Weichselian to Early Holocene vegetation development and fluvial adjustment in the Lower Rhine Valley, Germany

The role of climate change, glacio-isostasy and intrinsic characteristics of the river system

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Preface & Acknowledgements

This thesis provides the results of a graduation project forming part of the Master's programme Physical Geography of Utrecht University. Subject of study is the response of the lower Rhine and vegetation to climate change during the Weichselian and early Holocene. From earlier work it is known that other factors like glacio-tectonics influenced the lower Rhine simultaneously with climate change, so this study aims at gaining more insight in the individual effects of distinct forces (e.g. Busschers et al. 2007; Cohen et al. 2009). The project was executed as part of a combined Master's project in which also Holocene fluvial developments and middle Holocene flooding of the lower Rhine were investigated. The research makes part of the APEX project of the Delta-evolution program (Utrecht University, Department of Physical Geography) focusing on the development of the apex of the Rhine-Meuse delta during the Weichselian and Holocene. Dr Wim Z. Hoek, Dr Kim M. Cohen and Willem Toonen MSc from Utrecht University, The Netherlands, supervised the work.

Performing this research was not possible without the help of members of staff of the Department of Physical Geography. I would like to thank Kim Cohen and Gilles Erkens for recovering my *Kehrum Torfkuhle* peat core from five meters depth, for sharing insights and the interesting discussions. Kim, many thanks for your critical attitude when supervising my work and your great support during the period of thesis writing and colloquium preparation. I greatly acknowledge Willem Toonen for his support last year, especially during the period of field survey. Most of all, I am grateful to Wim Hoek, who initiated me into this research, supervised my work and helped me for many days with enormous enthusiasm in the field. Many thanks to Nelleke van Asch, for helping me with my palynological study, for giving me advice at moments when Wim was not available, and most of all, for her very good company at the university. I also like to thank Kees Kasse of the 'Vrije Universiteit Amsterdam' for his physical help and shared insights during his field visits. Michiel and Babette, I enjoyed our pleasant cooperation last year! Last but not least, I want to thank the farmers in the area of study for granting access to their property and Rita and Horst Heβelmann for a pleasant stay at their ferienhaus.

Anneleen Geurts
July 2011

PART I

Introduction, previous research and research strategy

1 Introduction

Multiple phases of pronounced climatic change characterise the late Weichselian and last glacial-interglacial transition in north-western Europe. During each phase of climate change, vegetation responded by changing its composition and rivers were forced to change their discharge regimes and sediment transfer and to adapt their morphology. The relationship between fluvial development and climate change has been subject of a large number of studies (e.g. Vandenberghe 1995, Blum and Törnqvist 2000; Schirmer 1995) and first attempts have recently been made to investigate this link in a more quantitative way (Erkens 2009, Busschers 2008, Van Balen et al. 2010). However, besides climate change and related alterations of for instance vegetation cover, soil stability and discharge regime, other so-called external (allogenic) factors are considered to be able to force a river system to develop in a certain direction. During Holocene times for instance, backfilling of the Rhine-Meuse delta in The Netherlands occurred as a consequence of sea-level rise which acted as a downstream control by increasing the base level of the system (Berendsen and Stouthamer 2001; Blum and Törnqvist 2000). On a similar time-scale of 10^3-10^4 yr, so-called glacio-isostatic movements of the Earth's crust to Pleistocene ice mass fluctuations are believed to have exerted influence on river functioning (Cohen 2003; Busschers 2008; Verschuren 2007). On longer time-scales (10^5-10^6), tectonics related to redistribution of intra-continental stress is considered to play an additional role (Cohen 2003; Schirmer 1995). Irrespective of the type of forcing, fluvial response is modulated and delayed by internal (autogenic) catchment-specific controls (Busschers 2008; Erkens 2009). Moreover, external forcing factors and internal control mechanisms generally acted simultaneously during Pleistocene times, leading to complex rather than linear fluvial responses (Blum and Tornqvist 2000). This idea becomes more and more realised in studies concerning the imprint of former external forcing events on the sedimentary record of fluvial systems.

This study aims to contribute to our understanding concerning fluvial and vegetational response to climate change and the potential impact of valley reach-specific controls and glacio-isostatic activity on river functioning. The area of study comprises a circa 30 km stretch of the LRV between Wesel (Germany) and the German-Dutch border (Nordrhein-Westfalen; figure 2). The research focuses on the late Weichselian to early Holocene time-interval, however, fluvial developments since the Saalian (de-) glaciation of the area are additionally considered (figure 1). The area suits well for this type of research since it is known to have experienced intense climatic fluctuations and differential glacio-isostatic uplift during the investigated time-interval. Moreover, the relative wide valley morphology in combination with the alluvial setting allowed the former Rhine to respond freely by mobilising its banks and relocating its channel belt. In turn, channel belt migration enlarged the chance for older deposits to remain preserved and for palynological records to become formed by infilling of the abandoned channel systems. Finally, as explained in chapter 2 of this thesis, the impact of sea-level change and long-term tectonics is considered to be negligible during this specific time-interval, leaving a smaller number of external factors to be unravelled in the sedimentary record.

In light of present-day climate change, a proper understanding of the way in which rivers react to climatic forcing is greatly needed to gain knowledge about how to protect intensively populated river valleys worldwide in the future. Moreover, research on the architecture of ancient fluvial deposits is of critical importance because these deposits function as repositories for hydrocarbons, ground water and other resources. Finally, a lot of uncertainties exist about the way in which vegetation will react to present day climate change and especially about the consequences for worldwide biodiversity, what makes studies on the vegetation-climate relationship in the past very valuable.

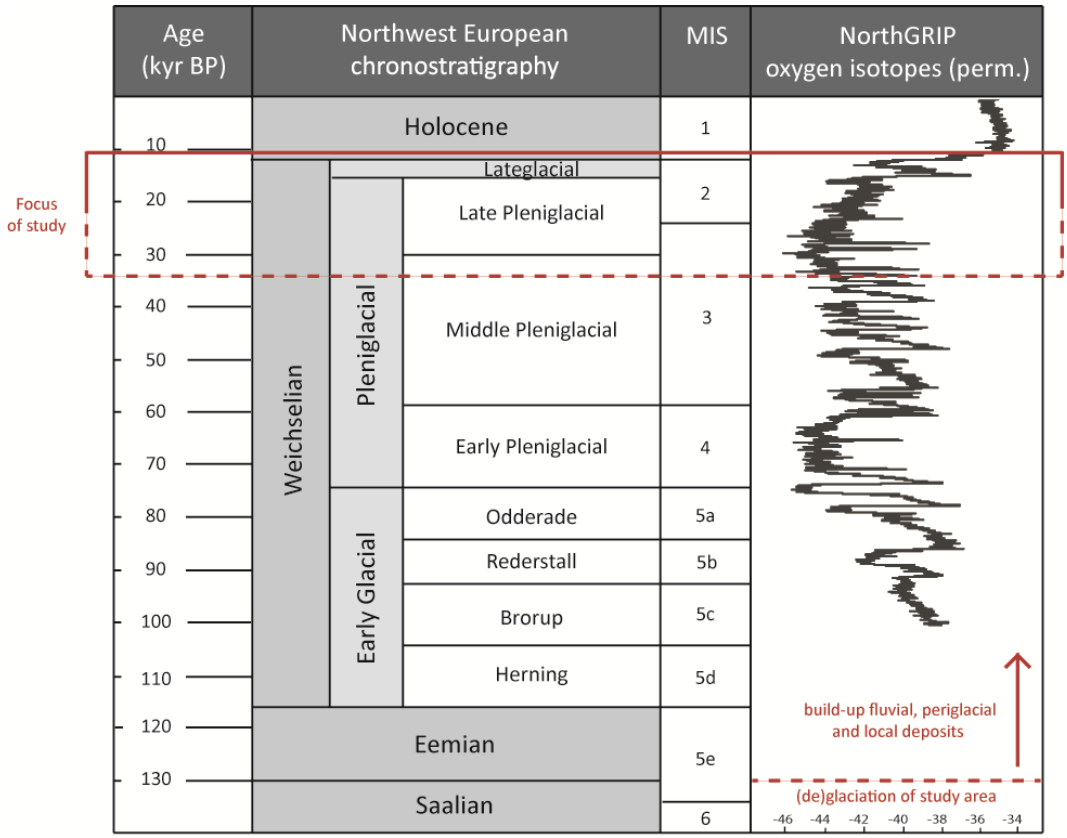

Figure 1 North-west European chronostratigraphy (Vandenberghe 1985; Van Huissteden and Kasse 2001) and climatic history for the last 140 kyr, as reflected by the oxygen isotope record of the NorthGRIP ice core (NGRIP Members, 2004).

Thesis objectives

Since only a limited amount of data was available from the area of study, the first objective was to investigate its sedimentary build-up and to construct a chronostratigraphic framework. Hereto, new field data were collected which were successively integrated with pre-existing data from different parts of the study area. More specific research questions concerning the chronostratigraphy are listed in section 3.3. A second objective was to study the sequence of geomorphological responses of the lower Rhine in time, on the basis of geometric and lithological properties of the different fluvial chronostratigraphic units. Thirdly, it was attempted to investigate the relative importance of different external factors by comparing the inferred fluvial history with climatological data, vegetation records, outcomes of studies on glacio-isostasy and the well-studied fluvial history of the Rhine-Meuse delta directly downstream. An additional objective

was to reconstruct vegetation development across the last glacial-interglacial transition on the basis of a newly derived sequence of palynological records.

The area of study contains the apex of the modern Rhine delta near its downstream end, following the sedimentological definition of the apex as the location where incision continues into aggradation in a downstream direction (near Rees, figure 2). The research area contains the bifurcation point between the northward directed former Oude-IJssel-Rhine course and the more westward orientated present-day active Gelderse Poort-Rhine. A third former Rhine branch, namely the Niers-Rhine valley system, diverges from the trunk valley approximately 20 km upstream of the study area (figure 2). Today, the area of study supports intensely used farming land, especially characterised by corn fields and pastures.

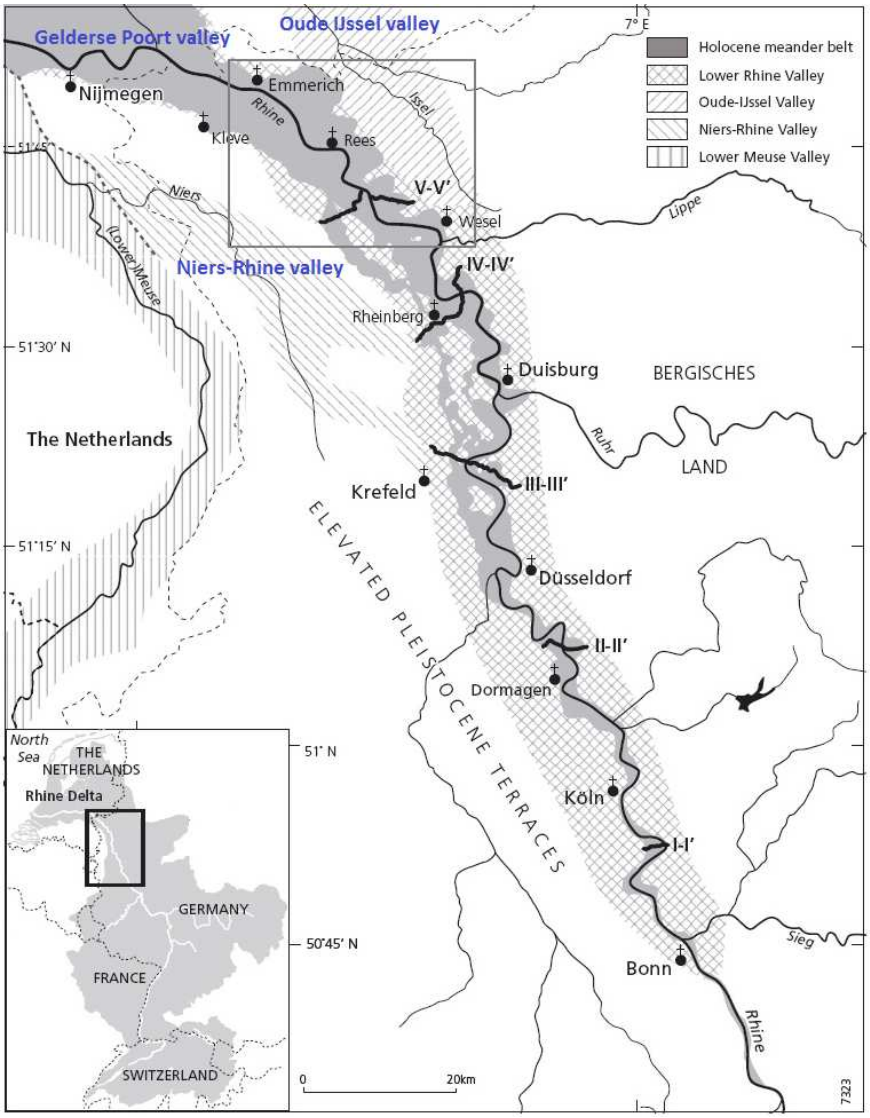

Figure 2 Setting of the area of interest in the downstream part of the Rhine catchment and the lower Rhine valley. The area of interest comprises the upstream part of the Oude IJssel and Gelderse poort valley systems. Together with the more upstream diverging Niers-Rhine valley, they form three branches of the lower Rhine (modified from Erkens 2009). Cross-sections I-V together span the whole lower Rhine valley and are provided by Erkens (2009) and Erkens et al. (2011). The most downstream ones, cross-sections IV and V, are located within and approximately 20 km upstream of the area of interest, respectively, and appeared to be very relevant and useful for present study.

Thesis outline

After this introduction, the thesis continues with three chapters which provide an overview of previous studies relevant for present study: Chapter 2 gives an overview of studies concerning the impact of external forcing and internal controls on fluvial dynamics. Chapter 3 lists previous research concerning the geology and palaeogeographic evolution of the study area in the downstream part of the LRV. In chapter 4, the general trend in Weichselian and early Holocene climate change and related vegetation developments is outlined for north-western Europe. In chapter 5, research approach and different research methods will be described, followed by two chapters with study results: Chapter 6 describes the sedimentary architecture of the lower Rhine valley on the basis of geological cross-sections and surface geomorphology. Additionally, sedimentary characteristics typical for channel-infillings of the LRV are analysed and described. Chapter 7 provides new palynological data obtained from six cores from the LRV. Regional biozones are defined and described and four cores are biostratigraphically correlated, together spanning the second half of the Lateglacial and early Holocene time-interval. In chapter 8, a floodplain chronology is constructed which forms an alternative to the pre-existing floodplain chronology of this area. Moreover, geomorphological response of the lower Rhine is analysed. A comparison is made with pre-existing results from the area itself and from valley reaches up- and downstream (chapter 9). Moreover, in this chapter, the potential link between climate change, glacio-isostasy and fluvial developments is discussed. In chapter 10, all data are synthesised, resulting in a reconstruction of the fluvial and vegetational history of the study area which is illustrated by a series of palaeogeographic maps. Chapter 11 lists the main conclusions of the research. In the final chapter, the author gives recommendations for future research (chapter 12). The thesis ends with a summary of the study results.

To improve the readability of this thesis, the study area is generally referred to as the lower Rhine valley (or LRV) despite the fact that it actually only comprises the most downstream ca. 30 km of the total lower Rhine valley (figure 2). Further, calendar years are used to express ages, unless stated differently.

2 The influence of external controlling factors and fluvial system-specific characteristics on river functioning

This chapter provides some theoretical background on the impact of external (or allogenic) factors on fluvial developments in general and how fluvial response is modulated by river system-specific characteristics (internal or authogenic factors). In line with the subject of study, this overview focuses on external forcing on time-scales of 10^3-10^4 and on alluvial rivers. This type of rivers is able to respond rapid on similar time-scales because of their mobile beds and banks (Blum and Törnqvist 2000). The following external factors are considered: climate change in combination with vegetation cover and soil stability (section 2.1), neotectonics and glacio-isostasy (2.2) and sea-level rise (2.3). Last section gives an overview of the additional role of intrinsic characteristics of the river system (2.4).

2.1 Climate change, vegetation cover and soil cohesion

Climate change forms an upstream controlling factor on river functioning by controlling river discharges and sediment supply (Blum and Törnqvist 2000; Berendsen et al. 1995). Especially on Quaternary (or Milankovitch) time-scales of 10^4-10^5 yrs, cyclicities within alluvial successions are generally interpreted as depositional responses to cyclic alternations in climatic conditions (Blum and Törnqvist 2000). Morphological and sedimentary expressions are for instance terrace staircases, braided-meandering changing planforms and coarse-fine alternating fluvial deposits (Lowe and Walker 1997). A number of theories exist which try to explain recorded fluvial developments by changing climatic conditions (a review is given by Vandenberghe 2003). Well-known is the *linearity principle*, which links directly and linearly climate to fluvial pattern and process (e.g. Büdel 1977, from Vandenberghe 1995). According to the linearity principle, glacials correspond to periods of aggradation by braiding rivers due to high peak discharges (snow- and (ground)ice-melt during spring), a sparse vegetation cover and a large sediment supply. This is in contrast to incision or the lack of depositional activity during interglacials when rivers have a meandering mode. Thanks to improved dating methods, enabling a more precise comparison between the timing of fluvial developments and climate change, it has become increasingly recognised that climatic perturbation often involves a multi-phase response (Busschers 2008). This is for instance included in the model proposed by Vandenberghe (1995) who describes a *non-linear fluvial response to climate change* due to a delayed response and vegetation composition to climatic changes. According to this model, instable conditions, characterized by a phase of erosion followed by a phase of sedimentation, coincide with transitions from glacial to interglacial climatic conditions and the other way around. Vandenberghe (1995) based this conceptual model on dated fluvial sediment sequences in the lower Meuse valley and Dinkel valley, in respectively the southern and eastern part of the Netherlands. The model is graphically showed in figure 3 and is described briefly here. Factors responsible for non-linear response but not related to climate are described in section 2.4.

A temperature drop at the beginning of each cycle causes lower evapotranspiration resulting in higher amounts of runoff and a larger river discharge. However, vegetation cover is persisting for certain time, keeping the soil relative stable. An increased discharge to sediment supply ratio induces river incision. As soon as the vegetation cover is disturbed, sediment yields increases and river aggradation starts. Aggradation takes place primarily by braided rivers, because of irregular river discharges and a large sediment supply, and continues to previous levels by infilling former channels. The later part of the glacial period shows more or less a balance between erosion and aggradation. When the temperature starts to rise at the transition towards an interglacial period, vegetation does not correspond directly causing the evapotranspiration to remain low and discharges to become relative high for a certain time. Because soil cohesion is recovering faster than vegetation cover, an increased ratio of discharge to sediment supply induces again river incision at the climate transition. Commonly this goes together with a meandering mode. As soon as evapotranspiration becomes high enough to reduce the ratio of discharge to sediment supply significantly, channel infilling starts for returning to the previous landscape. The later part of the interglacial is characterized by valley bottom stability and by soil formation (Vandenberghe 1995).

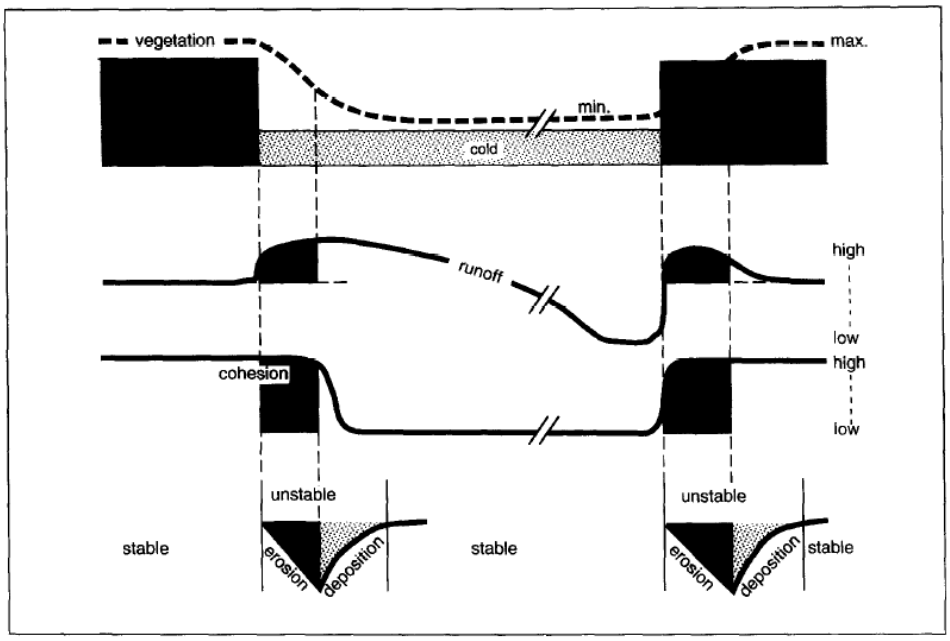

Figure 3 Model of fluvial stability-instability alternations in relation to climate and climate-derived parameters (Vandenberghe 1995).

The effect of the delayed response by vegetation and soil is thought to be reflected by a gradual transition from one fluvial state to another following a relative rapid climate change. For instance, the river Meuse in The Netherlands needed approximately 1300 yr to fulfill a transition from a braided system towards fully meandering system in response to Lateglacial climatic amelioration (Vandenberghe 1995; Tebbens et al. 1999). A more prolonged effect is suggested by Busschers et al. (2007) for the cooling limb of the cycle during the early Weichselian: persisting soil complexes are considered to have prevented large-scale regolith erosion for over 50 kyr, even under severe Early Pleniglacial climatic conditions.

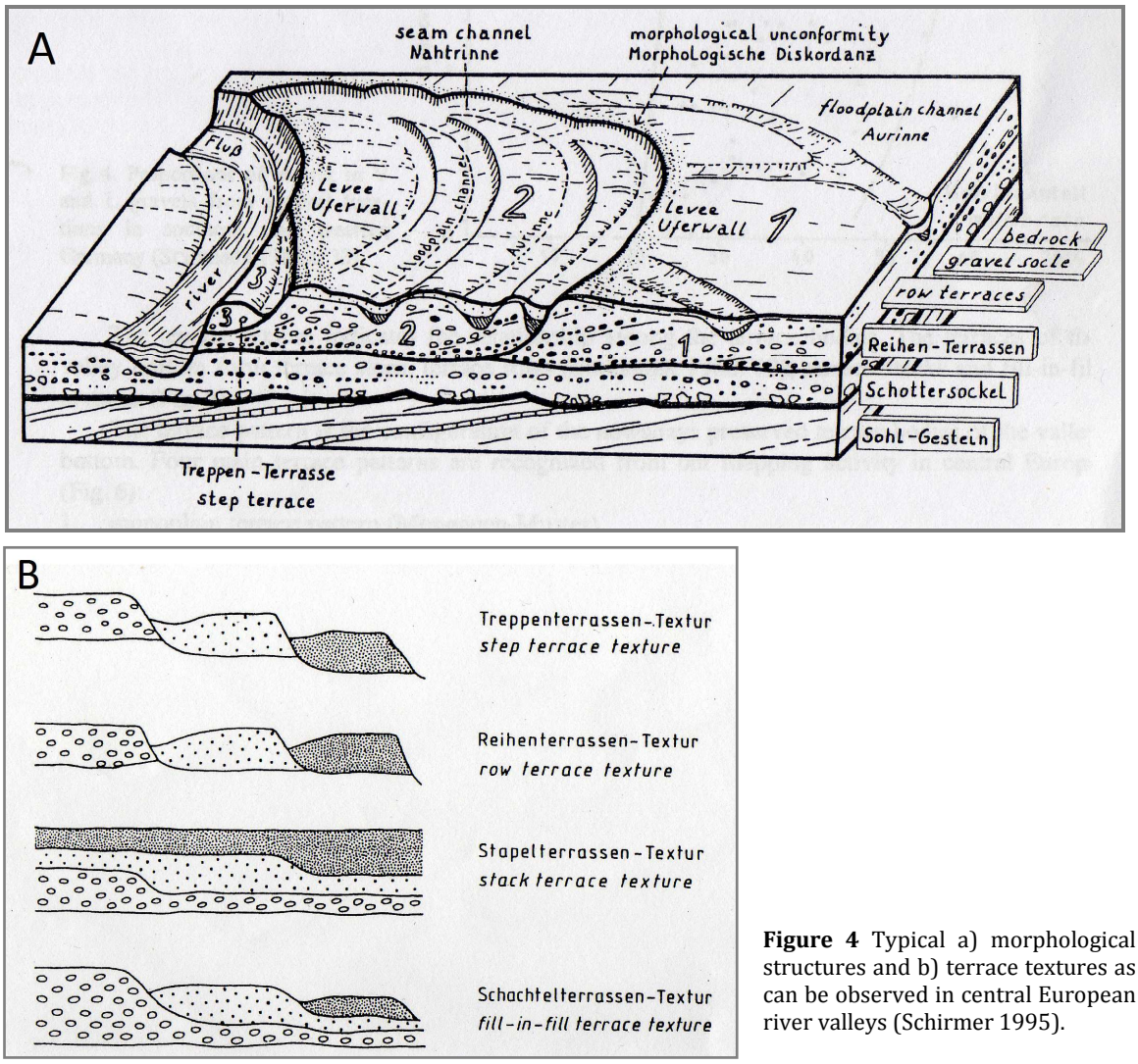

Figure 4 Typical a) morphological structures and b) terrace textures as can be observed in central European river valleys (Schirmer 1995).

Phases of river incision and aggradation do not only affect the vertical sedimentary profile of valley bottoms, but also act in a horizontal direction by means of lateral erosion and accretion. Depending on external forcing and internal control, vertical and horizontal alterations of the valley bottom in time, finally accumulate in distinct river terrace textures. Four types of terrace textures have been defined by Schirmer (1995) for a number of central European river valleys covering the Apline foreland (figure 4b). In a lateral direction, moreover, reworking of sediments might take place when there is no net loss or input of sediment (no net erosion or deposition). A terrace which is formed by lateral reworking of sediments, with a stable channel-base elevation in time, is called a *L-terrace* by Schirmer (1995) and is characterised by a strong vertical fining-upward tendency of the channel sediment (upward increase of sand:gravel ratio). A terrace which is formed by an aggrading system is called a *V-terrace*. The lateral and vertical sedimentary and erosional processes additionally result in a wide range of morphological features on top of the valley bottom, which are graphically showed in figure 4a.

2.2 Tectonics and glacio-isostasy

The basic principle behind the relationship between tectonic movement and river response is rather straightforward: Uplift of the river valley results in river incision, subsidence of the river valley in aggradation because the river systems aims at maintaining equilibrium. However, the impact of tectonics on river functioning is generally altered by the interaction with other external factors what makes it often difficult to quantify. For Quaternary studies, two types of crustal movements might be relevant, depending of the time-interval under consideration (Kiden et al. 2002): First, long-term movements on time-scales of 10^5-10^6 yr or longer, related to plate-tectonics, loading by overlying sediments and compaction of sediments. These movements could be considered linear at shorter time-scales or even negligible. Cohen (2003) quantified activity along the main faults of the tectonic rift structure of the Roer Valley Graben system in the south-central Netherlands. The most active faults are the Peel Boundary Fault and Tegelen fault, which are located ca. 40 and 30 km to the south-west of the area of study, respectively (figure 5). On the basis of displacements along the longitudinal profile of the Late Pleniglacial fluvial terrace, relative subsidence rates could be calculated for the last 15 kyr: 9-15 and 2-11 cm/kyr for these two faults, respectively. The area of study, however, is located outside this area of relative strong tectonic influence and experiences modest uplift. According to Klostermann (1992) tectonic control on fluvial developments is probably insignificant for time-intervals of a couple of ten-thousand years.

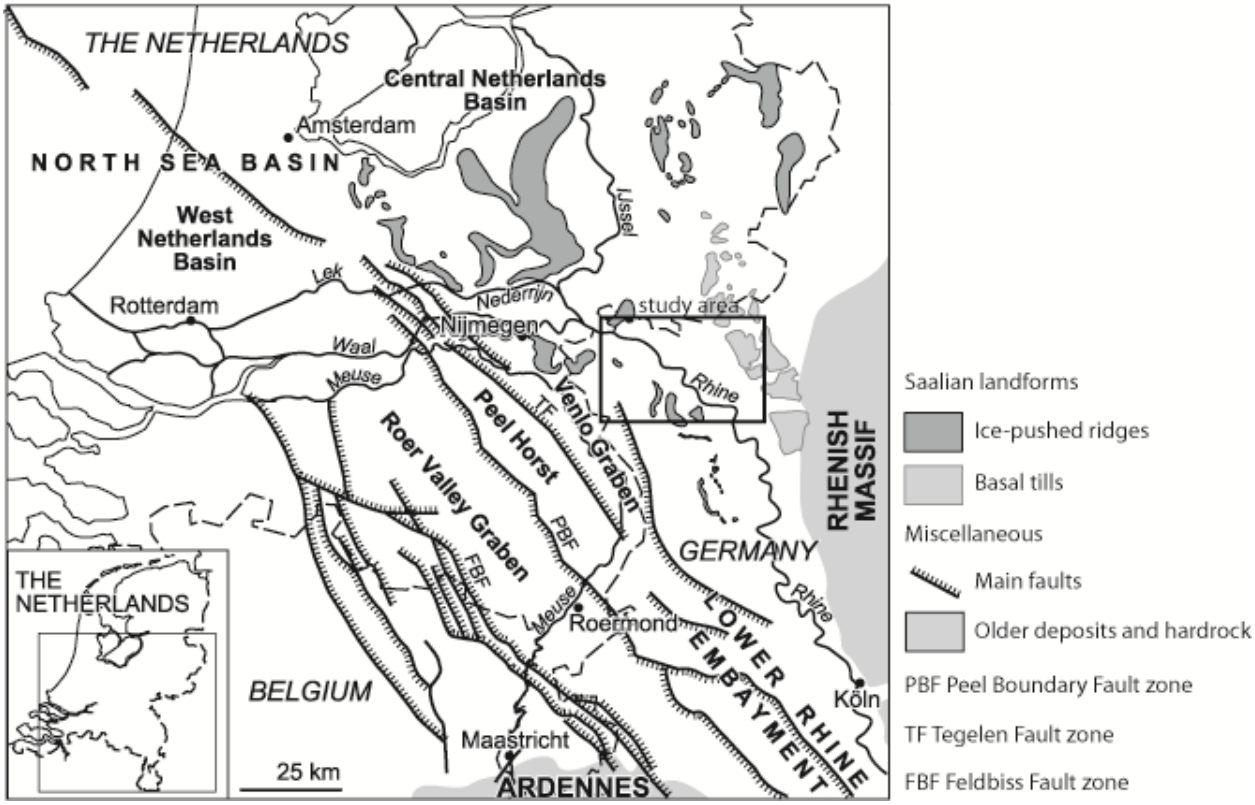

Figure 5 Tectonical structures in the southern-central Netherlands and adjacent regions in Germany and Belgium (Cohen et al. 2002). The area of interest is showed with a framework.

A second type of surface deformation occurs on shorter time-scales (10^3-10^4 years) and is called *glacio-isostasy*. It is the result of loading and unloading of the earth's crust due to the expansion and retreat of continental ice sheets (Lowe and Walker 1997). To compensate for crustal depression underneath the ice sheet, a zone of land fringing the ice-sheet is uplifted, creating a so-called *forebulge* or *peripheral bulge* (figure 7). Simultaneously, unloading and reloading of ocean basins due to capture and release of water by growing and melting ice sheets, respectively, takes place what is called *hydro-isostasy*. The combined effect of *glacio-hydro-isostasy* is still noticeable during interglacial periods following glaciations. Recent studies emphasize the importance of glacio-isostasy for fluvial processes of the Rhine during the Weichselian and early Holocene (Cohen et al. 2009; Busschers et al. 2007; Hijma and Cohen 2010). Forebulge updoming in run-up to the first and second glacial maxima of the Weichselian (circa 70 and 20 ka BP, respectively) followed by forebulge collapse probably influenced the position of distinct Rhine branches, the timing and direction of river avulsions and the processes of incision versus aggradation. During the Weichselian, the LRV was located on the southern flank of the joined forebulge of the British and Scandinavian ice sheets what resulted in a stronger uplift in the north compared to the south. The same applies for the Rhine-Meuse delta, for which Cohen (2003) hypothesizes that Pleniglacial differential uplift resulted in southward shifting of channel belts. A shift towards the north is thought to have happened during Lateglacial-Holocene forebulge collapse (figure 6).

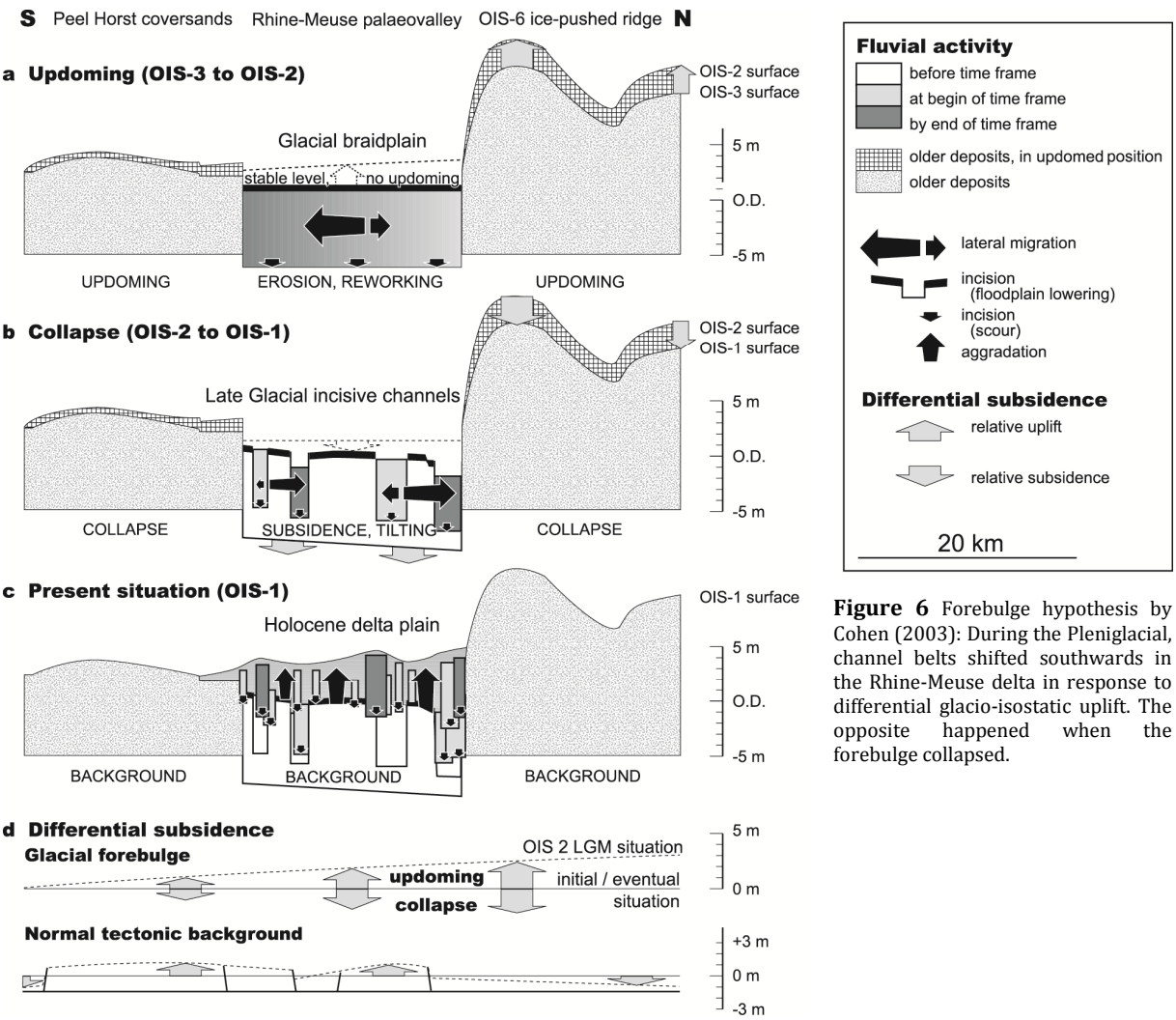

Figure 6 Forebulge hypothesis by Cohen (2003): During the Pleniglacial, channel belts shifted southwards in the Rhine-Meuse delta in response to differential glacio-isostatic uplift. The opposite happened when the forebulge collapsed.

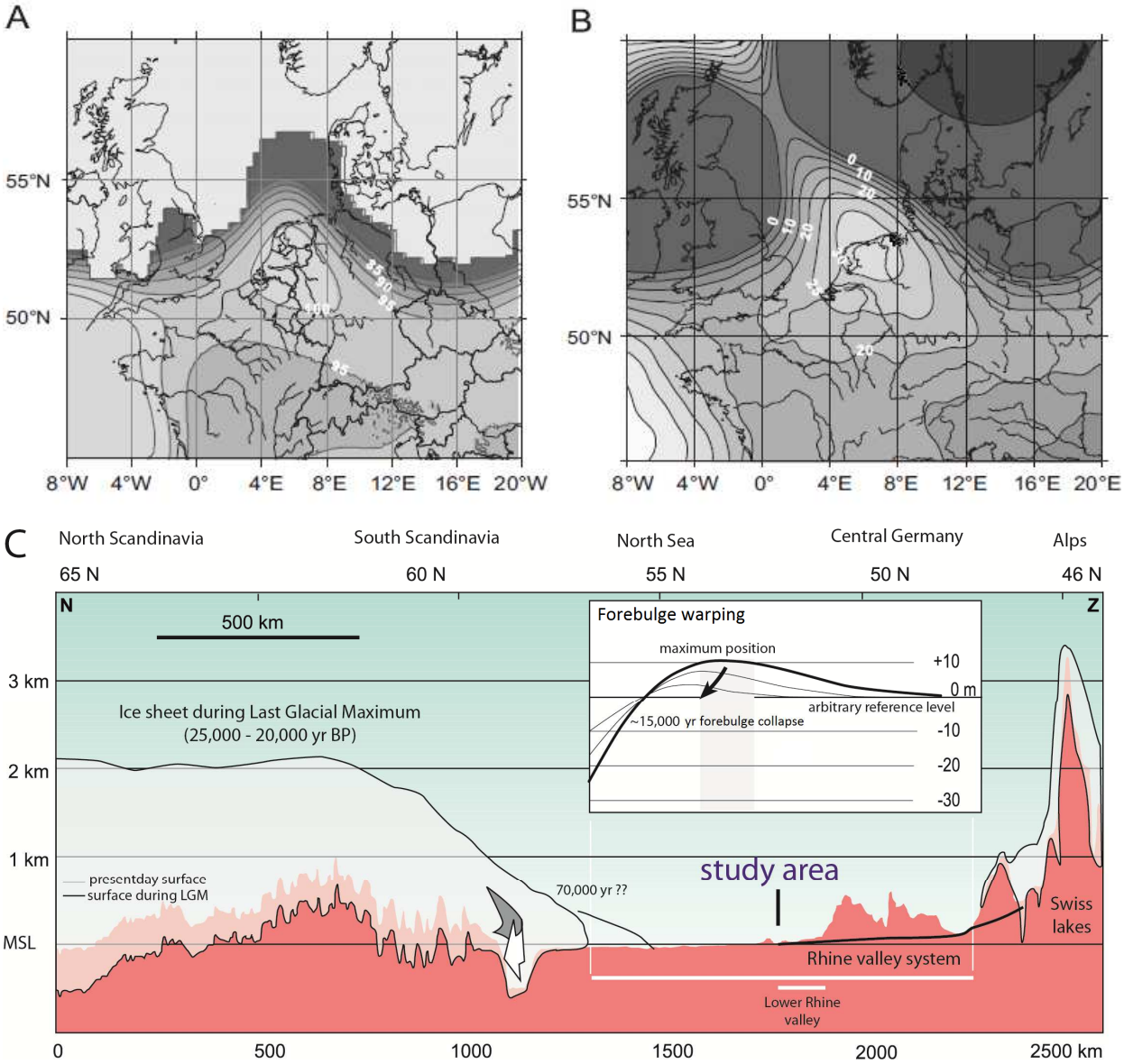

Figure 7 The position and dimensions of the forebulge around the Last Glacial Maximum (LGM, ~21 kyr BP), as modelled by; A) Peltier (2004), and B) Steffen (2006) (from Busschers et al. 2007). Figure C has been constructed by Cohen et al. (2009) and shows a side-view of the Scandinavian ice sheet in proportion to the Rhine catchment during the LGM and the position and dimensions of the forebulge.

Model outcomes in figure 7a and b, show the approximate extent and relative elevations of the last glacial forebulge area from plan view (Peltier 2004; Steffen 2006). In a north-south direction, a maximum difference in elevation in the order of ~ 5 m has been estimated for the whole Netherlands. According to Lambeck et al. (1998), the total uplift of the study area might have been more than 10 m relative to the non-uplifted area near the European Alps. A similar uplift magnitude is showed in figure 7c.

2.3 Sea-level rise

During the Pleni- and Lateglacial, sea level was at least 60 m lower than it is today, with a minimum relative sea level stand in the order of -120 m during the LGM. For the Rhine-Meuse delta, it is a well supported idea that fluvial developments were not directly influenced by eustatic sea level rise before the onset of the Holocene (e.g. Blum and Törnqvist 2000, Berendsen and Stouthamer 2001, Kasse et al. 2005). This can easily be explained by the fact that during most of the last glacial, the Rhine delta was located beyond the Strait of Dover, up to 800 km downstream from the present-day shoreline. According to Cohen et al. (2002), Lateglacial base level was controlled by hard rock in the Strait of Dover at circa 40-50 m below present sea level. In line with these studies, the impact of sea-level rise is considered to have been insignificant for fluvial evolution during Weichselian times in the LRV.

