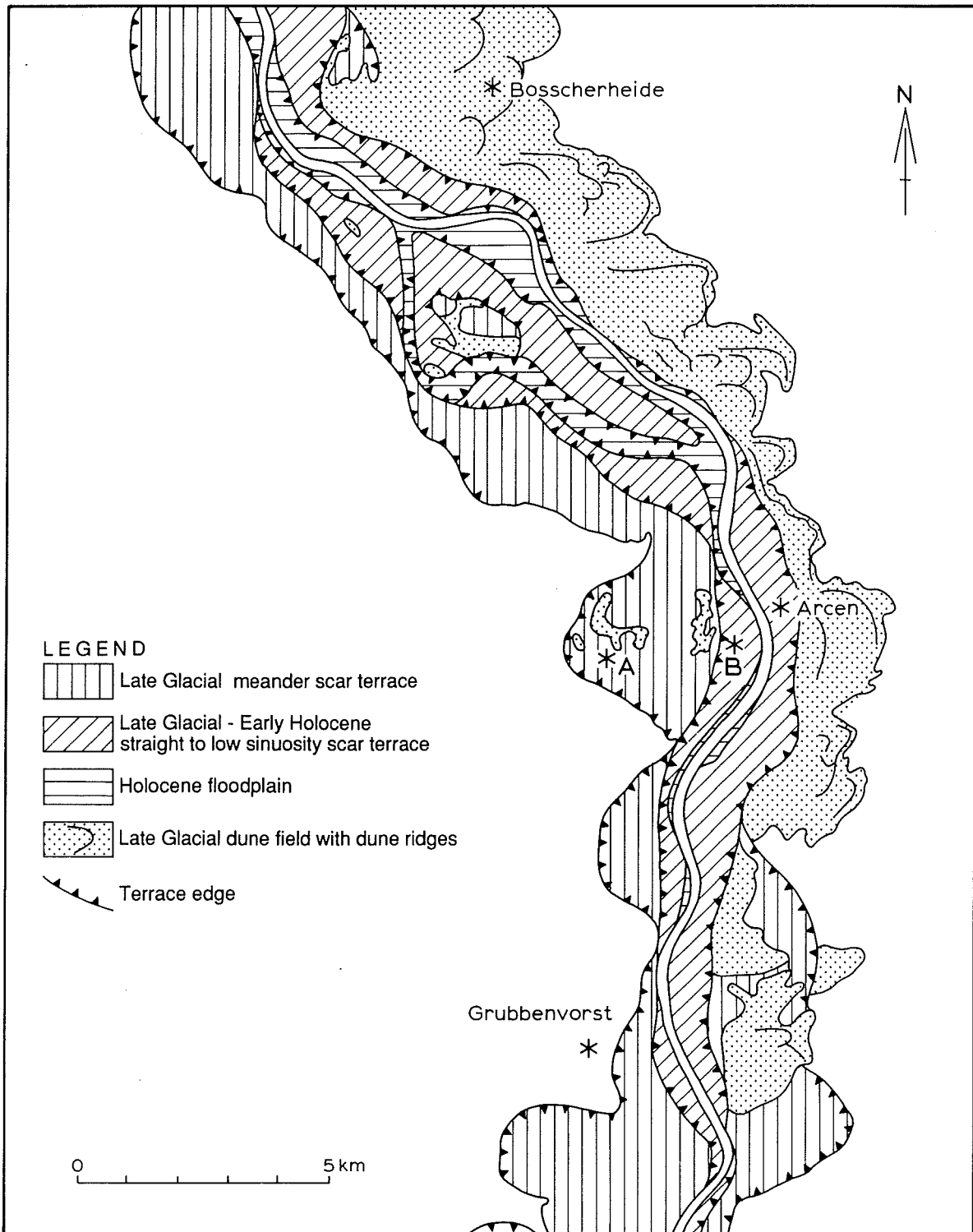


Fig. 6.3. Morphological map of the Late-Glacial and Holocene deposits in the Meuse valley (after Wolfert & De Lange, 1990). Excursion sites are indicated by an asterisk.



