Reconstructing palaeoflood discharges of the river Rhine during the Early and Middle Holocene.

Palaeoflood discharge estimations in an alluvial environment.

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Front page: A residual channel of the river Rhine.

Photo: M.M. de Molenaar.

Summary

The threat of river floods in the low parts of the Netherlands demands a detailed knowledge on flood behavior. Accurate flood records however only go back in time for roughly 100 years, while dikes are by law supposed to withstand floods that occur once every 1250 years (the design flood). The theoretical discharge of 16,000 m^3/s for a 1/1250 flood is determined by extrapolating the 100 years of data. This discharge number for the design flood is widely accepted and has become the standard for dyke design and safety analysis, ignoring the fact that extrapolating a time scale of 100 years 12.5 fold is way beyond what is statistically reliable. It is often also not realized that changes in climate and river training in those 100 yeas of recorded data have a profound effect on the outcome of the analysis. This is demonstrated by the recalculation of the 1/1250 year flood after the 1993 and 1995 floods of the rivers Rhine and Meuse, which resulted in a much larger discharge of 16,000 m^3/s for the design flood (the design discharge was 15,000 m^3/s prior to the recalculation)(Silva, 2001). Nor is realized that extrapolating a series of common floods to determine the magnitude of a rare flood does not take the natural behavior of rivers into account, it is possible that the design flood can never happen because of natural thresholds. For a more reliable determination of the design flood, the existing time series of 100 years needs to be lengthened.

Studies on this subject in the Rhine catchment are mostly focused on flood frequencies and flood severity and are based upon the damages done by each flood. Only one attempt is made by Herget & Meurs (2010) to estimate actual discharges during one flood event in 1342 AD at the city of Cologne, Germany. These studies make use of man made sources like city annals, journals, paintings and memorial stones. The drawbacks of using man made sources are that the availability of sources declines when going back further in time, sources my not be first hand or may be intentionally or unintentionally misleading.

In this research discharges of rare flood events are estimated using sedimentary records. The research is focused upon the part of the Holocene that has not seen noticeable human impact on the vegetation cover in the Rhine catchment as this allows for a more accurate

reconstruction of the environment in which the event took place. The research area lies just east of the Dutch – German border near the city of Rees. This area is located downstream from the last major tributary feeding water to the river Rhine and upstream from the bifurcation where the river Rhine splits into the river Waal, Lek and IJssel.

A 4 week field work campaign was carried out using hand operated coring devices. The description of the cores and the analysis of samples for Loss On Ignition analysis was used for determining palaeostage indicators (PSI) and the palaeotopography of the area. Palaeostage indicators that were found are: Slackwater deposits, flood layers in channel fills and pointbar ridges. These PSI's are an indication of the minimum water level that must have been reached. Scenarios were created to cover the reconstructed time period and conditions. The scenarios are used as input for a 1D hydrological model based upon the Chézy equation and a 1D hydrological model based upon the Manning's equation. This resulted in a lowest minimal discharge estimate for a flood event in the Late Subboreal or Early Subatlanticum of 13,200 m^3/s and a highest discharge of 18,700 m^3/s for the Early and Middle Holocene. The same volume of water passing trough the heavily trained river Rhine of today would result in most likely a shorter, steeper and higher discharge peak. In order to better translate the palaeo discharge to the present day situation further research is needed.

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1 Introduction

1.1 Background

"Leven met water" (Living with water) is the name of the government campaign started by the Dutch government in 2010 to raise the awareness of the general public on the risks and benefits of living in a country that found its origin and economical wealth in the large rivers flowing trough the heart, yet is always under threat by the very same rivers (www.nederlandleeftmetwater.nl, 2011). The actual risk for areas subjected to river flooding was assessed by Rijkswaterstaat (the Dutch waterways authority) (Rijkswaterstaat, 2005). In this assessment, chances of flooding of the area behind dikes caused by high water levels, tunneling/piping (sand below the dykes gets flushed away, undermining the dyke) and fail mechanisms (parts of the water protection that need active human involvement, like 'copures' (passages trough the dykes) that need to be closed) are determined, together with the economic and environmental damage that will be the result of a flooding. The outcome of this research shows that large parts of the Dutch river area are today subjected to a very high chance to flooding, often exceeding a statistical chance of more than one flooding event per century. (Rijkswaterstaat, 2005).

A major part of the flood risk is determined by the height the water reaches during peak discharges. Reliable data on water levels and discharges in the largest river in the Netherlands, the River Rhine, only go back 100 years in time. These data are recorded in the annals of the city of Nijmegen near the Dutch-German border (Lammersen, 2002) and are statistically analyzed in order to determine the discharge for a flood with a recurrence time of 1250 years (1/1250). This 1/1250 flood, often called the design flood, is set by the Dutch government as the standard for dike design, meaning that dikes are, by law, designed to resist water levels arising from this design flood (Silva, 2001). Statistically extrapolating such 'snapshot' data excludes the effects of changes in river training, climate, morphology and other natural influences on water levels, like thresholds that can have a limiting effect on the maximum water height downstream. For instance, when water levels rise above a threshold level after which a large area inundates, the water

retention in such an area could 'decapitate' the flood peak, causing the water levels downstream not to rise any further. Moreover, extrapolating only 100 years of measurements results in a large statistical uncertainty in the 1/1250 estimate. A single large event like a millennial flood, or the lack of one, will have a severe impact on the outcome of the estimate. Errors on flood discharge estimates by extrapolating data from a limited time series are generally considered to be in the range of 10% to 100%, depending on the quality of the rating curve and its extrapolation to large floods (Benito et al., 2004). The risk of potential loss of lives and the high economical losses when a dyke breaches are too high to depend on a statistical analysis of 100 years of flood records alone. On the other side, costs of achieving and maintaining a safety level for a flood that might never have happened or is physically very unlikely to be able to happen might be a waste of valuable resources.