2.4 Internal fluvial controls

Theoretically, spatial and temporal variations in fluvial response might be a direct consequence of external forcing. In reality however, they are often partially related to internal controls of the river system. By internal control is generally meant a process taking place within a river system in relation to for instance sediment or water supply, sediment transport and morphological adjustment of the valley (Van Balen et al. 2010; Erkens et al. 2011). Because these processes take time, time lags exist between initial forcing and ultimate result. Moreover, since changes in external conditions often trigger more than one proces at the same time, these processes interact and modulate each other. The result is a complex and delayed fluvial response (Vandenberghe 2003). Amongst others, the time-interval required for a certain transition depends on characteristics of the system, for instance catchment size, topography and subsoil lithology. Generally, the larger the system, the longer it takes until a new equilibrium is reached. Therefore, the model study by Van Balen et al. (2010) emphasizes the importance of a source-to-sink approach in modelling large river systems. Their study evidences the diachronic character of climatically triggered transitions, because these are first registered in the upper catchment and influence downstream reaches at a later stage. Further, Vandenberghe (2003) mentions the importance of the proximity of a system's state to threshold values. These conceptual thresholds have to be crossed before a certain response is recorded. Because of the existence of thresholds and required responding times, the magnitude and duration of external forcing events must be long enough to allow a river to react (Vandenberghe 2003).

The lack of an accurate chronostratigraphy and the fragmented character of sedimentary records make non-linear fluvial behaviour generally difficult to quantify. According to Van Balen et al. (2010), numerical modelling forms an appropriate alternative.

3 The geology and palaeogeography of the lower Rhine valley

3.1 Geological setting of the lower Rhine valley

The research area makes part of the mid-latitude Rhine catchment of north-western Europe, which measures a surface area of ~ 185 000 km^2. The Rhine has a total length of ~ 1320 km and flows from the Swiss Alps through Germany and the Netherlands towards the North Sea (figure 2). In the course of the Quaternary, the transport of sediment downstream the Rhine and other rivers (e.g. the Meuse), has resulted in the development of a major delta which hosts the Netherlands today. The present-day mean annual discharge of the lower Rhine is approximately 2200 m^3/s near the German-Dutch border. The area of interest is located near the downstream end of the lower Rhine valley (LRV) in the south-eastern corner of the North Sea basin. The upstream end of the LRV lies near Bonn (figure 2), where the river leaves the Rhenish Massif. This can be considered as the apex of the Rhine delta from a long-term geological point of view (Boenigk and Frechen 2006). In a northwest direction the LRV extends to approximately the Dutch-German border, which approximately coincides with the hinge zone between the uplifted LRV and the subsiding North Sea basin. As already mentioned in section 2.2, regional tectonic activity is controlled by southeast to northwest trending faults of the Roer Valley Graben rift-structure present in the border region of Belgium, Germany and the Netherlands. The last phase of tectonic rifting began during the Oligocene (circa 30 million years ago) and is still ongoing (Cloetingh et al. 2006; Schäfer et al. 2005).

During the main part of the Tertiary, the LRV was inundated by the sea and marine deposition predominated. However, in this near coastal marine environment a delta started to be formed by fluvial sediments brought in by the rivers Meuse and Rhine (Zagwijn 1989; Berendsen and Stouthamer 2001). The oldest fluvial deposits of the lower Rhine date back to the Miocene when this part of the river came into existence (Schäfer et al. 2005; Klostermann 1992). At that time, the Rhine was only a small river draining the Rhenish plateau towards the coastline that was located near Düsseldorf (figure 8a). The LRV was subject to relative subsidence with regards to the more intense uplifted hinterland, determining the position of the Rhine and its tributaries in time. Moreover, these tectonic developments led to step-wise increases of the drainage area by backward erosion and the process of stream capturing during the remaining part of the Tertiary (figure 8; Berendsen and Stouthamer 2001). At the onset of the Pleistocene (~2.5 million years ago), the modern catchment size was more-or-less achieved (figures 8c and d). The wide extending catchment including the Swiss Alps functioned as an enormous source of sediment during the Pleistocene (especially during glacial periods), resulting in the relative fast build-up of the Rhine delta (Berendsen and Stouthamer 2001). The coarse sediment delivered originates from the Rhenish Massif, the relative fine material from further upstream areas in central Germany (Schäfer et al. 2005).

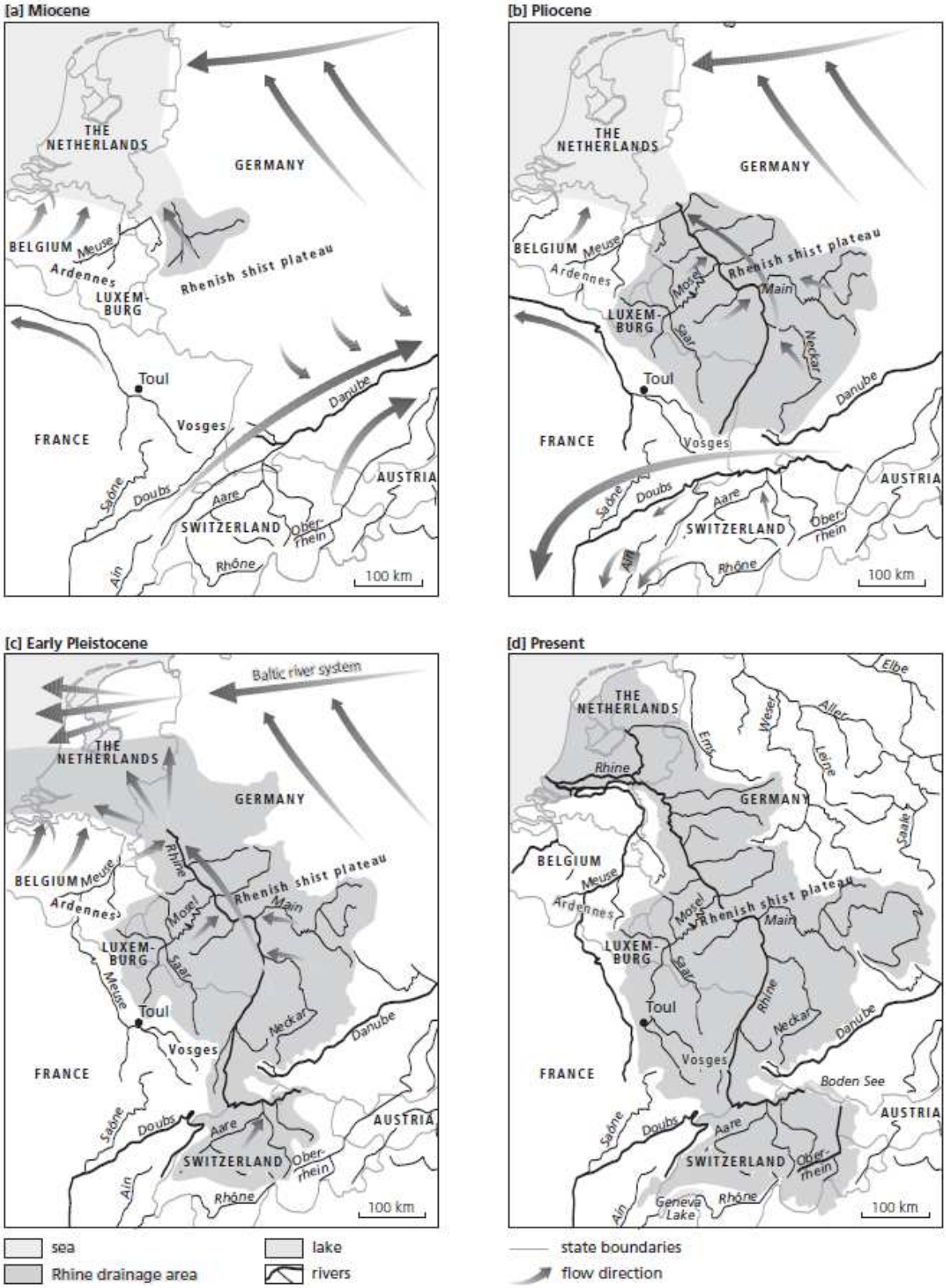

Figure 8 Increase of the drainage area of the river Rhine by stream capturing (Berendsen and Stouthamer 2001).

During the Quaternary, the course of lower Rhine has been determined by the interplay between the tectonics, glaciations and climate change (Klostermann 1992; Busschers et al. 2008; Kasse et al. 2005). Before the Saalian glaciation, the location of the Rhine was mainly determined by the location and relative movement of the distinct tectonic blocks present in the subsoil. Moreover, during this part of the Quaternary, the course of the Rhine shifted in a northeast direction due to tilting of the tectonic blocks towards the northeast (Klostermann 1992). This is illustrated by figure 9: At the onset of the Pleistocene the Rhine was draining through the Roer Valley Graben, but since the Waalian period (~1.5 million years ago) its course is positioned within the LRV.

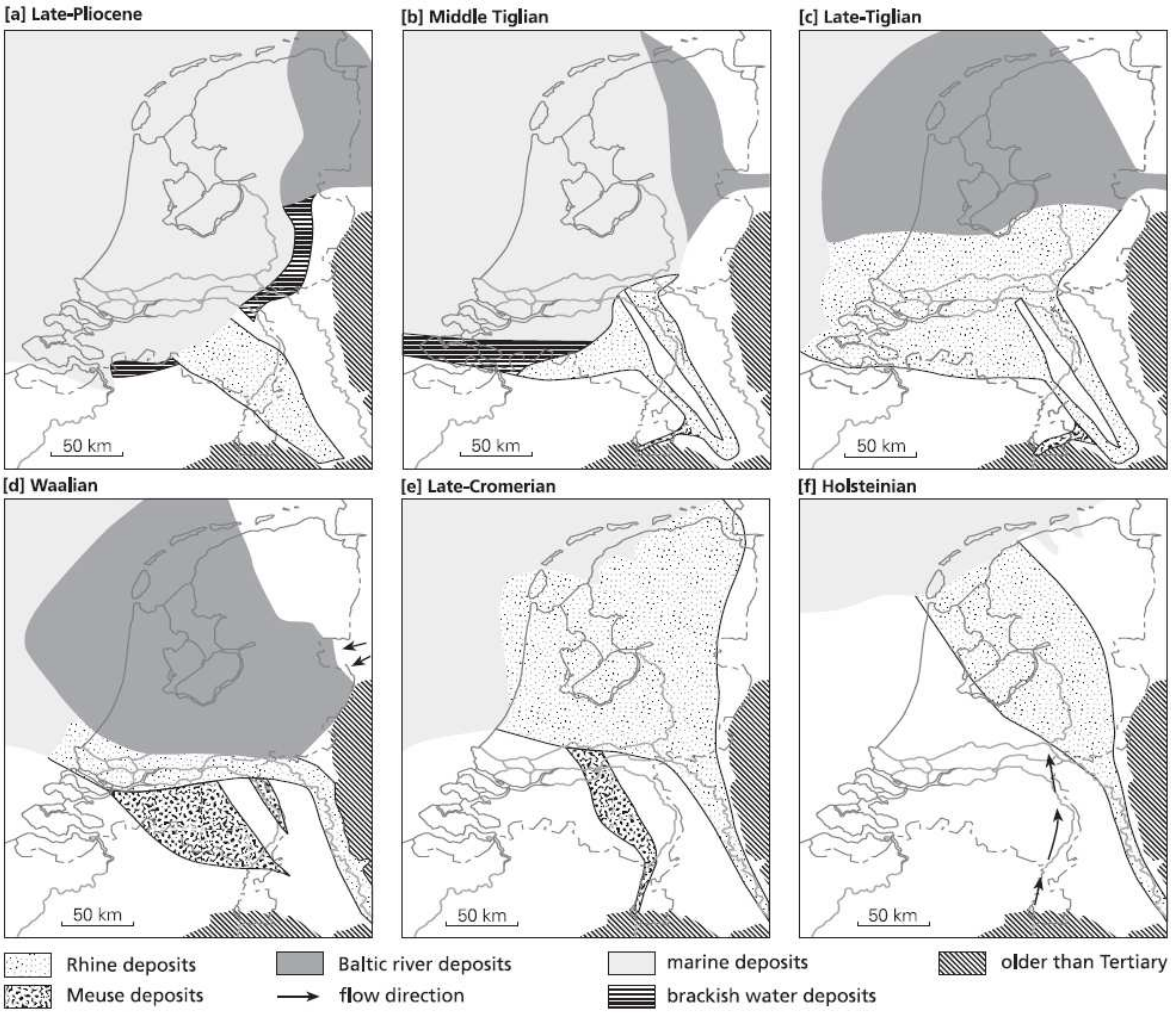

Figure 9 Palaeogeographic situation in the Netherlands during the early and middle Pleistocene (Berendsen and Stouthamer 2001).

3.2 Palaeogeographical developments in the lower Rhine valley since the Saalian glaciation

Drenthe glaciation and deglaciation during the Saalian

During the Saalian glacial period, severe climate cooling caused the Scandinavian ice sheet to expand extremely and to invade the northern part of the LRV. During the most recent and most extreme Saalian ice advance, the so-called 'Drente' or 'Older Saalian' glaciation (~200-130 kyr BP; marine isotopic stage 6 (MIS6)), the study area became covered with ice. The Saalian ice-sheet destroyed the relative flat geomorphology of the Rhine delta and produced an ice-limit morphology composed of ice pushed ridges (Nijmegen-Kleve-Düsseldorf), deeply scoured glacial basins and outwash fans, so-called sandur (figures 10a,b, 11 and 12; Busschers et al. 2008). The invading ice-sheet itself, but also the glacial deposits and morphology it created, forced the Rhine to re-organize its drainage pattern: The Rhine was diverted to the west, forming the Niers-Rhine valley south of the research area (figure 10a; Kasse et al. 2005; Busschers et al. 2008). The Niers-Rhine acted as an ice-marginal river transporting melt water originating from the Scandinavian ice mass and the glaciated Alps during maximum ice-sheet extent of the Drenthe stage (Klostermann 1992, Busschers et al. 2008).

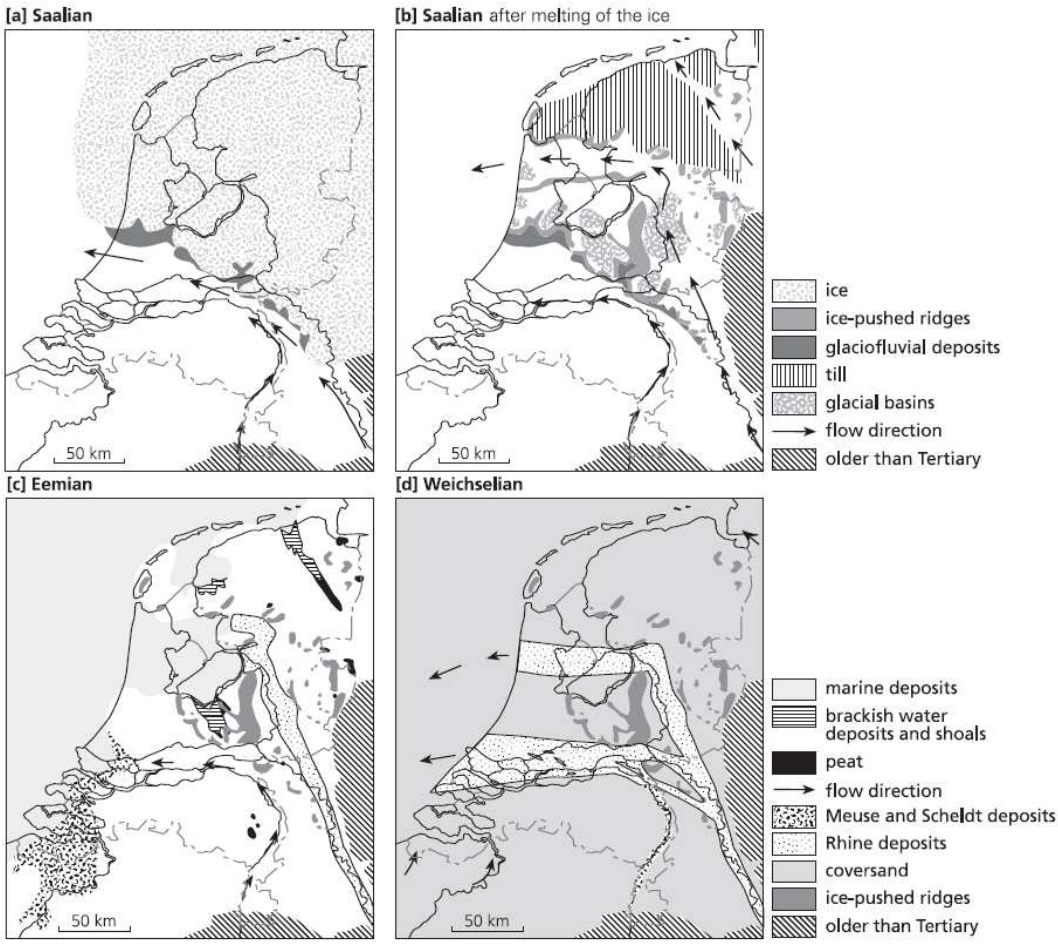

Figure 10 Palaeogeography of the Netherlands during the late Pleistocene (Berendsen and Stouthamer 2001, essentially after Zagwijn 1986).

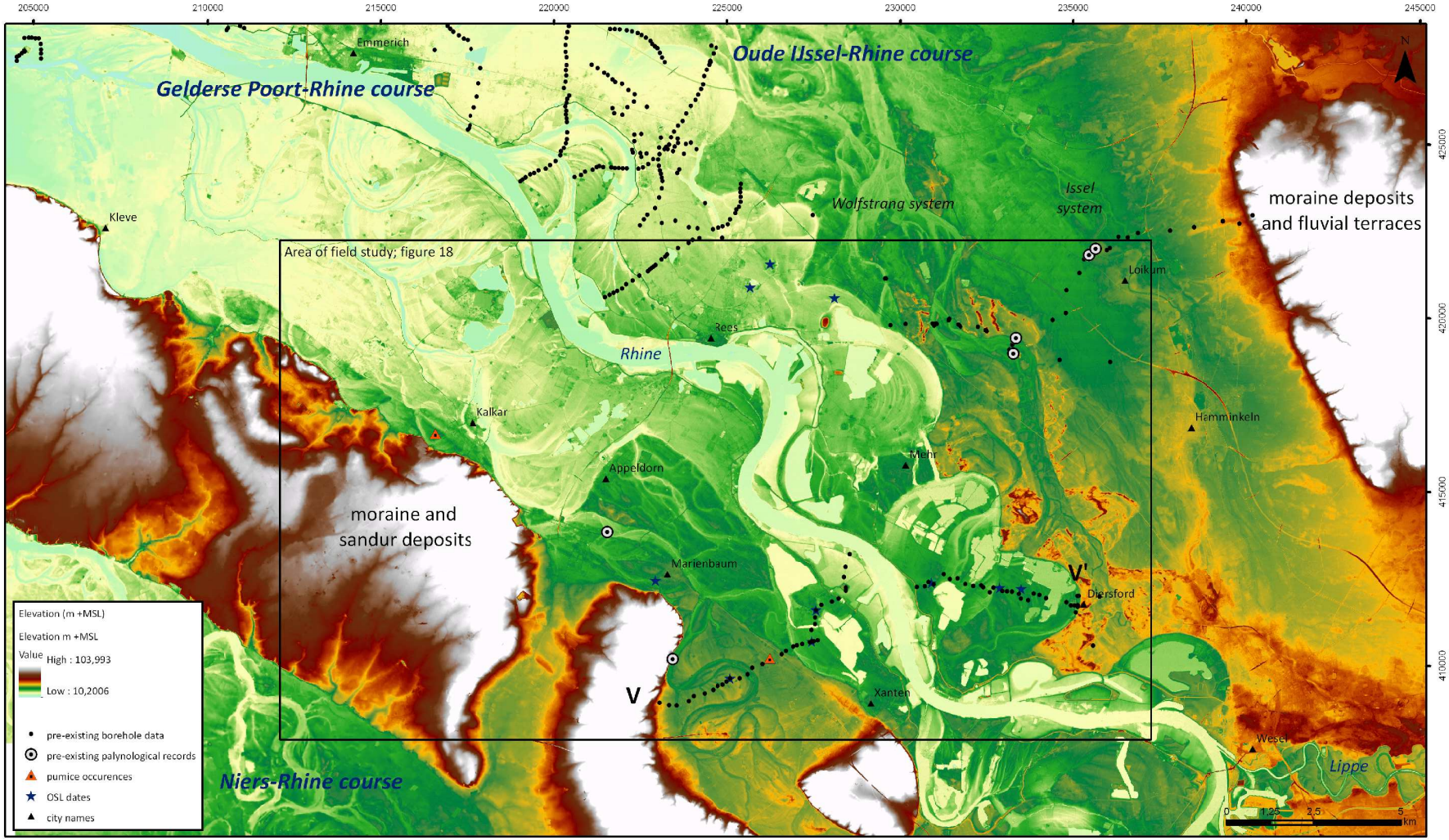

Figure 11: Digital elevation model of the area of interest, comprising the downstream part of the lower Rhine valley. The locations from which pre-existing evidence was retrieved are shown: Borehole data from Utrecht University (e.g. cross-section V-V' by Erkens et al. 2011), pumice founds, OSL-dates and palynological records.

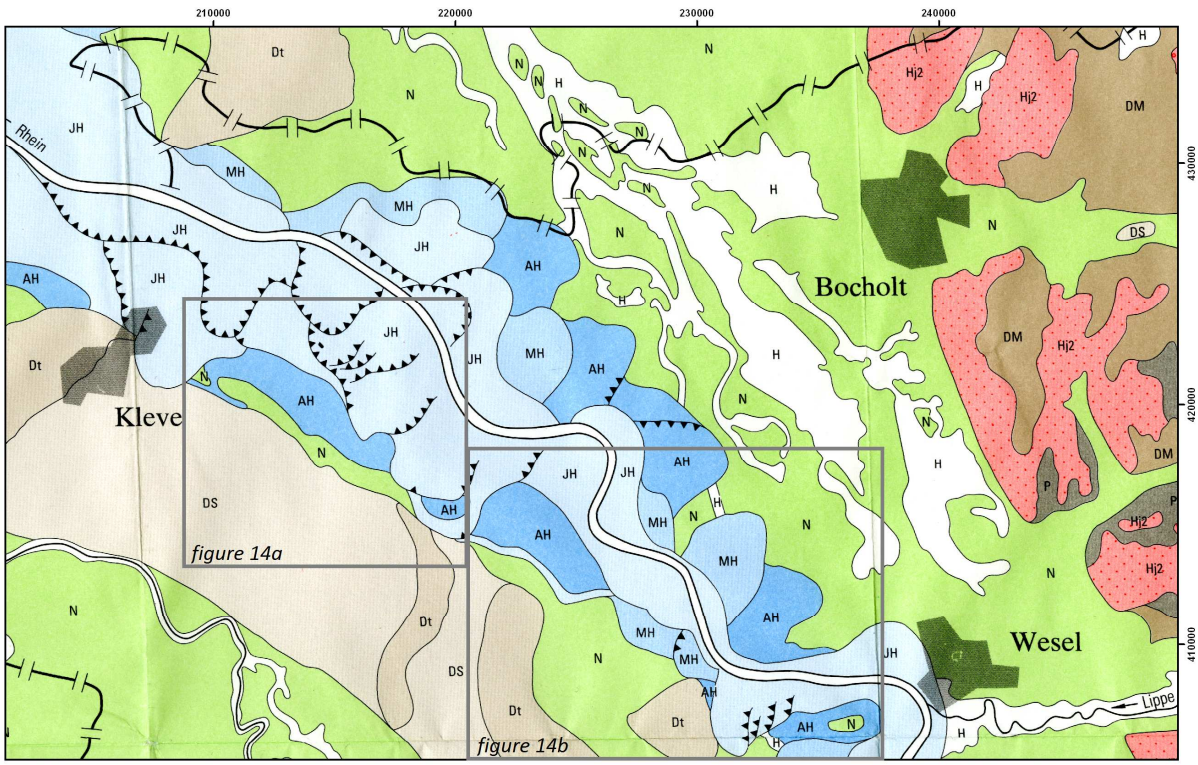

Figure 12 Fragment of the geological map of the lower Rhine valley comprising the area of study (Klostermann 1992). Codes of fluvial map-units: Hj2 = Hauptterrasse (Upper Terrace); N = Niederterrasse (Lower Terrace); AH = Altholozän (Early Holocene); MH = Mittelholozän (Middle Holocene); JH = Jungholozän (Late Holocene); H = Holozän, ungegliedert (Holocene, undifferentiated). Glacial and peri-glacial map-units are: DS = Sander (sandur deposits); Dt = Stauchmoräne (ice-pushed ridge deposits); DM = Grundmoräne (ground moraine deposits).

Eemian and early Weichselian

After the Saalian glaciations and deglaciation, remaining ice-marginal morphology continued to control the location of the river Rhine course and the width of its fluvial plains. During deglaciation, the river Rhine abandoned the Niers valley and regained its northern course through the Oude IJssel valley towards the deeply scoured glacial tongue basin which supports the IJssel valley today (Verbraeck 1984; Busschers et al. 2007; Cohen et al. 2002) (figure 10b). In the beginning of the Eemian interglacial, sea level rose up to levels a few meters above present-day sea level, the river Rhine kept its northern course through the Oude-IJssel valley (figure 10c). This course remained active during the Early Weichselian and Early Pleniglacial (Busschers et al. 2007; Cohen et al. 2009).

Middle Pleniglacial

During the early part of the Middle Pleniglacial period (~MIS3), an increase in sediment supply, presumably related to soil degradation and increased catchment erosion, resulted in pronounced aggradation in the Rhine-Meuse delta and in the apex region (Busschers et al. 2007). According to Kasse et al. (2005) this strong vertical build-up of the valley floor enabled the river to reoccupy the Niers-Rhine

valley. Furthermore, a new Rhine course came into being between the Montferland and Veluwe ice-pushed ridges: the so-called 'Rond-Montferland-Rhine' (figure 13; Van de Meene and Zagwijn 1978; Verbraeck 1984; Cohen et al. 2002). This westerly directed course developed due to the combination of aggradation and erosion of the glacial morphology. Between 60 and 40 ka BP, the Rond-Montferland-Rhine gradually took over the discharge of the IJssel-Rhine, causing the latter one to become abandoned before the onset of the Late Pleniglacial (Busschers et al. 2007; Cohen et al. 2009). Moreover, another western course developed crossing former ice-marginal morphology. This so-called Gelderse-Poort-Rhine developed at the expense of the Rond-Montferland-Rhine and the Niers-Rhine which simultaneously decreased drastically in size (figure 13; Van de Meene and Zagwijn 1978; Cohen et al. 2009; Verschuren 2007).

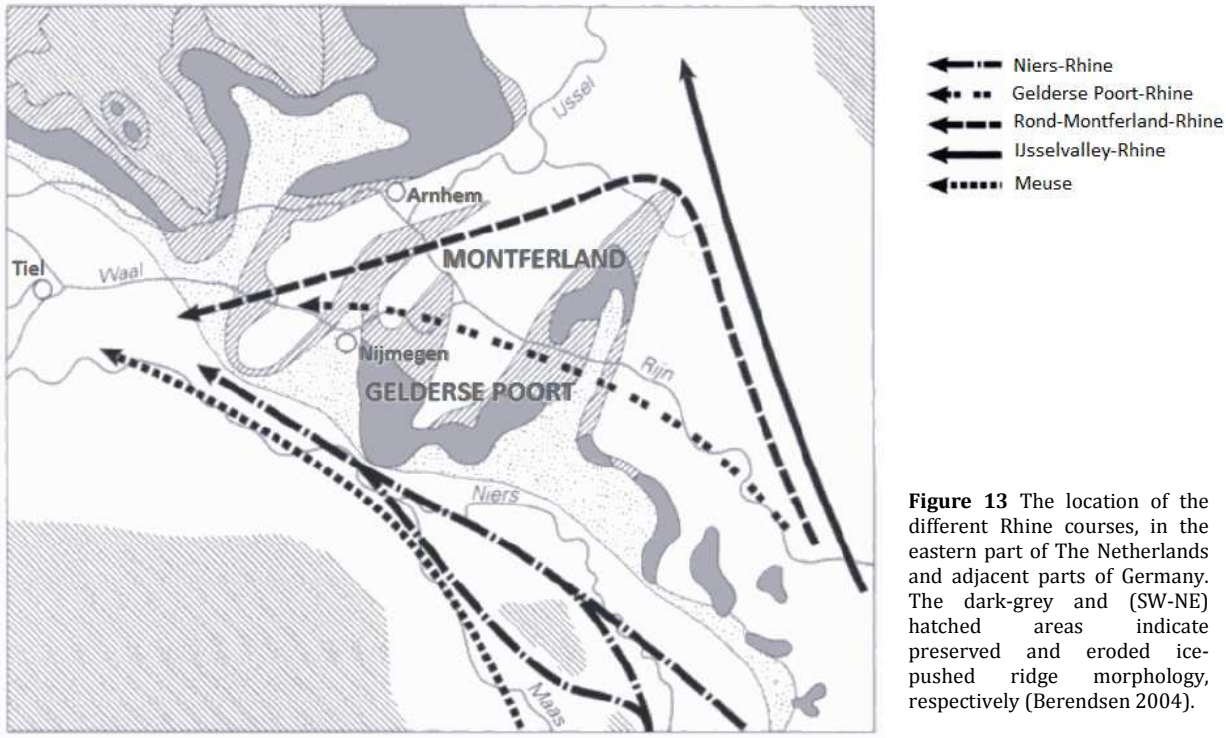

Figure 13 The location of the different Rhine courses, in the eastern part of The Netherlands and adjacent parts of Germany. The dark-grey and (SW-NE) hatched areas indicate preserved and eroded ice-pushed ridge morphology, respectively (Berendsen 2004).

Late Pleniglacial

During the Late Pleniglacial (30 000–14 700 BP; ~MIS2), the river Rhine evidently was in a braiding mode (morphology of abandoned surfaces preserved as river terraces), thought to be due to a climate-induced irregular discharge regime and large sediment supply. Aggradation led to the buildup of a broad braidplain which is commonly classified as the *Older Lower Terrace* (Dutch: Laagterras (Pons 1957) German: Ältere Niederterrasse, NT2 terrace (Erkens et al. 2011; Klostermann 1992)). The deposits belong to the Kreftenheye V formation, following the lithostratigraphic scheme of The Netherlands (De Mulder et al. 2003). The exact period of genesis of these deposits in the LRV is still uncertain but most accepted seems to be around the Last Glacial Maximum (LGM, ~21 cal ka BP, Erkens 2009; in agreement with OSL-dates collected downstream in the Netherlands, Busschers et al. 2007). The time-interval between ~17

and 13 ^{14}C ka BP is another period suggested for last aggradation, just before the onset of the Lateglacial (Schirmer 1995; Busschers et al. 2007). The fact that this floodplain level has been recognized in the Niers, Oude-IJssel and Gelderse Poort valleys, suggests that the river distributed its discharge over three river branches at that time (Kasse et al. 2005, Erkens 2009, Janssens 2010). In the Oude-IJssel and Niers-Rhine valleys, these braidplain deposits are flanked by older river terraces (Kasse et al. 2005; Van de Meene 1977; Verschuren 2007). The braidplain is broadest in the Gelderse Poort area, from which can be inferred that this valley supported the dominant Rhine branch during the Late Pleniglacial. However, the braidplain has been largely eroded by younger Rhine systems. A relative small remnant of this Pleniglacial braidplain has been recognized by Erkens (2009) to the southwest of the modern Rhine against the ice pushed ridge (south-westernmost tip of cross-section V-V' in appendix III). Northeast of the Holocene Rhine valley, this floodplain level is far better preserved, where it has been recognized by Erkens (2009; appendix III), Janssens (2010) and Greaves (2010). From palynological data from the upstream part of the Oude-IJssel valley, Janssens (2010) concluded that the braidplain got gradually abandoned during the Older Dryas (infilling of the *Eckerfeld* channel with loam, figure 17 in section 4.3) and became completely abandoned around the onset of the Allerød (transition from loamy into organic deposits).

Bølling – Allerød *interstadial complex*

Pronounced climatic amelioration marked the onset of the Lateglacial stage of the Weichselian, starting with the Bølling-Allerød interstadial complex (ca. 14.7 kyr BP). Permafrost gradually disappeared during the Lateglacial, first starting in the southern part of the Rhine catchment. The recovery of the vegetation led to more stable discharge regimes and lower sediment loads. The Rhine shrunk in active width; a transformation from a multichannel braided system into a single-channel meandering system started. In the sedimentary record of the LRV, Erkens et al. (2011) recognize the following sequence of responses which characterise this transition: First, river incision and abandonment of the NT2 terrace. Second, a change in fluvial style from braiding towards meandering. Third, contraction of flow into one single channel. The *Issel* and *Wolfstrang* meandering systems in the upstream part of the Oude IJssel valley (making part of the study area) presumably date back to this transition period and dissect the NT2 floodplain level (Janssens 2010). Other channel systems have been recognized by Schirmer (1990) and Schlimm (1984) in the upstream half of the LRV (Erkens et al. 2009). It concerns a number of secondary meandering channels which coexisted with the main Rhine channel, dissected the NT2 terrace and produced over bank deposits (Klostermann 1992) containing soils and peat of Allerød age (Jansen 2001; Erkens 2009). In the Niers-Rhine valley a 'transformation floodplain level' has been described by Kasse et al. (2005) which registered the same transformation of fluvial style. This floodplain level is characterized by relative straight channels (only partly meandering). From one of these channels the channel fill has been dated directly by radiocarbon dating and indirectly by pollen analysis to the Older Dryas-Allerød period (*Reinders* diagram; Janssens 2010; figure 17 in section 4.3), indicating that this channel system already became inactive during or shortly before the Older Dryas.

Younger Dryas stadial

During the Younger Dryas, the final pronounced cold phase of the Pleistocene, the river system within the central Rhine valley became braiding again (ca 12.8 kyr BP). Within the central Rhine branch, deposits of Bølling / Allerød age became largely eroded and the formation of a wide braidplain took place; the *Younger Lower Terrace* (Dutch: Terras X (Pons 1957) German: Jüngere Niederterasse, NT3 terrace (Erkens et al. 2011; Klostermann 1992)). The combined distribution of the NT2 and NT3 terraces is shown on the geological map constructed by Klostermann 1992 in figure 12 (map unit N). The deposits of the NT3 belong to the Kreftenheye VI formation, following the lithostratigraphy of the Rhine-Meuse delta (De Mulder et al. 2003). Climate-related drivers of the transition towards a braiding mode and the development of a wide braidplain were higher peak discharges, a reduced vegetation cover and an increased sediment supply due to more hill slope erosion. However, another factor might have played a role in the LRV: A major flood pulse after a catastrophic breaching of a pumice dam upstream of Bonn (Germany) created by the Laacher See eruption (Friedrich et al. 1999: 11063 ± 12 ^{14}C BP or 13-13.2 cal ka BP) (Litt et al. 2003). The flood might have destroyed older fluvial morphology, making the valley more susceptible for a braiding fluvial system when climate cooling started (Erkens 2009). The NT3 can be distinguished from the NT2 terrace because the NT3 deposits contain abundant river transported pumice. The fluvial deposits between 3 and 4 km in cross-section V-V' constructed by Erkens et al. (2011; appendix III) are classified as NT3 terrace because a small amount of pumice was potentially observed near the top of these in-channel deposits (G. Erkens, personal communication). A similar conclusion was drawn by Siebertz (1987) for the terrace level to the west of Kalkar (for locations, see figure 11).

Pumice founds in both the Niers-Rhine valley and Oude IJssel valley make evident that both branches must have still transported a small amount of the total Rhine discharge after the onset of the Younger Dryas (Kasse et al. 2005; Verschuren 2007). In contrast to the central Rhine branch through the Gelderse Poort, the threshold towards a braiding fluvial style was not crossed in the Niers-Rhine valley and the *Issel* system of the Oude-IJssel valley. On the contrary, the threshold was crossed in the *Wolfstrang* system of the Oude IJssel valley (Janssens 2010; Verschuren 2007). This system contains a floodplain level characterised by a braiding network of small channels. The infill of one of these channels, the *Schlederhorst* channel, was palynologically analysed by Janssens (2010) from which could be concluded that it became at least abandoned around the Younger Dryas-Holocene transition (figure 17 in section 4.3). The limited Rhine discharge in the *Wolfstrang* system inhibited the formation of a broad braidplain here in contrast to the Gelderse Poort valley.

Holocene

Both the Niers-Rhine and Rond-Montferland-Rhine became abandoned just after the onset of the Holocene, in favor of the central branch through the Gelderse Poort area (Van de Meene and Zagwijn 1979; Verbraeck 1984; Kasse et al. 2005). The latter one has remained the only active river branch from

that time onwards. Climatic amelioration around the onset of the Holocene led indirectly to incision, abandonment of the NT3 terrace and contraction of flow into a multi-channel system in the central Gelderse Poort branch.

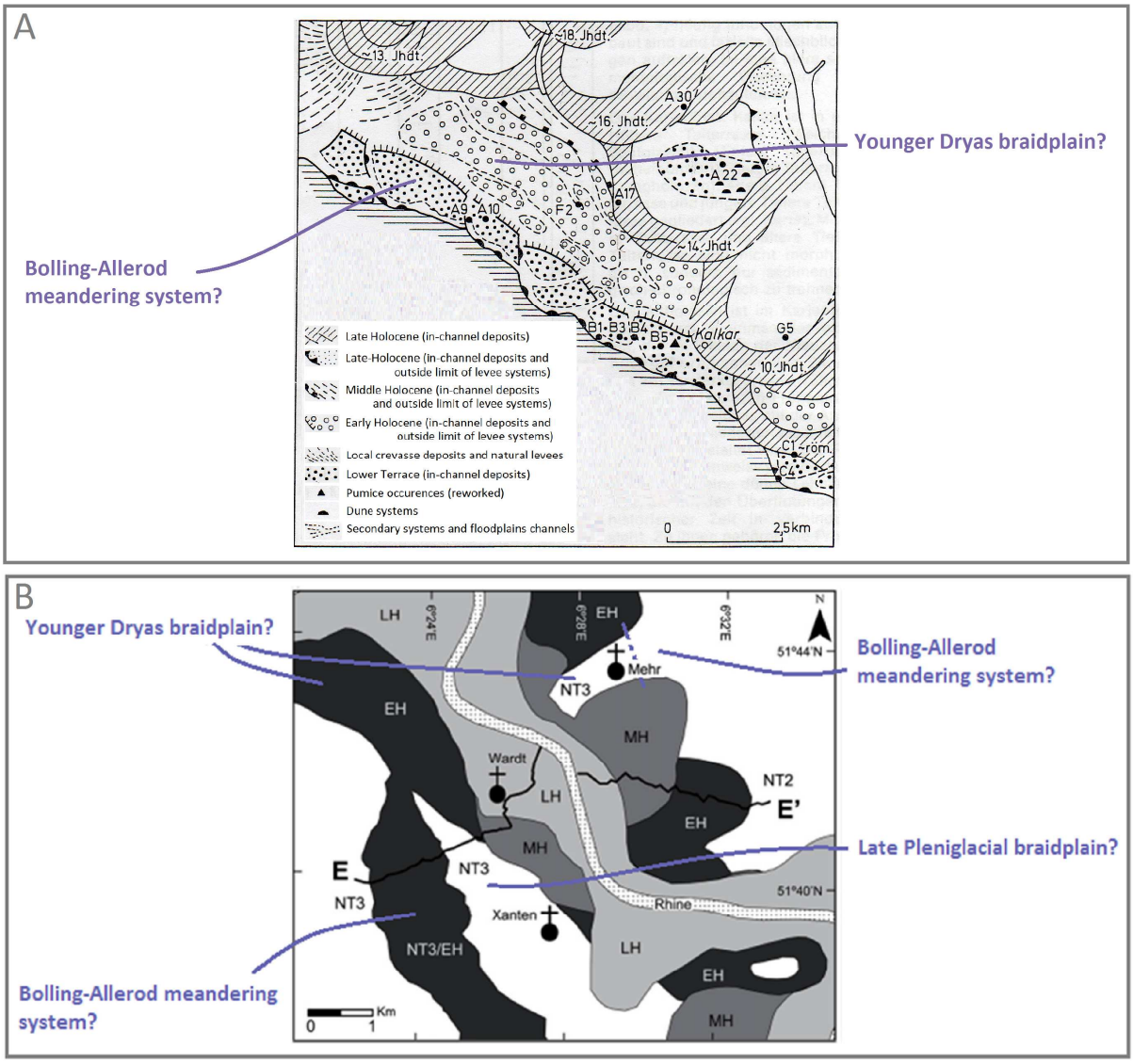

Figure 14 A) Geological classification of distinct geomorphological units by Siebertz (1987). B) Chronology of Weichselian terraces (NT2,3) and Holocene palaeo-meander groups according to Erkens et al. (2011, essentially after Klostermann 1992). Alternative ages hypothesised by the present author are additionally showed outside the originical figures (for more information, see section 3.3). For the location of both areas, see figure 12.