Fig. 6.4. Contour map in meters above sea level of the Late-Glacial terraces at the excursion sites 6A and 6B.

This pattern could reflect long term fluctuations in discharge during the lateral migration of the meander. Periods with high discharges led to high pointbars, coalescing with each other to form a higher pointbar complex. Periods with low discharges resulted in lower and smaller pointbars, together forming a low lying area between two high pointbar complexes.

However, coring in the meander inner bend (see fig. 6.4 for drill-hole locations) revealed that both the high and low areas are covered by eolian sediment. Therefore, the large-scaled ridge and swale topography is not of a pure fluvial origin, although the orientation of the morphology suggests that its genesis must, at least partly, be related to the fluvial system. The corings further indicate that the top of the fluvial sediments in the meander bend declines slightly to the northwest; this is in the direction of the channel migration, suggesting a simultaneous incision during the lateral migration. Until now it is unknown if the fluvial base (lag deposit) of the terrace shows the same trend in the northwestern direction.

Biostratigraphy and palaeoenvironment at Schuitwater

At two locations the deepest point of the palaeomeanders has been sampled for pollen analyses with the aim to obtain biostratigraphical and palaeoenvironmental information concerning the fluvial development. Pollen record SW 2B relates to the outermost meander scar at Schuitwater whilst SW 2A, located approximately 125 m northeast of SW 2B, derives from the more inward located palaeomeander (Fig. 6.5 and 6.6).

Boring SW 2B

The change from an active meander during which fluvial sands were deposited, towards an inactive or periodically active system occurs at the termination of the Bølling (362 cm) as appears from the high *Betula* values in the bottom part of the section. Here also reworked pollen of Lower Pleistocene or Tertiary deposits occur in the sediment (*Nyssa*, *Carya*, *Pinus haploxylon*, *Ulmus*).

In the overlying deposits up to 345 cm sand is virtually absent and the palaeomeander is fed by iron-rich seepage water leading to the deposition of siderite and FeS in a gyttja-like matrix. Shallow standing water in which algae thrive (*Botryococcus*, *Pediastrum*, *Spirogyra*) is concluded for this episode.

Regionally a herbaceous-rich vegetation is present dominated by Gramineae, *Helianthemum*, *Thalictrum*, *Artemisia* and various Compositae indicating a relatively dry steppe-tundra like vegetation. Biostratigraphically this episode can be correlated with the Older Dryas. *Hippophaë* is usually encountered in levels dated to the Older Dryas. Here *Hippophaë* possibly acts as a pioneer on the calcareous rich pointbar systems.

Resuming a dry regional vegetation coincides with a break in the fluvial activity and deposition of lacustrine sediments in the palaeomeander at Schuitwater. Possibly we are registering a period with a pronounced decline in the effective precipitation and diminished discharges in the fluvial regime.

From 345 cm up to 297 cm the sediment of the meander infill gradually becomes more sandy. *Polemonium*, *Linum anglicum*, Cruciferae and Chenopodiaceae are added to the steppe-tundra like vegetation probably emphasizing the

prevailing dry conditions.

At 297 cm all the dry elements in the vegetation show a distinct decline or become absent in the pollen record. *Pinus*, a dry continental species, drops and *Betula* and *Juniperus* show a distinct rise. Both latter species indicate a return to wet conditions and a simultaneous increase in the groundwater table. Discharges resume as indicated by the return of reworked species (*Nyssa*, *Carpinus*, *Alnus*) culminating in the deposition of a sandy gyttja between 278 and 282 cm. Locally aquatic species like *Potamogeton* and *Myriophyllum* are encountered. Following the 282 cm a hydroseral succession is registered in which at first the Cyperaceae play part followed by a strong local presence of *Betula*. The infill at this level consists of a coarse detrital gyttja. The *Betula* rise at 297 cm biostratigraphically coincides with the start of the Allerød.