The changing of the 1/1250 year flood size estimate after the 1993 and 1995 floods of the River Rhine are an example of the difficulties of extrapolating 100 years of data to much longer time scales. The 1/1250 year flood was estimated before the 1993 and 1995 flood events at 15,000 m^3/s at Lobith (where the River Rhine enters the Netherlands). After the 1993 flood and the 1995 flood with a peak discharge of 12,000 m^3/s (Silva 2001), the discharge estimate of the 1/1250 year flood was recalculated and raised to the new design discharge of 16,000 m^3/s.

This demands for methods to extend the 100 year record and to achieve estimates of extreme flood magnitudes. Therefore a research campaign was started by the University of Utrecht (the Netherlands) in 2010. This report is part of that research and will focus on reconstructing peak discharges and maximum water heights in pre-historic times at the apex of the river Rhine near the city of Rees, Germany.

1.2 Research objective and approach

The objective of this study was to estimate the magnitude of discharges of extreme flooding events during the Holocene (until the human influence on riparian vegetation

became substantial around 3000 – 4000 B.P. (Rösch, 1992; Bos & Urz, 2003; Houben et al., 2006; Welmoed, 2008; Middelkoop et al., 2010)) to extend the 100 year of records and to achieve estimates of extreme flood magnitudes. This was done using the analyzes of laminated channel fills and the spread of overbank deposits for reconstructing water levels and using a vegetation reconstruction and simple Chézy and Manning approaches to estimate discharges. In order to make such a reconstruction the following questions need to be answered:

(1) What levels did the water reach during maximum high water conditions and what was the timing of these events?

(2) What was the topography of the Lower Rhine Embayment at the time of major flood events?

(3) What vegetation was likely to be dominant on the flooded areas during the high water conditions? What is the hydraulic roughness of this vegetation and how does this influence palaeo discharge calculations?

(4) Can laminated channel fills on different terrace levels be linked to each other or to major flood events?

The first question needs to be answered since the water height above the bed is one of the inputs in the Chézy and Manning's formulas. The water height above the bed is determined by the actual water level and the elevation of the bed. For the latter, knowledge about the timing of the event is important, since the morphology and the elevation of the bed have changed over time. This change is described by the answer to question two. The answer to this question also results in channel roughness and slope estimates that are used as inputs for the Chézy and Manning's equation. Water level data were derived from palaeostage indicators (PSI's) (e.g. residual channel fills, slack water deposits and pointbar features) gathered during a four week fieldwork campaign. The palaeo geography was determined with the use of core descriptions from the fieldwork and previous research, a Digital Elevation Model (DEM) and datings (pollen analysis and OSL and ^{14}C datings in literature).

The answer to question three results in the last input that is needed. The vegetation roughness was estimated using literature and oral communications with experts in the field of (riparian) vegetation at the Alterra research institute (Wageningen University) and were cross-checked with pollen data from the channel fills in the research area (Geurst, in prep; van Munster, in prep; Bunnik (TNO), pers. Comm.). The actual roughness values were calculated using the USGS guide for determining Manning's roughness values (Acrement & Schneider, 1989) and by using a method by Baptist et al. (2007) for determining Chézy roughness values for vegetation. The final question, question four, was used to determine whether the proxies that were used to answer question one are the result of local or regional events. This was necessary as local events (for instance: A clastic layer in a channel fill caused by bank collapse or a falling tree that might be misinterpreted as a flood layer) might lead to incorrect water level estimates.

The answers to these questions were used as input for a 1D hydrological model based on Manning and the Chézy. The model was run for different scenarios, representing different time slices, uncertainties in palaeo channel dimensions and the seasonal differences in vegetation cover of the floodplains. The results are presented as (i) an estimate of a discharge that is the absolute realistic minimum discharge for the water required to reach the reconstructed water heights; (ii) a range of most likely discharges; (iii) a maximum possible realistic discharge for the water level elevation found.

2 Review of literature

2.1 The concept of palaeo hydrological flood research

To reduce risks associated with extreme floods, there is a critical need to increase the length of the flood records beyond that of the instrumental period (Thorndycraft, 2005). This can be done in a number of ways using palaeo-hydrology. Palaeo-hydrology is the study of past or ancient flood events that occurred prior to the time of human observation and/or direct measurements by modern hydrologic procedures (Baker, 1987). Published palaeohydrological studies can roughly be divided into two types: The first type focuses on flood frequencies and flood frequency distributions (e.g., Pinter et al., 2006; Glaser et al., 2008; Macklin & Lewin, 2008; Macklin et al., 2006; Vis et al., 2010; Brázdil et al., 1999; Benito et al., 2008; Glaser & Stangl, 2003; Glaser & Stangl 2004; Knox, 2000; Böhm & Weztel, 2006; Cyberski et al., 2006; Mudelsee et al., 2006; Thorndycraft et al., 2005; Brázdil et al., 2006; Minderhoud et al., 2010). The second type of research focuses on estimating actual discharges during palaeoflood events (e.g. Herget & Meurs, 2010; Sheffer et al., 2008)

Studies that focus on flood frequencies most often rely on documentary data. Examples of such data include: Narrative written sources (annals and chronicles), visual daily weather records, parish registers, personal correspondence, special prints, official economic records, newspapers, pictorial documentation, Stall-keepers' and market songs, scientific papers and communications, epigraphic sources and early instrumental records (Bràzdil et al., 2006). These sources all have to be carefully checked since they often are not of first hand, biased by the objectives of the publishers (in the hope of financial support), or uncertainties arising from not knowing the metric equivalent of the utilized unit, the exact location where the measurement was taken or the physical characteristics of the instrument itself (Glaser, 2010). Another factor is the availability of documented data, which is biased in time and space. Going back further in time, sources get scarcer and will often only cover more extreme events (Glaser & Stangl, 2003).