In the course of the eary Holocene, the Rhine developed a meandering course, as evidenced by OSL dating results of well-developed pointbar system in the upstream part of the area of study (Erkens et al. 2011). During the remaining part of the Holocene, the Rhine reworked a large part of the Weichselian braidplain terraces by lateral migration. The DEM in figure 11 shows a large number of abandoned meander bends of varying age which have been clustered by Klostermann (1992) and Erkens et al. (2011) into three groups which differ in their main period of activity and moment of abandonment, namely the early (~11.7-6 cal ka), middle (~6-2 cal ka) and late Holocene (~2 cal ka–present) (see figures 12 and 14b). For clustering,

they used cross-cut relationships, elevation, soil maturity (i.e. decalcification depth), OSL-dates and archeological and historical evidence. During the main part of the Holocene, the apex of the Rhine delta was located downstream of the research area. However, as a consequence of back-filling of the trunk valley (continuing long after sea level rise ceased, chapter 2) the apex moved upstream, reaching the study area approximately 3000-2000 years ago (Favier 2001). According to Erkens et al. (2011), human impact of prehistoric deforestation on sediment delivery is predominantly responsible for late Holocene back-filling of the Rhine valley.

3.3 Research objectives concerning the geology of the study area

From preliminary research of the DEM, a number of questions have arisen concerning the age and genesis of distinct morphological units in the way these are defined by pre-existing studies (Klostermann 1992; Erkens et al. 2011; Jansen 2001; Siebertz 1987). These questions have resulted in research objectives, which are listed in this section.

- The strong meandering channel system to the west of Xanten has been classified as a late Younger Dryas or earliest Holocene system by Erkens et al. (2011) on the basis of an OSL date of the pointbar system (10,910 ± 570 yr) and an early Holocene pollen signal contained by the lowermost channel fill (notation: 'NT3/EH'; compare figure 11 and 14b). On the DEM a younger fluvial system is visible, cross-cutting the meandering system (notation: 'EH' in figure 14b). This fluvial unit shows typical braidplain features on the DEM, namely; 1) a straight cross-cut boundary with the older meandering system, 2) a network of relative small and parallel channels (braiding pattern, except from local reworking by younger meandering systems), and 3) a relative flat surface topography without any pointbar structures. A continuation of these morphological units is visible directly to the northwest of Kalkar on the DEM. According to Siebertz (1987) and Erkens et al. (2011), this terrace level dates to the early Holocene (compare figure 11, 14a and 14b). The question that arises from the morphological observations is: If the meandering unit would have been active across the Younger Dryas – Holocene transition, what would be the age of the younger braidplain? One of the main objectives of the present study is to figure out the ages of both fluvial systems. It is hypothesised that the meandering level and braidplain date back to the Bølling-Allerød and Younger Dryas time-intervals, respectively, since a similar sequence of fluvial style transitions is recorded in other parts of the Rhine valley up- and downstream (section 3.2).

- On the other side of modern Rhine, a relative small terrace remnant, supporting the village 'Mehr' today, shows similar braidplain characteristics as the potential Younger Dryas braidplain just described. According to Erkens et al. (2011), this level was indeed formed during the Younger Dryas (NT3 terrace in figure 14). One research objective is to investigate whether these preserved braidplain fragments along both sides of the Rhine correlate to each other.

- The meandering system to the west of Xanten cross-cuts one or two older fluvial terraces which have been classified as NT3 (or NT2) terraces by Erkens et al. (2011) (figure 14 and appendix III). Since the meandering system is hypothesised to be of Bølling-Allerød age, these braidplains are expected to have been formed during the Late Pleniglacial. Solving the age-problem of the meandering system, will additionally provide insight in the age of the older braidplain.

- Two pumice discoveries (section 3.2; figure 11, 14a) reported by Siebertz (1987) and Erkens et al. (2011), are not in line with the hypothesised alternative classification described so far, since pumice is thought to be only admixed to Younger Dryas or (early) Holocene fluvial deposits (De Mulder et al. 2003). According to the alternative classification, these pumice granules were discovered on top of Late Pleniglacial and Bølling-Allerød terraces, in other words, which are thought to have been formed before (instead of after) the Laacher See volcanic eruption. The present study aims to unravel the discrepancy between the location of the pumice granules and the hypothesised ages of the distinct fluvial units.

- In line with the alternative classification, an early Holocene pollen signal contained by the channel fill of the meandering system to the west of Xanten, points towards a considerable time-lag between channel abandonment and the onset of channel infilling. This palynological record was mentioned by Erkens et al. (2011) and makes part of the present study (*Heesenhof* record; chapter 7). One research objective is to investigate the underlying mechanism responsible for this potentially present time-lag.

4 Weichselian – early Holocene vegetation development in north-western Europe

4.1 General vegetation succession during glacial-to-interglacial transitions

The north-west European general vegetation history since the end of the last glacial has been well studied and frequently described. The reconstructed vegetation history has been based on observed patterns in both time and space in a vast collection of European biostratigraphic records (e.g. pollen or macrofossil diagrams). By considering the ecology of the different plant taxa involved and comparing biostratigraphic records with independent climatic records (e.g. stable-isotope ratios), the observed vegetation developments could be interpreted as responses to (climate-induced) fluctuations in the physical environment. Iversen (1958) noticed a general trend of vegetation development during one quaternary glacial-interglacial cycle and proposed a simple model called the *interglacial cycle* (Birks 1986). The model describes different environmental and biological processes which are responsible for distinct phases of vegetation development. A brief description of the interglacial cycle is given below after Birks (1986).

The interglacial cycle model has been originally proposed by Iversen (1958), and was extended by Birks and Birks (2004). As shown in figure 15 it exists of a succession of phases, starting with a glacial period, the so-called *cryocratic phase*. Characteristic for this phase is a cold, dry continental climate and immature base-rich soils only enabling only pioneer, arctic-alpine, steppe species to survive. At the onset of an interglacial, temperature starts to rise and the *protocratic phase* begins. Basiphilous shade-intolerant herbs, shrubs and trees are able to settle on the unleached, fertile soils of low humus content and expand to form species-rich grassland, scrub and open woodlands. During this phase, competition is generally low because of plenty of sunlight, nutrients and space. Lakes generally support pioneer aquatic taxa. As soil formation continues, fertile brown-earth soils start to develop, initiating the *mesocratic phase*. During this phase climate is similar to or warmer than the climate of the protocratic phase. Temperate deciduous forests develop and shade-intolerant species become rare or totally disappear as a result of competition and habitat loss. Within the mesocratic phase, distinct stages of deciduous forest development can be recognized, reflecting contemporaneous climate change or delayed immigration of distinct tree species following the rapid climate amelioration at the beginning of the interglacial. Lakes are fertile, frequently supporting diverse biota. According to the original model of Iversen (1958), the mesocratic phase is followed by one retrogressive phase, the so-called *telocratic phase*, which is characterised by a declining trend of the temperature, heading on towards the next glacial period. Forests are opening up and conifer tree species become more dominant, together with heather and bogs growing on infertile, humus-rich podsols and peats. Locally, nutrient-demanding plant species typical for the mesocratic phase become extinct and light-demanding species become favored as a result of decreasing shade. Due to a high input of dissolved organic carbon into lake systems, leading to a so-called dystrophic state, the lake productivity

declines. In addition, Andersen (1964, from Birks 1986) introduced a fifth phase, the so-called *oligocratic phase*, which is positioned in between the mesocratic and telocratic phases and is also present in the version made by Birks and Birks (2004). It is a phase of soil retrogression, prior to climate cooling during the telocratic phase. At the end of the telocratic phase, forests disappear and the acid soils become destroyed as a result of frost activity and solifluction and the cycle begins again with a new cryocratic phase.

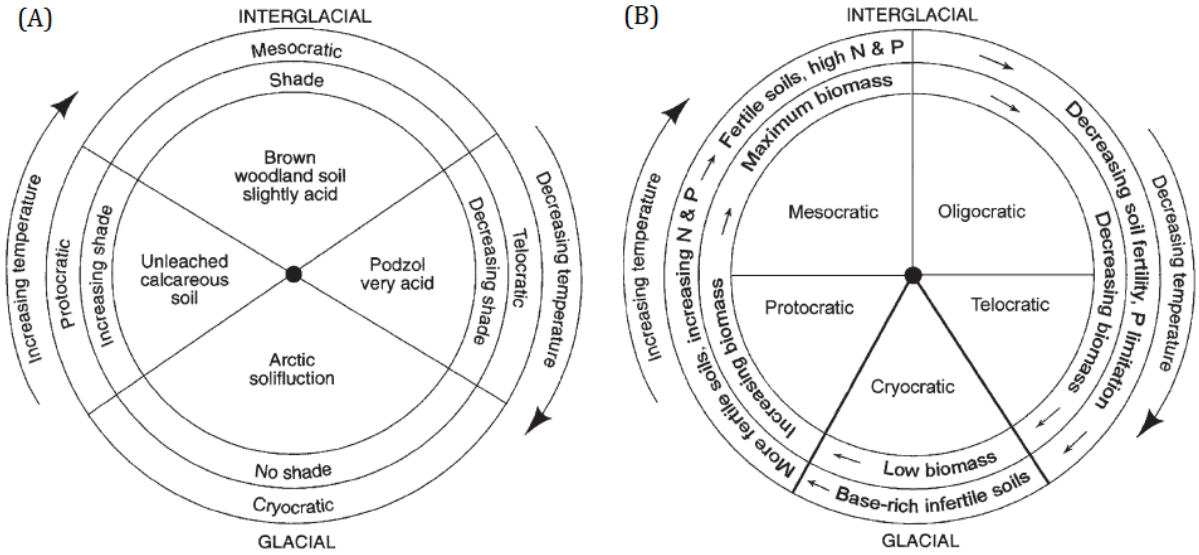

Figure 15 The interglacial cycle of changes in vegetation type, biomass productivity and soil conditions (Birks 1986, originally after Iversen 1958).

4.2 Lateglacial regional vegetation development in the lowland area of north-western Europe

In this section, the general Lateglacial vegetation history of the lowland area in north-western Europe is outlined (~The Netherlands, northern Belgium, and western Germany). It is a compilation based on the just described interglacial cycle (Birks 1986) and the well-documented and well-dated Lateglacial vegetation succession of The Netherlands (Hoek 1997a,b). The latter is based on 260 palynological records from lake and peat sediments from primarily The Netherlands, and an additional small number of cores from northern Belgium and western Germany, and includes a regional pollen zonations scheme and pollen distribution maps (figure 16a). Radiocarbon dated pollen zones have been calibrated and correlated to the GRIP oxygen isotope record (Hoek 2001; figure 16b). The timing of the distinct Lateglacial vegetation development as described in this section is based on this correlated and calibrated time-scale.

The last stage of the Weichselian Pleniglacial is characterized by a gradual temperature rise and forms the final part of the cryocratic phase. The vegetation cover, however, was still very open at that time. A more rapid and more intense temperature rise occurred around 15.5 cal kyr BP, accompanied by the beginning of organic accumulation and a rise in wormwood (*Artemisia*) (onset zone 1a; figure 16). The temperature

rise initiated northward migration of plants and animals from refugia in southern Europe. As a consequence of the distance between refugia and the areas suitable for colonization and limited migration speed, vegetational succession was generally lagging behind the climate signal. According to Hoek (2001), however, plants which were already present, did respond directly by increased flowering. Opportunistic heliophilous plant species settled first on the raw, base-rich mineral soils, for example Stone-breakers (Saxifragaceae), Alpine Sorrel (*Oxyria digyna*), Sheep's Sorrel (*Rumex acetosella*), Greater Plantain (*Plantago major*), Rock-rose (*Helianthemum*) and species belonging to the Chenopodiaceae. Also Cyperaceae and Gramineae expanded and several taxa with nitrogen-fixing nodules, like sea buckthorn (*Hippophaë rhamnoides*), were present. As soon as soils became more stabilized and humus and nitrogen started to accumulate, Crowberry (*Empetrum nigrum*), Juniper (*Juniperus communis*), Dwarf Birch (*Betula nana*) and willow species (*Salix* sp.) became increasingly important.

The onset of the Bølling interstadial, circa 14.7 kyr BP (GRIP record, Björck et al. 1998; 14.8 cal kyr according to Dutch zonation scheme; figure 16a), is marked by a rapid and intense temperature rise and marks the onset of the Weichselian Lateglacial. Moreover, it more or less coincides with the transition from the cryocratic phase towards the protocratic phase according to Birks (1986). Climate amelioration caused especially birch (*Betula*) to expand, but a lot of other species took their advantage (zone 1b). The assemblages of pioneer thermophilous aquatic plants which colonized the lakes at that time suggest a warmer climate than is indicated by the only sparsely forested land. This is presumably due to disequilibrium between tree distributions and climatic conditions (Birks 1986).

Around circa 14.1 cal kyr BP, the gradual temperature rise was interrupted by a period of relative cold conditions of about two centuries: the Older Dryas stadial (zone 1c; figure 16a). In western Norway and on the British Isles, however, there is no unambiguous evidence for a phase correlative with the Older Dryas, resulting in one long uninterrupted but variable Bølling-Allerød (Windermere) interstadial. In the remaining part of western Europe outside western Norway and the British Isles, the Older Dryas stadial is commonly characterized by an expansion of herbs typical for disturbed soils, increased erosion and inwashing of mineral material (Birks 1986). The opening of the landscape went together with a reduction in the extent of birch. In Dutch pollen records, higher percentages of willow pollen are observed, presumably attributable to willow shrubs.

The Older Dryas is followed by the relative warm Allerød interstadial, which started in the Netherlands around 14 cal kyr BP. Here, it has been subdivided in a phase of birch expansion (subzone 2a; figure 16a) and a phase with declining birch pollen percentages and pine becoming more dominant (subzone 2b, from 13.2 cal kyr BP onwards). The general trend is stabilization of the soil by an almost complete vegetation cover, and processes such as frost action, erosion, solifluction and aeolian activity becoming negligible. Soil maturation and humus accumulation began and organic sediments started to become deposited in lakes (Birks 1986). Locally, shade-intolerant herbs became extinct as a result of competition from taller, long-living woody plants. Similar to the situation of the Bølling, disequilibrium existed between the terrestrial vegetation and climate is suggested by the presence of thermophilous marsh and aquatic plant

species as great reed-mace (*Typha latifolia*) and common club-rush (*Schoenoplectus lacustris*). In north-western Europe, this Allerød period is additionally characterized by the introduction of aspen (*Populus tremula*), European rowan (*Sorbus aucuparia*) and bird cherry (*Prunus padus*) in the birch woodlands (Birks 1986).

In the Dutch pollen records pronounced opening of the vegetation starts around 13.0 cal kyr BP, as a consequence of a major and abrupt climatic deterioration at the onset of the Younger Dryas stadial. The well-develop Allerød birch-pine woodlands opened up or even disappeared, making room for dwarf-shrub heaths, open grass heaths and sedge-dominated communities (Birks 1986). Freeze-thaw and solifluction processes and aeolian activity made the landscape unstable again and locally destroyed or buried the humus soils that had been developed during the Allerød. Opportunistic plants like wormwood (*Artemisia*), stone-breakers (Saxifragaceae), greater plantain (*Plantago major*), rock-rose (*Helianthemum*) and species belonging to the Chenopodiaceae flourished on the open, mineral soils. In the Netherlands, the second half of the Younger Dryas is characterized by an increase in coverage by Ericales, especially crowberry (*Empetrum nigrum*).

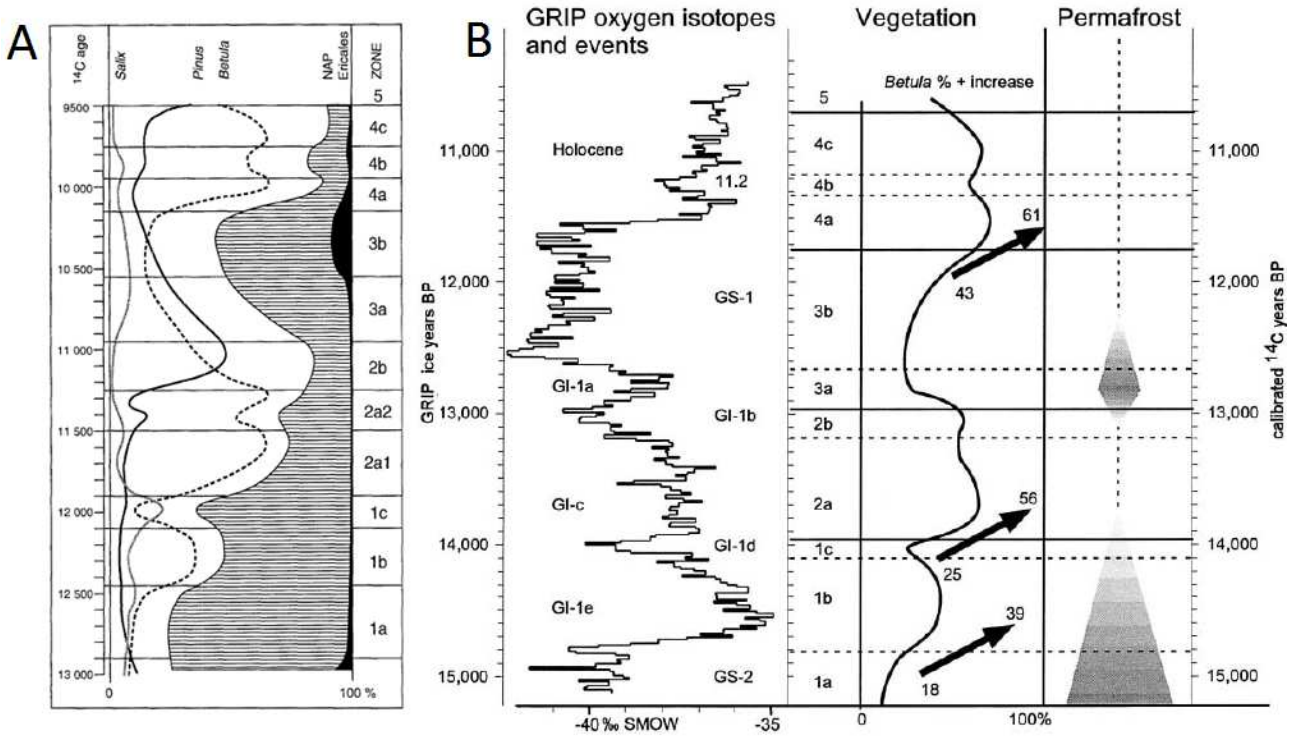

Figure 16 A) Generalised Lateglacial and Early Holocene pollen diagram for the Netherlands (Hoek 1997a,b). Note: This figure shows a radiocarbon time-scale. B) Correlation between the *Betula* pollen signal from the Netherlands (figure A) and permafrost occurrence with oxygen isotope events of GRIP on the INTIMATE calendar year timescale after Björck et al. (1998) (Hoek 2001).

The Younger Dryas ends with an expansion of birch forests again, marking the onset of the Holocene interglacial (~11.7 cal kyr BP). The climate of the Preboreal was still variable, as is reflected by an intitial phase of relative high abundances of juniper besides birch (*Friesland phase*, as defined by Behre 1966 and van Geel et al. 1981) followed by a phase with a somewhat more open and grassy vegetation (*Rammelbeek phase*, as defined by Hammen 1971; Van Geel et al. 1981; Hoek and Bohncke 2002). After these climatic

oscillations, birch forest expanded again and aspen (*Populus tremula*) became an important constituent of the vegetation (Hoek 1997a,b). Other constituents of the west European birch woodlands were probably willow, European rowan (*Sorbus aucuparia*), bird cherry (*Prunus padus*) and Guelder Rose (*Viburnum opulus*) (Birks 1986).

In the Netherlands, the beginning of the Boreal (~10.2 cal kyr BP) is marked by the immigration of the first thermophilous tree common hazel *(Corylus avellana*) and an expansion of pine (Hoek 1997a). The immigration, expansion and dominance of temperate deciduous tree species like lime (*Tilia*), oak (*Quercus*) and elm (*Ulmus*) and the development of fertile brown-earth forest soils marks the onset of the mesocratic phase of the interglacial cycle model (Birks 1986).

4.3 Late-glacial-early Holocene palynological records from the distinct branches of the lower Rhine

A recent MSc study by Janssens (2010) concerning fluvial developments of the Oude IJssel Rhine during the Lateglacial, comprised a palynological study of abandoned channel fills which are located in the study area of present study. The result was a sequence of four pollen diagrams which partly overlap in time spanning almost the whole Lateglacial period since the Older Dryas (figure 17). Unfortunately, because of a relative low sample resolution, it shows a relative flat pollen signal from which only major trends can be derived. Other Lateglacial palynological data are available from the most southerly located Rhine branch; the Niers-Rhine valley. The article from Kasse et al. (2005) shows a sequence of seven pollen diagrams overlapping in time and spanning the Lateglacial since the Older Dryas. Further, the pollen signal contained by the *Vethuizen* core (Verschuren 2007) from the Oude IJssel valley suggests that infilling started during the early Preboreal, probably around the end of the Rammelbeek phase (section 4.2).

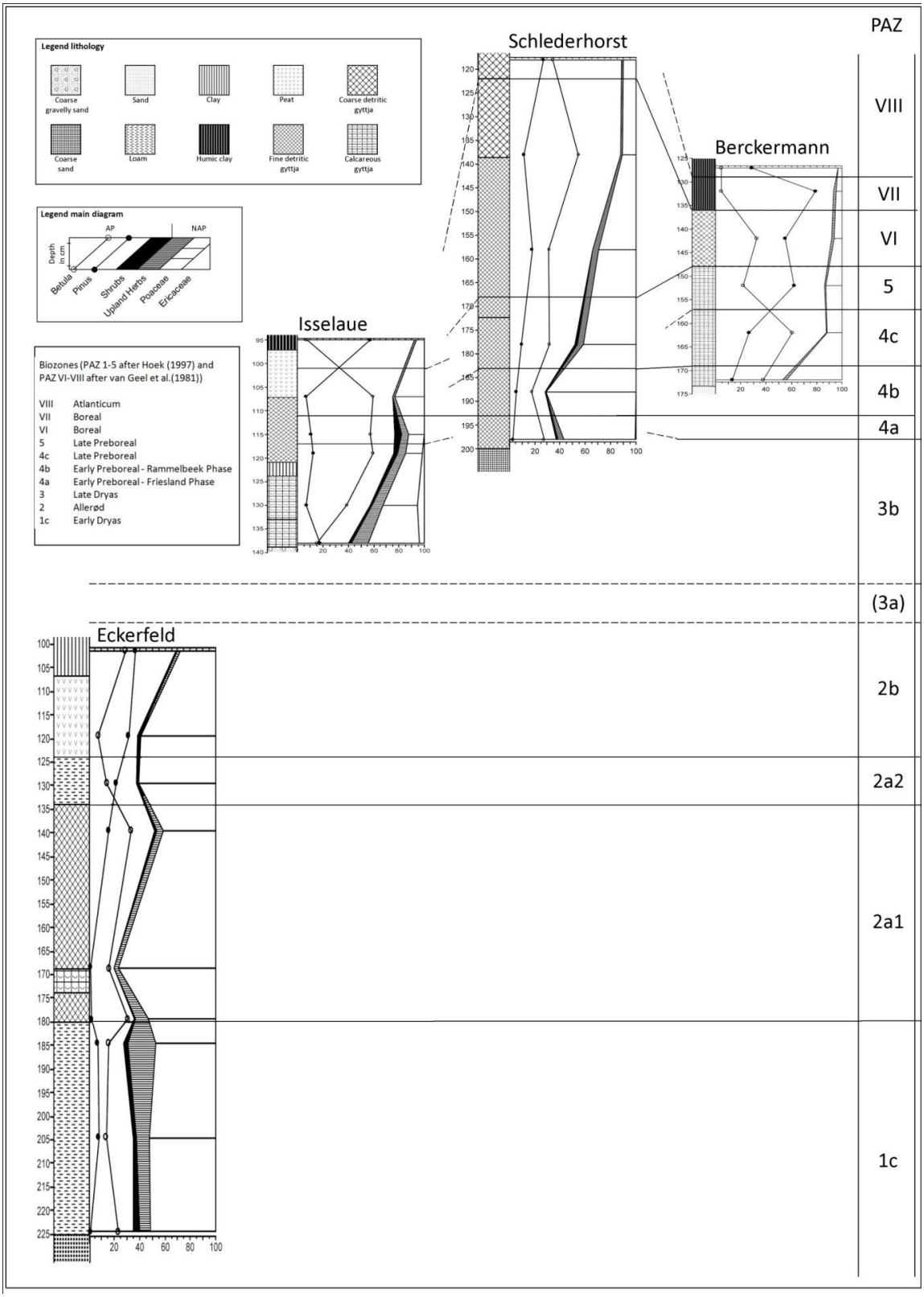

Figure 17 Biostratigraphic correlation diagram of four pollen records from the area of interest, as constructed by Janssens (2010). For core-locations of the distinct pollen records, see figure 18.

5 Research methods

5.1 Research strategy

In order to reconstruct the palaeogeography of the study area, a first step was to determine the (relative) age of the different floodplain levels that could be distinguished on the digital elevation model (DEM). The DEM is based on laser-measured altimetry data (grid size 1x1m; Landesvermessungsamt Nordrhein-Westfalen, Germany). Relative ages were primarily determined on the basis of cross-cutting relationships. In order to investigate the type of (mainly fluvial, but also aeolian and to a much lesser extent hill-slope and biological) underlying landscape-forming processes, the sedimentary build-up of the lower Rhine valley was studied by means of coring transects (second step). A more exact age of the distinct geomorphological units was obtained by correlating sedimentary and geomorphological units with the well-documented and well-dated fluvial history of the Rhine-Meuse delta. Additionally, pre-existing and newly collected biostratigraphically correlated channel fills, a low number of available OSL-dates from the LRV and pumice founds were used for age-control. The final step was to compare the constructed chronostratigraphy and fluvial history with climatic records in order to investigate the relationship between climate change (ice-mass fluctuations) and fluvial response and the related time-lags. Palynological results were additionally used for reconstructing regional and local vegetation successions. Moreover, local vegetation types were used as proxy for groundwater levels and river or flood activity during the time-interval of channel infilling.

5.2 Cross-sections

On the basis of the DEM, two large-scale transects were established across the lower Rhine valley, perpendicular to the main flow direction (cross-section I (A-A'-A") and II (B-B'-B"), figure 18). These transects intended to incorporate as many different geomorphological units as possible, especially those without any available data. It was aimed to investigate the sedimentary build-up of the valley, the lithological characteristics and dimensions of distinct units and the stratigraphic relationships between them. One additional detailed transect was established across a relative small secondary channel system (*Marienbaum* system; cross-section III (C-C'), figure 18).

The field study was conducted during a four-week period in September 2010, during which also the Holocene fluvial (flooding) history of the lower Rhine was investigated (Van Munster, MSc thesis in prep.; De Molenaar, MSc thesis in prep.). Exact borehole locations were decided on in the field. In total, 132 borings were performed on the Weichselian and early Holocene fluvial terraces in the LRV. Average borehole spacing and depth were ~300 m and 3-4 m, respectively. Augering was carried out with hand-operated devices (Edelman auger, Dutch gouge and Van der Staay suction corer).

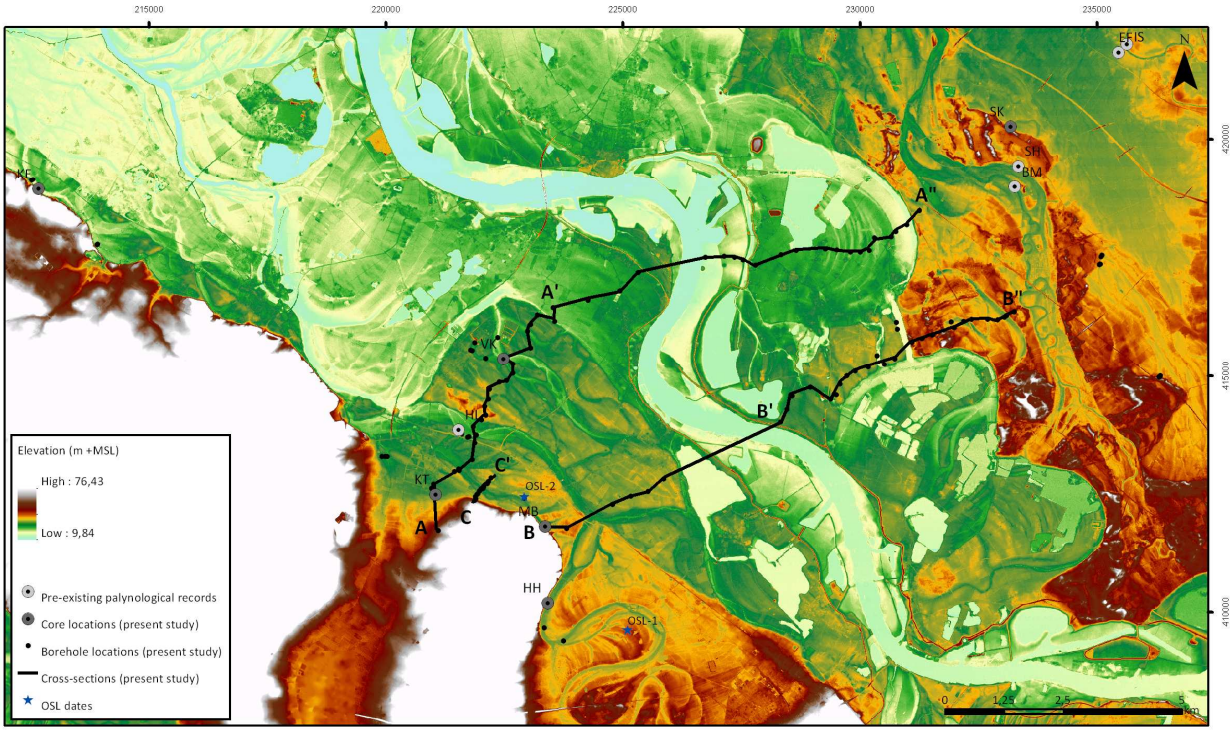

Figure 18 Digital elevation model showing the locations of the newly derived borehole data (black dots) and cores (grey circles; KT = *Kehrum Torfkuhle*, SK = *Sanders Kath*, VK = *Vosse Kuhle*, MB = *Marienbaum*, KE = *Klein Entenhorst*, HH = *Heesenhof*), and the position of cross-section I (A-A'-A"), II (B-B'-B") and III (C-C'). In addition, two relevant OSL-dating sites (blue stars) and core locations of already available pollen records or newly derived pollen records from other MSc studies are shown (white circles; Janssens 2010; Van Munster, in prep.; EF = *Eckerfeld*, IS = *Isselaue*, SH = *Schlederhorst*, BM = *Berckermann*).

Table 1 Available OSL dates from the area of study, most relevant for present study. Additional OSL-dates of Holocene pointbar systems are provided in appendix III.

Sample site in figure 18	Depth (cm)	Sedimentary unit	OSL-date (ka)
OSL-1	110	dune	11.72 ± 0.52
Erkens et al. (2011)	310	point bar	10.91 ± 0.57
OSL-2	50	dune	10
Tebbens (unpublished)	100	dune	11
	180	braidplain deposits	15

The borings were logged in the field at 10 cm intervals, in line with Berendsen and Stouthamer (2001). Coordinates and surface elevations of the borehole locations were respectively determined using a handheld GPS-device and on the basis of the DEM (sub-decimetre vertical resolution, accuracy <30 cm). Interpretation of the distinct sedimentary facies directly started in the field, distinguishing units of fluvial (in-channel, overbank and abandoned channel fill deposits, according to Berendsen and Stouthamer 2001) and aeolian origin or other locally produced deposits. For all transects, lithostratigraphic cross-sections were constructed which, together with the DEM based morphological unit mapping, formed the sediment architectural framework for the rest of this study.

5.3 Collecting sediment cores for lab analysis

Cores were taken at five selected sites for detailed sediment analysis in the laboratory and for pollen analysis using a "Bohncke-modified Livingstone Piston corer" (diameter: 6 cm). The selection of core locations was mainly based on the following criteria: 1) The abandoned channel systems had to make part of floodplain levels II, III or IV, because they are assumed to have become abandoned at the onset, in the course or at the end of the Lateglacial, after which infilling of the channel system could start. 2) The channel-infill must consist of totally organic or at least humic-rich sediment, since clastic fluvial sediment contains pollen from upstream. 3) The coring site has to be characterised by high groundwater levels today because it enhances the possibility that organic sediments have been preserved during the Holocene. Therefore, potential suitable locations were present near the foot of the ice-pushed ridges because of seepage-controlled groundwater-levels and underneath dune deposits which might create local slightly elevated water-levels.

Four cores were retrieved from residual channels fills and one from a peat layer at great depth (see figure 18). Location suitability was determined by prospective examination of the dimensions and sedimentary build-up of each palaeochannel, through making detailed coring transects. These prospective activities also identified where organic or humic sediments were thickest and sampling location was ideal and where breaks of 1 m core segments should be planned. The total length of collected core is circa 10 m of which circa 6.5 m has been analysed in the laboratory.

5.4 Loss on ignition

The weight percentage of organic matter in the core sediments was estimated using 'loss on ignition' (LOI). The standard protocol by Heiri et al. (2001) was followed. The cores were cut into two halves, described and photographed and samples with a constant volume of circa 0.5 cc were taken each centimeter as a continuous sequence. Thereby, no samples were taken across visible lithostratigraphic boundaries for retrieving a LOI signal which shows fluctuations as sharp as possible. The third step was to oven-dry the samples for circa 12 hours at 105 °C and to weigh them, giving values for DW_{105} (abbreviation for: dry weight 105 °C (in gram)). Finally, the organic matter within the samples was combusted in the oven for 4 hours at 550 °C. The weight of the samples after combustion is given as DW_{550} (in gram). The LOI values were calculated as a percentage using the following standard equation: LOI = $((DW_{105} - DW_{550}) / DW_{105}) * 100$ %.

5.5 Preparation of pollen samples

Pollen subsamples were extracted from the sediment with a small sampler of known volume (0.3 cc). For sediment of apparent low organic content, a double amount of sample was taken (0.6 cc) to ensure enough

pollen to be extracted. To each sample, *Lycopodium* spores were added (~2136 spores per sample) using a suspension based on *Lycopodium* tablets (~10679 spores per tablet). The pollen samples were treated as follows (protocol based on Faegri and Iversen 1989): The first step was decalcification by using 5% HCL solution. However, because four cores contained a negligible concentration of calcium carbonate in the sediments, this step was only performed for samples from one core. In order to remove humic acids and to disaggregate the sediment, all samples were heated at 70°C in 5% KOH and sieved over 200 µm. For acetolysis, the samples were boiled at 100°C in a mixture of acetic anhydride and sulfuric acid (in proportion 9 : 1). The gravity separation method was applied for removing mineral fragments by using a sodium-polytungstate liquid with a density of 2.0 kg/l. Finally, the samples were washed with alcohol and mounted on slides in a glycerine jelly. In total 135 samples were prepared, of which 90 have been analysed.

5.6 Pollen analysis

All samples were analysed for pollen and algae under the microscope using a magnification of at least 500 times. Several features were used for pollen grain identification (e.g. the general structure, size, shape, the number of apertures and sculpturing of the pollen walls) and this step was carried out with help of the determination key of Moore et al. (1991). Pollen was counted until a terrestrial pollen sum of 300-400 on average was reached. The suffix "type" has been used for pollen belonging to a group of different pollen species with similar morphological characteristics, following Moore et al. (1991). In addition, the suffix 'undiff.' means that family identification is certain, but one or more other morphological types within this family were distinguished and treated separately.

5.7 Presentation of pollen analytic data

The stratigraphical sequences of pollen counts were compiled into pollen diagrams by using *TILIA* (version 1.12, Grimm 1992) and *TILIA*-graph programs (version 2.0.b.5, Grimm 1991). Within these pollen diagrams, the different pollen and spore taxa were plotted against their stratigraphic depth, together with core lithology, measured LOI data and calculated pollen concentrations (based on *Lycopodium* counts). Because none of the cores have returned radiocarbon dates at the time of writing this thesis, no absolute pollen numbers could be calculated yet. The pollen data were expressed as percentages of a uniform pollen sum, representing the relative proportions of the different pollen species for each depth. Non-thermophilous trees, shrubs and upland herbs were included in the pollen sum. The tree species *Betula, Pinus, Salix* and *Populus* make up the arboreal taxa (AP) whereas the remaining upland species all together form the non-arboreal group of taxa (NAP). The Ericaceae family formed a third group included in the pollen sum. For Boreal and younger records, also thermophilous trees were included in the AP (e.g. *Corylus*) and a biostratigraphical correlation was made to regional pollen zones defined by Van Geel et al. (1980). The pollen and spores originating from (semi-) aquatic plants, ferns and mosses were excluded

from the pollen sum because they are of local origin. Further, all taxa have been classified into the groups 'trees and shrubs', 'upland herbs', 'semi-aquatic taxa', 'aquatic taxa' and 'ferns and mosses' and an additional group of algae.

In order to aid in the interpretation of the pollen diagrams itself and its comparison and biostratigraphical correlation with diagrams from other sites, the pollen diagrams were subdivided into distinct pollen zones. As defined by Gordon and Birks (1972), a pollen zone is 'a body of sediment with a consistent and homogeneous fossil pollen and spore content that is distinguished from adjacent sediment bodies by differences in the kind and frequencies of its contained fossil pollen and spores (from Birks and Birks 1980). Because the study area is located near the German-Dutch border, the subdivision has been based on the well-documented and well-dated regional vegetation development of the Netherlands as constructed by Hoek (1997a,b; 2001; see also section 4.2).

5.8 Reconstructing past vegetation from pollen

No one-to-one correspondence exists between the pollen assemblage and the real composition of past vegetation. The relative abundance of pollen types ending up within a sample is affected by several processes, including (species-specific) pollen production, dispersal, deposition and preservation (Janssen 1974). Moreover, the pollen assemblages consist of a local component and a regional component. For minimizing the possibility of incorrect interpretations, the vegetation reconstruction was preceded by a separation of the local and regional component. However, it has to be noted that some pollen types classified as semi-aquatic taxa actually might have been produced by upland species and the other way around. The most well-known example is grass pollen that is generally classified as upland component but can also originate from locally present grass species like *Phragmites* (reed). Therefore it has to be noted that the classification was only used for reasons of simplicity, but not used in a black and white way. Moreover, identification of the pollen grains was possible down to genus level, and sometimes only down to family level. Because identification down to species level was seldom possible, inferring of taxa specific ecological requirements was limited or impossible. However, from the review of Lateglacial palynological data in the Netherlands by Hoek (1997a,b) it is known which species are likely to have produced the Lateglacial fossil pollen in Dutch sediments enabling the reconstruction of regional vegetation composition at that time. For interpretation of local environmental conditions, ecological descriptions from Weeda et al. (2003) were used. As general for past ecosystem reconstructions, the underlying assumption of this study is that there has been little or no change in the ecological requirements and niches of organisms and communities since the time the fossil assemblage was laid down (Birks and Birks 1980). Besides the individual pollen species, the AP/NAP ratio was used as an indicator for changes in forest density. According to Zagwijn (1989), arboreal pollen percentages higher than circa 50%, indicate the presence of forest. Moreover, other abiotic proxies (geomorphological setting, lithostratigraphy, core lithology and LOI results) were additionally used for reconstructing past vegetation and environment.

5.9 Age control

So far, no samples have been radiocarbon dated and OSL samples taken yet. The time-interval contained by each core was estimated indirectly by correlating pollen zones from the lower Rhine valley with the Lateglacial regional zonation scheme as constructed by Hoek (1997a,b; 2001). Sedimentary units were indirectly dated by correlation with the well-dated sedimentary archive of the Rhine-Meuse delta (most up-to-date insights: Busschers 2008). An admixture of pumice within fluvial in-channel deposits was seen as an indicator for a Younger Dryas (or younger) age, as it originates from the Laacher See volcanic eruption just before the Allerød-Younger Dryas transition (see also section 4.2).

PART II

Research results

6 Sedimentary architecture of the lower Rhine valley

In order to study the spatial distribution of different sedimentary facies in the lower Rhine valley, three cross-sections were constructed based on circa 120 lithological borehole descriptions. The location and general architecture of each cross-section is described in section 6.1. Sedimentary units were distinguished on the basis of lithological characteristics, geometry, elevation, position in relationship to other units and the floodplain level they make part of. Because it appeared that the composition of the fluvial sediments in this part of the lower Rhine valley varies strongly at different spatial scales, the lithostratigraphic division is partly based on the fluvial styles of the distinct floodplain levels, which are inferred from the DEM. Sedimentological units are coded on the basis of inferred genesis and (relative) age and individual descriptions are given in section 6.2. In order simplify the readability of the cross-sections and the geological map, unit codes contain prefixes as listed below. In case of the fluvial units, code numbering starts with number F2 because there is one older fluvial unit (F1) in the study area which has been distinguished during other studies (Graeves 2010; Janssens 2010) but which has not been cored during present study. However, the floodplain level it makes part of (braided level I) is recognised on the DEM and has been described by Verschuren (2007), Graeves (2010) and Janssens (2010) in the upstream part of the Oude-IJssel valley. Section 6.3 provides an overall analysis of the distinct floodplain levels that can be distinguished in the area of study and which are shown on the newly constructed geological-geomorphological map. Calculated floodplain gradients are provided in section 6.4. The last section focuses on the type of channel fills which appear to be typical for the area of study (section 6.5).