Boring SW 2A

Registration in SW 2A starts where in SW 2B fluvial activity resumes, which coincides with the start of the Allerød. After the first rise a similar drop in the *Betula* and a concomitant rise in the *Juniperus* is registered. Subsequently fluvial activity in the palaeomeander fades out and a hydroseral succession comparable to the one in SW 2B occurs. The resurgence of *Betula* coincides with a change in the lithology towards a coarse detrital gyttja.

Halfway the *Betula*-phase of the Allerød a temporary wet interval is present, where *Nymphaea* cf. *candida*, a nordic species, occurs. This wet interval with *Nymphaea* also appears in the top of the analysed part of SW 2B.

Following this wet interval conditions became progressively drier as appears from the spread of ferns (monoletete psi.). Locally a carr peat develops indicating that the infill has reached a semiterrestrial character. Under the prevailing dry conditions *Pinus* gradually increases. This increase is temporarily interrupted by a subsequent wet phase (*Menyanthes*, *Utricularia*) before *Pinus* reaches dominance over *Betula* (the *Pinus*-phase of the Allerød). The lithology at this level changes into an amorphous peat indicating relatively dry conditions during the period of deposition. Possibly discharges reach a minimum during the *Pinus*-phase of the Allerød only to resume at the Allerød/Younger Dryas transition (see excursion site 5: Bosscherheide).