Other sources for flood frequency data that are less often used are fluvial records (e.g. Slackwater deposits, erosion lines, soil formation, morphology). These proxy do not suffer from the same uncertainties as the documented sources, however, they are open to misinterpretation, can be difficult to (accurately) date/correlate or might contain a hiatus due to erosion. It is rare that a single slackwater depositional site will preserve a relatively complete record of major floods on a river (Baker, 1986).

Studies of this type do not reach consensus on spatial and temporal frequency distributions. Where Macklin & Lewin (2008) reported a continental wide trend in European flood frequencies throughout the Holocene, which was put in perspective by Brázdil et al. (1999) and Glaser et al. (2010) who concluded that long term flood frequencies changed significantly over time within different river catchments, with little spatio-temporal coherence. The main coherence that was found is the increase in flood frequencies in periods with a cooler climate like the Little Ice Age (LIA) and at the onset of the 20th century; for other periods the results of the different studies differ considerably (Brázdil et al., 1999; Brázdil et al., 2006; Macklin et al., 2006; Glaser et al., 2010).

Studies focusing on palaeoflood discharge estimations (e.g. Sheffer et al., 2008); mostly use the same sources as studies focused on flood frequency distributions and thus suffer from the same drawbacks. These studies make use of one dimensional hydrological models and palaeo stage indicators (PSIs) (e.g. Herget & Meurs, 2010; Thorndycraft, 2005; U.S. Department of the Interior, 2003; England Jr. et al., 2010). Natural palaeo stage indicators are for example: Slackwater flood deposits, eddy bars, deposits in tributary mouths and flood erosion lines (Benito et al., 2004; Sheffer, 2008), but documented sources also provide an indication for palaeo stages (Herget & Meurs, 2010). Palaeo stage indicators include indicators for the minimum water level reached (often the case with natural PSIs), for the exact water level reached (often marks in buildings, documents or flood erosion lines) or for non-exceedence bounds (e.g. undisturbed soils, documents) (Benito et al., 2004; England Jr. et al,. 2010).

Hydrological modeling of palaeofloods requires the estimation of the hydraulic characteristics of the river reaches, such as slope, roughness and wetted cross sections (Benito et al., 2004). In the case of non-alluvial rivers, it can be assumed that river morphology has remained stable on the relative short time scales of the Holocene. In that case, the wetted cross section can be very accurately determined using the current morphology combined with water levels derived from the PSIs (e.g. Benito et al., 2004; Thorndycraft, 2005; U.S. Department of the Interior, 2003; England Jr. et al., 2010). In case of alluvial rivers the river morphology can not be assumed to be stable. Therefore a careful reconstruction of the cross section at the time of the palaeoflood needs to be obtained. This requires assessment of for example incision rates, past vegetation characteristics, channel location/dimensions and floodplain morphology. Uncertainties in these reconstructions cause uncertainties in the flood discharge estimates. For this reason, most palaeoflood discharge studies focused on bedrock river reaches, while comparatively few studies were performed in alluvial river valleys (e.g. Herget & Meurs, 2010). Uncertainties in roughness determinations and water-surface elevations relative to PSIs increase errors in discharge estimates in both alluvial and no-alluvial rivers (Webb & Jarrett, 2002). Baker (1987) concluded that, despite these uncertainties, palaeoflood magnitude determinations using PSIs can be greatly superior to historic flood discharge data.

2.2 Palaeohydrological studies in the Rhine catchment

Palaeoflood studies in the Rhine catchment mainly focused on flood frequency reconstructions (e.g.: Brázdil et al., 1999; Glaser & Stangl 2004; Glaser et al., 2010; Minderhoud et al., 2010; Pintern et al., 2006; Tol & Langen, 2000), although some also include magnitude indications. Only Herget & Meurs (2010) used PSIs for estimating actual discharges during the 1342 AD flood event in the German city of Cologne. Studies in the Rhine catchment on flood frequencies are almost solely based on documented data with their inherent drawbacks and advantages as described in the previous section, with the exception of Minderhoud et al. (2010) who based their analysis on sedimentary proxies (channel fills). The drawbacks of using documented data become apparent when

the result of the various studies are compared. They show very little consensus on the timing of periods with a higher flood frequency. For instance, Tol & Langen (2000) show an increased frequency of high magnitude floods between 1450 AD and 1700 AD and a very low frequency for the period around 1300 AD. Herget & Meurs (2010) did however find a number of sever floods in both periods, Glaser & Stangel (2004a, 2004b) report a period of relatively low flood frequencies between 1450 AD and 1550 AD. Finally Glaser et al., (2010) find a very low flood frequency in the River Rhine for the period between 1750 AD and 1800 AD and a period of high flood frequencies between 1800 AD and 1900 AD. The variability in outcomes of these studies makes a link to climate fluctuations uncertain, though most studies tend to agree on a rise in flood frequency during periods of a cooling climate (Little Ice Age) and after the onset of the twentieth century (Brázdil et al., 1999; Glaser & Stangl 2004; Glaser et al., 2010; Pintern et al., 2006; Tol & Langen, 2000). The rise in flood frequency in the twentieth century can likely be linked to the development of highly efficient drainage networks in German rural areas for the benefit of agriculture. These drainage systems deliver runoff in high volumes and high rates to stream networks (Pinter et al., 2006).