G: glacial deposits
S: hill-slope deposits
F: fluvial in-channel deposits
O: overbank deposits
E: aeolian deposits
V: peat deposits, outside residual channels
R: residual channel fills, both organic and clastic

6.1 Description of the cross-sections

Cross-sections I (part A-A') and II (B-B') are of relative great length, respectively spanning ~5.5 and 4.5 km of the lower Rhine valley perpendicular to the average flow direction (figure 18). Cross-section I, however, has an extension towards the other side of the valley across the Holocene floodplain (part A'-A"; appendix II). The most interesting part of the first transect (part A-A') is positioned to the southwest of the modern Rhine near the village 'Appeldorn' (figure 20). The southern tip reaches ice-marginal topography, namely slope-wash and original deposits of the flank of the ice-pushed ridge (units G1 and S1). The overall

architecture is dominated by two fluvial units (F3 and F4) which are present near the surface. A third fluvial unit (F2) is present along the total length of the transect at approximately 4-5 m depth. A very interesting unit is V1, a peat layer which has been recorded at one site but which is approximately present everywhere in this area beneath unit F2 (Klosterman 1989). Further, isolated dunes (unit E1) are sporadically present on top of successively overbank deposits (units O1-O3) and units F3 or F4. The most interesting part of cross-section II (B'-B") is present on the other side of the modern Rhine (figure 21 and figure 18). It is positioned near the village called 'Mehr', circa 3 km upstream of cross-section I on average. A similar distance is in between the second transect and the further upstream located cross-section V as constructed by Erkens et al. (2011, see appendix III). Similar to cross-section I, fluvial units F3 and F4 predominate near the surface, unit F2 is present at 4-5 m depth. Again, overbank deposits and younger dune morphology cover these fluvial in-channel deposits. From both cross-section I and II can be derived that unit F3 has a slightly higher surface elevation that unit F4, a difference in the order of 0.5-1 m.

A third much shorter (~0.7 km) and more detailed cross-section is constructed across the channel system near the foot of the ice-pushed ridge near the village '*Marienbaum*' (cross-section III (C-C'), figure 18). The purpose was to acquire more insight in the complicated build-up of the fluvial sediments in this part of the valley and the degree of local diversity. The cross-section is given in figure 19, showing the steep slopes of the ice-pushed ridge on the far left, which are covered with periglacial hill-slope sediments. Unit F3 is located against the foot of the ice-pushed ridge and unit F4 in the central part of the transect.

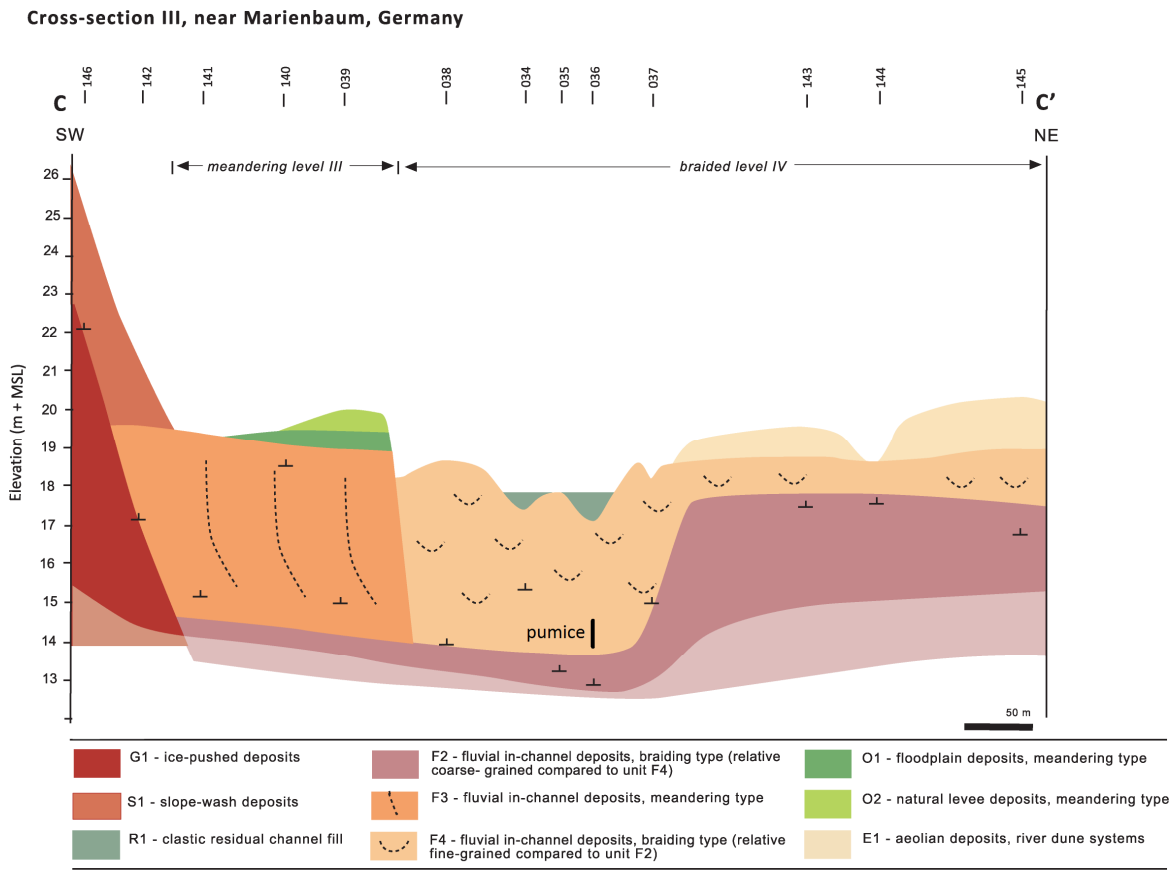

Figure 19 Cross-section III across the *Marienbaum* channel system. For location, see figure 18.

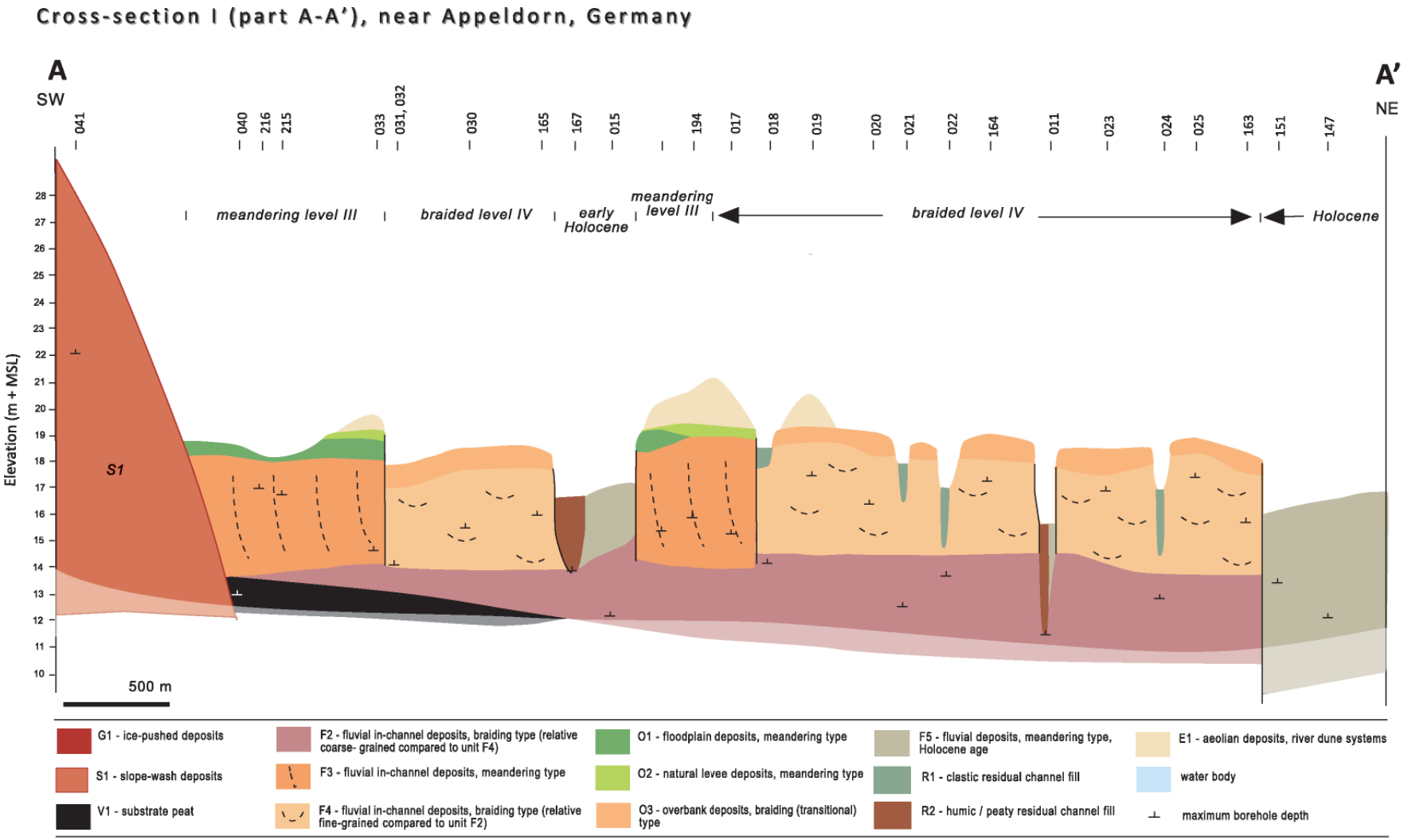

Figure 20: Part A-A' of cross-section I to the southwest of the Rhine, near the village Appeldorn. For location of cross-section, see figure 18.

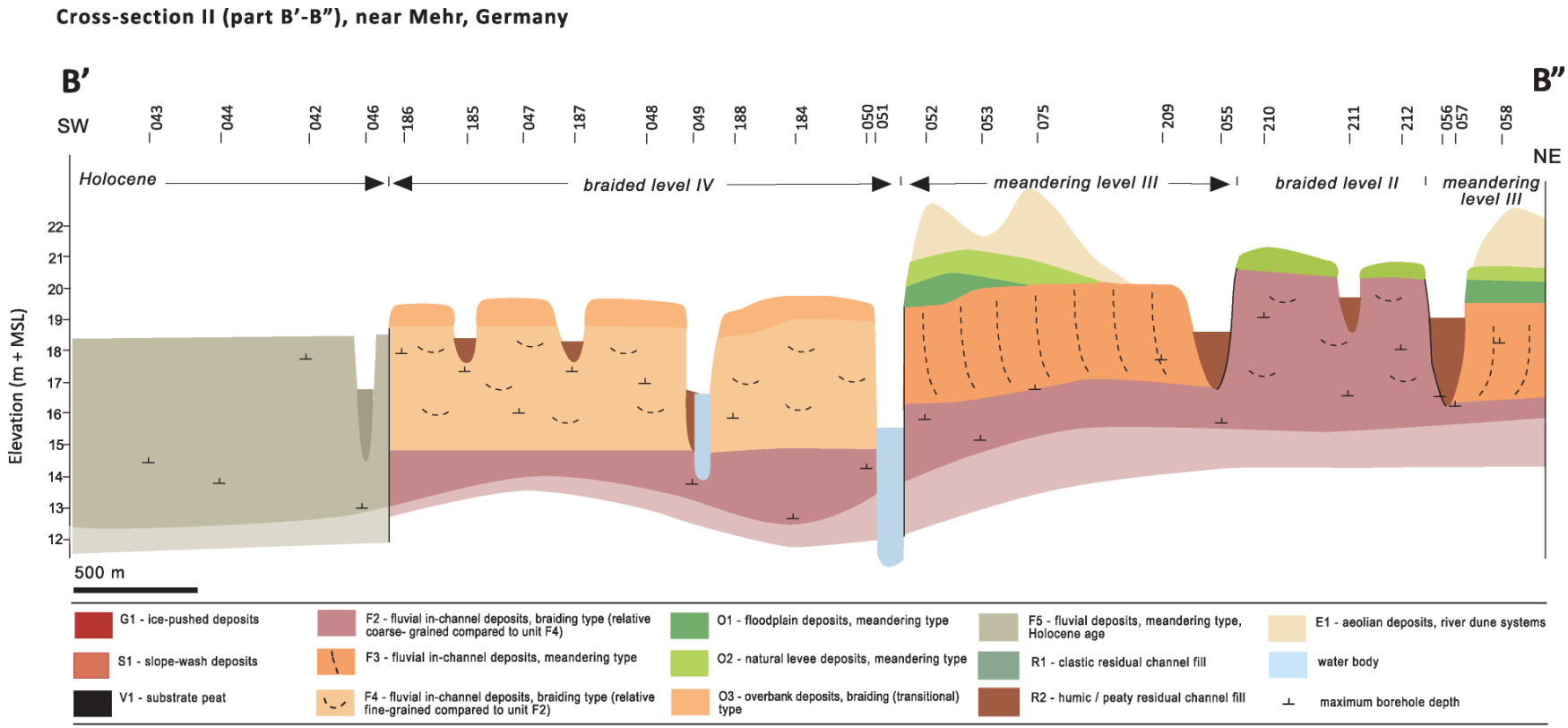

Figure 21: Part B'-B" of cross-section II to the northeast of the Rhine, near the village Mehr. For location of the cross-section, see figure 18.

6.2 Description of the sedimentary units

Unit G1 – Ice-pushed deposits

The top of unit G1 is observed at maximum borehole depth at three sites in cross-section I and III. Compared to other sedimentary units that are present in the lower Rhine valley, unit G1 is very coarse-grained and consists of gravelly sand and sandy gravel. The distribution of this unit is mainly based on the DEM, because the deposits make part of a ridge topography. Unit G1 is interpreted as glacial deposits making part of the ice-pushed ridges to the south of the lower Rhine valley. From many studies it is known that these ridges have been formed by the Saalian ice sheet (e.g. Klostermann 1992). According to the Dutch lithostratigraphic classification scheme of De Mulder et al. (2003), this unit belongs to the Drente Formation.

Unit S1 – Slope-wash deposits

Unit S1 has been distinguished at two sites in the south-western tips of cross-section I and III. This unit is located near and on top of the flanks of the ice-pushed ridge and covers the glacial deposits belonging to unit G1 or adjacent fluvial terraces units. Compared to unit G1, the sediment is much better sorted, has a smaller grain-size on average and contains less gravel. Layers of relative coarse gravelly material interchange with sandy layers and very fine-grained silty layers. Unit S1 is thought to have been produced by snow or ice-meltwater streams on the slopes of the ice-pushed ridge and to consist of reworked and transported glacial deposits. Following De Mulder et al. (2003), this unit is of periglacial origin and belongs to the Boxtel Formation.

Unit V1 – Substrate peat

Unit V1 has only been distinguished at one site (borehole 040) at 4.5 m depth. It comprises a dark-brown/grey sandy peat layer of at least 40 cm thick with a few thin whitish sand layers near its upper boundary. At the borehole location, it is covered by a sandy fining-upward sequence classified as unit F3. Based on stratigraphic position, elevation and a scan of its pollen content, the peat is presumed to make part of a widespread peat which's presence is showed by geological cross-sections N-O and L-M from Klosterman (1989). Because of its great extent and high organic content, it has probably been formed under interglacial or interstadial conditions in the floodplain area of a former Rhine system. According to Klosterman (1989) this peat was formed during the Eemian and early Weichselian. The pollen content of this layer has been analysed and is described in chapter 7 (*Kehrum Torfkuhle* core).

Unit F2 – Fluvial in-channel deposits, braiding type

Unit F2 consists of poorly sorted medium- to coarse-grained gravelly sands or sandy gravels with a grey to light-brown colour. The gravel content is generally above 10% and the gravel is relative coarse compared to other gravelly facies described later on. This unit is typically encountered in the deepest part of borings or its upper boundary formed the level beyond which coring by hand was not possible to continue, due to the coarseness of the material. Only at one location (boreholes 210 and 212 in cross-section II), this unit is thought to be present near the surface at circa 20.5 m +MSL. Except from this location, the top is located around 15-16 m +MSL in the western part of cross-section I and at circa 16-17 m + MSL in the eastern part of cross-section II. It is generally covered by unit F3 or F4. The dominance of coarse-grained sands, a high gravel concentration and small-scale internal fining upward sequences, point towards deposition in a braiding river system (vertical aggradation type (V gravel / In German: V-Schotter) according to Schirmer 1995). In the Netherlands, this unit would be part of the Kreftenheye Formation (Verbraeck 1984; De Mulder et al. 2003; Busschers 2008).

Unit F3 – Fluvial in-channel deposits, meandering type

Unit F3 is dominated by light brown-grey fine- to medium-grained sands which are well or moderately sorted. Gravel content is normally very low, or slightly elevated for intervals of maximum twenty centimeters. In several borehole descriptions, one well-developed fining-upward sequence was documented spanning the full extent of the sand body and with a gravelly layer of circa 10-20 cm at the base. In other cases, a couple of slightly smaller fining upward sequences were observed. The total thickness of this unit ranges from 3.5 to 4.5 meter and its top is located at circa 18.5-19.5 m and 20 m + MSL in cross-section I and II, respectively. Furthermore, it is located on top of unit F2 and is itself covered by overbank deposits (O1 or O2) and occasionally dunes (E1). At some sites, the cores penetrated the entire sandy succession and reached the so-called channel-lag: gravelly deposits at the base of meandering channel deposits, which indicate the largest depth of channel scour during lateral migration (Erkens et al. 2011). Despite the fact that it is covered by overbank deposits everywhere, the surface morphology of unit F3 is clearly visible of the DEM and shows a large number abandoned meanders, sometimes containing a scroll bar and swale topography. In combination with the high degree of sorting of the material and long fining-upward sequences, there seems to be no doubt that it has been created by a former meandering river system with a channel depth of circa 3-4 m (lateral accretion type (L gravel / In German: L-Schotter) according to Schirmer 1995).

Unit F4 – Fluvial in-channel deposits, braiding type

Unit F4 consists predominantly of moderately or poorly-sorted fine- to coarse grained gravelly sands, however, the degree of sorting and coarseness of the material varies largely for different sites or depths.

Also the color of the sediment varies widely; however, on average it has a somewhat darker (browner) appearance than other poorly sorted sediments in this area (e.g. unit F2), possible due to a slightly enhanced admixture of organic material and/or iron-coatings covering the individual grains. Determination of the base of unit F4 is tentatively based on sedimentological indications (dm-scale F4 sequences). An estimate of 14-15 m + MSL is given for cross-section I, estimating the thickness of this unit to be in the order of 3 to 4 meter. Based on the inferred chronostratigraphy (see next section) and cross-sections from the upstream part of the Oude-IJssel valley (Greaves 2010; Janssens 2010) it seems plausible that it covers unit F2 over its full section width. Because the material occasionally contains pumice granules (e.g. boreholes 036 and 195) it equals the Kreftenheye VI subunit of that formation (Verbraeck 1984; De Mulder et al. 2003). Despite its blanketing with overbank deposits (unit O3), its irregular surface morphology is visible on the DEM, with internal elevation differences up to 1.5 m. It shows a braiding channel pattern that, however, has been partly reworked by younger incising meandering river systems. In combination with the coarseness and poorly sorting of the sediment development by a braiding river system is inferred (V gravel type, Schirmer 1995).

Unit F5 – Fluvial deposits of Holocene age, meandering type

Sediments belonging to unit F5 have a wide range of characteristics but on average a relative large silt fraction compared to sedimentary facies described so far. However, unit F5 is distinguished primarily on the basis of the DEM, because its surface is located at a significant lower level than adjacent units and shows very well-developed meandering channels with associated point-bar topography. The difference in surface elevation of the floodlevel build-up of unit F5 and adjacent floodplains is typically between 1-2 m. This unit of clear Holocene age, as confirmed by e.g. Klostermann (1992) and Erkens (2009), is out of scope of this research and will not be described in further detail. For more details, see Van Munster (MSc thesis, in prep.)

Unit O1 – Floodplain deposits, meandering type

Unit O1 is a compacted sandy clay or sandy loam layer on top of unit F3. It has a (light) grey-brown color, a thickness of circa one meter and is relative rich in manganese. On average, its base is located at 19-20 m and 18-19 m +MSL in cross-section II and I, respectively. In one part of cross-section II (boreholes 210 and 212) it is located approximately half a meter higher and it covers unit F2. Based on the high clay content of unit O1 and its position on top of in-channel deposits of meandering facies F3, it is concluded that it represents floodplain deposits of a former meandering river system. Periodically, the area flooded and clay particles settled from suspension as soon as the water became stagnant. In line of the classification scheme of the Mulder et al. (2003), this clayey layer represents the Wijchen Member (defining the top of the Kreftenheye Formation).

Unit O2 – Natural levee deposits, meandering type

Unit O2 consists of lightly loamy or silty fine- to medium-grained sands (matrix grain size 210-420 μm) with a color varying from light-brown to light-grey. Generally, it shows a fining-upwards trend in coarseness of the material. It is either located directly on top of meandering facies F3 or unit O1 is in between them. This unit is distinguished at two and four sites in cross-section I and II, respectively. Because of a higher silt content at the expense of clay compared to unit O1, unit O2 is thought to have been deposited more proximal to former river channels in the form of natural levees. Because both units have more or less the same elevation, they are considered to be formed by the same meandering system.

Unit O3 –Overbank deposits, braiding (or transitional) type

Sedimentary unit O3 has many lithological characteristics in common with unit O2: It consists of modestly loamy fine- to medium-grained sands and has a thickness of approximately one meter. It is distinguished from O2 on the basis of stratigraphic position: Unit O3 covers unit F4 over its full extent, in contrast to unit O2 which is present on more local scale on top of unit F3. Unit O3 represents floodplain deposits primarily formed by the braiding system responsible for unit F4. The sediment making up unit O3 is called 'Auensande' by Schirmer (1995), making part of the V terrace fluviatile series.

Unit E1 – Aeolian deposits, river dune systems

The distribution of sediments belonging to unit E1 can be easily determined both in the field and on the DEM, because of their dune-morphology. These dune sediments consist of well-rounded and well-sorted very fine- to medium-grained sand (predominantly 150-300 μm). Unit E1 is generally located on top of one of the overbank facies (O1,O2 or O3) or directly on top of in-channel deposits belonging to unit F3. In the area of field-study, the dunes have a height of 2 m on average and a maximum height of circa 4-5 m. The fact that they almost exclusively occur on the north-eastern side of former river courses, points towards formation by dominantly south-western winds which transported sand out of the non-vegetated river systems. These so-called 'river-dunes' make up the Delwijnen Member in the Boxtel Formation in the lithostratigraphical scheme of The Netherlands (De Mulder et al. 2003).

Units R1 and R2 - Residual channel fills, organic and non-organic

Units R1 and R2 comprise sediments which are locally present in abandoned river channels. In the area of interest, these channel fills predominantly consist of non-organic or lightly humic loamy and clayey sediments, often containing thin sand layers. These sediments are coded as unit R1. Humic or highly organic channel fills (unit R2) are rare in the area of study, probably because many abandoned channels were temporally reactivated during floods for long time after abandonment (see chapter 10). The

channels are located in sand bodies belonging to units F2, F3 or F4. Typical characteristics of the residual channels-fills in the area of study and the sedimentary processes which are thought to be responsible for them, are discussed section 6.5.

6.3 Geological-geomorphological map and floodplain level descriptions

In the cross-sections, three floodplain levels are distinguished above the Holocene meandering floodplain. Based on cross-cutting relationships and elevation differences, the relative chronology of the different floodplain levels has been determined for the lower Rhine valley. From old to young the relative chronology is as follows: braided level II, meandering level III and braided level IV. In the north-eastern part of the area of interest, at least one additional older level can be distinguished on the DEM and is called braided level I. This section restricts to a description of the different floodplain levels and an interpretation of their way of genesis. Thereby, the description of braided level I is purely based on its characteristics as visible on the DEM and synthesized results from other studies (Verschuren 2007; Graeves 2010; Janssens 2010). A more absolute chronology is constructed on the basis of indirect (pumice, pollen) and direct (OSL) dating results, and is presented and discussed in chapter 8. However, in order to enhance the coherence of this thesis, the inferred periods of formation of the distinct floodplain levels make already part of the subtitles in this section.

A geological-geomorphological map has been constructed for the downstream part of the lower Rhine valley and the upstream part of the Oude-IJssel valley and is shown in appendix I. New insights have been synthesized with results from Klostermann (1986, 1992), Jansen (2001), Erkens et al. (2011), Janssens (2010), Graeves (2010), Verschuren (2007) and Favier (2001). The DEM was used to extrapolate the distinguished floodplain levels to areas from which no data are available.

Braided level I – Middle Pleniglacial

The oldest braidplain level known from along the study area is found in the very northeast near the hills of the 'Bergisches Land'. Almost no fluvial morphology is visible on the DEM since this level is blanketed by a circa 0.5 m thick layer of coversand and is partly reworked by the younger *Issel* and *Wolfstrang* river systems (Janssens 2010). Only at high magnification, channel structures become visible showing a braided network of relative straight and small channels. From Graeves (2010) and Janssens (2010) it is known that this floodplain level is build-up by sediments of typical braiding facies, successively covered by a loamy/silty layer of circa 10-20 cm and coversand with a thickness of half a meter on average. Near Loikum (figure 11), the top elevation of the braidplain deposits is at circa 18 m +MSL and shows elevation differences up to 1 or even 1.5 m (Graeves 2010; Janssens 2010). Janssens (2010), however, treats the most eastern part of this floodplain level as another even older braidplain. Unfortunately, the thickness of the coversand in this eastern part of the valley (at least 3 m) limits the detection of more than one

braidplain, so there is no conclusive evidence for an extra floodplain level. Therefore, in this study the fluvial sediments on this side of the valley are treated as one level (figure 22). In a downstream direction, it continues into the Oude-IJssel valley, linking up to mapping by Verschuren (2007, map unit F2) where it is also covered with aeolian sand. From the latter study it is known that this braiding level has a top elevation of circa 12.5 m +MSL, a few kilometers upstream from Ulft (appendix I). Because the base of the sand body making up braided level I was nowhere reached, only a minimum thickness of 2.5 m can be provided for the area near Ulft.

Braided level II – Late Pleniglacial

A second younger braidplain is distinguished in the lower Rhine valley southwest of braided level I. Relative large fragments of braided level II are present at the surface along both sides of the modern Rhine river, near Hamminkeln (NE of the Rhine) and Xanten (SW of the Rhine) (figure 11). The rest of the braidplain has been partly reworked by younger river systems leaving sediments only preserved at depth. The braidplain body consists of sedimentary unit F2 and makes part of cross-sections I, II and III (figures 19-21). Where the braided level has been spared from lateral erosion it is covered by a 0.5-1 m thick layer of overbank deposits (units O1 or O2). Further, braidplain II is locally covered by large dune systems of parabolic type (unit E1).

Near Xanten, braided level II is present as part of a secondary channel system that is closely surrounded by ice-marginal topography (figure 11 and 22). Here, Erkens et al. (2011) describes two braidplain remnants on both sides of a younger meandering system and classifies them as NT2 and (probable) NT3 terraces (appendix III). The northern terrace remnant was identified as (probable) NT3 terrace because pumice was encountered in the top of the channel sands. The top of the sand deposits is circa 20 and 19.5 m +MSL for the southern and northern remnant, respectively. Apart from the pumice admixture in the northern fragment, the elevation difference seems not significant when realising that elevation differences within a braidplain surface can be over one meter and when knowing that the elevation of the southern remnant is based on one single borehole description. Moreover, pumice was observed in a very small amount near the very top of the channel sands at only one location (G. Erkens, personal communication). Therefore, due to the lack of strong evidence for a difference in age, both remnants are thought to make part of the same floodplain level, namely braided level II. On the DEM, these braidplains are flat areas with a small number of non-winding and relative narrow abandoned channels. The fact that no dense network of braiding channels appears from the DEM is probably due to the location of these terrace remnants outside the main trunk valley of the Rhine and in the lee of elevated ice-pushed deposits, prohibiting intense fluvial activity in this area.

On the other side of the river Rhine, braidplain II appears much better preserved and is distinguished as a clear level between braidplain I and the Holocene floodplains on the DEM (figure 22). Unfortunately, overlying large-scale river dune systems make it impossible to determine the precise distribution of

braidplain II along this side of the river Rhine. The area west of Hamminkeln forms an exception; a well-developed braided channel morphology is present there. The orientation of the channels and the north-eastern limit of braidplain II as derived from the DEM, point towards a westward re-directed fluvial system compared to braidplain I. In the same area, another striking morphological feature related to braided level II is visible on the DEM: Braided level II has a higher surface elevation than the older braided level I while it also cross-cuts level I. The cross-cut boundary is clearly visible as a straight and sharp contact between both floodplain levels (figure 22). The underlying mechanism which is presumed to be responsible for this morphological feature is discussed in chapter 9.

Braidplain II is thought to form the eastern tip of cross-section V from Erkens et al. (2011), where the top of sand-deposits has an elevation of circa 21.5 m +MSL. Approximately 5 km downstream, an isolated terrace fragment is distinguished in the eastern part of cross-section II (of present study), where the top of the sand deposits is located at circa 20 m +MSL. The base of sedimentary unit F2 has nowhere been reached so only a minimum thickness can be determined of 7 m on the basis of its lowest and highest occurrence in cross-section II (figure 21).

*Meandering level III – Bølling-*Allerød *interstadial complex*

In the lower Rhine valley, one meandering floodplain level is distinguished outside and at a higher elevation than the Holocene meanders. It is the oldest meandering level in the area, coded as meandering level III to reflect its relative age compared to the older and younger braidplain levels (I and II, and IV, respectively). In the Oude-IJssel valley northeast of the Holocene Rhine and in a relative narrow stroke of land against the ice-pushed ridge southwest of the Holocene Rhine, a large number of high-sinuosity meandering channel fragments is visible on the DEM (figure 22). These channels have a width in the order of 100-200 m. By comparison with the dimensions of the Holocene meanders and the present-day Rhine, it seems likely that these meanders were formed by a multi-channeled river system (Cohen 2003; Erkens et al. 2011). In the area to the west of Xanten, the system show a well-developed scroll bar system in combination with relative large channel dimensions. Moreover, here it is clearly visible that the meandering system cross-cuts braided level II. On the other side of the Rhine, meandering level III is almost entirely covered with river dunes.

The sandy in-channel deposits making up meandering level III are classified as unit F3 and often reveal a clear fining-upwards tendency and a gravelly channel-lag at the base. As shown in cross-sections I, II and III (figures 19-21), and cross-section V from Erkens et al. (2011; appendix III) the thickness of the meandering facies unit F3 varies between 3 and 4.5 m. It is covered by overbank deposits belonging to units O1 and O2 and locally by river dunes (unit E1). In a downstream direction, the top of the sand deposits decreases from circa 19.5 +MSL in cross-section V from Erkens et al. (2011; appendix III) towards circa 19 and 18 m in cross-sections II and I, respectively, along the south-western side of the

Rhine valley. On the other side of the Rhine, an average top elevation of circa 19.5 m +MSL appears from cross-section II.

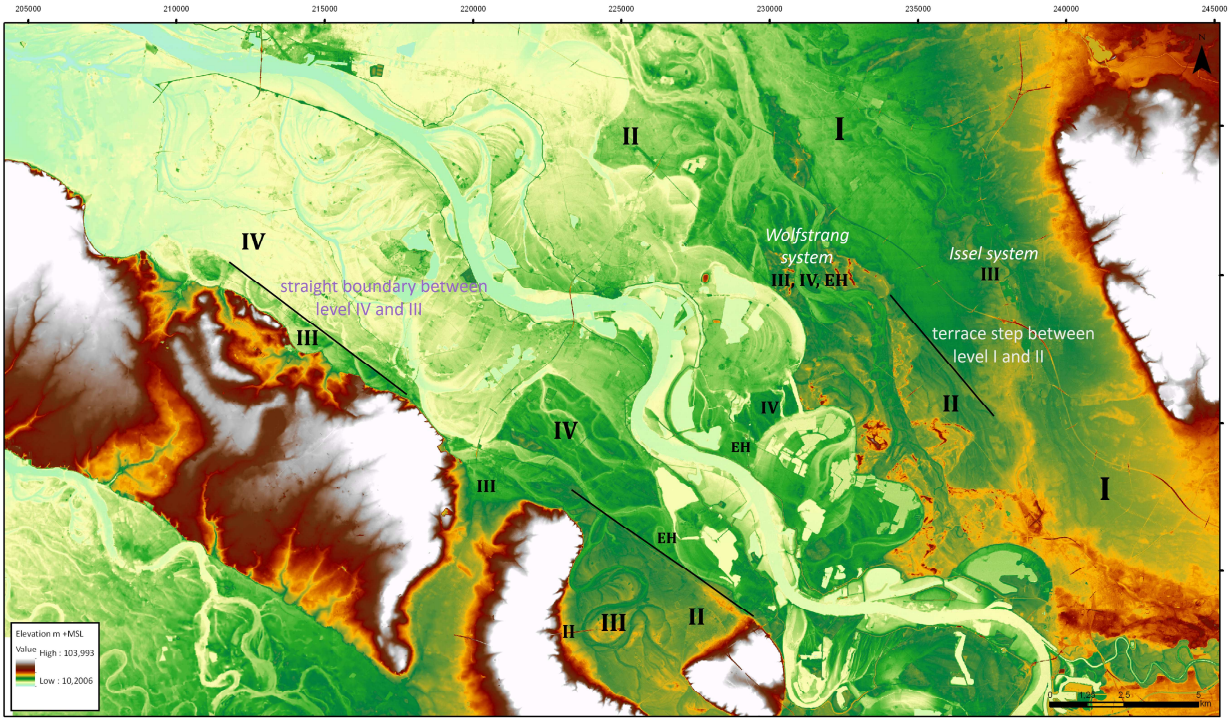

Figure 22 Digital elevation model of the area of study, showing a subdivision of the valley bottom in distinct floodplain level fragments (I-IV).

Braided level IV – Younger Dryas stadial

The youngest braided level that can be distinguished in the lower Rhine valley, occurs at a significant lower elevation than the older levels. Despite the fact that the largest portion of this youngest has been laterally eroded by Holocene meandering activity, the stratigraphic position of the preserved fragments indicates braidplain IV to have dissected braidplain II and III over the axis of the valley, in the same way as the Holocene meander belt. In the vicinity of cross-section I and II, the level has two terrace remnants, supporting the villages 'Appeldorn' and 'Mehr' to the southwest respectively northeast of the Rhine (figure 18 and 22; appendix I). The relative flatness of their surfaces in combination with the braiding channel patterns point towards deposition by a braiding river system. Moreover, the very straight cross-cut boundary with older deposits (for over ten kilometres between Kalkar and Xanten) points towards a braided river style. Also the poor sorting and relative coarseness of the deposits point towards a similar genesis. These sediments (unit F4) are almost everywhere covered with loamy overbank deposits (unit O3). Braided level IV dominates the eastern respectively western half of cross-sections I and II, and makes part of a relative small channel system in cross-section III. Internal surface elevation differences are in the order of 1-2 m and the top of sand deposits is elevated at approximately 19, 18.5 and 18 m in transects I, II and III, respectively. The average thickness of the sand body is circa 4 m. Especially the terrace supporting Appeldorn has partly been reworked by younger and slightly meandering secondary floodplain channels

which incised into the terrace surface. In the Oude IJssel valley, braided level IV is present as one or a couple of parallel channel belts of 400-600 m wide, together making up the *Wolfstrang* system (figure 22). The really active channels are even smaller, of which one channel fill was palynologically studied by Janssens (2010) and Greaves (2010) (*Schlederhorst* channel, compare 18 and 22). In the *Wolfstrang* system, deposits of braidplain IV have been partly reworked by a younger much smaller meandering channel (*Berckermann* channel, sampled by Janssens 2010 and Greaves 2010). From the very limited dimensions of braided level IV in the Oude-IJssel valley, it can be concluded this river system has transported only a very small discharge in the past.

6.4 Floodplain level gradients

Average floodplain gradients have been calculated using a linear regression over data points of top of in-channel-deposits, including additional data from two upstream cross-sections (Erkens et al. 2011) and two downstream cross-sections (Janssens 2010; Verschuren 2007). For each point, the distance towards the line Xanten-Diersford was measured, which is located in the upstream part of the study area, perpendicular to the average flow direction (see figure 11). Since the area of study contains the bifurcation point between two Rhine courses, a (rough) distinction was made between data points from different (lateral) parts of the LRV. Three zones were distinguished: the (south-)western, central and (north-)eastern part of the valley. Due to the fragmented character of the preserved floodplains and the limited number of boreholes it was, however, not possible to calculate three gradients for each floodplain level. Except from meandering level III, only one single gradient within one lateral zone could be determined for each floodplain level. The resulting gradients are showed in figure 23a and the zones are additionally mentioned.

Because of the low number of top-of-sand (channel belt sand) data, gradients were also estimated on the basis of surface elevations from the DEM. Along a straight line parallel to the presumed flow direction (this is directed towards the Oude IJssel valley for braidplain I) highest points of the floodplain surfaces have been determined (e.g. pointbars, braid bars) which have not been disturbed by man. Hereby, it is assumed that deposits overlying the different floodplain levels have a similar thickness everywhere and do not influence the terrace gradient. The results are showed in figure 23b, and again, for each gradient the lateral zone is mentioned.

All gradients appear to be in the range of 27 to 30 cm/km, independent of data point type (top of sand or surface), except from one outlier: Braidplain I appears to have a slope of 39 cm/km on the basis of top of sand elevations. Since this trendline is only based on elevations at two distances from the line Xanten-Diersford (cross-sections from Janssens 2010 and Verschuren 2007), one incorrect elevation already has an enormous impact on the calculation. Because all other gradients appear to be very similar, a gradient of 39 cm/km is thought to be too high. Reconsideration of both cross-sections brought forward one other option: braidplain I (= unit F2a according to Janssens 2010) might correlate to unit F3 instead of unit F2 in

cross-section C from Verschuren (2007). This provides a more expected gradient of 29 cm/km. On the basis of lithostratigraphic relationships, however, this alternative correlations seems unlikely, so further research is required in this part of the Oude IJssel valley.

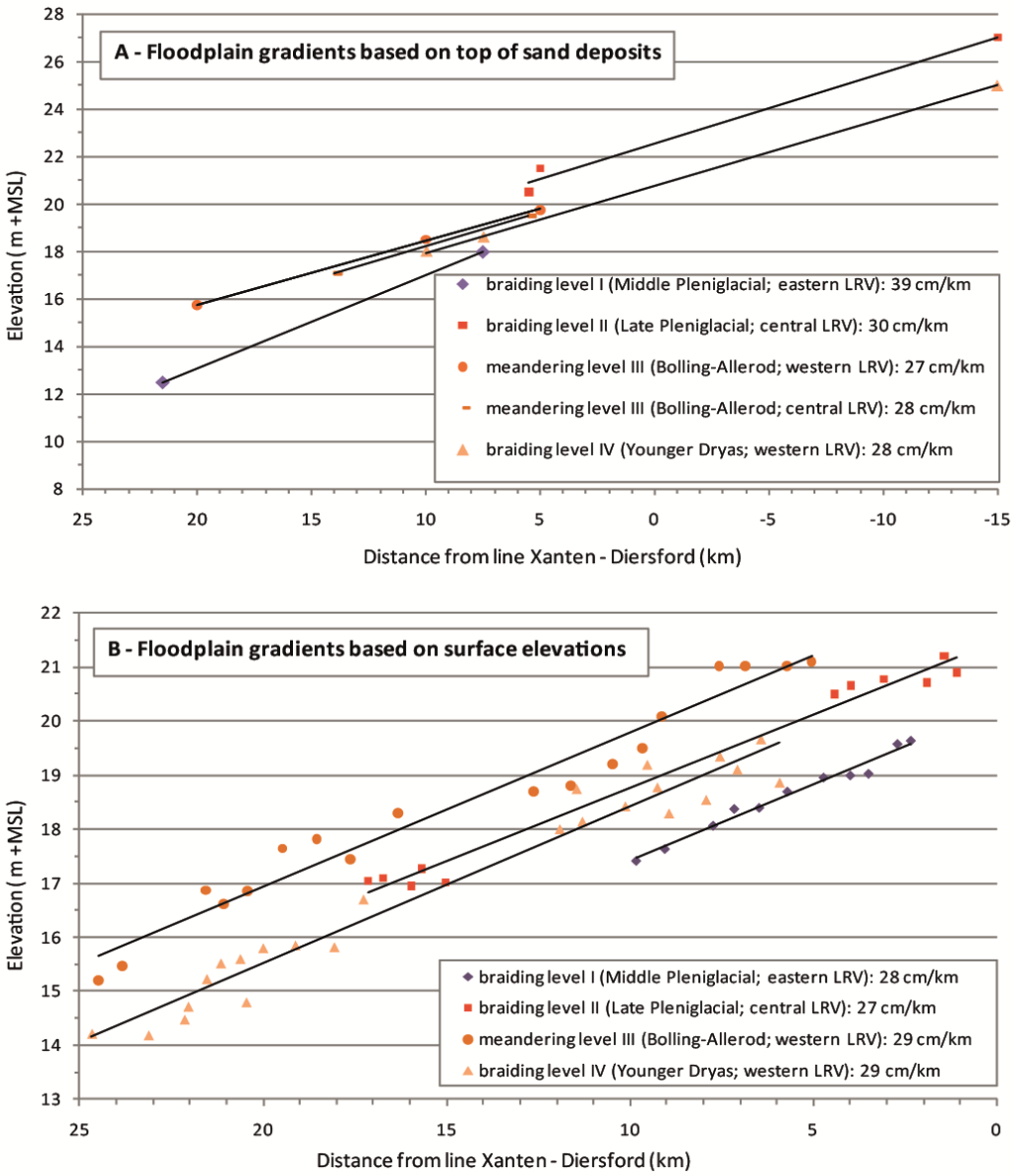

Figure 23 Calculated floodplain gradients on the basis of top-of-(channel-belt) sand deposits (A) and modern surface elevations (B). For the construction of figure A, additional data from Erkens et al. (2011), Verschuren (2007) and Janssens (2010) were used. For interpreting the gradients, the approximate location of the lines along which the gradients have been calculated is given; in the central, eastern or western part of the LRV.