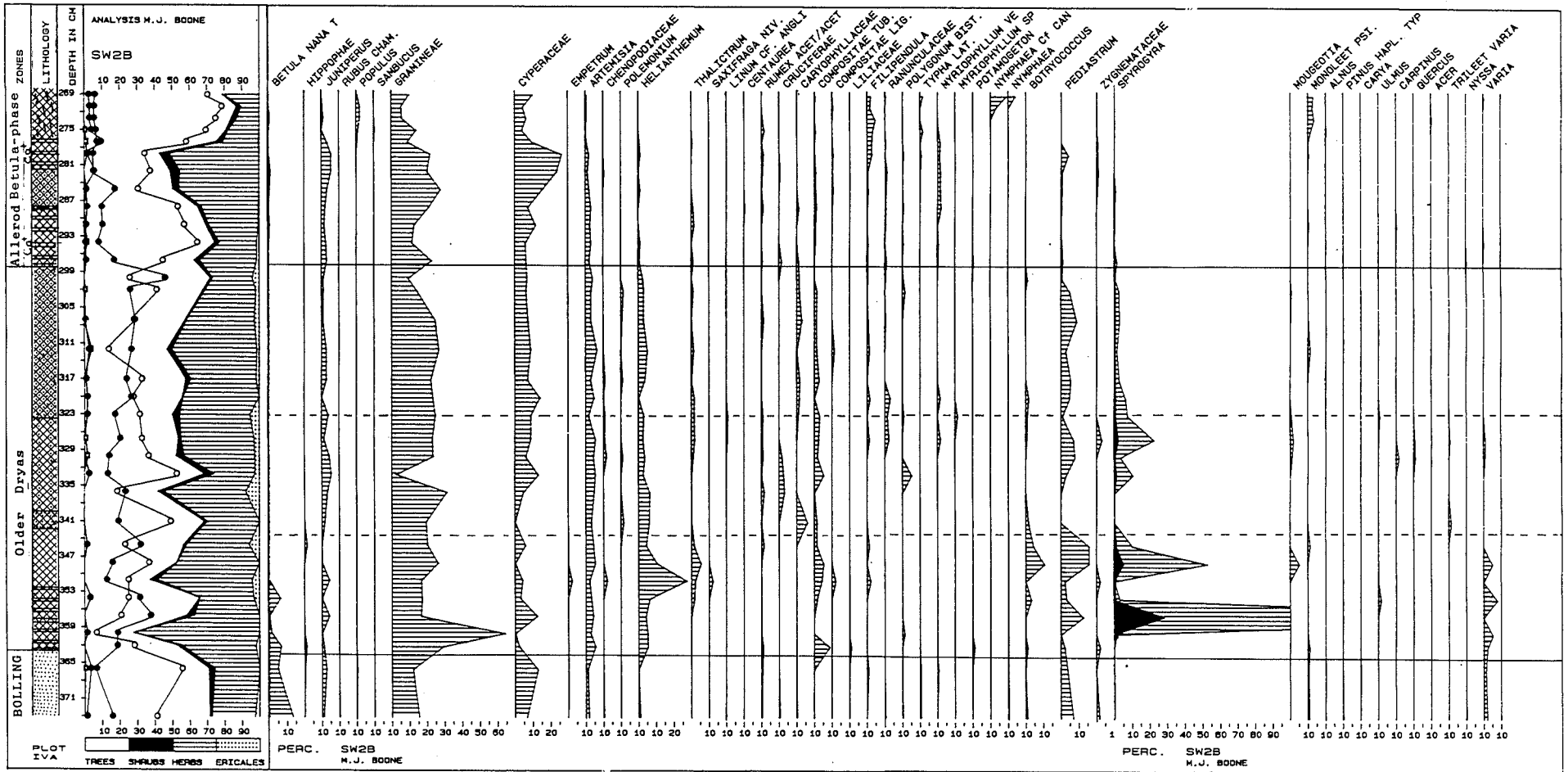
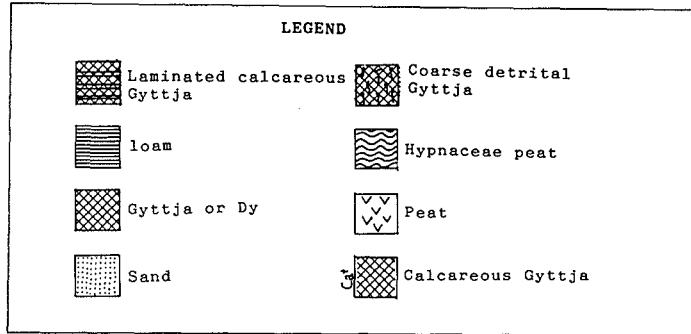
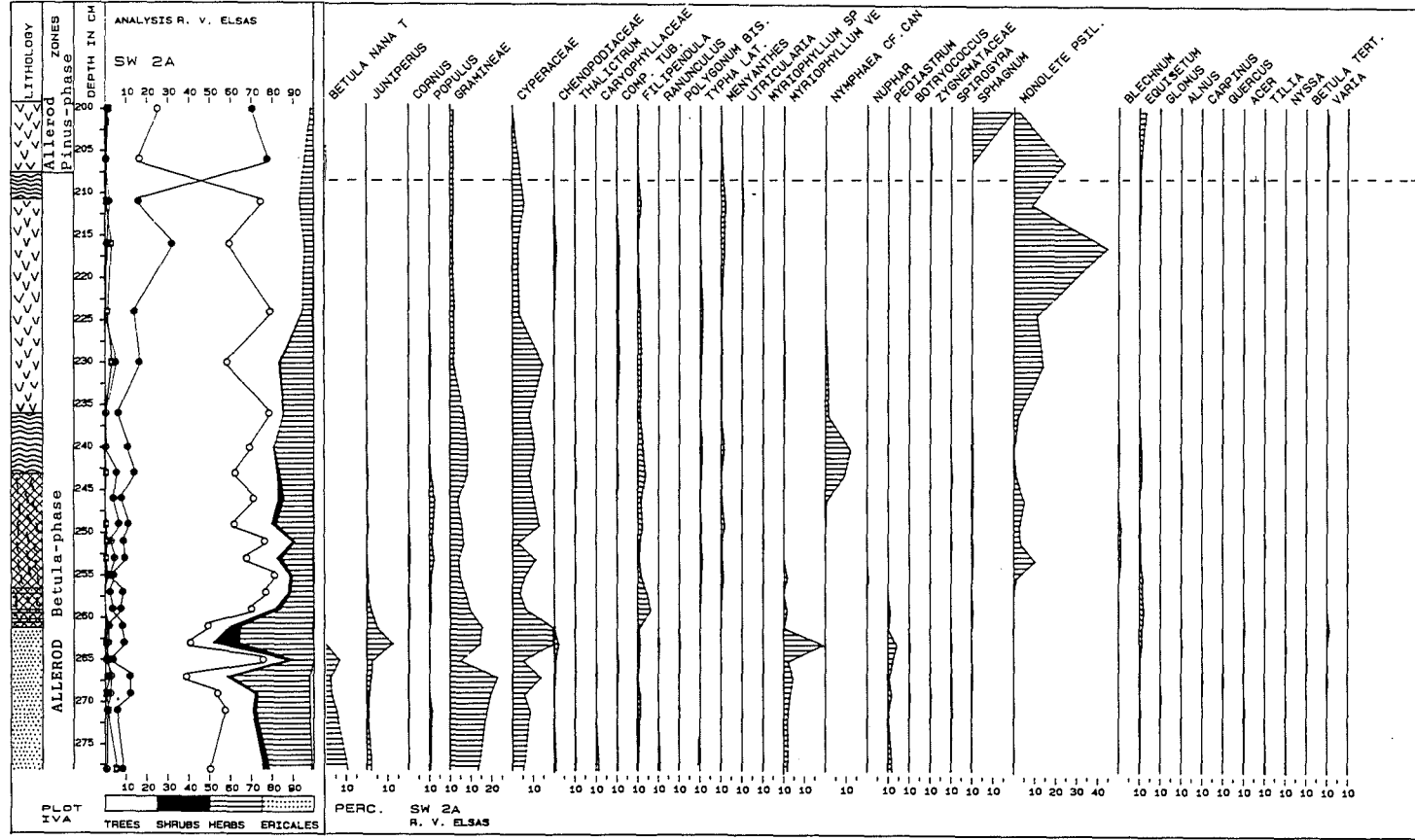