Besides flood frequency distributions over the last millennium, the flood frequency distribution of the river Rhine over the seasons is also still open to debate. Brázdil et al. (2006) found that 50% of the floods in the last 500 years in the River Rhine occurred during summer, mainly in July, 23% of the floods occurred during the winter months, mainly in December. Mudelsee et al. (2006) state that the overall risks of floods in winter (November – April) is approximately 3.5 times higher than the summer flood risk. Winter flood risk peaked around 1760 and 1860, summer flood risk peaked at around 1760. Finally Pinter et al., (2006) conclude that peak flows along the upstream half of the River Rhine generally occur during late spring to summer, whereas peak flows tend to occur during winter along the downstream half of the river. While spatial and temporal differences in the focuses of the studies might slightly differ, it is obvious that no consensus has been found on this subject. However, these studies do make it clear that flood events are not restricted to one season and can happen both in the winter/early spring and in the summer.

The studies into flood frequencies mentioned in the previous paragraph all rely on documentary data. As of now, only Minderhoud et al. (2010) performed a study on Rhine flood frequencies solely based upon sedimentary proxies. This study is focused on the channel fill of a residual channel of the River Rhine near the city of Rheinsberg (Germany) that was abandoned circa 5000 BP. This channel fill has acted as a sediment trap, trapping suspended sediment each time the active Rhine flooded the residual channel, while the period in between floods is market by the deposition of organic material. The result is a long sequence of layers with varying organic matter content. Minderhoud et al. (2010) studied this sequence in the channel fill and constructed an age/depth relationship after which they could reconstruct the flood occurrences by simply counting the layers with a particularly low organic content, representing the sedimentation of a flood layer (Fig 1). The deviation of organic content in these layers from the mean organic content was used for estimating flood magnitude. The results show a period between 2500 BP and 2900 BP with little to no flood activity, however, this might also be related to the distance of the residual channel to the river in this period. As the River Rhine was very active in this area during the studied period, the location of the main channel frequently shifted. At times when the active channel was located farthest away from the residual channel, less flood signals are to be expected in the channel fill, because during smaller floods, the sediment might not reach the location of the channel fill.
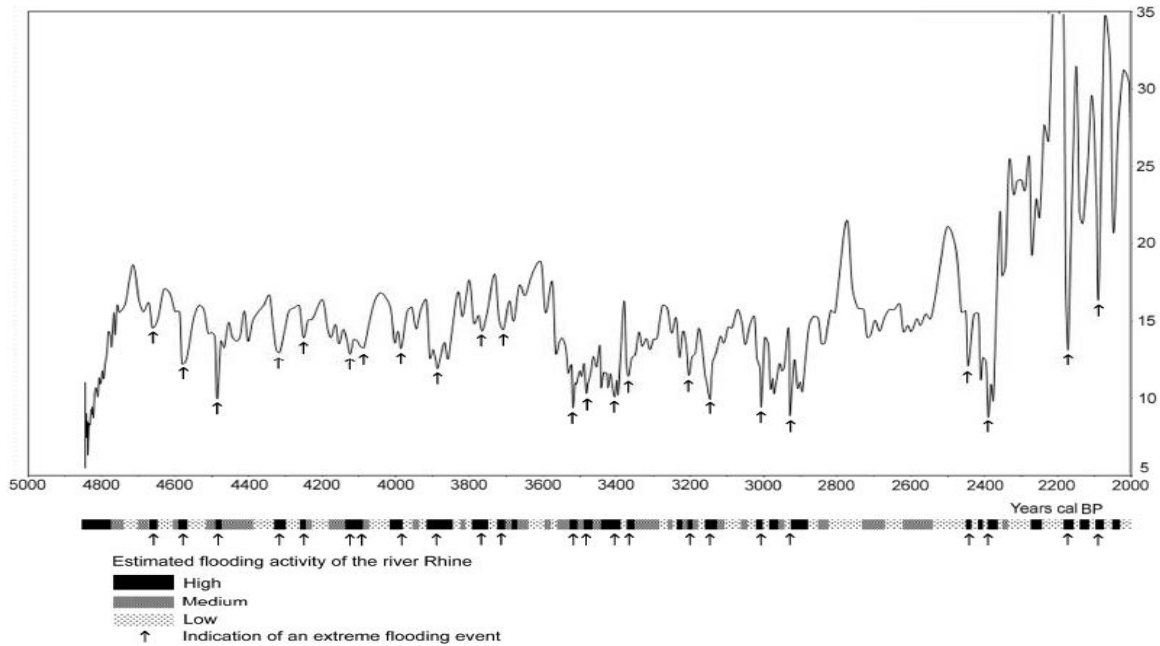
Figure 1: LOI analysis of the Rheinberg residual channel fill (Minderhoud et al., 2010).

Herget & Meurs (2010) have performed the only study as of yet reconstructing palaeo or historic discharges in the River Rhine. They focus upon the catastrophic flood event in 1342 AD at the city of Cologne (Germany). Using documented data the environment and water levels at the time of the event were reconstructed. This reconstruction was used to determine both the part of the valley cross section that was inundated as the distribution of the hydraulic roughness along this cross section. Hydraulic roughness is caused by the drag water experiences as it is forced to flow around objects like trees and houses. The determination of the roughness, combined with the slope, the water height and the wetted perimeter (the part of the cross section flooded by water) enabled them to use the Manning's formula (section 2.3.1) for reconstructing the peak discharge for the historic flood level of 1342 AD. For this event they found a discharge range of 18,800 m^3/s to 29,000 m^3/s, with a most likely discharge of 23,800 m^3/s.