6.5 Residual channel fills

In section 6.2 it was already mentioned that residuel channel fills in the area of study predominantly comprise non-organic sediments. Especially the relative shallow channel systems of braidplains II and IV appear to have filled-up with clastic sediments, containing no or only a very low fraction of (washed-in) organic material. The meandering channels formed during the activity phase of floodplain level III contain an infill of successively (sandy) loam covered by gyttja or peat. The cross-section from the *Heesenhof* site (figure 24) shows this characteristic infilling that has been observed in at least three other well-developed meandering systems of level III.

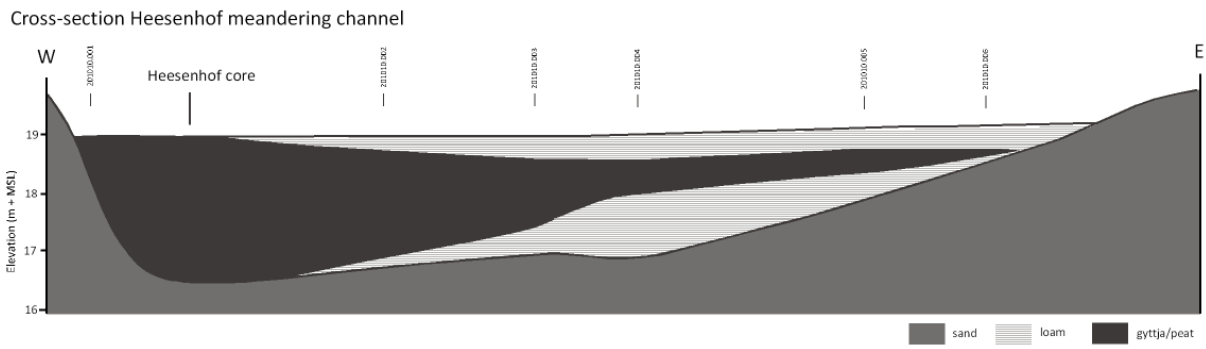

Figure 24 Cross-section through the meandering channel system near *Heesenhof*, to the west of Xanten. For location see figure 22. The uppermost clay layer, is presumably of late Holocene age.

The difference between the thickness of the clayey-organic channel fill and the thickness of the meandering channel-belt body, respectively ~2.25 and 4.5 m for the *Heesenhof* meander, suggest gradual abandonement of the system, resulting in initial shallowing of the channel with sand, before final abandonment. Possibly also reactivation during floods produced by younger Rhine systems might have resulted in an additional input of sand. Moreover, gradual abandonment and flooding activity at a regular basis is reflected by clayey channel fill facies, for instance up to circa 2 m in the *Heesenhof* meander. This seems to be especially the case for the preserved abandoned meanders making part of floodplain level III.

7 Palynological results and core descriptions

An original research aim was to reconstruct the regional vegetation development across the last glacial-interglacial transition, based on the pollen content of a sequence of residual channels fills from the LRV. Unfortunately, only one out of four potential Lateglacial channel fills turned out to contain a Lateglacial signal, despite the fact that core locations were selected diligently based on preliminary research (see section 5.3 for the criteria for core location selection). Despite the unexpected outcome than no pollen record spanning the whole Lateglacial could be constructed with the obtained pollen diagrams, new insights developed concerning the fluvial dynamics of the lower Rhine and its flooding activity during the Lateglacial and early Holocene, which are discussed in part III of this thesis. Pollen analysis was performed on five cores; four retrieved from residual channel fills of different river terraces and one from a peat layer at depth covered by fluvial sediments (*Kehrum Torfkuhle* core).

7.1 Coring sites, core descriptions and loss-on-ignition

Kehrum Torfkuhle (KT)

Initially, it was only intended to perform a boring at the *Kehrum Torfkuhle* site as part of main transect I (figure 18). The site is located near the point of outflow of a large meltwater valley through the ice-pushed ridge. The direct surroundings of the site have a relative flat surface morphology and the surface soil is very organic (figure 25). At the core location, the uppermost circa 4.5 m of sediments consists of well-sorted fine- to coarse-grained sand, characterised by a long fining-upwards trend (sedimentary unit F3 in figure 20). The lower 30 cm of this sequence is poorly sorted coarse-grained sand mixed with gravel and with an orange iron-oxide color. The lower boundary of this mineral sequence is very sharp, where there is a transition towards dark brown peat. According to Klostermann (1989) this peat layer has and Eemian age and has a wide distribution in the downstream part of the LRV. A core of approximately 35 cm of peat was taken for pollen-analysis, for estimating the period of time it has recorded. The core has a relative homogenous appearance despite of a couple of thin sandy layers of approximately 1 cm thick. This is also showed by the loss-on-ignition results (see figure 28). Moreover, almost no organic macro remains were present and the whole core was poor in calcium-carbonate.

Marienbaum (MB)

The *Marienbaum* coring site is located at the upstream 'dead-end' of the *Marienbaum* channel system as is showed by the DEM (figure 18 and 25). Cross-section III is located across the same channel system but circa 1.5 km downstream (figure 19). The core is taken at a site where the channel is very close located to the ice-pushed ridge and peat formation is favoured by upwelling water. Today the soil is (still) very humid in the pastureland at this site.

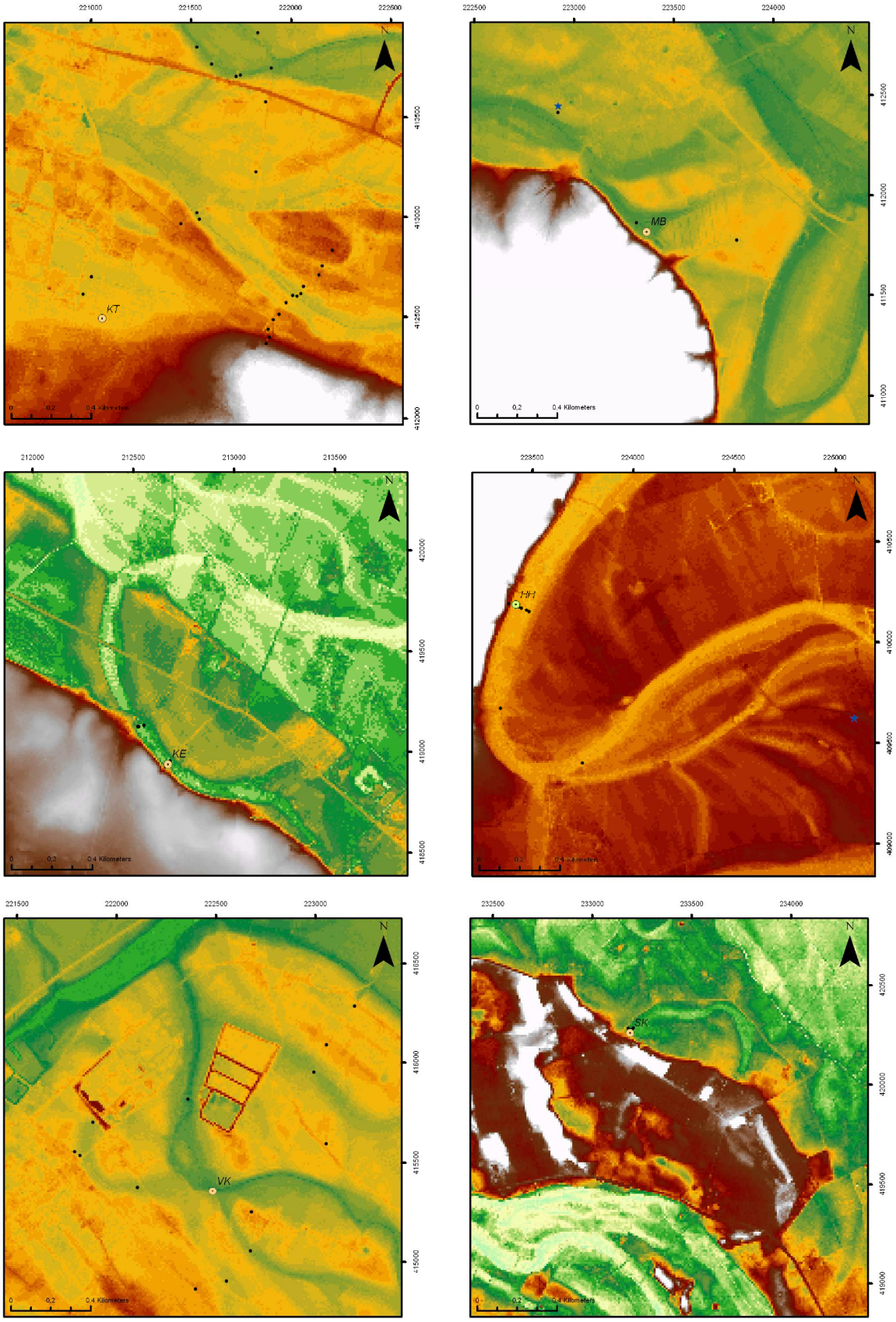

Figure 25 Detailed DEM showing the six core locations from where data were gained for palynological study. From left to right and from top to bottom: *Kehrum Torfkuhle* (KT), *Marienbaum* (MB), *Klein Entenhorst* (KE), *Heesenhof* (HH), *Vosse Kuhle* (VK) and *Sanders Kath* (SK).

A core of almost two meters was retrieved from 180 to 365 cm depth. Down to circa 180 cm depth, dark brown peat with a very loose structure and coarse plant material is present, which covers relative compacted homogenous peat down to approximately 230 cm. Beneath this peat, there is a grey-brown (lightly greenish) humic clay layer of almost 20 cm thickness. The remaining part of the core consists of laminated light-grey sandy humic clay containing a couple of sand layers (1 - 4 cm thick). The laminated profile is relative fine, with laminations in the order of 0.1 to 1 cm thick. Near the bottom of the core, at 360 cm depth, there is a sharp transition towards fine- to medium-grained sand deposits of Rhine-characteristic heterogenous mineralogy and color. The total core is poor in calcium-carbonate and the loss-on-ignition results are showed in figure 29.

Klein Entenhorst (KE)

The *Klein Entenhorst* site is located in the southwest of the study area in a palaeomeander near the foot of the ice-pushed ridge (figure 18 and 25). Similar to the *Marienbaum* coring site, the soil is very humid as a consequence of upwelling groundwater, favoring an alder-dominated swamp vegetation today. The peaty channel fill has been partly excavated. Therefore, coring was performed on a narrow longitudinal embankment where the total channel fill had been spared. A segment of 40 cm was sampled which was retrieved from 300 to 340 cm depth. The total core (from surface to 340 cm depth) appeared to contain a low calcium-carbonate content but to be highly organic as showed by the loss-on-ignition results in figure 30. Coarse-grained gravelly sand deposits are present at the bottom of the core, directly covered by approximately 4 cm's of very fine-structured and laminated gyttja. The transition between both sediment types is very sharp, and the uppermost centimeter of sand is coloured by iron-oxide (figure 26). From the gyttja upwards, there is a gradual trend from fine gyttja-like sediments towards peaty material with an increasing abundance of coarse plant remains. At circa 313 and 307 cm depth, there is a transition towards sedge-dominated peat and partly oxidised black peat, respectively.

Figure 26 Photograph of the lowermost part of the *Klein Entenhorst* core (343-294 cm depth), showing the sharp transition between gravelly sand (with iron-oxide top layer) and fine laminated gyttja. Base of the channel to the right.

Heesenhof (HH)

A couple of years before the onset of present study, a core was taken from the abandoned meandering channel to the west of Xanten, near the small village called *Heesenhof* (figure 18 and 25). The channel system is approximately 150-200 m wide as appears from the DEM and makes part of meandering level III. The coring site is located near the foot of the ice-pushed ridge. The pollen content had been analysed by students (Reconstructing Quaternary Environments course 2008, Utrecht University, under supervision of Dr W.Z. Hoek) and shows an early Holocene signal followed by a hiatus and a middle-late Holocene signal. A core length of approximately 125 cm was analysed: from 110 to 235 cm depth. The maximum coring depth was determined by the presence of gravely coarse-grained sand, directly covered by 20 cm of (sandy) clay. Between approximately 215 and 140 cm, the core consists of calcareous gyttja, incorporating at least one clay layer of ~2 cm. From 140 cm onwards, the core comprises peat. No LOI data are available for the *Heesenhof* core. A cross-section through the channel system is shown in figure 24.

Vosse Kuhle (VK)

Compared to the coring sites described so far, the *Vosse Kuhle* site is located more to the central part of the Rhine valley, and makes part of cross-section I (figure 20). A core was retrieved from a well-developed meandering channel system which is lightly incised in the surrounding river terrace (figure 18 and 25). Today, this channel system mainly supports pastureland. A core of almost one meter was taken from 290 to 380 cm depth. Coring was finished at 380 cm because of the presence of gravel sediments, from which a very sharp transition towards the organic core sediments can be concluded. The core contains greyish laminated clayey siderite gyttja deposits with loss-on-ignition values in between 0 and 40% (figure 32). Material with a relative high siderite (iron-carbonate) content appear yellow, as is shown in figure 27. Because the gyttja is rich in calcium-carbonate, these LOI values reflect the total weight percentage of organic carbon and calcium-carbonate. Alterations in organic and siderite content and the sandiness of the material, result in a laminated profile (figure 27). At several levels, small whitish shells (~1-2 mm) or shell dust were observed.

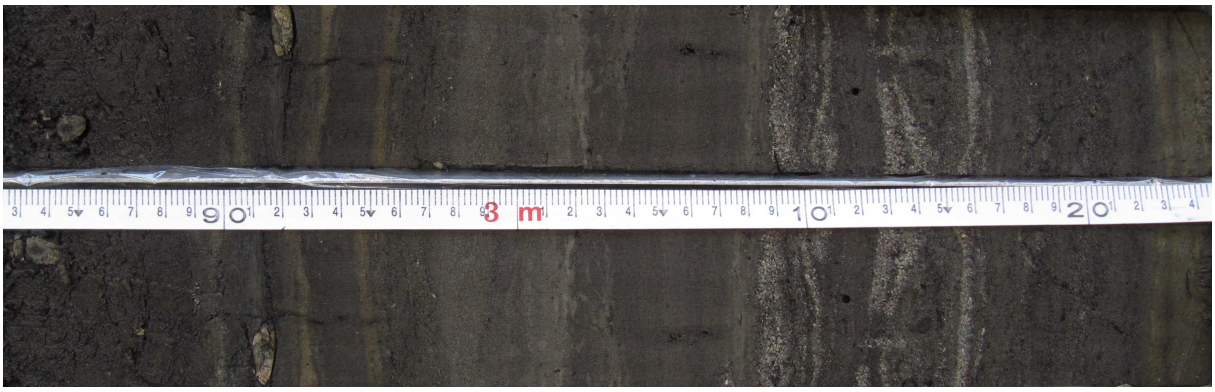

Figure 27 Photograph of a part of the *Vosse Kuhle* core (283-324 cm depth), showing laminated siderite gyttja. Base of the channel to the right.

Sanders Kath (SK)

The *Sanders Kath* core was taken from a palaeomeander which is likely to have been part of a former Rhine branch following the Oude-IJssel-Rhine course. Due to the formation of dunes along the eastern banks of a younger channel system, channel fills are expected to be (partly) preserved beneath aeolian sand deposits (figure 18 and 25). A core of approximately one meter was taken from beneath the dune deposits. However, only between 105 and 125 cm depth, there are peat sediments characterised by a high organic carbon content, as is showed by the LOI results in figure 33. The peat is located on top of lightly humic sandy loam. Upwards from the peat layer, the core comprises (very) sandy peat deposits. All sediments have a low calcium-carbonate content. Pollen samples were only taken from the peat layer and the underlying loamy deposits.

7.3 Pollen zonation and biostratigraphical correlation

In the Lateglacial and Holocene diagrams from the lower Rhine valley, regional pollen assemblage zones (PAZ) were distinguished on the assumption that they reflect ecological events which take place synchronously over a wide area, e.g. the total Netherlands and the adjacent lowland area across its borders. PAZ definitions are based on major shifts in the dominant regional pollen types and the AP/NAP ratio. In order to enable biostratigraphical correlation with other diagrams from the region and for indirect dating of the sediments, definition, description and coding of the biozones has been adapted to existing regional pollen zonation schemes; Lateglacial and early Holocene PAZs are in accordance with the regional pollen zonation scheme provided by Hoek (1997a,b) and the middle and late Holocene PAZs with a Holocene zonation scheme for northwest Germany (Van Geel et al. 1980 after Overbeck 1975). The *Kehrum Torfkuhle* pollen record has been compared to long-term Quaternary vegetation signals provided by De Mulder et al. (2003). The biostratigraphical correlation with these (reference) studies is discussed in section 7.4. However, in order to improve the readability and coherence of this thesis, the inferred time periods make already part of the subtitles in this section. The abbreviation 'LRV' within the biozone codes, stands for Lower Rhine Valley. The pollen diagrams are showed in figures 28-33.

Zone LRV-0: Eemian

Zone LRV-0a is characterized by high *Alnus* pollen percentages (20-40%) and a slightly lower relative abundance of *Pinus* (20-30%) and *Betula* (~20%). *Corylus* and *Picea* percentages vary between 0 and 5%, and *Quercus* between 0 and 10%. Sporadically, *Juniperus*, Ericales, *Populus*, *Fraxinus* and *Carpinus* pollen are recorded. Gramineae percentages are in the order of 15-30%. An abrupt decrease in *Alnus* down to only a few percent and abrupt increase in pine up to 60% marks the LRV-0a/0b boundary. It is accompanied by an increase in *Picea* pollen percentages up to 5-10%. With a similar magnitude, pine and *Picea* percentages decrease again across the LRV-0b/0c contact. Because the pollen assemblages

correspond best to the Eemian pollen signal provided by De Mulder et al. (2003), the *Kehrum Torfkuhle* peat layer is presumed to have been formed during the Eemian interglacial (in line with Klosterman 1989). Because this study focuses on the Weichselian-early Holocene interval, this pollen record will not be considered in more detail.

Subzone LRV-2b: Pine-phase of the Allerød *interstadial*

Subzones LRV-2b to 4a are recorded in the *Marienbaum* core (figure 29). Subzone 2b is characterized by overall high *Pinus* percentages. A distinction is made between subzone LRV-2b1 and LRV2b2. Subzone 2b1 shows *Pinus* values of at least 45% and maximum values of 60-70% are reached. *Pinus* is almost totally responsible for AP percentages varying between 50 and 80% because the pollen percentages of *Betula*, *Salix* and *Populus* together do not exceed 10%. The oldest part shows relative stable values for *Empetrum* (1-3%), which disappears from the record towards the top of this subzone where *Juniperus* becomes relative more abundant (up to 5-10%). Especially the uppermost half of subzone 2b1 shows an admixture of pollen produced by thermophilous tree species (*Picea, Alnus, Corylus, Quercus, Fraxinus, Carpinus* and *Ulmus*). The discontinuous character of these pollen signals in combination with the high clastic fraction of the sediment point towards reworked pollen. The effect of redeposition of sediment is also clearly visible near the upper contact of this subzone and within the lowest couple of centimeters of the record, where Gramineae peaks and minimum pine percentages coincide with the presence of sand. Another characteristic feature for subzone 2ba is the high diversity in upland herbs. Pollen taxa which are almost continuously recorded in considerable percentages (~5%) are *Artemisia,* Ranunculaceae, *Aster* type, Compositae liguliflorae and *Rumex*.

Subzone LRV-2b2 is characterized by slightly lowered pine percentages compared to subzone 2b1, in the order of 40%. Due to elevated birch and willow percentages, up to 30 and 10%, respectively, AP reaches maximum values of almost 80%. This subzone shows low *Juniperus* pollen percentages, even down to zero, and relative high *Empetrum* percentages up to 5%. Similar to subzone 2b1, reworked pollen of thermophilous tree species is recorded. Since this subzone coincides with an organic carbon content of almost zero, it cannot be excluded that the distinction between subzone 2b1 and 2b2 is the consequence of a different input of sediment and admixed pollen from upstream. Many of the upland herb taxa of subzone 2b1 are not recorded in subzone 2b2.

Zone LRV-3: Younger Dryas stadial

The boundary between zone 2b and zone 3 as a whole, is based on a pronounced decrease in pine pollen in contrast to NAP, grasses, birch, juniper and *Artemisia* which become relative more abundant. In whole zone 3 reworked pollen is recorded, however, in smaller percentages compared to zone 2b. Subzone 3a is characterized by relative stable NAP and grass pollen percentages around 40-45% and 30%, respectively.

Birch gradually increases from 30 to 40% and juniper shows percentages around 10% on average. After a temporal disappearance of *Populus* in subzone 2b2, it is recorded again in small amounts (1-2%). In subzone 3a, *Calluna* enters the record and *Empetrum* starts to increase again, however, together producing only 3-5% of the pollen sum. Very characteristic for zone 3 as a whole is relative high and stable *Artemisia* percentages in the order of 10-15%. Furthermore, taxa such as *Aster* type, *Saxifraga, Sanguisorba*, Cruciferae and *Helianthemum* are well-represented.

Calluna and *Empetrum* become more abundant in subzone 3b, in which they are responsible for 5-10% of the total pollen sum. Near the upper contact of subzone 3b, *Artemisia* and NAP pollen reach maximum values of circa 15% and 60%. *Betula* gradually decreases from 30 towards 15%. Juniper pollen remains relative stable at 8-12%. The upland herb vegetation incorporates similar taxa as in subzone 3a, however, species belonging to the families Ranunculaceae and Compositae liguliflorae and to the genera *Rumex*, *Potentilla, Rosaceae* and *Sanguisorba*, become relatively more abundant.

Zone LRV-4: Preboreal (together with LRV-5)

The lower contact of zone 4 is based on a strong increase in *Betula* pollen percentages from circa 20 towards 70%. Within subzone LRV-4a pine percentages further decrease and stabilize around 5%. Together with the strong rise in birch pollen, AP values increase up to 75 or even 80%. *Juniperus* and *Populus* percentages vary between zero and 10% and 3%, respectively. Gramineae, *Artemisia* and Ranunculaceae pollen gradually decrease. Overall, the diversity in upland herb taxa declines rapidly across the zone 3-4 transition.

High birch pollen percentages come to an end at the transition from subzone 4a to 4b. Subzone 4b is recorded in the *Marienbaum* and *Klein Entenhorst* records and is characterized by a low in birch pollen and related AP, both in the range of 20-40% (figure 29 and 30). In the *Marienbaum* record the lower boundary of subzone 4b coincides with a strong increase in LOI values from circa 40 up to 80% and for the *Klein Entenhorst* site it coincides with first build-up of gyttja deposits. A peak in grass pollen is recorded, reaching a maximum of circa 50-65%. *Juniperus* values are negligible low in the *Marienbaum* core but are around 5% for *Klein Entenhorst*. Small peaks of circa 1-2% occur near the top of subzone 4b for *Salix*, *Populus* and *Empetrum*. No upland herb taxa are recorded except from Cruciferae, *Rumex*, *Anthemis* type, *Cirsium, Plantago*, *Artemisia* and Ranunculaceae.

The lower contact of subzone 4c is marked by an increase in birch pollen percentages up to more than 40%, at the expense of especially grasses which decrease towards circa 20% or sometimes even become absent at the *Heesenhof* site. This subzone is especially well recorded at the *Klein Entenhorst* and *Heesenhof* sites where it comprises respectively ~30 and 65 cm of core length. Birch and pine percentages fluctuate between 40 and 70% and between 15 and 35%, respectively (figure 30 and 31). *Salix* and *Juniperus* are together respectively responsible for up to 10% of the pollen sum. Near the end of subzone

4c, some poplar pollen is recorded at the *Klein Entenhorst* site (~1%). Similar herbs as mentioned for subzone 4b are recorded in subzone 4c.

Zone LRV-5: Pine-phase of the Late Preboreal

Except from its lower contact, subzone LRV-5 is not represented in the LRV cores provided by present study. The lower boundary is marked by a rapid increase in pine pollen up to more than 60% at the expense of birch.

Zone LRV-6: Boreal (together with LRV-7)

Only the upper contact of zone LRV-6 is presumed to be incorporated in the lowermost part of the *Vosse Kuhle* record (figure 32). Van Geel et al. (1980) distinguish a sixth zone (VI $_{NWD}$; NWD = northwest Germany) which is characterized by high pine percentages and the appearance of *Corylus* pollen in the record. According to the same study, *Corylus* gains high values during the following biozone (VII $_{NWD}$). Near the bottom of the *Vosse Kuhle* core there is a strong increase in *Corylus* from circa 15 towards 45%, which is thought to resemble more or less the just described zone transition following Van Geel et al. (1980).

Zone LRV-7: Boreal (together with LRV-6)

Zone LRV-7 is well recorded at the *Vosse Kuhle* and *Hohe Leye* sites (figure 32). A core from the latter location is palynologically analysed by Van Munster (MSc thesis, in prep.). Whole zone 7 is characterized by relative high *Corylus* percentages in the range of 40-60% and stable AP percentages around 90%. Sporadically, very low percentages (< 1%) of *Populus*, *Juniperus*, *Salix* and Ericaceae pollen is recorded. Zone 7 is the first zone (except from zone LRV-0) that incorporates relative high values in combination with continuous signals for thermophilous tree species. For subzone 7a this group is represented by *Quercus*, *Ulmus*, *Fraxinus* and *Alnus*. Pine and birch pollen percentages are relative low compared to older zones, namely in the order of 15 and 10%, respectively. In the lower part of subzone 7a several upland herb species are recorded (e.g. *Galium*, *Rumex*, *Aster* type, Compositae liguliflorae, *Artemisia*, *Potentilla* and Caryophyllaceae) but this group of taxa almost totally disappears from the record towards the top of subzone 7b. The regional pollen assemblage of subzone 7b is very similar to the one of subzone 7a despite the fact that *Tilia* enters the record.

Zone LRV-8: Atlantic

Zone LRV-8 is recorded in the channel infill of the small palaeomeander called *Sanders Kath* (figure 33). It shows high values for *Tilia*, *Alnus* and *Pinus* of approximately 20, 25 and 30% on average, respectively,

resulting in continuous high AP values of circa 90%. Percentages lower than ~3% are observed for *Betula*, *Ulmus*, *Salix*, *Picea*, *Populus*, *Quercus*, *Juniperus* and Ericaceae. *Corylus* shows stable values in the order of 10 to 15%. Another very characteristic feature for zone 8 is the total lack of pollen produced by upland herbs taxa.

Zones LRV-9 and LRV-10: Subboreal and Subatlantic

The contact between LRV-8 and 9 is marked by a decline in relative abundance of *Tilia* and *Ulmus* pollen and a enormous increase in *Alnus* up to 50% of the pollen sum. A decline of *Tilia* and *Ulmus* pollen is described by Van Geel et al. (1980) for the transition between zone VIII_{NDW} and IX_{NDW}. However, because *Alnus* trees are restricted to areas with moisture (often seepage) conditions, fluctuations in *Alnus* pollen percentages do primarily reflect changes in its local abundance rather than alterations of the regional (upland) vegetation composition. This implies that the sharp decline in *Tilia* might be indirectly the consequence of the enormous rise in *Alnus* pollen because a relative abundance scale is used. Therefore, the correlation of the LRV-8-9 boundary with the zone boundary defined by Van Geel et al. (1980) is ambiguous. The enormous increase in upland herbs diversity reflects the disturbance of the vegetation by man. The impact of farming activity becomes even more visible in zone LRV-10, with a sharp increase in grass and *Plantago* pollen.

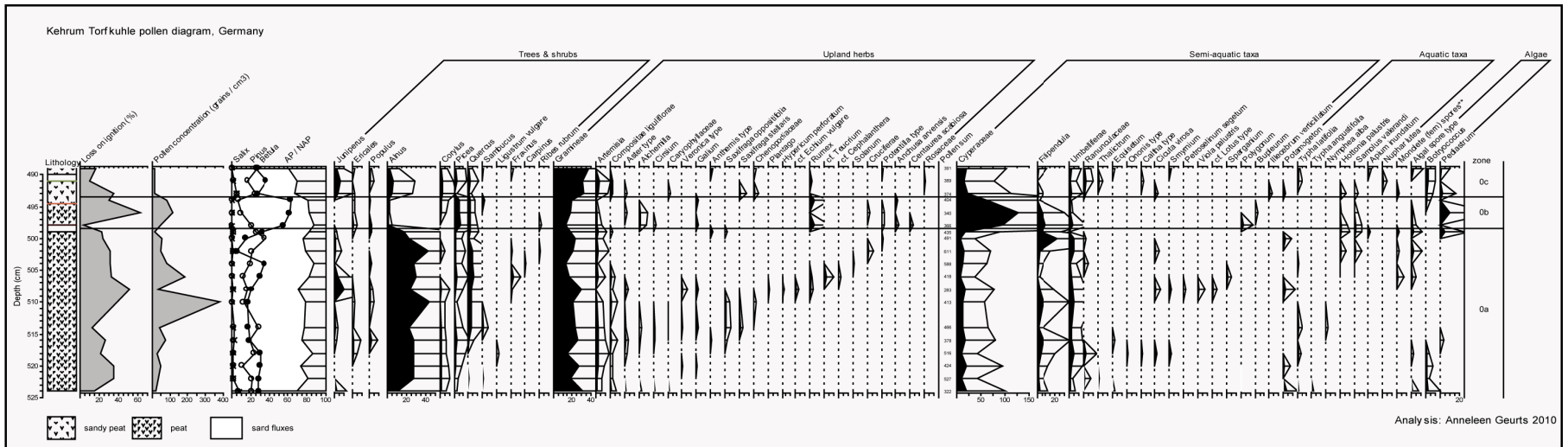

Figure 28: Kehrum Torfkuhle pollen diagram (abbreviation: KT) from a peat layer at 5 m depth underneath fluvial deposits making up meandering level III. For core location, see figures 18 and 25.

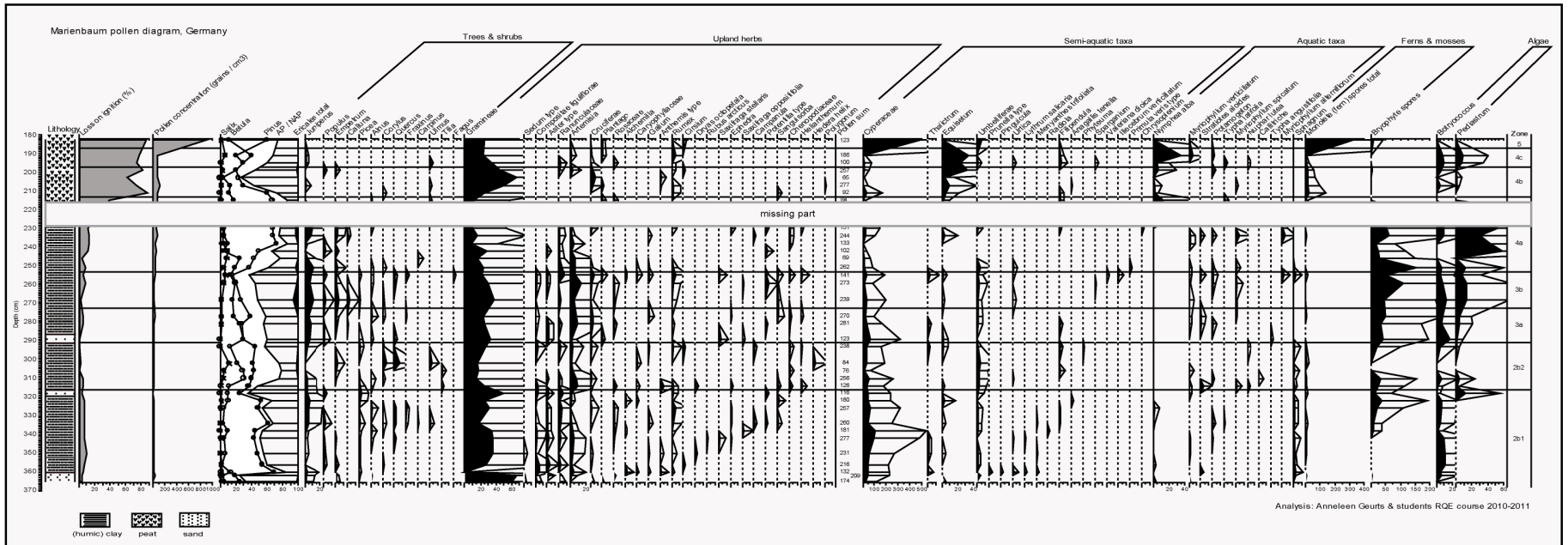

Figure 29: Marienbaum pollen diagram (abbreviation: MB) from an abandoned channel system making part of meandering level III. For core location, see figures 18 and 25.

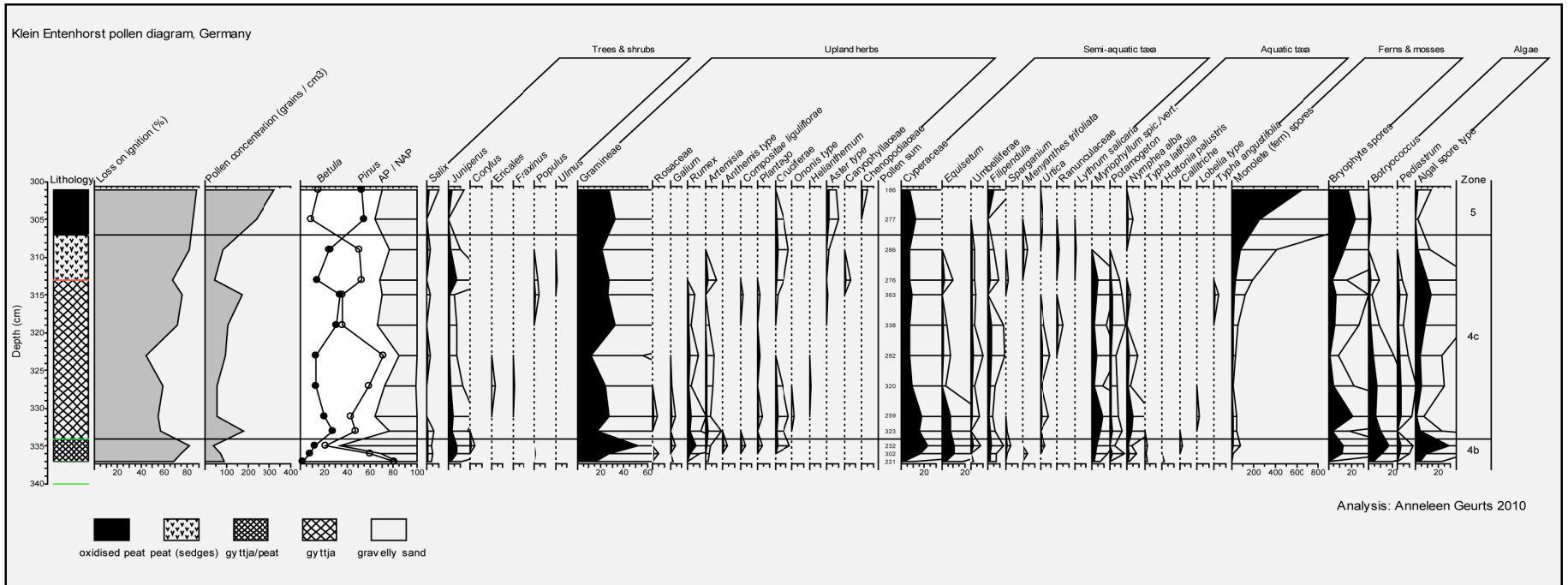

Figure 30: Klein Entenhorst pollen diagram (abbreviation: KE) from an abandoned channel system making part of meandering level III. For core location, see figures 18 and 25.

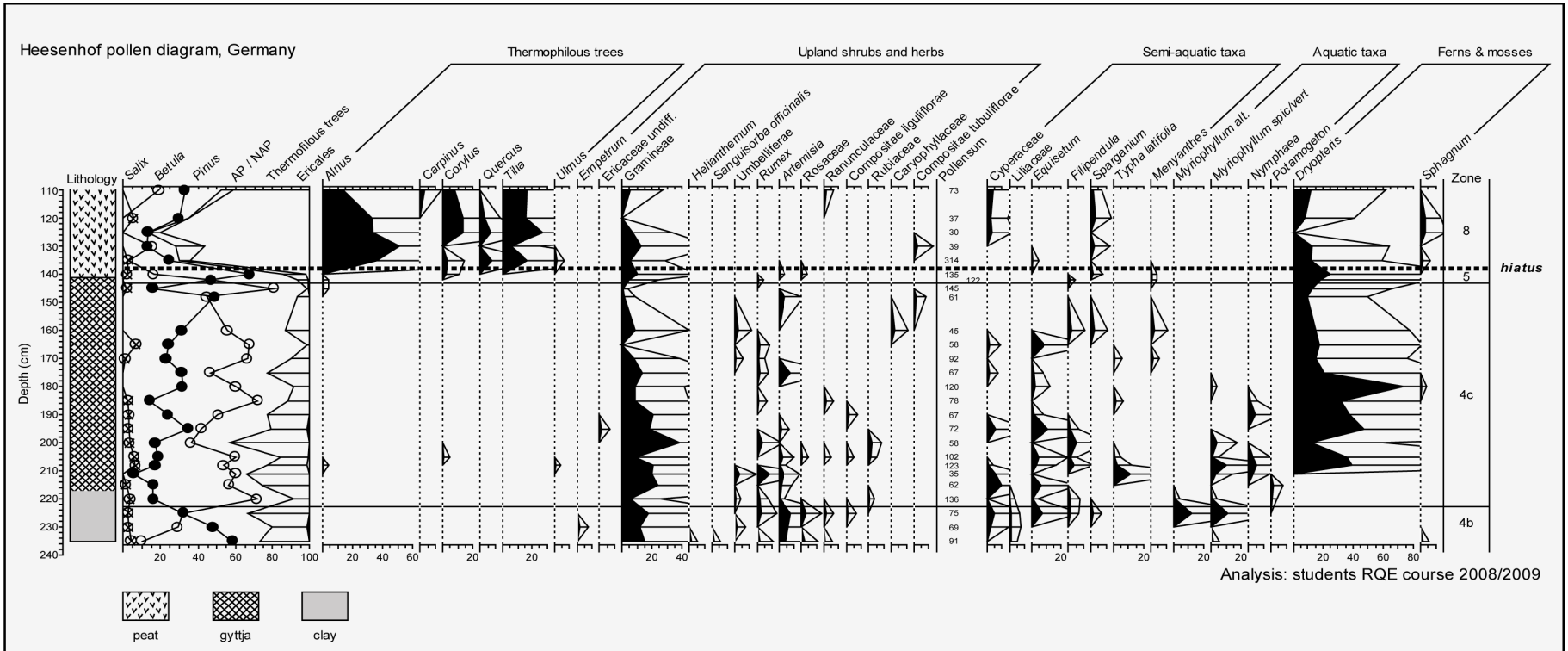

Figure 31: Heesenhof pollen diagram (abbreviation: HH) from an abandoned channel system making part of meandering level III. For core location, see figures 18 and 25.

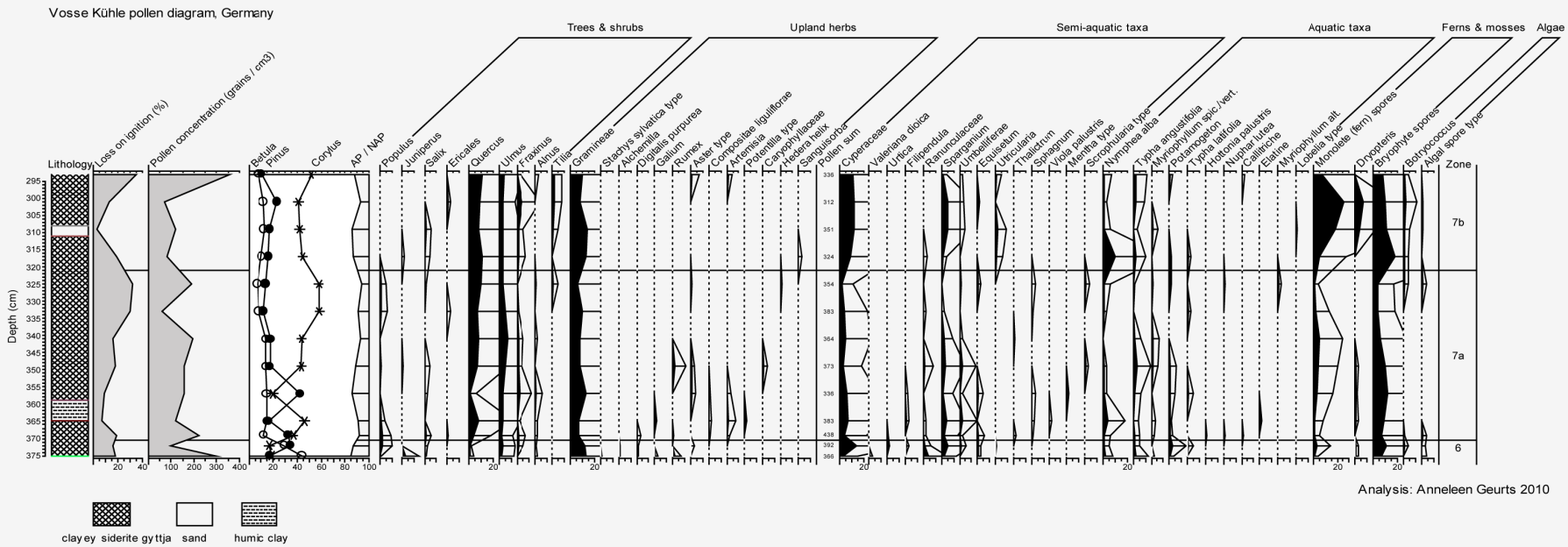

Figure 32: Vosse Kuhle pollen diagram (abbreviation: VK) from an abandoned channel system of early Holocene age which has a slightly incised position within braiding level IV. For core location, see figures 18 and 25.