Fig. 6.5 Pollendiagram of Schuitwater (SW) 2B

Fig. 6.6 Pollendiagram of Schuitwater (SW) 2A



The lower terrace level declines from 18 m in the south to 14 m in the north of fig. 6.3. It is characterized by its straight, elongate paleofloodplain with straight to low sinuous scars (fig. 6.3: excursion stop B: Aastbroek). Straight scars occur especially along the terrace edge to the higher meander scar terrace. Towards the present river the scars develop into low sinuosity scars, more or less conform the actual low sinuosity river course.

At excursion point B the lower terrace is 15 to 16 m above sea level, which is approximately 5 m below the neighbouring meander scar terrace (fig. 6.4). The altitude differences on this terrace, between the low lying scars and intermediate higher areas, are 1.5 m. The higher areas cannot be individual pointbars, since they are too large and too widely spaced (3 bars over 600 m). They could represent several higher pointbars, coalescing to form larger bars; the lower areas between the large-scale bars can be channels or complexes of low pointbars. The large compound bars are clearly conform the present river course, which proves the fluvial origin. Eolian sediments have not been found on this terrace.

The low sinuosity of the bars and intermediate scars and the position on the inner bend of the present river, give the impression of side attached bars in a more or less straight paleofloodplain. The side bars developed perhaps due to long term variations in discharge during the lateral migration of the channel to the east. Although there are differences in elevation between the bars and the depressions in between, the tops of subsequent bars appears to be on the same level. This indicates that the high water level was fairly constant during subsequent bar formation. However, coring revealed that the base of the lower terrace at point B dips rather steeply from c. 3.25 m below the surface close to the terrace edge to c. 6.5 m below the surface close to the river. This dip of the base of the terrace points to a strong vertical incision during the lateral migration of the river channel. As the top of the terrace is found at a more or less constant level it is concluded that during the formation of the lower terrace, the river morphology changed from a broad and shallow braided channel into a narrow and deeper low sinuosity meandering channel, like the actual river course.