2.3 Hydrology: Calculating discharges.

The field of palaeohydrology is a combination of a geological and a hydrological approach. The geological approach is aimed at finding the parameters (e.g. roughness,

palaeogeography, water levels) which are used in the hydrological approach to calculate the actual discharges. Two of the most often used hydrological approaches in studies are the approach by using the Manning's equation and by using the Chézy equation. This section gives a short overview of both methods to estimate flood magnitudes as done in this study.

2.3.1 Discharge calculations using the Manning equation

The equation as developed by Manning to calculate flow velocities (and thus discharges) for hydraulically rough is most commonly used in researches because of its simplicity (Eq 1) .

$$Eq.\ 1:\ V_p = \frac{R_p^{2/3} S^{1/2}}{n}$$

In which V_p is the mean flow velocity in m/s, R_p the hydraulic radius of the water flow during the flood event, S is the slope of the energy line and n is the coefficient of roughness, specifically known as Manning's n. The hydraulic radius can also be defined as the cross sectional area (A) of flow divided by the wetted Perimeter (P); Eq. 2.

$$Eq.\ 2:\ R_p = \frac{A}{P}$$

The greatest difficulty in applying the Manning formula is in determining the roughness coefficient n, since there is no exact method for selecting the n value. To help with selecting appropriate n values, the United States Geological Survey (USGS) (Arcement & Schneider, 1989) developed methods for determining n values for channels and for floodplains. The n value to be used is the sum of the n values of individual drag causing factors. For channels the n value can be computed by using equation 3 (Arcement & Schneider, 1989):

$$Eq.\ 3:\ n=(n_b+n_1+n_2+n_3+n_4)m$$

n_b Is a base value for a straight, uniform, smooth channel in natural materials, n_1 is a correction factor for the effect of surface irregularities, n_2 is a value for variations in shape and size of the channel cross section, n_3 is a value for obstructions, n_4 is a value for vegetation and flow conditions and m is a correction factor for meandering of the channel. The n value for floodplains can be computed using the same formula (Eq. 3), with n_b as the base value of n for the floodplain's natural bare soil surface, n_1 is a correction factor for the effect of surface irregularities on the floodplain, n_2 is a value for variations in shape and size of the floodplain cross section (assumed equal to 0), n_3 is a value for obstructions on the floodplain, n_4 is a value for vegetation on the floodplain and m is a correction factor for sinuosity of the floodplain, equal to 0 (Chow, 1959; Arcement & Schneider 1989; Anderson et al. 1996; Herget & Meurs, 2010).

2.3.2 Discharge calculations using the Chézy equation

Antoine de Chézy published the still often used Chézy formula for calculating flow velocities in open channels (Eq. 4):

$$Eq.4:\ u = C\sqrt{Ri}$$

In which u is the depth average velocity, C is the Chézy value representing roughness ($m^{0.5}/s^{-1}$), R is the hydraulic radius and i is the energy slope. In rivers with a large width/depth ratio the hydraulic radius is replaced by h which represents the water height above the bed. Note that a higher Chézy value represents a lower roughness. The White-Colebrook formula gives the Chézy value for hydraulically rough flow:

$$Eq.\ 5:\ C = 18\log\left(\frac{12R}{K_n}\right)$$

In which K_n is the Nikuradse equivalent roughness. In natural channels the K_n value can be replaced by 3 times the D_{90} of the grain size of the bed (Anderson et al., 1996; Baptist et el., 2007; Chow, 1959; Liu, 2001; Van Rijn, 1984), resulting in:

$$Eq.\ 6\text{:}\ C = 18\log\left(\frac{12R}{3D_{90}}\right)$$

During a flood, as is researched in this study, water flow is no longer restricted to the channel, as the floodplains are inundated as well. While the roughness in channels is mainly determined by the friction of the bed, this friction is usually negligible on floodplains compared to the roughness exerted by the vegetation to the overbank flow. It is therefore necessary to establish a separate equation for flow over vegetated floodplains; this has been attempted in several studies (e.g., Campana, 1999; Darby, 1999; Wilson, 2007; Ebrahimi et al., 2008). After comparing multiple approaches in different studies on this subject, Baptist et al. (2007) state that in the case of submerged vegetation the flow resistance can be described as:

$$Eq.\ 7\text{:}\ C_r = \sqrt{\frac{1}{\frac{1}{C_b^2} + \frac{1}{2g}C_D mDk}} + \frac{\sqrt{g}}{\kappa}\ln\left(\frac{h}{k}\right)$$

In which Cr is the representative Chézy value for vegetation, C_b the representative Chézy value for the bed roughness, C_D the bulk drag coefficient, m number of vegetation individuals expressed as cylinders per square meter, D the stem (cylinder) diameter, h the water depth, g the gravitational acceleration (9,81 m/s^2) and κ is a scaling factor for the logarithmic flow profile, the so called Karmann's constant 0.41. The first term in the equation describes the flow trough the vegetation and the second term describes the flow over the vegetation. In the case where the vegetation is not fully submerged, this term is equal to zero. (Keijzer et al., 2005; Baptist et al., 2007).

2.4 Vegetation reconstruction for the Lower Rhine Embayment

The discharge for a flood event can be calculated by adding the discharge through the channel (Equations 1 and 4) to the discharge over the floodplain (equation 1 and 7). In order to use equation 7 and for determining the n value for equation 1 a detailed knowledge of the vegetation density on these floodplains is needed. For the period between the start of the Holocene and 3000 yr. B.P. the vegetation on the floodplains of the River Rhine has seen little to no human influences (Bos et al. 2003, Bunnik, pers. comm.). This research will thus assume that the vegetation on the floodplains consists of natural vegetation. As completely natural vegetation is presently non-existent in the Lower Rhine Embayment, this research will rely on literature data and expert opinions.