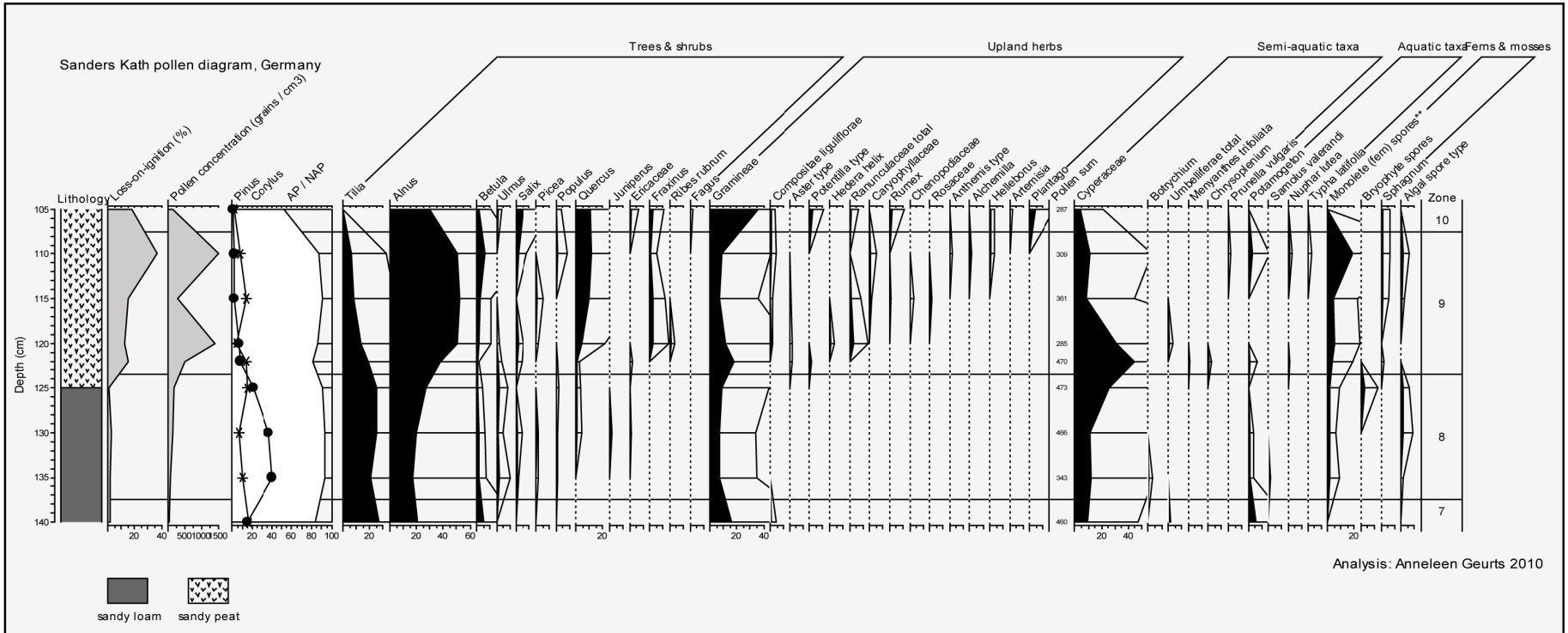

Figure 33: Sanders Kath pollen diagram (abbreviation: SK) from an abandoned channel system probably formed by the meandering but multi-channeled Rhine responsible for meandering level III. For core location, see figures 18 and 25.

7.4 Bio- and time-stratigraphic correlation

The pollen diagrams showing a Lateglacial or early Holocene signal could be correlated with the well-dated regional zonation scheme for The Netherlands (Hoek 1997a,b; Hoek 2001). These pollen diagrams are *Marienbaum*, *Klein Entenhorst* and *Heesenhof*, which show an overlap. The early-middle Holocene *Vosse Kuhle* diagram could be correlated with the less well-dated PAZs defined by Van Geel et al. (1980). Biostratigraphic correlation has resulted in a presumed chronostratigraphic framework which is showed in figure 34. It is emphasized that no strong conclusions can be drawn from this chronostratigraphy because of the lack of radiocarbon dates.

Subzone LRV-2b can be correlated with subzone 2b of the regional zonation scheme which is described as the *Pinus* phase of the Allerød, lasting from 13.2 to 13.0 kyr BP (Hoek 2001). According to Hoek (1997a,b) the beginning of this zone is characterized by a strong increase in pine pollen percentages to more than 20%. Despite a similar strong increase at the bottom of the *Marienbaum* core, this marked trend is not thought reflect onset of this biozone, because it is recorded in sand and is probably old reworked pollen. Therefore, based on the presumed chronostratigraphy, channel infilling of the *Marienbaum* system is expected to have started in the course of biozone 2b, somewhere in between 13.2 and 13.0 kyr BP.

Subzones LRV-3a and 3b can be correlated with subzones 3a and 3b as defined by Hoek (1997a,b), respectively. Together, they are considered to be equivalent to the Younger Dryas, lasting from 13.0 to 11.7 kyr BP (Hoek 2001). This is based on an increase in *Betula*, grass and herb pollen percentages at the expense of pine. The correlation of LRV-3b is based on slightly elevated values of Ericaceae pollen.

Subzones LRV-4a, -4b, -4c and -5 are considered to reflect similar and contemporaneous vegetation developments as described by Hoek (1997a,b) for biozones 4a, 4b, 4c and 5, respectively, together representing the Preboreal period of the early Holocene. The lower contact of subzone 4a is marked by a enormous expansion of tree birch, resulting in pollen percentages over 70% in the *Marienbaum* core. This first Holocene period of climatic amelioration is known as the Friesland phase (Van Geel et al. 1981; subzone 4a, 11.7-11.3 kyr BP; Hoek 2001). A temporal vegetational revertence is reflected by a minimum in birch pollen (5-20% in LRV-4b) which is considered to be equivalent to the so-called Rammelbeek phase (Van Geel et al. 1981; 11.3-11.2 kyr BP; Hoek 2001). *Betula* pollen percentages rise again in zone 4c and dominate over pine and non-arboreal pollen (11.2-10.7 kyr BP). Zone 5 comprises the pine phase of the Late Preboreal, during which pine dominates over birch (10.7-10.2 kyr BP).

Thermophilous tree species appear in the biostratigraphic record from zone LRV-6 onwards, starting with an expansion of *Corylus*, followed by *Quercus*, *Ulmus*, *Fraxinus* and *Alnus* (LRV-7a), *Tilia* (LRV-7b), *Picea* (LRV-8) and *Carpinus* (LRV-9). These biozones correlate with the Boreal and younger Holocene stages (after 10.2 kyr BP) (Hoek 1997a,b; Van Geel et al. 1980).

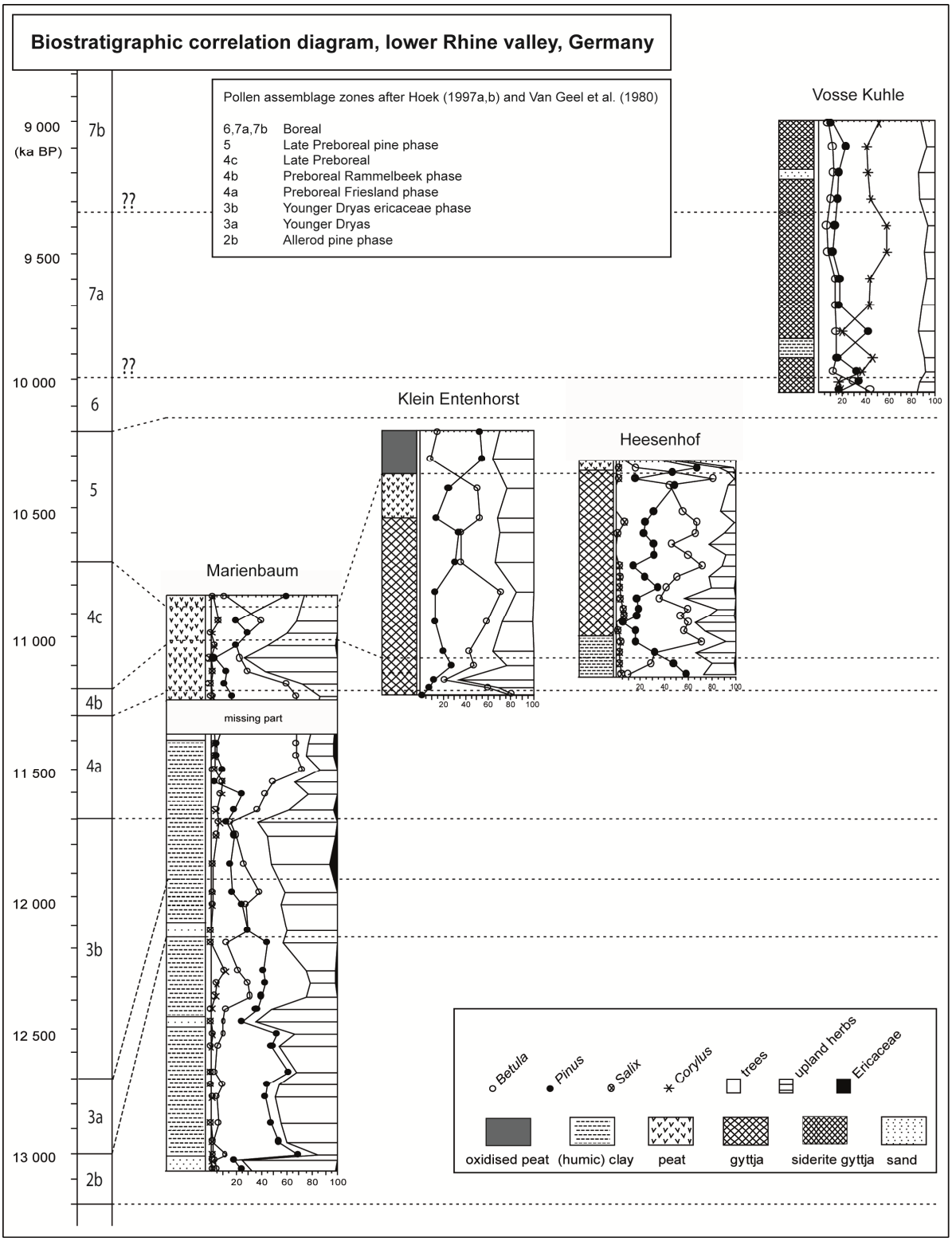

Figure 34 Synthesis of four Lateglacial – Early Holocene pollen diagrams from the lower Rhine valley.

7.5 Local vegetation succession and hydrology

Pollen assemblages produced by plants growing on the sampling site itself, are used for reconstructing former local vegetation composition (for limitations of this method, see section 5.8). Because it concerns taxa which are growing within or along the shores of the former lake, the local vegetation composition in turn provides information about the local palaeohydrology. In this section, local vegetation successions and hydrological developments are reconstructed for each core location, in line with the chronostratigraphic framework described in section 7.4. Ecological descriptions of the different plant taxa are primarily based on Weeda et al. (2003).

Marienbaum site

At the *Marienbaum* site, infilling started during the Allerød. By that time, pine predominated the pollen record, from which a more exact timing can be inferred; during the pine-phase of the Allerød which dates to ca. 13.2 – 13.0 kyr BP (Hoek 2001). The very loamy character of the sediments with sand intercalations indicates that the channel still became flooded on a regular basis. From the combination of very low LOI values and interstadial conditions of high-biological productivity can be concluded that the input of clastic fluvial sediment was high and a significant portion of pollen was added from upstream. However, the sharp transition in pollen assemblage composition across the boundary between the basic sand layer and overlying loam (~360 cm; figure 29), suggests a considerable increase in the local/upstream pollen ratio. Therefore, the composition of the semi- and full aquatic pollen taxa is assumed to have been predominantly locally produced and will be described here.

Significant flooding activity near the end of the Allerød (LRV-2b) probably resulted in a very sparse aquatic vegetation, dominanted by *Myriophyllum verticillatum* and *Potamogeton* in the pollen record. In much lower percentages and less continuous, *M. spicatum*, *Nymphea alba*, *Nuphar lutea*, *Stratiotes aloides*, *Typha latifolia* and *Callitriche* are recorded. This pollen assemblage indicates open water conditions with water depths of at least 1-2 m (Hannon and Gaillard 1997) and eutrophic to mesotrophic conditions, resulting from the combined effect of the nutrient-rich fluvial environment and the location within the seepage zone of the ice-pushed ridge. In the less deeper parts of the residual channel, a sedge-swamp vegetation is present, dominated by horsetails, ferns and taxa belonging to the (cyper)grasses and Umbelliferae family. The phase of channel abandonment might have been accompanied by an overall decrease in water table, causing an increasing part of the shore zone to become aerated. Together with strong seasonal water level fluctuations, a net release of nitrogen in the humic topsoil due to aeration likely favored *Urtica* and *Lythrum salicaria*, and later on also *Filipendula ulmaria.* Moreover, the shores were peaty as a consequence of *Sphagnum* growth, favoring *Menyanthes trifoliata* in the submerged part of the shore zone and *Thalictrum flavum* more upland.

The Younger Dryas biozone (LRV-3a,3b) shows lower reworked pollen percentages (thermophillous tree pollen) pointing towards a still ongoing decline in fluvial activity. More stagnant water conditions probably favored algae growth (*Pediastrum, Botryococcus*) and a more abundant macrophyte vegetation, including *Myriophyllum verticillatum, M. spicatum*, *Stratiotes aloides*, *Typha angustifolia* and *Potamogeton*. This assemblage indicates a water depth of at least 1 m (Hannon and Gaillard 1997). Under severe environmental conditions, *Sphagnum* was not able to reproduce and totally no spores ended up in the sediments. Overall, biodiversity in the shore zone decreased and only *Thalictrum* flavum, *Urtica* and species belonging to the horsetails, (cyper-)grasses and Umbelliferae sustained. One new arriving species was *Sparganium* near the end of the Younger Dryas. Similar aquatic and semi-aquatic pollen assemblages are recorded in the first phase of the Holocene (LRV-4a), except from the settlement of *Valeriana dioica, Illecebrum verticillatum, Primula veris* type in the riparian zone. One marked difference is the (probably climatically-induced) strongly enhanced algae productivity reflected by high *Pediastrum* percentages. The upper contact of the zone represents the transition from the Friesland phase towards the Rammelbeek phase (LRV-4a -> 4b) and is accompanied by a change from humic loamy sediments towards peat. This sediment compositional change is believed to be related to strong early Holocene river incision and the end of flood activity in the *Marienbaum* residual channel and an overall increase in biomass productivity. The latter results from the combined effect of climate amelioration and a reduction in water depth favoring a riparian vegetation type producing a larger volume of organic material per unit of time. As a consequence of seepage, however, local water level remained high enough to support macrophyte vegetation dominated by *Nymphea alba* during the remaining part of the Preboreal.

Klein Entenhorst site

In contrast to the *Marienbaum* site, the *Klein Entenhorst* channel became abandoned abruptly near the beginning of the Preboreal during the Rammelbeek phase (zone LRV-4b). This is reflected by a sharp transition from sand towards highly organic gyttja. During the Preboreal, a hydroseral succession took place from predominantly open water conditions and gyttja formation towards a swamp environment with peat formation. The is additionally showed by a gradual decline in macrophyte abundance consisting of *Myriophyllum spicatum/verticillatum, Potamogeton* and *Nymphea alba* and a gradual disappearance of algae. On the contrary, mosses and ferns strongly expand. The riparian vegetation composition remains relative constant, comprising cypergrasses, Umbelliferae, *Filipendula*, *Urtica* and Ranunculaceae. Probably related to the build-up of organic sediments, horsetails gradually diminish in relative abundance.

Heesenhof site

Similar to the *Klein Entenhorst* site, the *Heesenhof* meandering channel became totally abandoned in the very early beginning of the Holocene, namely during the Rammelbeek phase (zone LRV-4b). Also a very similar hydroseral succession is registered, showing a gradual decline in aquatic taxa percentages

including *Nymphea alba*, *Potamogeton, Myriophyllum alterniflorum and M. verticillatum/spicatum* and a relative constant shore vegetation comprising *Filipendula*, *Typha latifolia*, (cyper-)grasses and ferns. Maximum water depth was at least 1 m (Hannon and Gaillard 1997). *Sparganium, Menyanthes trifoliata* and *Sphagnum* become more abundant in the course of the Preboreal, in contrast to *Equisetum* which gradually disappears out of the record.

Vosse Kuhle site

In contrast to the three channel systems just described, the *Vosse Kuhle* channel is located more near the centre of the LRV. It is thought to have been originally formed during the Younger Dryas as part of a braidplain IV, but was deepened and altered into a slightly meandering system during the early stage of the Holocene. Based on the biostratigraphic correlation diagram in figure 34, this transition took place during the first ca. 1500-1700 yrs of the Holocene. Around the onset of the Boreal continues fluvial activity stopped and channel infilling started. For another period of approximately 800 yrs, however, the channel system remained flooded on a seasonal basis or kept on transporting a very small portion of the Rhine discharge. This is reflected by the clayey gyttja sediments with relative low LOI values and numerous thin sand intercalations. Despite a distance of 3 km from the ice-pushed ridge morphology, a high siderite concentration in the core evidences seepage conditions. Ferrous iron contained by upwelling groundwater must have entered the clayey gyttja sediments under oxygen-free conditions and successively precipitated with carbonate ions in the form of siderite ($FeCO_3$; Van der Perk 2006). The dominance of *Sparganium* along the shores of the channel indicates that the channel likely supported more or less continuously slowly moving water. *Typha angustifolia*, *T. latifolia* and *Potamogeton* also thrive well in this type of environment and *Nymphea alba* might have been restricted to sheltered water areas near the shores. Moreover, all these species/genera prefer a thick clayey and humic substratum at the bottom of the channel, what corresponds to the clayey gyttja making up the core. Moreover they point towards a mesotrophic to eutrophic and slightly basic to neutral conditions. Other semi-aquatic plants making part of the shore vegetation during the Boreal belong to the (cyper-)grasses, Ranunculaceae, Umbelliferae, *Equisetum*, *Thalictrum*, *Mentha* type and *Scrophularia* type families and genera. Zone LRV-7b is different from 7a based on the strong expansion of ferns, mosses, cypergrasses and *Sparganium*, and the settlement of *Utricularia* in the submerged channel. *Utricularia* is typical for the ongoing hydroseral succession towards shallower water depths, possibly accompanied by slightly acidification of the environment. Due to ongoing fluvial incision, also flooding activity ultimately stopped (roughly around 9 kyr BP, figure 34) as indicated by a replacement of the clayey gyttja by highly organic peat at ~290 cm depth (above the samples in the pollen diagram; figure 32).

Sanders Kath site

The small *Sanders Kath* channel system is thought to have been originally formed during the Allerød. Similar to the *Heesenhof* and *Klein Entenhorst* site, it contains loamy sediments which are believed to have been formed during Younger Dryas floods (figure 24). Despite the fact that it became totally abandoned around the onset of the Holocene, channel infilling did not start until the Atlantic. Channel depth was too small to support an open water environment as indicated by the lack of semi-aquatic and aquatic taxa (figure 33). Presumably, it was a sedge-swamp vegetation type that resisted at least until man settled in the area of study (LRV-9/10).

PART III

Interpretation, discussion and conclusions

8 Floodplain chronology and incision versus aggradation

8.1 Floodplain chronology

Despite the fact that this study has not generated any new OSL or radiocarbon dates yet, it has resulted in new insights into the fluvial history of the lower Rhine valley. In combination with already available dates, these new insights have led to a new chronostratigraphic framework for the study area, which forms an alternative to those provided by Klostermann (1989, 1992), Jansen (2001), Erkens et al. (2011) and Siebertz (1987). In this section, additional sources of evidence are listed which provide a more absolute floodplain chronology than the one presented in section 6.3. The chronology is depicted in figure 35 and is included in the geological-geomorphological map in appendix I.

- The position of the fluvial deposits within a Saalian maximum glaciation glacial morphology testifies an age younger than the moment of maximum glaciation for all fluvial deposits and local deposits in between or on top of them.

- The oldest fluvial floodplain in the lower Rhine valley which is present at the surface is formed by a braiding river system (braidplain I, section 6.3). A buried Eemian distal floodbasin peat layer at more than four meter depth (relative to the top of braidplain I), indicates that braidplain I was formed later in time, likely during the Weichselian. Moreover, the wide spatial distribution and thickness of the deposits point towards a laterally eroding and vertically aggrading river system which has been active under glacial climatic conditions. Widening of the braidplain in time in turn explains the non-preservation of a late Saalian deglaciation terrace.

- From several radiocarbon and OSL dates it can be concluded that the youngest meandering channel belt in the valley is of Holocene age (Erkens et al. 2011). From the morphological dimensions of the oldest dated Holocene pointbar sequences in the LRV can be concluded that strong lateral migration and pointbar development started at least around 10 ka BP (pointbar at 13-14 km in cross-section V in appendix III). Around 9.0 ka BP, the lower Rhine became a single thread river after abandonment of the last secondary channel system (Erkens et al. 2011, pointbar at 3 km in cross-section IV in appendix III). This means that floodplain levels I to IV all have been formed at least before ~10 ka BP, and date to the Weichselian Pleniglacial, Lateglacial or earliest Holocene.

- In the Rhine-Meuse delta, the Niers-Rhine valley, the Oude-IJssel valley and the lower Rhine valley upstream of the study area, a braiding-towards-meandering transition is recorded during the early Lateglacial (Kasse et al. 1995; Berendsen et al. 1995; Kasse et al. 2005; Schirmer 1990; Cohen et al. 2002; Busschers et al. 2007; Hijma et al. 2009; Verschuren 2007). According to

Erkens et al. (2011) a similar and contemporaneous transition occurred in the downstream part of the LRV. No meandering deposits of Pleniglacial age are known from the Dutch or German lowlands and such also seem very unlikely to have been formed given the periglacial environmental conditions of that time. Therefore, meandering level III is likely of Bølling-Allerød age. The younger and braiding type floodplain level IV must have been formed during the Younger Dryas. Because a Younger Dryas braidplain is indeed known to be present in the Rhine-Meuse delta (NT3 terrace or Terrace X), this geomorphology-based scenario seems plausible and is partly confirmed by Verschuren (MSc study, 2007).

- The following sources of dating evidence confirm the chronology inferred so far: First, an OSL date in a dune system on top of the well-developed pointbars near Xanten gives a late Younger Dryas age of 11.72 ± 0.52 kyr BP (cross-section V, appendix III). Second, a channel system making part of floodplain level IV (*Marienbaum* system) is nested within this meandering floodplain level and contains abundant pumice near its base. Third, a small palaeomeander making part of floodplain level III has become abandoned in the course of the Allerød (*Marienbaum* pollen diagram, figure 34). There are also a couple of OSL dates, palynological records and locations of pumice admixture which correspond with preliminary age estimates of Erkens et al. (2011), who attribute a Younger Dryas or early Holocene age to meandering level III in the region near Xanten. However, their chronology totally neglects the presence and morphological expression of braided level IV and alternative interpretations of these dates can be provided (chapter 10).

- An Allerød pollen spectrum within a soil profile on top of braided level II indicates this terrace to have abandoned before the Allerød (Jansen 2001). This palynological signal presumably indicates that braidplain II was active during the Late Pleniglacial, and possibly also during the early Lateglacial. This is confirmed by an OSL date of circa 15 ka BP from a site where these deposits occur near the surface (compare location of dating site OSL-2 in figure 18 with the location of cross-section III (figure 19)). Since a Late Pleniglacial braid belt surface, dating to ~15 ka BP is also present in the Rhine-Meuse delta downstream (Pons 1957; Berendsen and Stouthamer 2001; Berendsen et al. 1995), braidplain II is indeed likely formed during Late Pleniglacial times.

- Dates relevant for age determination of braidplain I are not available from the lower Rhine valley. The orientation of the channels into a northwest direction strongly suggests the Rhine to have flowed through the Oude-IJsel valley. The coverage with thick aeolian sand deposits and the presence of deep cryoturbation structures in the Rond-Montferland extension of this terrace, testify to a Middle Pleniglacial age for floodplain level I. This is confirmed by the OSL-dated timing of the avulsion of the main Rhine branch towards the Gelderse Poort area around 35-30 ka BP (Busschers et al. 2007), resulting in the final abandonment of braidplain I.

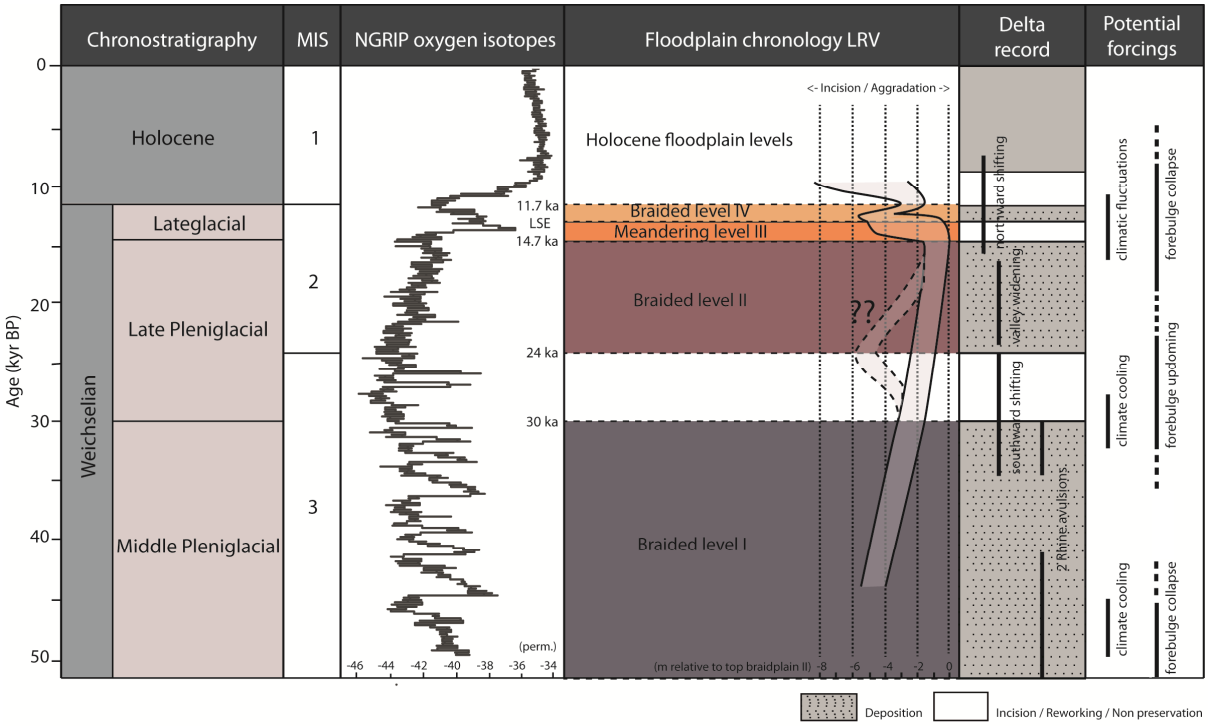

Figure 35 Weichselian floodplain chronology for the downstream part of the lower Rhine valley, based on correlation with dated sedimentary units in the central Netherlands and IJssel valley (Busschers et al. 2007; Hijma et al. 2009). An incision – versus – aggradation curve is added on the basis of top-of-sand elevations provided in section 8.2. Hereby, elevations are relative to the top of braidplain II. The left (or lowest) line reflects estimated channel base. Therefore, the thickness of the area in between them, shows estimated channel depth. The NGRIP oxygen isotope record is additionally showed as climatic signal (NGRIP Members 2004). Marine isotope record after Bassinot et al. (2004). Moreover, the timing of different potential external forcing factors is showed, as derived from Busschers (2008).

8.2 Incision versus aggradation

Elevation differences between the calculated trendlines in figure 23a, provide additional information on net channel incision or aggradation. Successively lower elevations of level II to IV point towards continuous (net) incision during the Lateglacial. Floodplain level III has a circa 1 m lower top elevation than level II in the central part of the LRV. Floodplain level IV is circa 0.5 m lower than level III in the west. Furthermore, the top of braidplain II is located at least 2 m higher than the top of level I, as measured in the central and eastern part of the LRV, respectively. Because lateral tilting is thought to have had altered floodplain elevations, these values cannot simply be used as a magnitude of incision or aggradation. Because the first two values (level II compared to III, and level III compared to IV) have been calculated within one lateral zone, these values are expected to be reliable estimates of actual incision. The third value was calculated on data from the central and eastern part of the LRV and is expected to have been influenced by differential tectonical movements.

In table 2, for three cross-sections (I, II of the present study and cross-section V of Erkens et al. 2011 (appendix III)), the elevation of the top of each river terrace is given. Braidplain I, however, is not included since this braidplain makes no part of these cross-sections. The elevations are average values, primarily based on the highest parts of each floodplain (top of pointbars or braid bars) if possible. Despite the fact

that these values were also included in figure 23a, this table is purely based on what is showed by the cross-sections itself, without any correlation or regression influences.

Table 2 This table shows the average elevation of the top of the in-channel sands making part of the different floodplains, as derived from three cross-sections (* cross-section V is constructed by Erkens et al. 2011, appendix III). The differences in top elevation provide an estimate for the amount of net incision that has taken place between two successive moments of terrace abandonment.

	Cross-section I		Cross-section II		Cross-section V*	
Terrace	Elevation (m +MSL)	Net incision (m)	Elevation (m +MSL)	Net incision (m)	Elevation (m +MSL)	Net incision (m)
Braidplain II	-		20.5		20.0	
				~ 0.5		~ 0.5
Meandering level III	18.5		20.0		19.5	
		~ 0.5		~ 1.0		
Braidplain IV	18.0		19.0		-	
						~ 2.0
Early Holocene	-		-		17.5	

Both table 2 and figure 23a are used for estimating the net amount of incision or aggradation that occurred for the area of study. Table 3 shows the results. The table shows the estimated differences in floodplain elevations between two floodplain level couples. The first line in the table, for instance, says that net aggradation took place in the order of 1–1.5 m, during the time-interval between the abandonment of braidplain I and the abandonment of braidplain II. These estimates are also graphically depicted in figure 35.

Table 3 Net incision or aggradation estimates for four floodplain couples (for explanation, see text).

Couple of terraces	Process	Net incision / aggradation (m)
I - II	Aggradation (floodplain level increase)	~ 1 – 1.5
II - III	Incision (floodplain lowering)	~ 0.5
III - IV	Incision (floodplain lowering)	~ 0.5 – 1.0
IV - EH	Incision (floodplain lowering)	~ 2.0

9 Discussion

9.1 Floodplain chronology

A striking result of the present study is the qualitative similarity of the late Weichselian-early Holocene fluvial history to the reconstruction of Erkens et al. (2011) and Siebertz (1987), who based this on a different chronological classification of the landforms. Starting from pre-existing classifications (conceptually after e.g. Klostermann 1992), Erkens et al. (2011) investigated the sequence of geomorphological responses to external forcing across the Weichselian-Holocene transitional period and the total Holocene. The discrepancy between both classifications, however, throws new light upon the timing of fluvial response and related time lags. This seems relevant for coming studies concerning quantification of fluvial response.

Considering the late Weichselian to earliest Holocene time-interval, the difference between both classifications is simple: 1) Erkens et al. (2011) do not seperately distinguish units of Bølling-Allerød age, 2) do divide braidplain II in fragments of Late Pleniglacial and Younger Dryas age, and 3) do classify both meandering level III and braiding level IV as early Holocene floodplain levels. Their classification is based on cross-sections across the LRV (figure 2), a number of OSL-dates and palynologically dated channel fills. Despite the fact that these sources of evidence seem to confirm the classification of Erkens et al. (2011) at first glance, some of them are believed to have been interpreted incorrectly. The alternative floodplain chronology is principally based on the geomorphology and sedimentary architecture of the valley and an assumed chronostratigraphic correlation to the well-dated and well-known fluvial history of the Rhine-Meuse delta in The Netherlands (chapter 8).

One OSL date used by Erkens et al. (2011) has been recovered from the top of the pointbar system to the west of Xanten. It provides an early Holocene age; 10,910 ± 570 yr BP (site OSL-1 in figure 18, table 1). In combination with an OSL date from an overlying dune system of 11,720 ± 520 yr BP, a kind of average age was deduced of late Younger Dryas/early Holocene for the total sequence. The potential presence of a small amount of pumice in the very top of the surrounding older terrace (G. Erkens, personal communication) seems to confirm this, so does an early Holocene channel fill of the palaeomeander (*Heesenhof* pollen diagram). In light of new sedimentological data, however, this argumentation seems rather unreliable. First of all, the OSL dates are stratigraphically in the wrong order, while more than 2 meters of sediment and even a very distinct sedimentary unit (1.5 m of loamy floodplain deposits) was in between them. Second, the presence of pumice granules was not certain. Even if it was present, it was only a small amount in the very top of the terrace deposits. Third, the palynologically dated channel fill only provides a minimum date for the moment of channel abandonment. Numerous studies show that the lower Rhine and many other northwest European rivers changed from braiding towards meandering in the beginning of the Bølling-Allerød interstadial (e.g. Berendsen et al. 1995; Berendsen and Stouthamer

2001; Verschuren 2007; Kasse et al. 2005; Janssens 2010; Graeves 2010; Busschers 2008; Schirmer 1995; Tebbens et al. 1999), a development which is also expected by Erkens et al. (2011) to have occurred in the area of study. However, the latter study suggests that the Bølling-Allerød channel belt has been relatively confined and has been totally eroded during the formation of the wide NT3 terrace (braidplain IV). Based on the just described couple of OSL dates, they therefore presume an early Holocene age for the meandering system to the west of Xanten, despite the fact that the late Younger Dryas age of the dune system points towards a pre-Younger Dryas age. Their classification implies that the secondary Rhine system must have fulfilled a transition from a braiding system towards an intense meandering and pointbar-developing system more or less instanteously following Younger Dryas-Holocene climate warming. The following sources of sedimentological and geomorphological evidence, however, make such a rapid development very unlikely: First, the basal part of the channel fill consists of thick sand and loam deposits what indicates that it took time to fulfill total abandonment of the channel, followed by a time interval with periodical reactivation (section 6.2). Second, reactivation seems to be confirmed by a number of small channels within the older meandering structure and the dead-ending diverging channels which are respectively indicative for much smaller volumes of water and an oppositely directed flood-flow within the system (figure 36). Third, the negligible amount of organic carbon contained by the basal loamy channel-fill deposits, makes it more likely that the first phase of channel-infilling occurred under periglacial Younger Dryas conditions, than under interglacial conditions during the early Holocene.

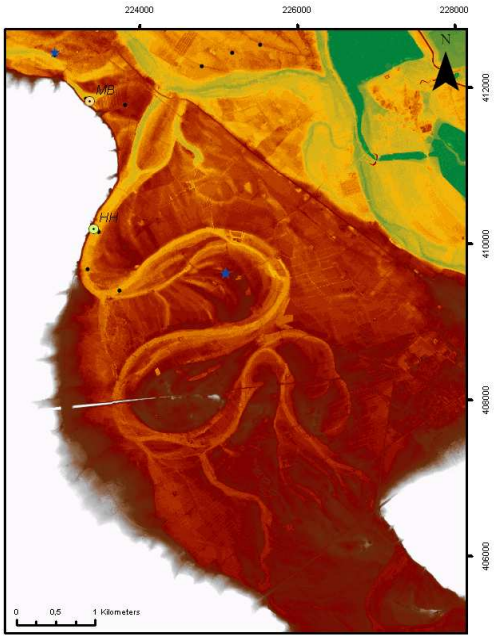

Figure 36 DEM showing the meandering channel system to the west of Xanten, which is thought to have been partly reshaped during Younger Dryas floods.

The pumice granules were observed in the uppermost sediments of the terrace remnant classified as NT3 by Erkens et al. 2011 and as braidplain II (=NT2) by present study (appendix III). A similar pumice discovery is reported by Siebertz (1987) on top of braidplain II near Kalkar (figure 11, section 3.2). This is not unexpected in line of the alternative chronology: By the time of the Laacher See eruption (~Allerød-Younger Dryas transition) the top of the active channel belt was located at only a slightly lower elevation than the top of abandoned braidplain II. Surge-like floods due to a sudden collapse of the pumice dam upstream would therefore have certainly flooded braidplain II within a distance of at least several kilometers. This flooding probably resulted in the deposition of pumice on top of the Late-Pleniglacial terrace.

An unambiguous explanation for the early Holocene OSL date of the pointbar system cannot be provided here. One option is that the pointbar and overlying dune system indeed date back to the late Younger

Dryas/early Holocene and were formed within a short time-interval: Reactivation of the lower end of the meandering system ('slackwater'-like) at the end of the Younger Dryas might have resulted in reworking of the top of the Allerød pointbar sediments. This might have been possible because by this time aggradation had resulted in a similar floodplain level height between braidplain IV and meandering level III. An option which seems to be more likely is that some type of OSL-dating error has resulted in a date which is too young. Potential sources of error might be incorrect clay- (possibly downward transport of clay from the overlying floodplain deposits) and moisture content (groundwater fluctuations) estimations, which can lead to a significant, underestimation of age.

It is hypothesised by Erkens et al. (2011) that the secondary meandering system to the west of Xanten (cross-section V from the same study, appendix III) is connected to the meandering system at 5 km in cross-section IV near Rheinberg (appendix III; figure 37). For both cross-sections a different sequence of floodplain levels is observed in the secondary channel system: Simply stated and ordered from high towards low top elevations, the sequence consists of braiding–meandering–braiding-meandering and of braiding–braiding–meandering–meandering in cross-section V and IV, respectively. Because the youngest meandering system in cross-section IV dates back to the early Holocene (based on 2 OSL dates) and pumice is present within (and not only on top of) the youngest braidplain, the interpreted ages by Erkens et al. (2011) seem plausible. This implies that the oldest meandering level is also of early Holocene age and no Bølling-Allerød level has been preserved near Rheinberg. Therefore, it is unknown whether or not an upstream continuation of the Xanten meanderbelt was originally present near Rheinberg. Another option is that the Bølling-Allerød meanderbelt diverges from the main valley near the villages Borth and Wallach (figure 37).

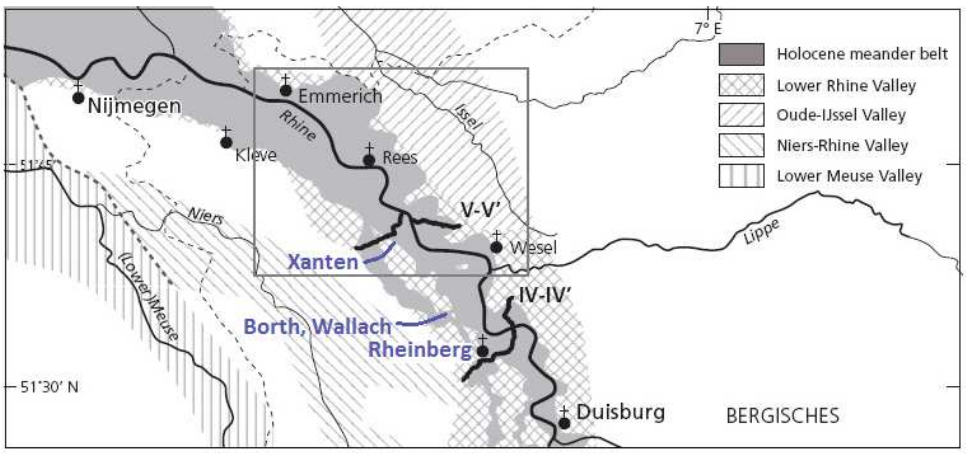

Figure 37 Fragment of figure 2, showing the Bølling-Allerød secondary system near Xanten, which is thought to have originally diverged from the trunk valley near Rheinberg or near Borth and Wallach.

Three OSL dates (L.A. Tebbens, unpublished) from the *Marienbaum* channel system (table 1; sample-site OSL-2 in figure 18) were taken from a relative high located floodplain level with a surface characterized by a typical 'bar-like' topography (figure 38). At the sample location, the sediment was logged during present field study and appeared to show a similar sedimentary sequence as present in the north-eastern part of

cross-section III (figure 19); fine-grained and very well-sorted sand on top of medium- to coarse-grained gravelly sand. The lowermost OSL sample (~15 ka) was taken from the coarse-grained unit which presumably was formed during the Late Pleniglacial. The upper two samples (~10 and 11 ka) were taken from the fine-grained facies which probably represents late Younger Dryas or earliest Holocene dune deposits. Unfortunately, the shallow occurrence of the Late Pleniglacial braidplain deposits does not exclude one of the floodplain classifications, because this braidplain might have been locally spared from intense erosion by either Lateglacial or early Holocene river systems. The development of the typical 'bar-like' topography remains uncertain: The morphology suggest a fluvial genesis, however, it is present at a higher level than the active Younger Dryas braidplain IV. Therefore, these 'bars' are thought to be dunes, which's morphology is thought to have been 'smoothed' during Younger Dryas floods.