Lithology

The corings reveal that the two terraces have a different lithological build-up. The higher terrace with meander scars is characterized by a very thick (7.5 m) fining-upward sequence, formed by lateral migration of the channel and accretion on the meander inner bend. The fining-upward sequence consists of 2 m gravelly, poorly sorted, medium to coarse sand (300-850 μm) at the base overlain by a transitional bed of circa 1 m of moderately sorted, fine to medium sand (150-300 μm). This coarse grained lower part was formed by strong tractional currents on the meander channel bottom and lower part of the inner channel slope.

The upper 4.5 m of the fining-up are moderately or more often well sorted fine sands (105-210 μm) with thin sandy silt beds (1-13 cm), which increase in number and thickness towards the top. Some smaller fining-up sequences, separated by erosional boundaries, are present within this fine sand unit, probably reflecting reactivation surfaces of the meander inner bend during high discharges. The fine-grained, well sorted upper part was deposited by weaker tractional currents or it settled from suspension on the upper channel slope. The silt beds

reflect high water levels on the upper pointbar slope with local weak currents and deposition of fines from suspension.

The fluvial sediments are covered by moderately sorted fine sand, often with coarser sand grains and fine gravel (2-3 cm). The morphology, the lateral and vertical homogeneity and the horizontal to low-angle cross-bedding indicate an eolian origin. The thickness varies from 0 to 4 m depending on the morphological position.

The lower terrace level reveals laterally a larger heterogeneity than the higher terrace level (preliminary results). The almost straight paleochannel scar along the edge of the terrace contains a circa 1 m thick clay-layer, directly overlying medium sand and gravelly sands. The lag deposit of the channel is found at c. 3.25 m below the surface.

In the direction of the present river the base of the terrace declines. Here the fluvial sequence is characterized by a fining-upward, which resembles the sequences of the higher meander scar terrace. The lag deposit occurs at 6.5 m below the surface. The base of the sequence consists of 2 m medium to coarse, gravelly sand, which is overlain by fine to medium sands (1.5 m). The upper 3 m consist of clayey fine sand. The top of the section is modified by soil formation. The reddish brown colour and the clay illuviation in the Bt horizon indicate the presence of a Luvisol.

Eolian phases

At least two eolian "phases" are recognised in the Late-Glacial river morphology (fig. 6.3). The oldest phase is found on the higher terrace level east of point A. The meander inner bend is covered by a 0 to 4 m thick unit of eolian sediments, which are more extensive than illustrated in fig. 6.3. Regarding the westerly winds during the Late-Glacial (Maarleveld, 1960; Schwan, 1988), these eolian sands could originate theoretically from the extensive Pleniglacial coversand area west of the paleomeander scar. However, there are no indications for eolian sediments within the scar, nor dunes migrating into and crossing the paleochannel. Furthermore, if this had been the case, then the stagnant paleochannels should have been filled with sand instead of gyttja and peat. Therefore, it is suggested that the eolian sand originated from within the meander belt itself. Two possibilities then exist: eolian activity occurred after the meander belt was formed or it occurred during the lateral migration of the meander.

If the eolian deposition took place after the formation of the terrace, then the eolian sediments were derived by local deflation from the top of the fluvial sediments. The volume of the blow outs should then be more or less equal to the amount of sand in the eolian accumulations. However, a large part of the meander inner bend east of point A is covered with eolian deposits and blow-outs eroded into the fluvial substratum have not been found. In the deepest blow-out in the dune area, the silty beds, which occur at the top of the fluvial fining-up sequence, are found immediately below the surface, illustrating that no fluvial sand has been eroded from the subsoil to build the dunes.