In northwestern Europe riparian vegetation is divided into two groups; softwood forests (*Salicion albae,* Fig. 3) and hardwood forests (*Ulmenion carpinifoliae*) (Rademakers & Wolfert; 1994). Softwood forest consists of fast growing vegetation and is highly resistant to floods, high sedimentation rates and long periods of inundation, while the hardwood forest needs more temperate, less dynamic, conditions and a relatively low groundwater level. The composition of species in these forests depends on the severity of these conditions and on soil composition. Figure 2 represents a schematic representation of a typical profile lowland river valley. The profile is subdivided in zones based on these conditions.
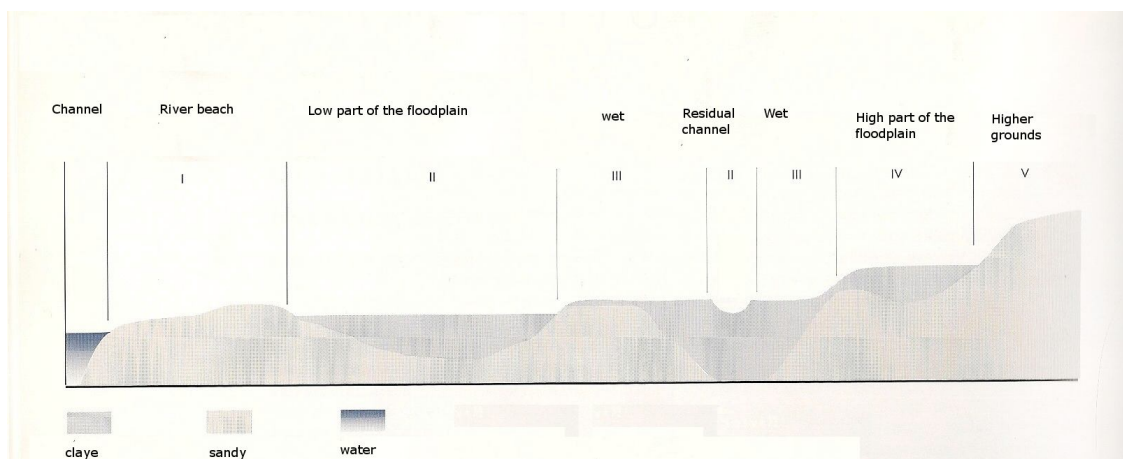

Figure 2: Schematic visualization of riparian vegetation zones (after Wolf et al., 2001)

Zone 1 (Fig. 2) consists of the river shores. Inundation occurs more than 60 days per year, clay content is low (below 10% in the top 0.25m) and nutrients are readily available. *Populus nigra* (Black Poplar) and *Salix alba* (White Willow) are the dominant tree species, *Artemisia vulgaris* (Mugwort or Common Wormwood) and *Salix viminalis* (Common Osier) the dominant shrub species. Zone 2, the low parts of the floodplain and the filled residual channels with a high clay content (larger than 10% in the top 0.25m). It is more than 60 days per year inundated and nutrients are readily available. Here, *Salix alba* is the dominant tree species and *Salix viminalis*, *Salix triandra* (Almond Willow) and *Crataegus monogyna* (Common Hawthorn) are the dominant shrub species. Zone 3, the moist floodplain that is 10 – 60 days per year inundated, with a high (larger than 10% in the top 0.25m) clay content and readily available nutrients; *Salix alba* is still the dominant tree species here and *Salix viminalis*, *Salix triandra*, *Salix cinerea* (Grey Willow) and *Crataegus monogyna* are the dominant species in the shrub zone. Zone 4, the high floodplain that is inundated 1 – 10 days per year at most, with a relatively low ground water level (more than 1.2m below surface) and a high (more than 10% in the top 0.25m) clay content, *Fraxinus excelsior* (Ash) and *Quercus robur* (English Oak) are the dominant tree species, while Sambucus nigra (Elder) and *Crataegus Monogyna* are the dominant shrub species in this zone. Zone 5 induces the levees and river dunes which are inundated less than 10 days per year, have a low groundwater level (more than 1.2m below field), a low (smaller than 10% in the top 0.25m) clay content, while nutrients are more scarce, *Fraxinus excelsior*, *Quercus robur*, *Ulmus minor* and *Populus canadensis* are the dominant tree species and *Sambucus Nigra* and *Crataegus monogyna* are amongst the dominant shrub species along with other shrubs, such as *Corylus avellana* (Common Hazel) and *Prunus serotina* (Black Cherry). It should however be noted that *Populus canadensis* found it's origin in France in 1750 after cross breeding. (Wolf 2001; Oral communication with Alterra).

All mentioned shrubs (with the exception of *Artemisia vulgaris*), hard wood and soft wood tree species lose their leaves in autumn and regenerate new leaves in spring to form a dense foliage, thus showing seasonal variety in density. *Artemisia vulgaris* dies in autumn, leaving a dense concentration of dead, rigid, leafless stems. (Weeda et al., 1985).