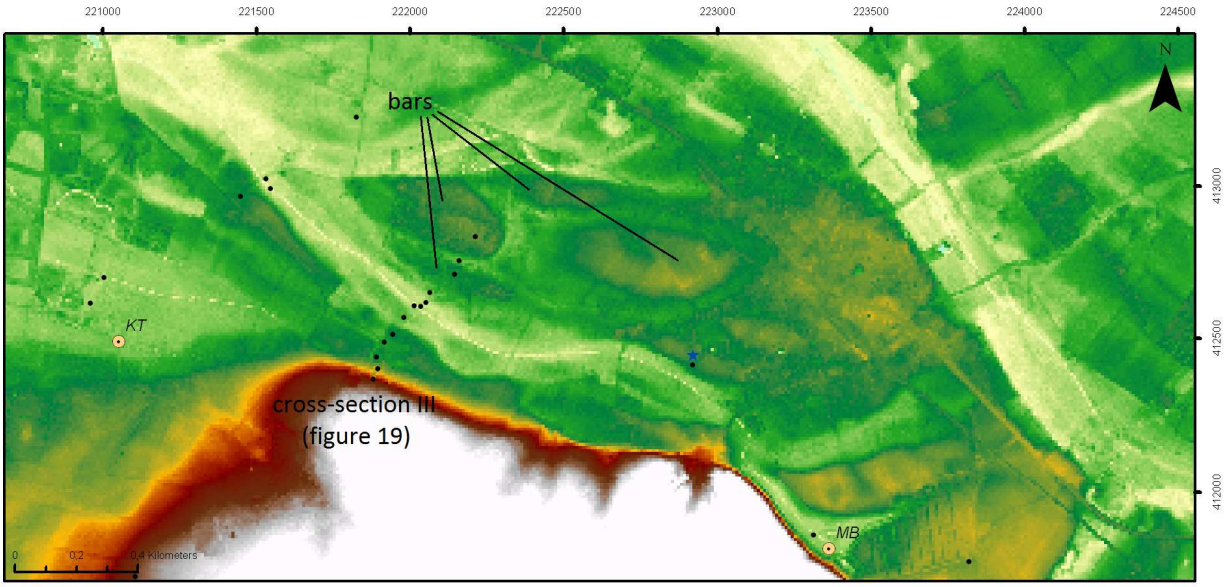

Figure 38 Digital elevation model showing the 'bar-like' topography bordering the *Marienbaum* system (see text).

Due to data limitations in the north-eastern (Oude IJssel) part of the study area, a large area is classified as braidplain I. Graeves (2010), Janssens (2010) and Verschuren (2007), however, make a distinction between two braiding terraces in this area and more downstream in the Oude-IJssel valley. This distinction goes back to Van der Meene (1977), who distinguishes two river tracks (A and B) flowing respectively in the eastern and western part of the Oude-IJssel valley. Because the fluvial geomorphology is concealed beneath a thick layer of coversand, it was not attempted to make a distinction between these units purely on the geological-geomorphological map. Both terraces have probably been formed during the Middle Pleniglacial (chapter 8).

In section 6.3, it is mentioned that an elevation difference in the order of 1-2 m exists between braidplains I and II to the west of Hamminkeln, along a cross-cut boundary of circa 5 km and level II having the heighest elevation. This feature might simply be the consequence of aggradation by the younger system resulting in the build-up of natural levees up to a higher elevation than braided level I. These levee systems (possibly helped by Pleniglacial aeolian deposits) successively prevent the river system to

develop a distributary towards the lower and older floodplain level I, despite the high dynamics of a braiding fluvial system. Differential glacio-isostatic uplift across the LRV is thought to have played an additional role in directing the river system to the west.

9.2 Relative influence of external and internal controls on fluvial evolution

For studying the relationship between external forcing and fluvial response, a precise and accurate chronostratigraphic framework is required concerning the formation of the different geomorphological-sedimentary units. Because of the low number of available OSL dates from the area of study, the age control is limited. However, since the fluvial developments could be linked to the well-dated fluvial history of the Rhine-Meuse delta (Busschers 2008) a chronostratigraphy was inferred which is used to discuss the potential link between distinct external forces and fluvial response in this section.

During the Middle Pleniglacial, strong aggradation in the LRV occurred due to an increased sediment supply to the river system triggered by intense regolith erosion under cold climatic conditions (Busschers et al. 2007). Aggradation was favoured by an increase in accommodation space as a consequence of subsidence of the valley bottom related to the collapsing MIS4 forebulge. Net aggradation was in between 4-8 m and was produced by a braiding river system draining through the Oude-IJssel valley and building up braidplain I. From a large number of OSL dates from the Rhine-Meuse delta, it is known that this phase of aggradation took place between ca. 80 and 30 ka BP (figure 35; Busschers et al. 2007). The fact that no aggradation is recorded for the Early Weichselian and much colder Early Pleniglacial, is supposed to be related to the persistence of soil complexes in the Rhine catchment and the fact that it takes considerable time to transport sediment (especially the coarse material) from the upper catchment downstream (Busschers et al. 2007; Busschers 2008; Van Balen et al. 2010). Anyhow, it implies a time lag in the order of 20-40 kyr between first intense climate cooling and river response and suggests an important role for soil profiles in stabilizing surfaces. In the course of the Middle Pleniglacial, aggradation is believed to have enabled the Rhine to cross the Saalian ice-pushed topography and to relocate its course between 60 and 40 ka BP towards the Rond-Montferland track. Between 35-30 ka BP, a second avulsion took place towards the Gelderse Poort area. The timing of the second avulsion coincides with a phase of strong ice-mass build-up (Waelbroeck et al. 2002 and Mangerud 2004, see appendix IV). Since this avulsion requires a strong deflection of the Rhine towards the west (see chapter 10), forebulge updoming and related differential uplift of the LRV is thought to have played a major role in initiating this development. It probably led to erosion of the upstream side of the Gelderse Poort-ice pushed ridge and, ultimately, to the development of the Gelderse Poort valley.

In the sedimentary archive of the Rhine-Meuse delta, no fluvial sediments date back to the time-interval between ca. 30 and 24 ka BP (Busschers et al. 2007). On cross-sections from the same study, it is visible that Late Pleniglacial deposits (<24 ka, unit B5) have an incised position into Middle Pleniglacial deposits (>30 ka, unit B3,4), suggesting that a phase of river incision occurred during 30 and 24 ka BP (Busschers

et al. 2007). Sedimentary units making up braidplain I and II in the LRV, are considered to be time-equivalents of sedimentary units B3,4 and B5 from Busschers et al. (2007). Since borehole data retrieved in this study do not provide insight in the vertical position of the base of braidplain II in relation to braidplain I, it is unknown whether the Rhine kept an aggrading mode during the considered time-interval or started incision (figure 35). According to Busschers et al. 2007 two forcing factors might have produced fluvial incision: First, climate instability surrounding the MIS3/2 climate cooling, and second, uplift of the valley bottom as a result of forebulge upwarping. Because contemporaneous southward shifting of channel belts is recorded in the central Netherlands, the latter one is regarded to have played a dominant role (Busschers et al. 2007; Cohen 2003). If the inferred chronostratigraphy for the study area is assumed to be correct, a similar contemporaneous southward deflection is also recorded over here, reflected by the position of braidplain II in the south of the area of study and the preservation of braidplain I in the northern areas (appendix I). Since also (partial) northward channelbelt shifting is observed during the time-interval of forebulge collapse (Late Pleniglacial–Lateglacial interval), glacio-isostatic uplift is expected to have resulted in a similar contemporaneous phase of fluvial incision in the LRV during the ca. 30-24 ka BP time-interval (figure 35).

In the delta, a new phase of aggradation is dated between ca. 24 and 16 ka BP which is reflected in the area of study by the formation of Late Pleniglacial braidplain II. Because crustal upwarping is believed to have continued during the greatest part of this time-interval, climatically-induced sediment supply likely became dominant over glacio-isostatic forcing again (figure 35). Aggradation ultimately resulted in a floodplain surface elevation of circa 1-2 m above the surface of braidplain I. Despite net aggradation compared to braidplain I, the Rhine did not re-activate this braidplain in the Oude-IJssel valley what suggests a significant differential uplift of the northern part of the LRV compared to the southern part. Unfortunately, the limited data availability and restricted extent of the study area do not allow a reliable calculation of a differential uplift perpendicular to the longitudinal axis of the studied valley reach. However, it is believed to have been more than glacio-isostatic model outcomes do suggest. These models suggest a slope of ca. 10m/500km = 2cm/km for the south-western flank of the forebulge, at the time of maximum ice-sheet extent (figure 7). In the area of study, more differential uplift in the order of at least 1 m is hypothesised to have been necessary for preventing the Rhine to re-enter the 1-2 m lower floodplain level of the Oude-IJssel valley. This would indicate a forebulge slope of at least 5 cm/km (rough estimate). A similar magnitude for north-south tilting is suggested by Cohen (2003) on the basis of fluvial terrace geometry in the central Netherlands. Since the modelled dimensions of the forebulge do not allow such sharp slopes, it might have been a regional deviation of the average slope. Such a regional effect might be related to the nearby positioned Roer Valley Graben structural system. The distinct faults are structural weaknesses in the lithosphere along which also glacial tectonics might have been expressed (Cohen 2003).

The lower Rhine is thought to have started a transition from a braiding towards a meandering river system already during the latest stages of the Late Pleniglacial (e.g. triggered by climate warming ~17 ka BP) or immediately following climatic amelioration around the onset of the Lateglacial (~14.7 ka BP). In the LRV, the initial response comprised flow contraction (deepening channels, see figure 35) and the

formation of a high-sinuosity meandering system, in response to climate-induced alterations of the Rhine catchment: The gradual disappearance of permafrost, enabling precipitation and snow melt to infiltrate into the soil, and a relative increase in rainfall at the expense of snowfall, caused peak discharges to diminish despite an increase in effective precipitation (Berendsen et al. 1995). The Allerød pollen signal (*Marienbaum* core) recorded by a well-developed pointbar system in the study area, indicates that the stage of intense lateral migration, lateral reworking of older deposits and point bar development was at least reached somewhere during the Allerød. Since it takes considerable time for large rivers like the Rhine to fulfill the total transition, the transition is thought to have started at least around the onset of the Bølling. Whether the lower Rhine was also able to fullfill the final step towards a single channel system during the Allerød is unknown, because the major Allerød channel belt has been eroded by younger river system. Floodplain lowering started not directly following climatic amelioration but somewhere in the course of the Allerød, what can be explained by the existence of a time-lag between climate-induced reduction of sediment supply towards the river system and the onset of incision downstream due to a certain volume of sediment that is still available in the channel network (Van Balen et al. 2010). After a temporal disturbance related to the Laacher See eruption, sediment depletion might have caused a continuation of incision during the early part of the Younger Dryas (figure 35). Since Allerød incision is also observed in the Meuse system (Tebbens et al. 1999), the just described source-to-sink hypothesis from Van Balen et al. (2010) better explains this phase of incision than a direct effect of climatic instability characterising the onset of the Younger Dryas according to Vandenberghe (1995). However, a transition towards aggradation again (figure 35) in the course of the Younger Dryas, is still likely related to enhanced sediment supply due to climatic deterioration. Forebulge collapse might have had an additional role, by providing accommodation space.

Rhine activity in the Oude-IJssel totally stopped during the Late Pleniglacial or was only represented by a very small distributary of the Rhine (Verschuren 2007; Cohen et al. 2009). During the Lateglacial, fluvial activity increased, presumably as a consequence of still ongoing forebulge collapse (lateral tilting of the valley bottom towards the northeast). Moreover, this development is thought to represent the process of northward shifting of the Rhine system. However, due to a considerable gradient advantage towards the west, the Rhine did not fully re-enter the Oude-IJssel valley, but kept its course through the Gelderse Poort area.

During the early Holocene, the transition towards meandering might have proceeded faster than during the Bølling-Allerød because of the more rapid vegetation recovering as a consequence of a denser vegetation cover in the Rhenisch Massif area during the Younger Dryas than during the Late Pleniglacial (Erkens et al. 2011). In the LRV, early Holocene river incision became significant across the transition between the Friesland and Rammelbeek phase (~11.3 ka BP; biozone contact LRV-4a/4b in figure 34). This is evidenced by the final abandonment (ending of floodactivity; see section 6.5) of three meanders making part of floodplain level III and from which a core was taken: Simultaneously, loamy flood sediments become replaced by peat in the *Marienbaum* channel system and channel infilling starts at the *Klein Entenhorst* and *Heesenhof* sites (figure 34). Two other palynologically analysed channel systems

provide information about the rate of early Holocene fluvial evolution: The Vosse Kuhle (present study) and Hohe Leye (Van Munster, in prep.) channel systems which are imbedded in the Younger Dryas braidplain (figure 18). Both systems are thought to represent early Holocene secondary channels, reflecting the initial phase of flow contraction and meandering. Both systems show a meandering pattern, but the Hohe Leye system additionally shows first signs of lateral migration, pointbar formation and lowering of the floodplain level (figures 39 and 20). The pollen content of the basal parts of the channel-fills testifies to an early Boreal and late Boreal moment of channel abandonment for the Vosse Kuhle and Hohe Leye sites, respectively. In other words, the difference in fluvial morphology reflects the rate of fluvial response during a time-interval of almost one millennium. However, the nearby presence of a much larger and well-developed pointbar system, OSL-dated to 10-9 ka BP (cross-section V Erkens et al. 2011; appendix III), shows the relevance of channel dimensions for the rate of fluvial response. Around the Boreal-Atlantic transition, the Rhine probably contracted towards one single-channel, at the expense of smaller secondary systems.

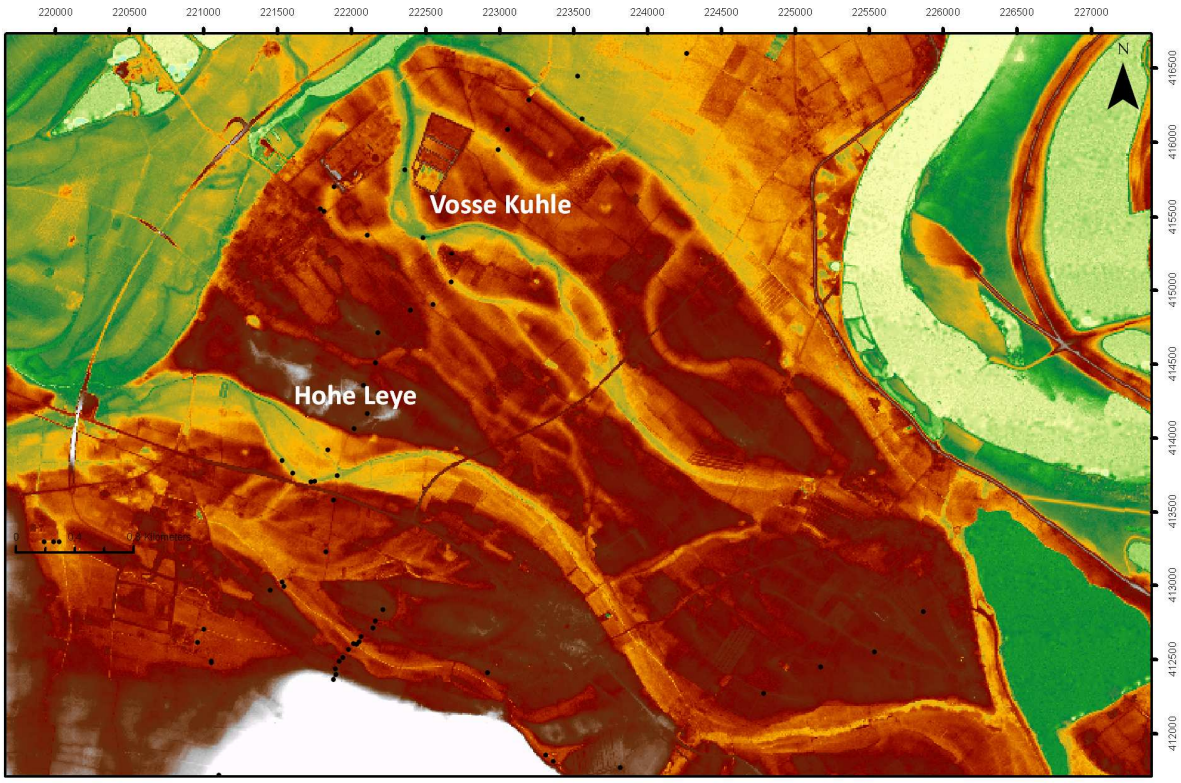

Figure 39 Digital elevation model of the *Hohe Leye* and *Vosse Kuhle* channel systems, from which cores were taken. The morphological differences between both systems in combination with the palynologically-determined timing of channel abandonment provide information concerning the rate of morphological response of these secondary systems to Early Holocene climate change (see text).

9.3 Gradual channel abandonment and flooding activity in abandoned channel systems

Very characteristic for the area of study are abandoned channel systems which are partly or largely filled up with clastic material. A basal infilling with sand in the palaeomeanders making part of floodplain level

III, is reflected by a difference in thickness (depth) of the non-sandy infilling and the depth of the surrounding pointbar system (section 9.1; e.g. *Heesenhof* channel system). It points towards a gradual decrease in discharge instead of an abrupt channel abandonment, and related shallowing of the system. On top of the sand-deposits, generally a thick loam layer occurs. The geometry of this loam layer (figure 24), in combination with the lack of admixed organic material (e.g. gyttja), testify to periodical reactivation of these channels at the time of flooding by younger river systems. As soon as renewed river incision had taken place, flooding activity definitely stopped and organic infilling could start. Only during exceptionally high waterlevels, the systems became drowned again, however, each time only producing a relative thin layer of clastic material within a predominantly organic sequence.

In case of the preserved Bølling-Allerød meanders along the southwest side of the Rhine valley, their location close to the Younger Dryas braidplain made them vulnerable for reactivation. Since strong incision characterised the early stages of the Younger Dryas (figure 35), reactivation of the Allerød meanders probably mainly occurred during the very early part of the Younger Dryas or during the final stage after a phase of aggradation and returning in the direction of former floodplain elevations. Aggradation and associated widening of the Younger Dryas braidplain, moreover, caused continuous erosion of the Bølling-Allerød terrace, keeping the entrances of the palaeomeanders open. After a second phase of strong incision across the Younger Dryas-Holocene transition, also periodical reactivation stopped and prolonged organic infilling could start. In this way, it can be explained why an early Holocene pollen record instead of a Lateglacial one is incorporated within residual Allerød channel morphology.

Pollen analysis of a channel fill of braidplain I shows that this braiding system became abandoned in the course of the Older Dryas (*Eckerfeld* pollen diagram from Janssens 2010; for location see figure 15). Despite the fact that the accurate age of braidplain I is still unknown from OSL-dates, it is considered to have become abandoned much earlier in time and a considerable time-lag exists between the moment of channel abandonment and the onset of channel infilling.

10 Palaeogeographic reconstruction and vegetation history

In this chapter a reconstruction is given of the development of the lower Rhine valley since the Saalian deglaciation, thereby focusing on the Weichselian Pleniglacial and Lateglacial time-interval (circa 60 to 10 ka BP). In order to illustrate this reconstruction for different time-slices, palaeogeographic maps have been constructed. However, it is noted that the depicted distribution of especially the older fluvial systems is far from certain because these terraces have been preserved in a very limited extent or have become covered by younger dune topography.

10.1 Eemian, Early Weichselian and Early Pleniglacial (~130–60 ka BP / MIS5 and 4)

During the optimum of the Eemian interglacial (~MIS5e, ~130-115 ka BP), north-western Europe experienced mean annual temperatures of 1-2°C higher than at present (Lowe and Walker 1997, after Guiot et al. 1989). The lower Rhine was in a meandering mode (Verbraeck 1984; Busschers et al. 2007) and the channel belts were probably concentrated in the north-eastern part of the study area because of a gradient advantage in the direction of the IJssel valley (figure 40). The south-western part of the study area likely functioned as a large-scale overbank environment supporting a dense vegetation cover (Van de Meene and Zagwijn 1979). In the course of the Eemian a similar vegetation development took place as during the Holocene interglacial, starting with a dominance of pine and birch species which gradually became replaced by deciduous trees (Berendsen 2004). Based on stratigraphic position, geological cross-sections by Klostermann (1989) and pollen content, the peat layer containing the *Kehrum Torfkuhle* pollen record is thought to have been formed during the final stage of Eemian (De Mulder et al. 2003). At that time, the upland area supported mixed forests consisting of *Quercus*, *Betula*, *Pinus* and *Corylus*. The floodbasin area was dominated by a swampy vegetation type, supporting *Alnus* trees, (cyper-)grasses, *Filipendula*, Umbelliferae and a wide range of other semi-aquatics. *Corylus* is believed to have grown underneath the *Alnus* trees at relative dry locations in the floodbasin. The large-scale development of more than one meter of peat required relative high groundwater levels for creating swampy conditions in the floodbasin.

Marked climatic deterioration characterised the MIS substage transition 5e/5d which is referred to as the onset of the Early Weichselian (Martinson et al. 1987). Mean annual temperatures dropped in the order of 3-4 °C (Lowe and Walker 1997, after Guiot et al. 1989). The Early Weichselian spans a time interval of more than 50 kyr and forms a transitional period from full interglacial conditions during the Eemian towards full glacial conditions. The latter became established at the onset of the Early Pleniglacial (~70 ka BP, MIS4), when a new drop in mean annual temperatures took place of circa 4-5°C and continuous permafrost developed on large-scale in north-western Europe (Lowe and Walker 1997). The vegetation reconstruction by Behre (1989), points towards the predominance of an open steppe vegetation type.

According to Busschers (2008), the fluvial record in the IJssel valley downstream of the LRV indicates that the input of fresh sediment into the river system remained low during the Early Weichselian and Early Pleniglacial due to the persistence of (Eemian-Early Weichselian) soil complexes in the catchment area. The lower Rhine probably mainly kept on reworking older deposits, under a snowmelt discharge regime. At the end of the Early Pleniglacial, the lower Rhine started its gradual avulsion towards the west, using the so-called Rond-Montferland track (Busschers et al. 2007; Cohen et al. 2009), immediately downstream of the study area. The avulsion away from the IJssel valley is thought to have taken circa 20,000 year.

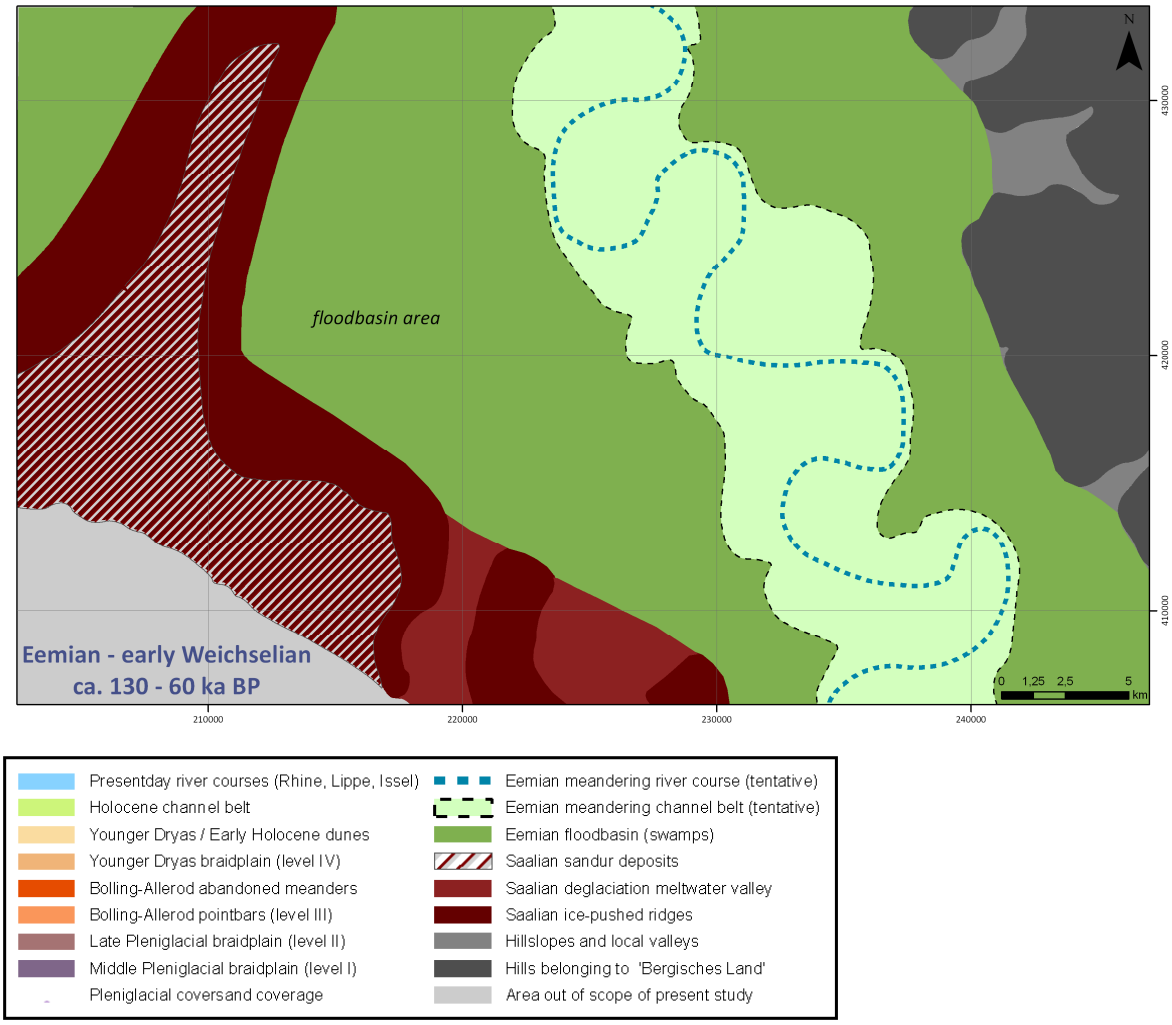

Figure 40 Palaeogeographic map for the Eemian – early Weichselian time-interval, and legend for all palaeogeographic maps.

10.2 Middle Pleniglacial (~60–30 ka BP; MIS3)

In the course of the Middle-Pleniglacial, soil complexes and vegetation cover in the Rhine catchment are thought to have sufficient deteriorated to allow large scale catchment erosion to occur, resulting in an increased sediment input into the river system and the onset of aggradation (section 9.2; Busschers et al. 2007). Vertical aggradation in the apex area of the delta enabled the Rhine to change its course gradually towards the Rond Montferland pathway, between 60 and 40 ka BP (Van de Meene 1997; Verbraeck 1984;

Busschers et al. 2007; Cohen et al. 2009). During the same period, vertical aggradation is supposed to have enabled to Rhine to re-enter the Niers valley which started to function as a secondary Rhine system (Kasse et al. 2005). In the lower Rhine valley, this phase of aggradation led to the build-up of braidplain I which finally reached a braidplain elevation of circa 3-5 m higher than the Eemian floodplain levels. Since this braidplain is believed to have been active for a period of circa 20-30 kyrs, it is expected to have been very wide, comprising the total width of the Oude IJssel valley (figure 41). Around 35-30 ka BP, braidplain I gradually abandoned as the result of a new avulsion of the Rhine away from the Rond-Montferland-course and towards the Gelderse Poort area (Verbraeck 1984; Van de Meene and Zagwijn 1979; Busschers 2008; Cohen et al. 2009). This avulsion required the creation of a new valley system, for which in turn erosion of the ice-pushed ridge between Kleve and Emmerich was necessary. Based on the orientation of the erosional flanks of the upstream side of the Montferland ice-pushed topography, lateral erosion by the Rhine took place in a north(-northeast)ward direction (figure 42). Additional erosion might have taken place along the downstream side of the ice-pushed ridge between Kleve and Emmerich. Erosion at the upstream side requires a strong deflection of the Rhine channel belt towards to south in the area of study which is thought to have become possible as a consequence of forebulge updoming (section 9.2). The timing of forebulge updoming is supported by British-Scandinavian ice-sheet reconstructions which provide evidence for a strong increase in ice volume following circa 40-30 ka BP, reaching a maximum ice-sheet extent around 30-25 ka BP (Houmark-Nielsen and Kjaer 2003).

10.3 Late Pleniglacial (~30 ka – 14 700 yr BP; Late MIS3 and 2)

The onset of the Late Pleniglacial, circa 30 ka BP, is characterised by a climate transition towards extreme severe conditions and the development of continuous permafrost in large parts of the Rhine catchment (Vandenberghe 2001). In a great portion of the periglacial area of north-western Europe, tundra and shrub-tundra vegetation types became replaced by barren steppe (Behre 1989). By this time, the Rhine is believed to have fulfilled the relocation of its main course towards the Gelderse Poort area (Busschers 2008; Cohen et al. 2009). In the Oude IJssel and Rond-Montferland valley systems, braidplain I became partially or totally abandoned and subject to cryoturbations and niveo-eolian coverage (Van de Meene and Zagwijn 1979; Verbraeck 1984; Verschuren 2007). By that time, the Rhine was located in the southwest of the study area where braidplain II started to develop (figure 43). Braidplain II in the study area presumably correlates with the NT2 terrace or Older Lower Terrace (German: Älteren Niederterrasse; Dutch: Laagterras) as distinguished downstream in the Netherlands (Berendsen and Stouthamer 2001), more upstream in the lower Rhine valley (Erkens et al. 2011) and in the Niers-valley (Kasse et al. 2005). It remains uncertain whether the Rhine kept an aggrading mode across the MIS3/2 transition or started incision as is recorded in the sedimentary archive downstream in the Rhine-Meuse delta (section 9.2). Over there, a phase of strong incision precedes renewed aggradation from ~24 ka BP onwards. Moreover, according to Busschers (2008) and Cohen (2003) the combination of channel incision with southward shifting of channel belts in the central Netherlands suggests a continuing role of glacio-isostasy (mirrored by aggradation and northward migration lateron).

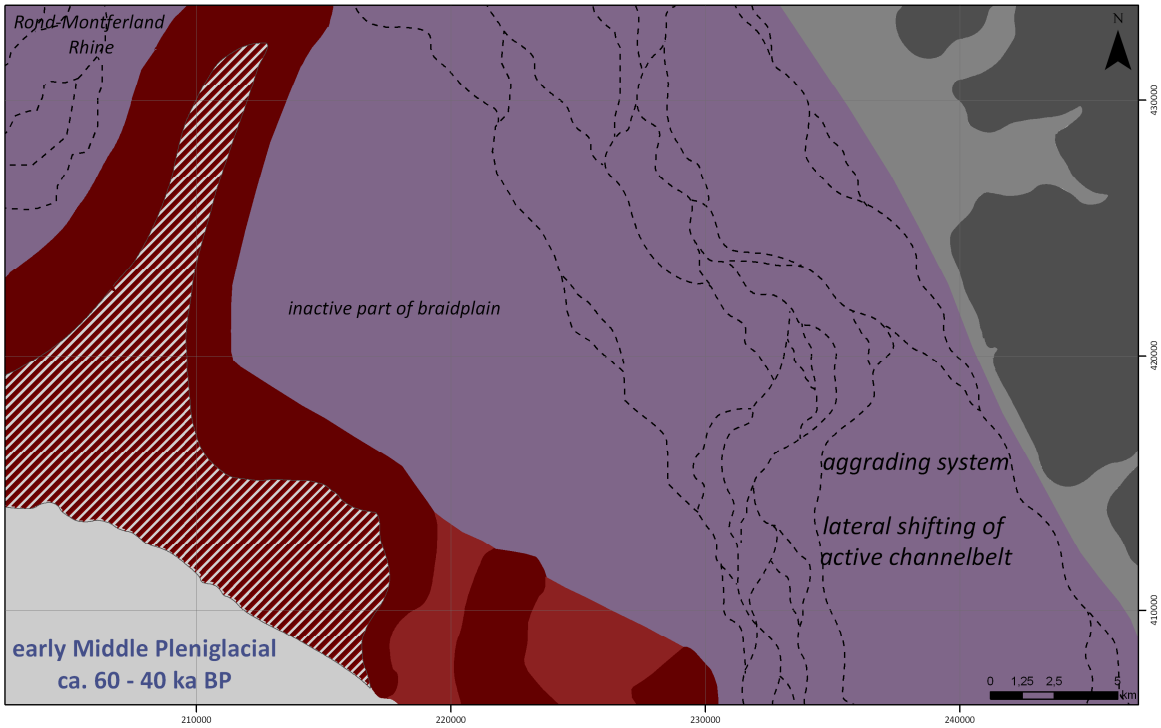

Figure 41 Palaeogeographic map for the Middle Pleniglacial period. For legend, see figure 40.

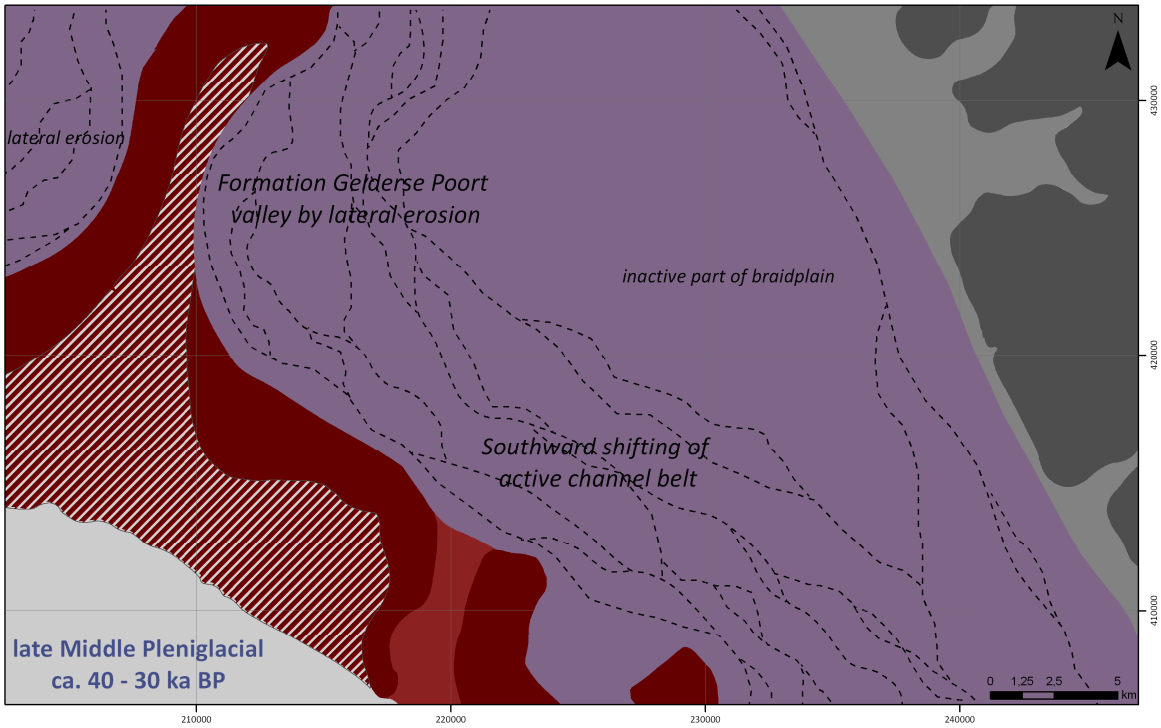

Figure 42 Palaeogeographic map showing the way erosion might have occurred during the formation of the Gelderse Poort-Rhine course. For legend, see figure 40.

A comparison between braidplain I and II on the DEM shows that overall an increase in floodplain level took place (at least for the part of braidplain II that preserved), evidencing net aggradation in the order of 1-2 m during the Late Pleniglacial. Based on the onset of aggradation in the delta downstream, aggradation in the study area is presumed to have started at least ~24 ka BP. Aggradation might have started slightly (<1000 yr, Van Balen et al. 2010) earlier because of its location nearer the sediment source area (Rhenish Shield, according to Busschers 2008). Overall, aggradation is supposed to have resulted from an increased sediment yield due to the climate-induced combination of a sparse vegetation cover with intense ground disturbance by periglacial processes in the Rhine catchment. On top of braidplain I and inactive parts of braidplain II, coversand deposition took place during (especially the latest stages of) the Late Pleniglacial (figure 43; Verbraeck 1984; Verschuren 2007; Kasse 2002). The deposition of coversand, a very low organic productivity and poor preservation conditions explain the virtual absence of Late Pleniglacial organic infills within abandoned channel systems of braidplain I.

Following the moment of maximum ice-sheet extent between 30-25 ka BP, the northwest European ice-sheets started melting. The visco-elastic rebound of the earth's crust started almost immediately. Despite the fact that the former forebulge area is still subsiding today, the highest rebound rates were experienced before the onset of the Holocene (Lambeck 1995). Forebulge collapse caused north-eastward tilting of the valley bottom an increase in fluvial activity in the northern part of the LRV and in the Oude IJssel valley (section 9.2).

10.4 Bølling – Allerød interstadial (~ 14 700 – 12 800 yr BP or LSE; Late MIS2)

The ice-core record shows that rapid climatic amelioration occurred within only a few years at the onset of the Lateglacial circa 14.7 ka BP (Björck at al. 1998). Fossil Coleoptera assemblages in Britain suggest a rapid increase in mean annual temperature from circa -8 up to +7°C within only a couple of centuries (Atkinson et al. 1987). In north-western Europe this warming is reflected by the replacement of a heliophilous herbaceous vegetation by dwarf shrubs communities and later on by forest (Behre 1989, chapter 4). In the lowland areas of The Netherlands, vegetation directly responded to climate change by a gradual development of birch forest during the Bølling (Hoek 2001). Highest temperatures were reached during the Bølling, followed by a prolonged cooling trend towards the Younger Dryas thermal minimum. This trend is characterised by a number of climatic oscillations, with the Older Dryas stadial being the most pronounced one (Lowe and Walker 1997). Despite gradual cooling, temperatures remained high enough for an ongoing development of forest in the lowland areas, culminating in relatively closed forest with a predominance of pine at the end of the Allerød (Hoek 1997a,b).

Triggered by climatic change around 14.7 ka BP or already during the late Late Pleniglacial (e.g. ~17 ka BP), the lower Rhine abandoned braidplain II and started to shift from a braiding into a meandering mode. Near-instantaneous, a transition in fluvial style and flow contraction took place, causing a gradual reduction in the number of active channels and in the width/depth ratio of the remaining channels.

Lateron followed by lateral migration, pointbar formation and ongoing contraction of flow (section 9.2). The duration of the total interstadial period (~1800 yrs) was probably too short to complete the final step of contraction of flow into one single channel (section 9.2). Incision did not start until the Allerød, as a consequence of a time-lag between climate-induced reduction of sediment supply towards the river system and the onset of incision downstream due to a certain volume of sediment that is still available in the channel network (section 9.2). In the Meuse, a phase characterised by high-sinuosity channels is dated to the Bølling-Allerød transition (Huisink 1997; Tebbens et al. 1999). A transition towards incision, meandering and reworking of older sediments is observed in the Rhine-Meuse delta downstream (Cohen 2003; Busschers et al. 2007). Channel belts in the Niers-Rhine and lower Meuse valleys started a similar transition towards a meandering mode in the course of the Bølling, however, under the conditions of a gradually reducing discharge (Kasse et al. 2005; Kasse et al. 1995; Berendsen and Stouthamer 2001; Tebbens et al. 1999).

Soil development occurred on top of the abandoned braidplain II, as inferred from an Allerød pollen spectrum in a soil profile studied by Jansen (2001). The presumed Allerød part of the *Marienbaum* record shows that pine forest with an undergrowth of grasses and some *Empetrum* probably covered the ice-pushed ridges. Relative high juniper and *Artemisia* pollen percentages and the presence of *Ephedra* species in the final stage of the Allerød are indicative for a decreasing soil stability and more aeolian activity, related to climate change towards colder and drier conditions. The active floodplain area and the surrounding older terraces probably supported a more open vegetation cover compared to the upland area. Relative dry parts of the floodbasin were probably characterised by poplar and birch trees and herb species such as *Rumex*, Ranunculaceae, Compositae liguliflorae and Cruciferae.

Whether the Oude-IJssel valley did become totally or only partially abandoned during the Late Pleniglacial is not known for sure (Verschuren 2007) but the presence of a couple of slightly meandering systems and a number of individual palaeomeanders evidences that it was certainly active during the Bølling-Allerød interstadial period (figure 44). One of these systems, the so-called *Wolfstrang* system (Janssens 2010; Graeves 2010), diverges from the main Rhine valley towards the northeast. Due to strong ongoing channel downcutting in the Gelderse Poort valley, however, the Rhine kept its course towards the west and only a small portion of its discharge flowed into the *Wolfstrang* system (section 9.2). More to the northeast an even smaller meandering system was present that successively dissects the Late and Middle Pleniglacial terraces. This system is called the *Issel*-system because it supports the modern river *Issel* (Janssens 2010; Graeves 2010). In an upstream direction the channel belt deflects into the modern Lippe valley, suggesting that it was fed by the upland area of the Lippe during the Lateglacial and not by the river Rhine.