Since there is no evidence for an eolian supply from outside the area, nor for a supply from the subsoil, one must conclude that these eolian sediments on

top of the terrace east of point A were formed during the lateral migration of the meandering river channel. The comparable grain-size of the eolian deposits and the upper channel slope deposits support this hypothesis. The following mechanism has to be considered. During bankfull discharge, probably in spring due to melting of the snow cover and low evapotranspiration, fresh sediment was deposited on the meander upper inner bank. During the following low discharge this barren sediment on the upper slope was deflated by the prevailing westerly winds and deposited on top of or just behind the point bar. Since the fluvial ridge and swale morphology of the meander inner bend is still roughly visible, it is assumed that the windblown sand was stopped by vegetation on the point bars. If the point bars had been unvegetated as well, then the fluvial morphology should have been obscured by the eolian deposits, which is not the case. The low dune morphology itself is an additional argument in favor of eolian deposition in a vegetated landscape.

As the start of the meander scar infill could be dated in the Bølling, it is clear that the lateral channel migration and the connected deflation and dune accumulation took place during the Bølling and/or shortly before the Bølling, during the Oldest Dryas period. This Bølling-Oldest Dryas eolian sedimentation represents an older "phase" than the Younger Dryas eolian deposition discussed below.

On the east side of the Late-Glacial Meuse valley large dune complexes occur, lying on the Late-Glacial higher terrace or older fluvial deposits (fig. 6.3; see excursion site 5: Bosscherheide).

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

At the base of the dune sediments Late-Glacial organic sediments and peats are present, which offer the opportunity to date the start of the eolian activity (Vandenberghé et al., 1991; see excursion site 5: Bosscherheide). From the top of the peaty layer, characterized by an alternation of moss-laminae and eolian sand-laminae, a C14 dating of 10.500 ± 60 BP was obtained, which places the overlying dune body in the late Younger Dryas period.

The eolian sediment of this phase is fine to medium grained, which is coarser than the Bølling/Oldest Dryas eolian sediments at point A. Because of the prevailing southwestern winds, the source area was west of the dune field. Since the higher Late-Glacial meander scar terrace on the east side of the Meuse is covered also by these eolian sands, it was previously thought that the lower Late-Glacial straight to low sinuosity scar terrace was the source of the dune sand (Wolfert & De Lange, 1990). However, there is no evidence for eolian activity on this lower terrace. There are no eolian sediments on the lower terrace surface and the ridge and swale morphology is clearly of fluvial origin, since it conforms the present river course. Therefore, the eolian sediments must have been derived from a not-preserved Younger Dryas paleofloodplain. The fluvial morphology of the lower terrace is probably (slightly) younger than the Younger Dryas eolian phase.

It is likely that the eolian sediment was blown out of the Younger Dryas floodplain during periodic low discharges. Only the sediments that were deposited outside the floodplain, on the Late-Glacial meander scar terrace or older fluvial units, were preserved. The eolian sediment within the floodplain had a low preservation potential, since it occurred on top of the fluvial sediments and therefore was

easily eroded during subsequent periods of high discharges. If the lower terrace level had already been present during this deflation phase, then the fluvial morphology should have been destroyed and a deflation lag should have been developed on the terrace surface.

The vast amount of eolian sand in the extensive dune field on the eastern bank of the Meuse points to a recurring process of fluvial reworking and eolian deflation of the floodplain. From these alternating processes it is concluded that during the late Younger Dryas the Meuse had a more intermittent character than during the previous Late-Glacial periods. The larger fluctuations in discharge can be attributed to the climatic deterioration during the Younger Dryas. The lower evapotranspiration and the larger volumes of snow melt water will have led to larger peak discharges. As a result the Late-Glacial meandering river course changed into the straight Younger Dryas course.

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