Shrubs form a dense sub layer of roughly one meter in height during spring and summer. In winter, with the exception of the dead stems of *Artemisia vulgaris*, this dense layer completely disappears. The tree density of the *Salicion albae* is around 275 individual stems per hectare for a fully developed forest. The height reached by these individuals under average circumstances is around 24 to 32 meters for Willow and 32 to 40 meters for Poplar. Average stem diameter at breast height is 0.4 m for both Willow and Poplar. (Wolf et al., 2001; Weeda et al., 1985-1994; Oral communication with Alterra). These values represent average healthy natural communities with no major recent disturbances. However, disturbances on the banks of a natural river are frequent, this dynamic environment creates a complicated pattern of patches of fully grown forests and pioneer vegetation of greatly varying extent. Especially *Salicion albae* is subject to frequent flooding, during which dead wood piles up, young specimens get flushed away and large parts may be destroyed by ice when the flood occurs in winter or early spring. The fully grown (climax) forest (a) will still have some bush like Willow and a dense brush layer. The pioneer vegetation (b) consists of a very dense and almost impenetrable bush layer of Willows with almost no underbrush. The intermediate mixed forest (c) will have elements of both the climax and the pioneer forest. The ratio between these forest types is roughly estimated a : b : c = 2 : 1 : 1 (Oral communication with Alterra).

Figure 3: A Salicion albae forest near Neerijnen, the Netherlands. Photo: M.M. de Molenaar

The habitat of *Ulmenion carpinifoliae* is less dynamic due to less frequent flooding as a result of the greater distance and elevation to the river. The more stable environment leads to a higher ratio of patches with a fully grown (climax) forest (a) as opposed to the pioneer forest (b) and the intermediate mixed forest (c). The ratio between these forest types is roughly estimated a : b : c = 5 :1 :2. It should be noted that there are no strict boundaries between vegetation and forest types, transition is nearly always gradual and at places interrupted (Weeda et al., 1985-1994; Wolf et al., 2001; Oral communication with Alterra).

3 Methods

3.1 Research area selection

The research area is situated around the delta apex of the river Rhine near Rees, Germany (Fig 4). This area is selected for multiple reasons: This research requires an area with a considerable number of river terraces with residual channels for sediment trapping. For an accurate water height estimation it is important that the difference in elevation between the terraces is small (see section 2.1).
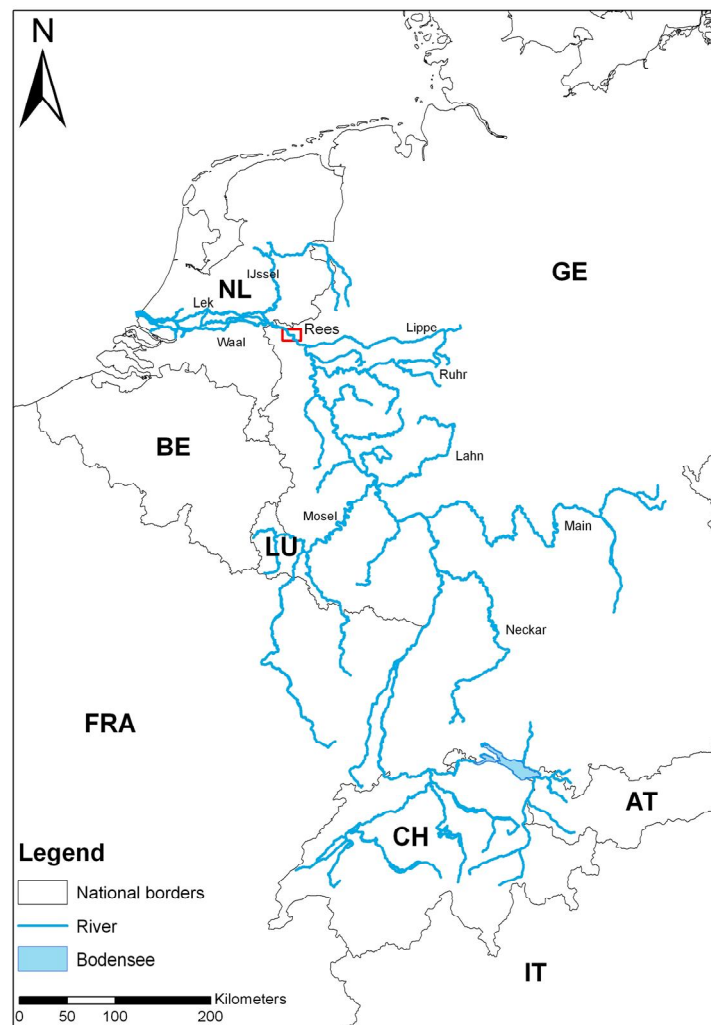

Figure 4: The Rhine with its tributaries. The fieldwork area near Rees is indicated by the red square.

Another requirement is a downstream location from the last significant tributaries, the rivers Lippe and Ruhr (Fig 5). The highest measured discharges for these rivers are 396 m^3/s for the river Lippe and 906 m^3/s for the river Rurh (Landesumwelambt Nordrein-Westfalen, 2002).
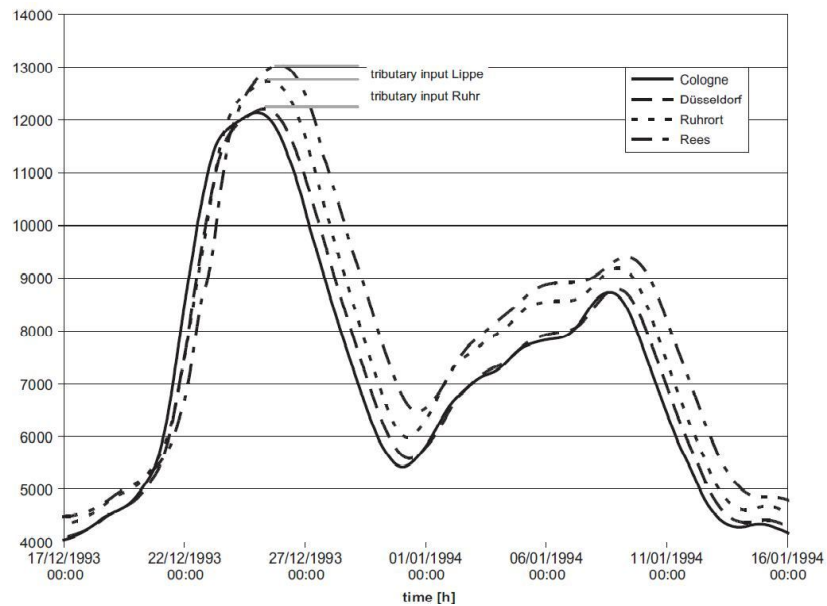