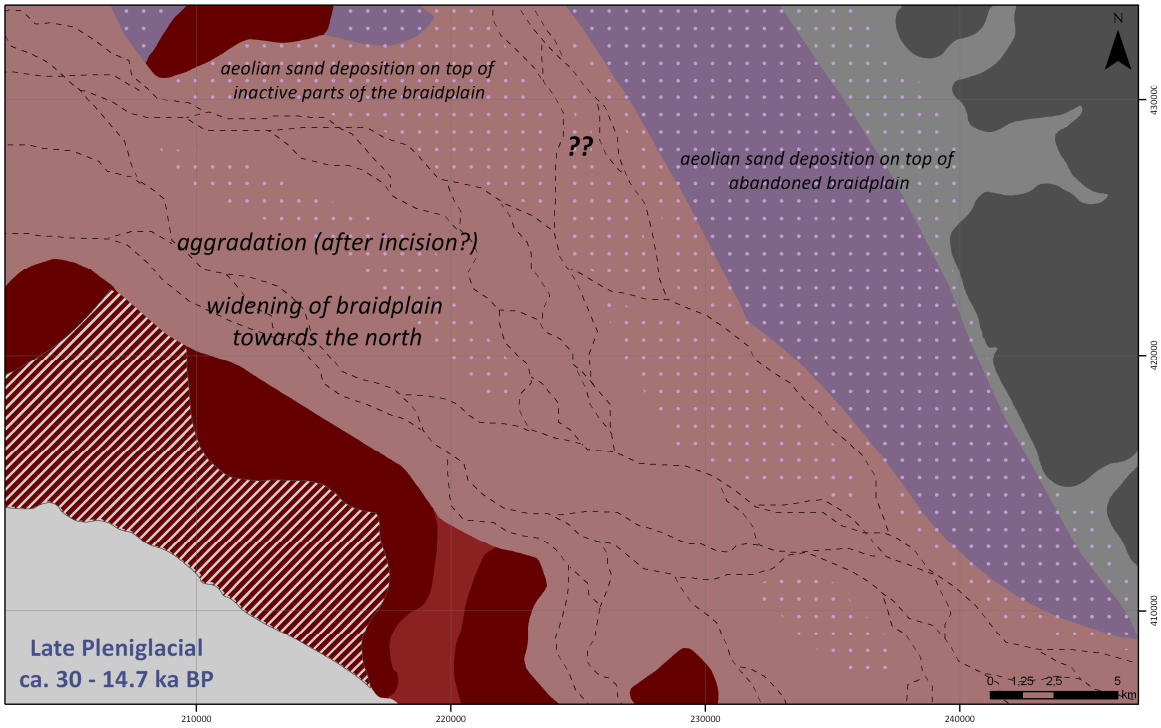

Figure 43 Palaeogeographic map for the Late Pleniglacial period. For legend, see figure 40.

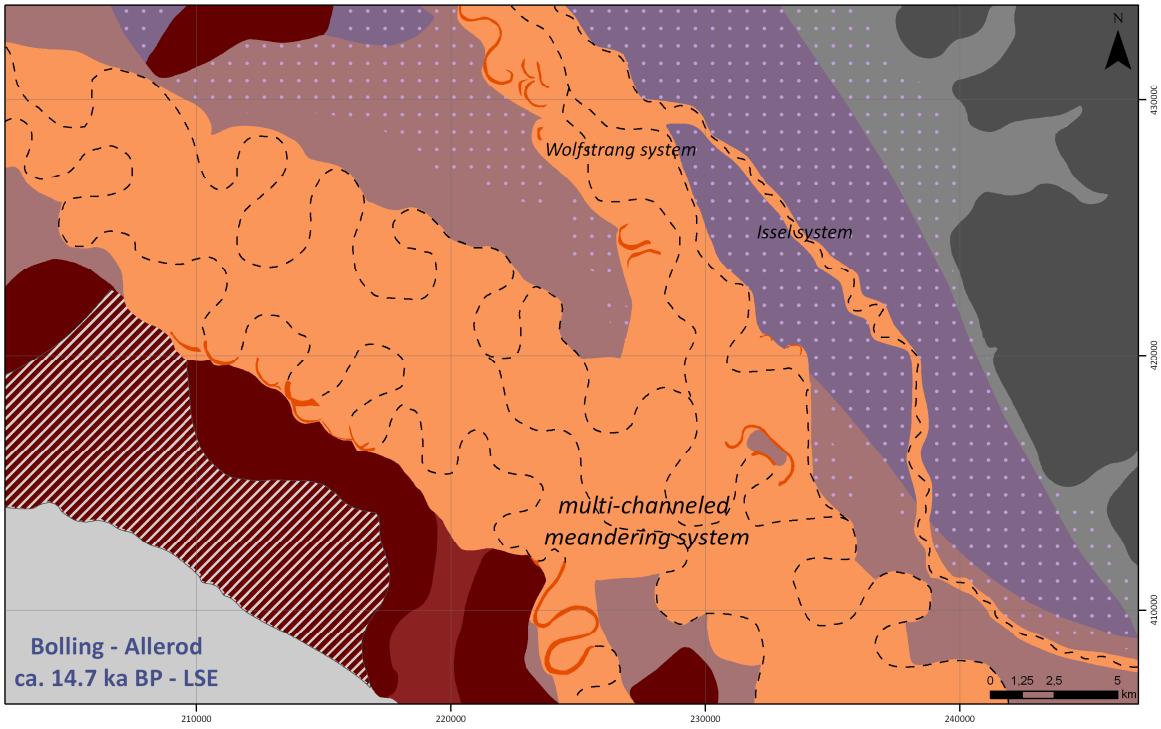

Figure 44 Palaeogeographic map for the Bølling-Allerød interstadial period. For legend, see figure 40.

During the Late Pleniglacial, northeastward tilting of the valley bottom enabled the Rhine to re-enter the Oude-IJssel valley during the Lateglacial. Tilting in combination with a floodplain level difference of 1-2 m between floodplain levels I and II created a (local) gradient advantage towards the Oude-IJssel valley. This explains why the *Issel* system successively cross-cuts braidplain I, braidplain II and braidplain I again (figure 44). By this time, infilling of the *Eckerfeld* channel (abandoned braidplain I channel, figure 18) starts; During the Older Dryas, the channel becomes filled with loamy floodplain deposits originating from the nearby Issel system. In the course of the Allerød, ground-water levels become high enough for organic infilling and preservation (figure 17).

Around 13.2-13.0 ka BP, just before the onset of the Younger Dryas cooling, the volcanic eruption of the Laacher See occurred (LSE) leading to the development of a pumice-dam across the Rhine valley circa 150 km upstream of the study area. Breaching of this dam caused a release of an enormous volume of water with admixed pumice and tephra in the form of one or several surge-like floods (Litt et al. 2003). These floods probably partly destroyed former fluvial morphology of the lower Rhine valley. Because of the very short time interval between the LSE and the onset of the Younger Dryas cooling, it seems unlikely that the Rhine regained its meandering style in time, before climatic cooling forced a transition towards a braiding mode. Therefore, it is suggested that the turning point between the activity phases of meandering level III and braidplan IV should be taken as the moment of the LSE.

10.5 Younger Dryas stadial (~ 12 800 or LSE – 11 700 yr BP; Late MIS2)

Between approximately 12.8 and 11.7 ka BP Europe experienced renewed cold conditions during the Younger Dryas stadial. Mean annual temperatures dropped to around -5°C, forests became largely replaced by shrub-tundra vegetation and discontinuous permafrost conditions returned (Atkinson et al. 1987; Behre 1989). In the Netherlands, a regressive vegetation development took place towards a more open landscape covered by individual forest stands separated from each other by areas dominated by shrub-tundra vegetation. The shrub-tundra was characterised by dwarf willow, dwarf birch and a variety of heliophilous herbs species (Hoek 1997a). Extreme dry conditions during the second half of the Younger Dryas caused an expansion of heather species, e.g. *Empetrum* and *Calluna* and the development of river dunes.

Presumably directly at the beginning of the Younger Dryas, the lower Rhine started to adjust its fluvial style and channel belt dimensions to cooler climatic conditions. However, despite a near-instantaneous climate-related increase in sediment supply, incision continued during the early Younger Dryas due to the time-interval required for transporting the material to downstream areas (section 9.2). In the course of the Younger Dryas, the multi-channeled river started aggradation and the development of braidplain IV (figure 45). Because of the LSE just before the onset of the Younger Dryas, pumice is incorporated in the basal part of the braidplain IV deposits. At the end of the Younger Dryas, a floodplain surface elevation is reached of circa 0.5-1 m lower than the top of the Allerød terrace. The dimensions of preserved Younger

Dryas channels show that these were approximately 1 m deep and 50 m wide. Downstream in the Rhine-Meuse delta, a similar switch to a braiding mode is observed (Pons 1957; Berendsen et al. 1995). There, the so-called Terras X was formed which is considered to be equivalent to braidplain IV of present study and the NT3 terrace or Younger Lower Terrace as distinguished by Schirmer (1995) and Erkens et al. (2011) in the lower Rhine valley (German: Jüngere Niederterasse). In contrast, the threshold towards a braiding mode was not crossed in the Niers-Rhine valley (Kasse et al. 2005). Figure 45 also shows that the reconstructed width of braidplain IV is considerably smaller than the Late Pleniglacial braiding floodplain area. This difference is the consequence of the relative short duration of the Younger Dryas, providing a limited amount of time for aggradation and widening of the braidplain.

Except from the time-interval of early Younger Dryas fluvial incision, the elevation difference between the abandoned meandering level III and the active braidplain IV is thought to have been small enough to enable flooding of level III channel systems (section 9.3). Periodically reactivation of these systems during periods of high waterlevels inhibited the formation of organic channelfills. Instead, loam was deposited in the meandering systems. Reactivation came to an end at the onset of the Holocene, when a renewed phase of river incision started. By this time, channel infilling was able to start in meandering systems experiencing relative high groundwater levels in the seepage area near the ice-pushed ridge (e.g. *Klein Entenhorst* record).

Probably in the early stages of the Younger Dryas, the *Issel* channel belt became inactive and the Lippe redistributed its discharge towards the *Wolfstrang* system. This inferred timing of abandonment is based on the lack of any geomorphological signs of fluvial adjustment to climate change in the *Issel* channel belt. Moreover, pollen analysis of a core from the channel-infill indicates that the system lost its activity at least halfway the Younger Dryas (*Isselaue* record, figure 17, Janssens 2010). Similar to the main Rhine channel, distributaries in the *Wolfstrang* system changed towards a braiding mode.

To the northeast of the active Younger Dryas braidplains, large scale dune systems developed. High fluvial dynamics inhibited vegetation growth on top of the channel bars, resulting in a continuous source of sediment available for aeolian processes. The parabolic shape of these river dunes indicates that vegetation was present at the older terraces which could intercept wind-blown sand. High aeolian activity is known to have characterised especially the second part of the Younger Dryas from circa 12.7 to 11.7 ka BP (Berendsen et al. 1995; Hoek 2001). In the *Marienbaum* record, the onset of this phase probably coincides with the expansion of *Empetrum* at the lower contact of zone LRV-3b.

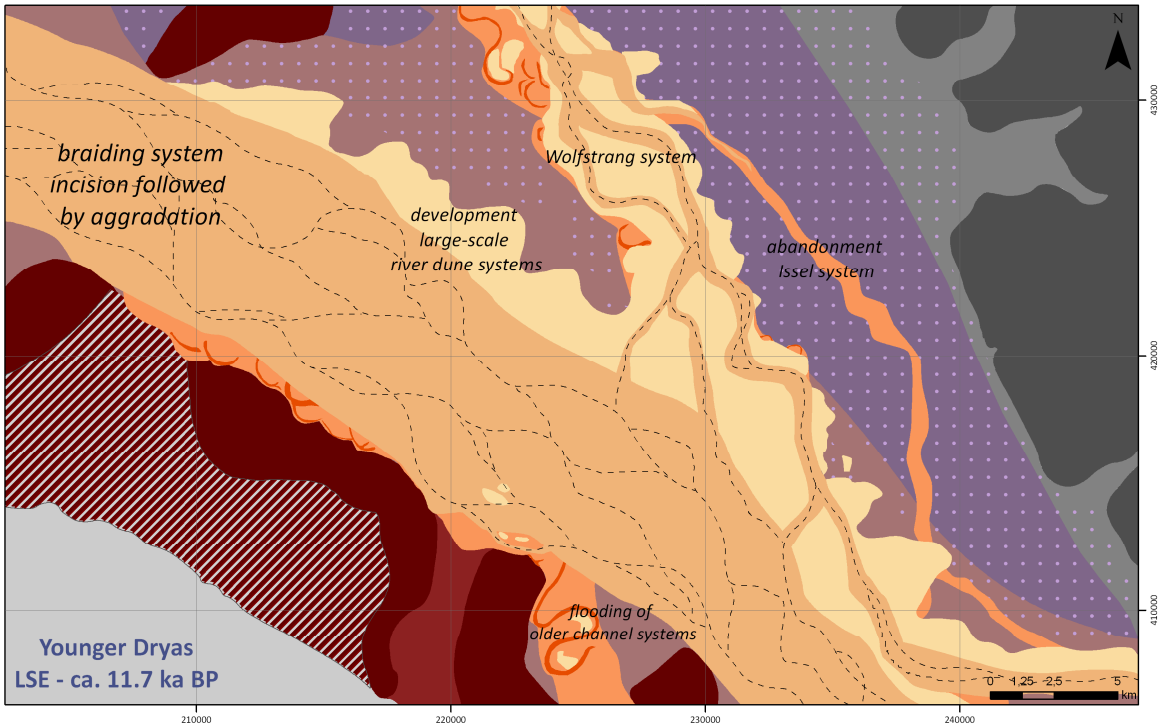

Figure 45 Palaeogeographic map for the Younger Dryas period. For legend, see figure 40.

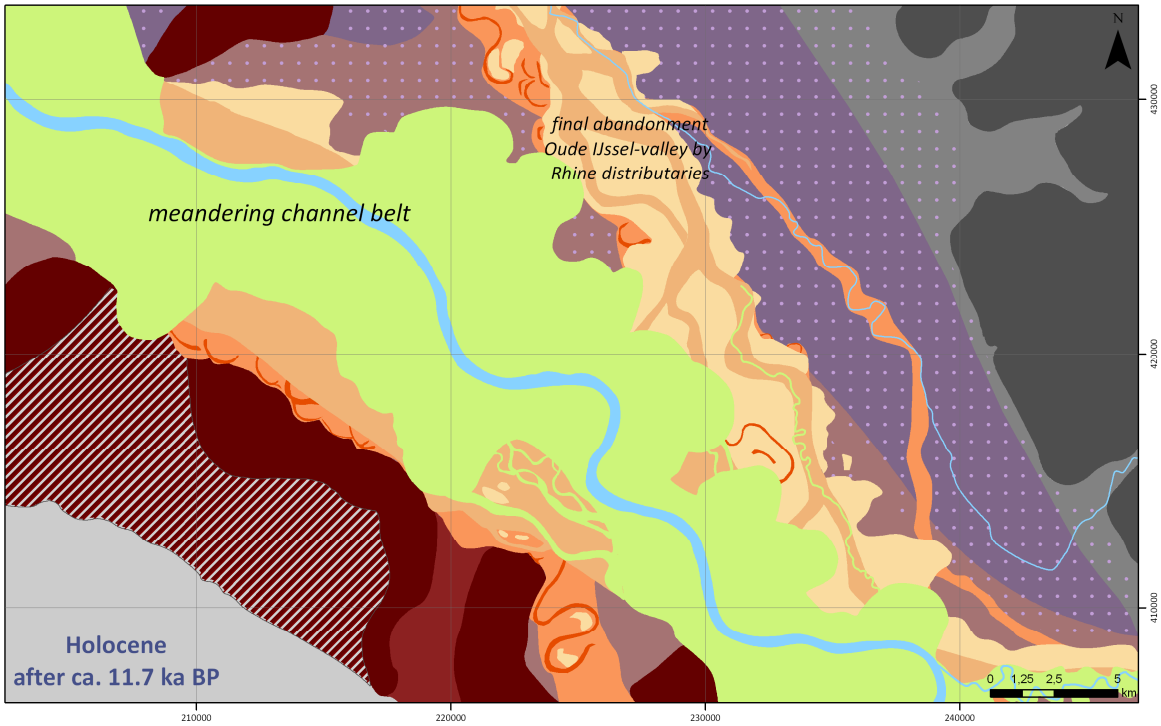

Figure 46 Palaeogeographic map for the Holocene period. For legend, see figure 40.

In the course of the Younger Dryas, the vegetation cover gradually opened up as reflected by the gradual decrease in the AP/NAP ratio in the *Marienbaum* record. Especially pine suffered from climate cooling, but remained dominant over birch in the pollen record. Open pine forests persisted at the ice-pushed ridges with an undergrowth of heather and grasses. Juniper, *Artemisia*, *Helianthemum* and *Ephedra* flourished on relative open spots with unstable soils within the forest. Birch and poplar probably characterised the tree-shrub layer in the more humid floodbasins along both sides of the active channel belt. A lot of herb species were favored by unstable soil conditions and the input of fresh fluvial sediment in the floodplain area, for instance Chenopodiaceae, *Rumex* and *Galium*. Sedges, horsetails, Ranunculaceae, Cruciferae, Umbelliferae, *Filipendula* and *Thalictrum* probably grew in the more swampy parts of the floodplains.

10.6 Holocene (from ~11 700 yr BP onwards / MIS1)

The Holocene interglacial started with a rapid increase in mean annual temperature from circa -5°C to over 5°C within only a few years (Atkinson et al. 1987; Björck et al. 1998). Probably related to a strong increase in vegetation cover, the lower Rhine started flow contraction and incision into braidplain IV (Berendsen et al. 1995). Around the onset of the Rammelbeek phase (~11.3 ka BP) this had resulted in the final abandonment of channel systems making part of meandering level III. More stable discharge regimes caused a transition towards a multi-channeled meandering river system (section 9.2). At least around 9 ka BP, the river system contracted towards one single channel and started to develop wide pointbar systems (Erkens et al. 2011).

During the remaining part of the Holocene, the Rhine kept a meandering mode and reworked a great portion of the older floodplain levels under relative stable climatic conditions and without any significant impact of glacio-isostasy (appendix I). During this time-interval, especially human impact appears to have become relevant for fluvial evolution in the LRV (Erkens 2009). Middle- and Late Holocene fluvial developments in the downstream part of the LRV are out of scope of the present study, but are studied in detail by Van Munster (MSc study, in prep.) and De Molenaar (MSc study, in prep.).

11 Conclusions

This study highlights the importance of climate change, glacio-isostasy and intrinsic characteristics of the river system for fluvial evolution of the lower Rhine during the Weichselian-early Holocene time-interval. Climate change additionally forces the vegetation to change its composition, which in turn affects catchment characteristics and fluvial development. The main conclusions are as follows:

- Differential glacio-isostatic uplift of the downstream part of the lower Rhine valley caused lateral shifting of the active Middle Pleniglacial braid belt towards the south(-west). This deflection enabled the Rhine to create a new pathway through laterally eroding the moraine deposits in the Gelderse Poort area.

- Late Pleniglacial to Lateglacial forebulge collapse and related tilting of the valley bottom, caused aggradation and widening of the braidplain in a northward direction. Moreover, it led to an increase in fluvial activity in the Oude-IJssel valley. Since a gradient-advantage existed for the Gelderse Poort-Rhine pathway, forebulge collapse did not result in a full return towards a Middle-Pleniglacial fluvial-geographical setting (in other words, no avulsion back to the Oude-IJssel valley).

- The lower Rhine started to switch from a braiding to a meandering fluvial style in response to climate warming around the onset of the Bølling-Allerød interstadial complex. This transition is recorded both for the Gelderse Poort channelbelt as for the distributary in the Oude-IJssel valley. Fluvial response started with contraction of flow, abandonment of the braidplain, channel incision and meandering, lateron followed by lateral migration and the formation of pointbars. The final stage of a single-channel system was almost or totally reached at the end of the Allerød.

- Glacial climatic conditions of the Younger Dryas favoured a transition towards a braiding river system; however, breaching of the Laacher See pumice-dam is expected to have initiated this transition. Either as a consequence of climatic instability or delayed sediment depletion of the river system (in response to Late Pleniglacial–Lateglacial climatic amelioration), a phase of channel incision characterised the early stages of the Younger Dryas. High sediment supply caused a next phase of aggradation and braidplain widening by lateral erosion and reworking of older terraces.

- Early Holocene climate change initially resulted in flow contraction and abandonment of the Younger Dryas terrace level. Flood-activity in Bølling-Allerød channel systems finally ended around the Friesland-Rammelbeek transition (~11,300 yr BP), from which is concluded that

channel incision had become significant by that time. It took the Rhine until the end of the Boreal (~9000 yr BP), to change from a multi-channel towards a single-channel system. The formation of pointbar systems and floodplain level lowering started earlier in time, probably around the Preboreal-Boreal transition (~10,000 yr BP).

- In the downstream part of the lower Rhine valley, channel systems making part of abandoned river terraces kept an important role in draining water during periodical floods. This caused partial infilling of the channels with clastic material and initially inhibited infilling with organic (autochtonous) material. The latter process could not start until significant river incision resulted in final abandonment of these older channel systems. Therefore, considerable time-lags (up to several millennia) exist between initial channel abandonment and the onset of organic channel infilling, making pollen records of Lateglacial age very rare in the lower Rhine valley.

12 Directions for future research

The reconstructions and chronostratigraphy provided in this thesis almost totally rely on correlation with the well-documented and well-dated sedimentary record from the Netherlands. Despite the fact that this reference record is presumed to be rather accurate, it remains an indirect way of dating including a list of assumptions. Therefore, future research in the area of study should incorporate OSL dating of distinct geomorphological-sedimentological units. Most essential for the chronostratigraphy is the age of the terrace remnant supporting the village Appeldorn today, believed to be a Younger Dryas braidplain IV. Establishing the age of this unit is expected to be descisive to one chronostratigraphic framework; that of Erkens et al. (2011; essentially after Klostermann 1992) or the alternative proposed by this study. Only at one location, near the base of the sand body, pumice was found. Because this unit is hard to core by hand, OSL dating is recommended for dating this unit. Moreover, caution should be taken with interpreting pumice-evidence. Pumice near the top of a unit might be deposited there during Younger Dryas or early Holocene floods. In this area only a considerable amount of pumice which is really admixed to the braidplain deposits is believed to have been formed by Younger Dryas river systems. OSL samples from braidplain II have already been collected by W.Z. Hoek and C. Kasse and dates are expected soon. If in-channel deposits from the north-eastern part of the area are additionally dated, a more complete correlation to the Dutch chronostratigraphy can be made, what will provide a more funded basis for future research on the character and timing of fluvial response and underlying processes.

For using biostratigraphic correlation as an indirect dating method, it had to be assumed that vegetation developments took place at the same time all over north-western European lowlands. Because it is known this was certainly not the case and these developments were spatially and temporally diachronous, this assumption weakens the bio-chronostratigraphy of this area. Even the records that were used for correlation lack an exact independent chronological framework: The Lateglacial vegetation scheme of the Netherlands totally relies on radiocarbon dates, concerning a period of time which incorporates several plateaux in the radiocarbon-calibration curve (Hoek 2001). Therefore, an independent area-specific bio-chronostratigraphic framework based on radiocarbon dates is recommended for future research concerning the timing and character of vegetation development in this area.

Further research concerning the magnitude of differential uplift due to glacio-isostasy should concentrate on the downstream part of the area of study, approximately around the line Emmerich – Ulft. Over there, the valley morphology is relative wide enabling the Rhine to respond freely by relocating its course. Moreover, the wide extent of the Middle and Late Pleniglacial fluvial deposits over here will result in a more reliable quantification. A similar type of study should focus on the downstream part of the Oude-IJssel valley (near Doesburg) which contains similar characteristics and the results from both areas should be combined. For this type of study, top and base elevations of distinct sedimentary units must be determined. Because of the thickness of the sediments, automatic augering techniques are recommended.

Summary

In the downstream part of the lower Rhine valley (between Wesel and Emmerich, Germany), at least four floodplain levels of pre-Holocene age can be distinguished, primarily based on plan view channel patterns, top elevation of sandy channel deposits and lithological characteristics. First a relative floodplain level chronostratigraphy was constructed on the basis of cross-cut relationships as inferred from digital elevation models and both new and pre-existing cross-sections. Second, this relative chronology was correlated to the well-dated sedimentary record of the Rhine-Meuse delta (Busschers 2008; directly downstream). The result was a new chronostratigraphy for this area, which forms an alternative to the pre-existing chronostratigraphy provided by Erkens et al. (2011) and Klostermann (1992). Factors which are believed to have predominantly influenced river functioning during the Weichselian are climate and climate derived characteristics of the river catchment, glacio-isostatic uplift and subsidence and time-lags between sediment-input- and discharge-related alterations of the upper catchment and the moment of registration downstream.

During the Eemian and early Weichselian, the Rhine was a meandering river that drained through the Oude-IJssel and IJssel-valley towards the north. Broad floodbasin areas on both sides of the river supported swampy thermophilous forest during the Eemian, later on replaced by more open shrub-herb vegetation under glacial conditions. The Rhine changed towards a braiding mode around the beginning of the Middle Pleniglacial. A long period of climatically-induced regolith erosion and irregular discharge regimes caused intense sediment accumulation to occur in a wide floodplain that extended all over the area (braidplain I). As a result of gradual forebulge updoming, the active channel belt shifted towards the south around the Middle Pleniglacial-Late Pleniglacial transition. Moreover, it resulted in a phase of strong incision. Despite still ongoing forbulge updoming during the first half of the Late Pleniglacial, renewed aggradation started building up braidplain II. Lateglacial climatic amelioration pushed the river system towards a meandering mode while at the same time the number of channels and channel width/depth ratio's declined, culminating in a high-sinuosity meandering system at the end of the Bølling-Allerød interstadial period. During the same period, glacial vegetation became replaced by open birch-pine forests. As a consequence of a delay in (originally climatically-induced) sediment depletion of the river system, incision did not start until the end of the Allerød. Incision continued during the first stage of the Younger Dryas, followed by a return towards a braiding mode and aggradation. This return was related to renewed degradation in vegetation cover and enhanced regolith erosion in the catchment area upstream and more irregular discharge regimes. Fluvial incision by meandering rivers dominates the early Holocene, when interglacial conditions returned and prolonged forest development could start.

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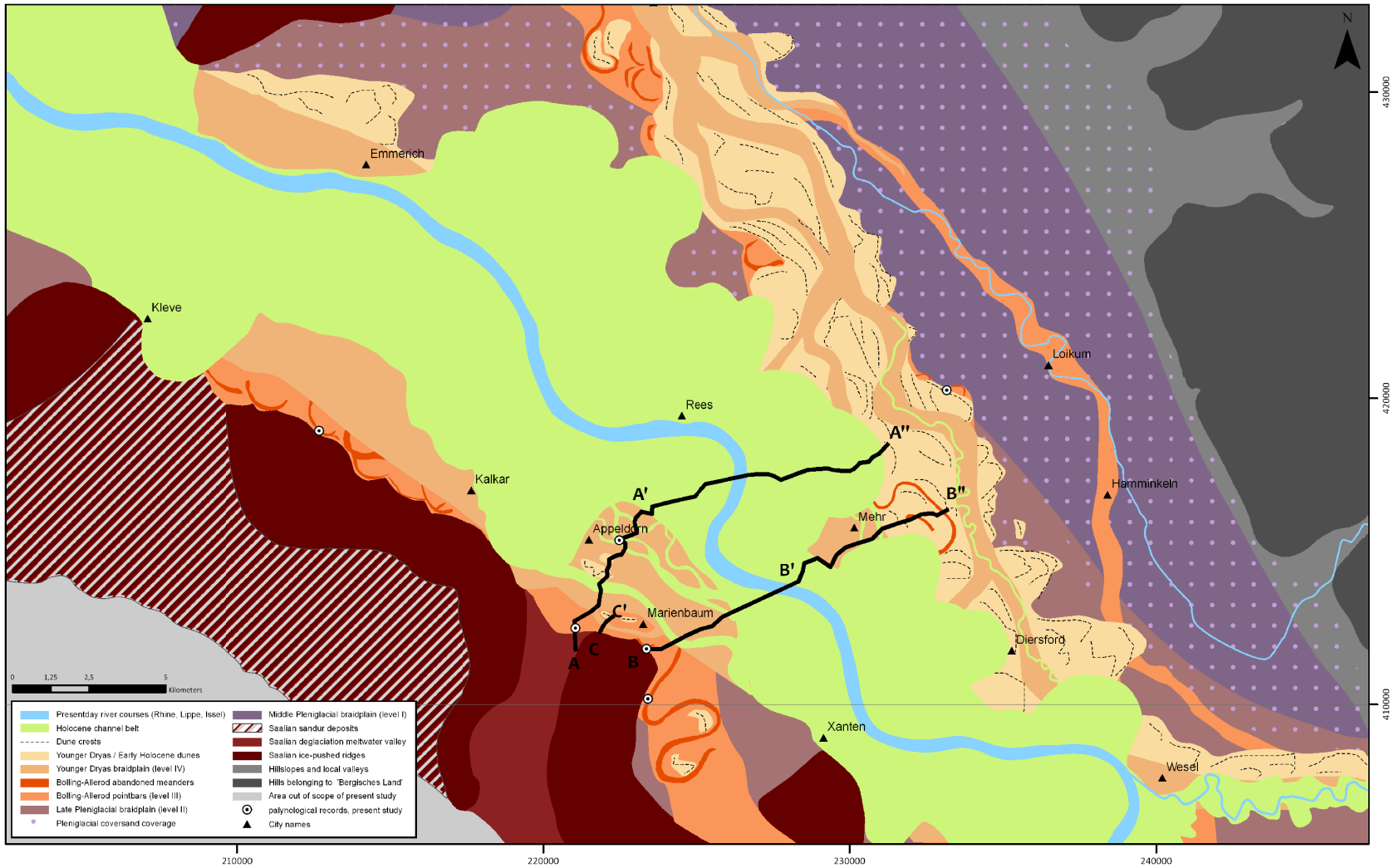
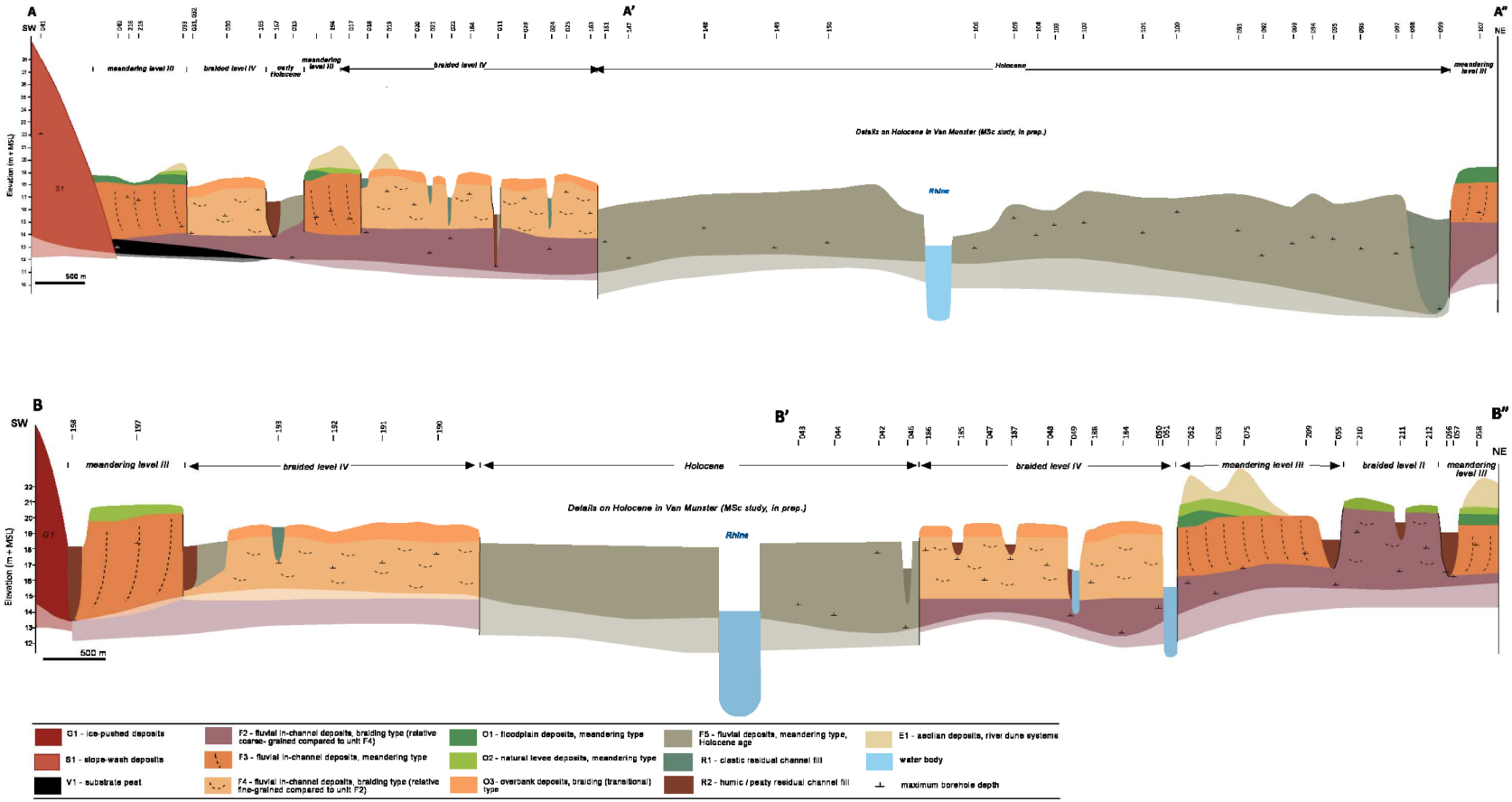
Appendix I: Geological - Geomorphological map of the lower Rhine valley, Germany

Geological map of the Lower Rhine valley as constructed by Geurts (2011).



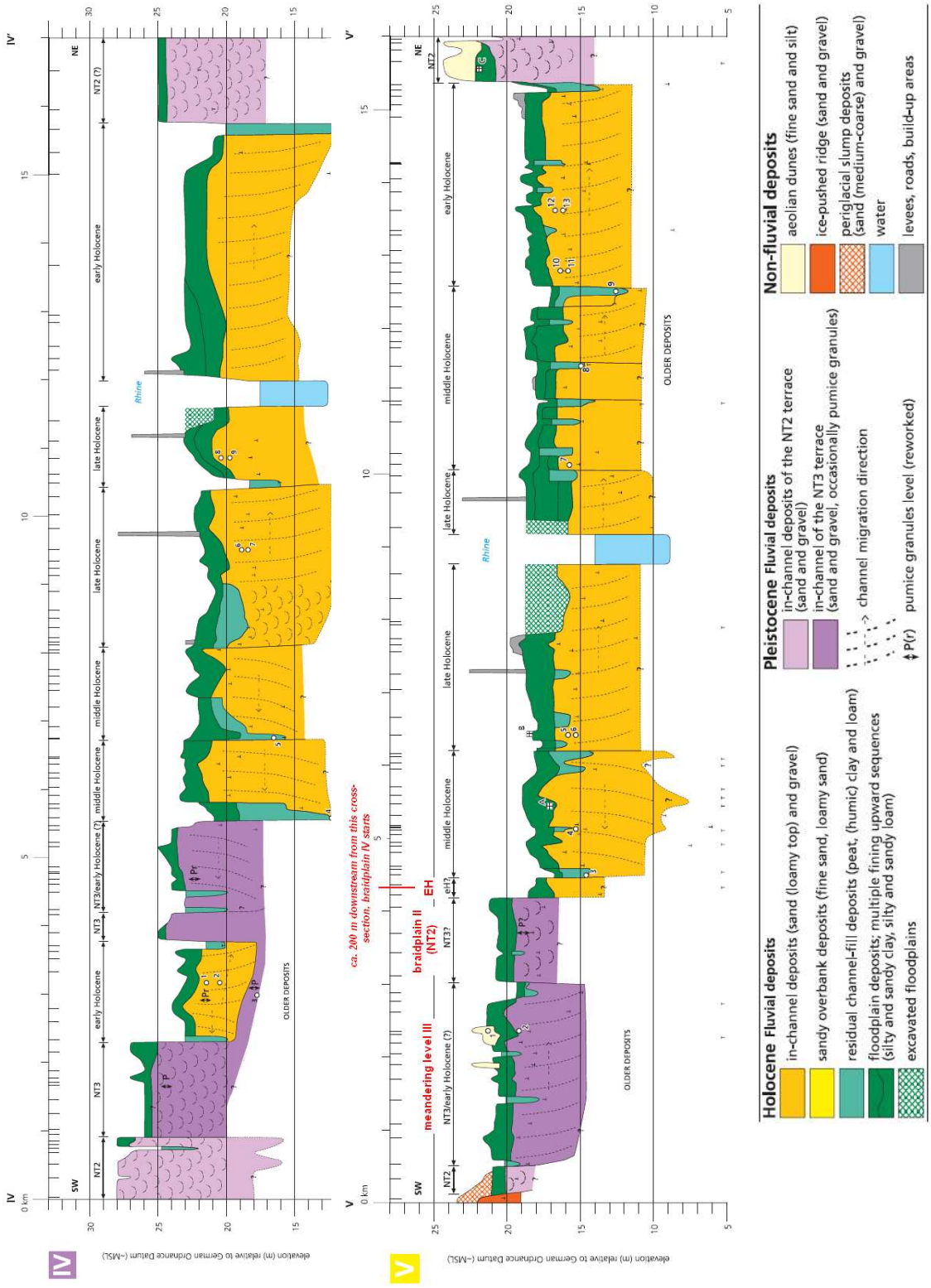
Appendix II: Cross-sections I and II from the lower Rhine valley,

Cross-sections I (A-A", near Appeldorn) and II (B-B", near Mehr) from the lower Rhine valley as constructed by Geurts (2011).

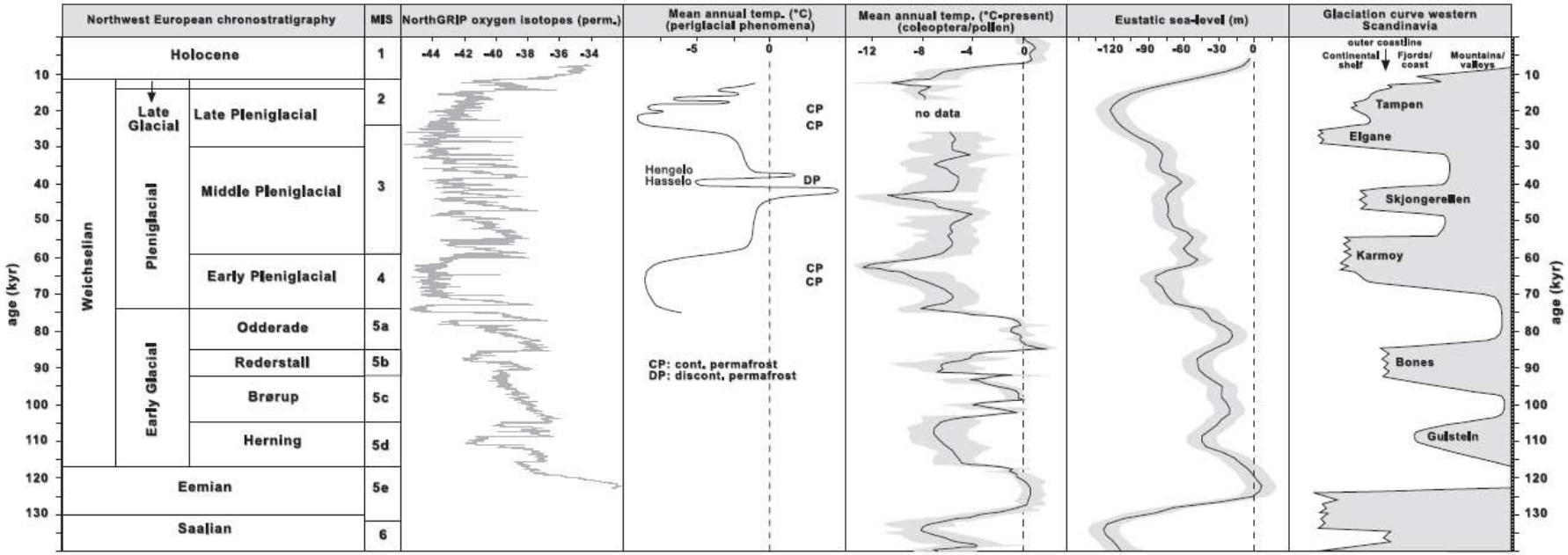


Appendix III: Cross-sections IV and V (Erkens et al. 2010)

Cross-sections V-V' and IV-IV' in the southernmost part and upstream of the research area, respectively (Erkens 2009; Erkens et al. 2011; figure 2). In red, the floodplain classifications in line with the present study are shown.



Appendix IV: Compilation of northwest European climatic signals (by Busschers et al. 2007)



From left to right: Chronostratigraphic framework (Vandenberghe 1985; Van Huissteden and Kasse 2001), marine isotope record (Bassinot et al. 1994), NorthGRIP oxygen isotope record (NGRIP 2004), Pleniglacial mean annual temperatures (Vandenberghe et al. 2004), temperature reconstruction based on the La Grande Pile record (Guiot et al. 1989), mondial eustatic sea-level position (Waelbroeck et al. 2002) and the Scandinavian glacial history (Mangerud 2004).