Fig 5: The contribution of the Ruhr and Lippe tributaries to the 1993 flood in the River Rhine (Apel et al, 2009)

The coincidence or lack of discharge peaks in the tributaries and the main Rhine can have a considerable impact on the downstream peak discharge, as discharge peaks will be significantly higher when the discharge peak in the main river passes at the same moment as the discharge peak of the tributary is added to the main stream. For example: the 1/35 yr. flood of the river Rhine in 1993 started as only a once in two years discharge at the Alps, at the same time the Neckar tributary experienced a 1/50 yr. discharge. After further contribution by the Main and the Nahe tributaries the flood wave in the Rhine developed into a 1/35 yr. flood wave (Silva et al., 2001). Finally, it is of importance that the floodplain is neither too wide, too small nor too irregularly shaped. A floodplain that is too wide, like in the delta, will mean that a considerable increase in discharge will not lead to a noticeable increase in water depth over the floodplains. On the other hand, a floodplain that is too narrow will probably imply that the river has eroded much of the

evidence needed for making palaeo reconstructions of the wetted perimeters. The valley should also be not too irregularly shaped. An irregular shaped valley will cause extra turbulence induced drag as the water needs to flow around or over the obstacles and this will influence the water height, for this reason it is also important that there are no obstructions or bifurcations just upstream or downstream of the research area as the backwater effect caused by this will have consequences for the water height. Therefore, it is important that the research area was not located too close to the downstream bifurcation near the present day Pannerdens Kanaal near the Dutch city of Nijmegen. The area around Rees ticks all boxes; the succession of river types during the Last Glacial - Interglacial transition has caused accelerated and interrupted intervals of incision, resulting in multiple terraces, all with a different mean elevation. The cross section of the river is narrow enough for the water to reach several of these terrace levels during high water conditions, yet there still is enough evidence of previous main channel locations for palaeo cross section reconstructions.

3.2 Coring and sampling

For this research a total of 208 corings with a combined depth of 642.1 m were drilled. The corings were used for analyzing the subsoil composition, reconstructing palaeo environments (e.g., the location of the river, channel dimensions) and analyzing flood proxies in residual channel fills.

The locations of the corings were selected along several transects across residual channels to determine their age and dimensions, and along two valley wide transects to gather data about slopes and morphogenesis. The locations were chosen with a relative distance of roughly 200 meters, depending on the morphology; core locations were chosen at either ridges or channels still visible in the field. This was done to prevent confusion because of samples taken at the transition zone between two systems or morphological features, as samples taken in transitional zones might contain features of multiple systems. This might result in an erroneous interpretation of the samples. (Fig 6; Appendix 2)
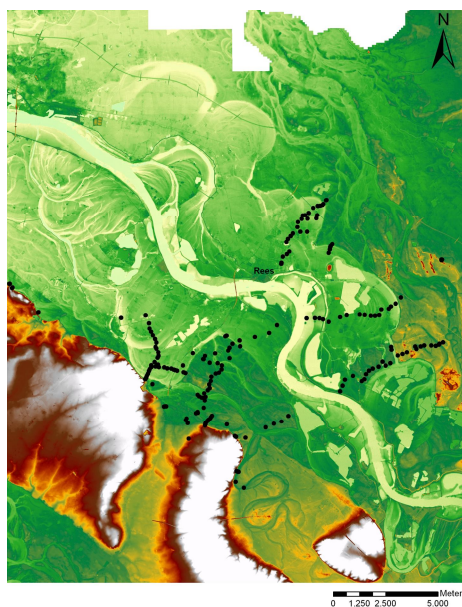
Figure 6: Coring locations

The corings were all drilled by hand using the Edelman auger, Dutch gouge, Van der Staay suction corer and Boncke piston corer. The Edelman auger was used for unconsolidated sediment above groundwater level. The Dutch gouge was used in non-consolidated cohesive materials below groundwater level like clays, gyttjas and peat. The Van der Staay suction corer was used in sandy and gravely materials (with a grain size up to 4 centimeters) below groundwater level. The Boncke piston corer is a modified Livingston piston corer that was used for retrieving sample material of a length of up to 1 meter and a diameter of 0.1 meter. It was used in unconsolidated, cohesive materials below groundwater level like clays, gyttjas and peat.

Cores taken in the field by the Edelman auger, the Dutch gouge or the Van der Staay suction corer were sampled in the field and were described according to the system developed by Berendsen & Stouthamer (2001). The sampling was done every 10 centimeters and the texture, gravel content, organic matter content, Ca and Fe content, oxidation/reduction and other noticeable details (like disturbances, the presence of anthropogenic materials and micro sedimentological characteristics such as lamination thickness) in the lithology were described. The cores taken with the Boncke piston corer were taken to the laboratory of the department of Physical Geography at Utrecht

University for further analysis. The Boncke piston corer was in this research solely used for taking samples in residual channels since these offer the best possibility of long and detailed flood records. The exact locations were selected on base of the depth and fill of the channel. These locations were: The Mariënbaum residual channel, the Hohe Ley residual channel, the Vosse Kuhl residual channel, the Schloss Bellinghoven residual channel, the Haus Groin residual channel and the Heeren Bril residual channel (Fig 7).
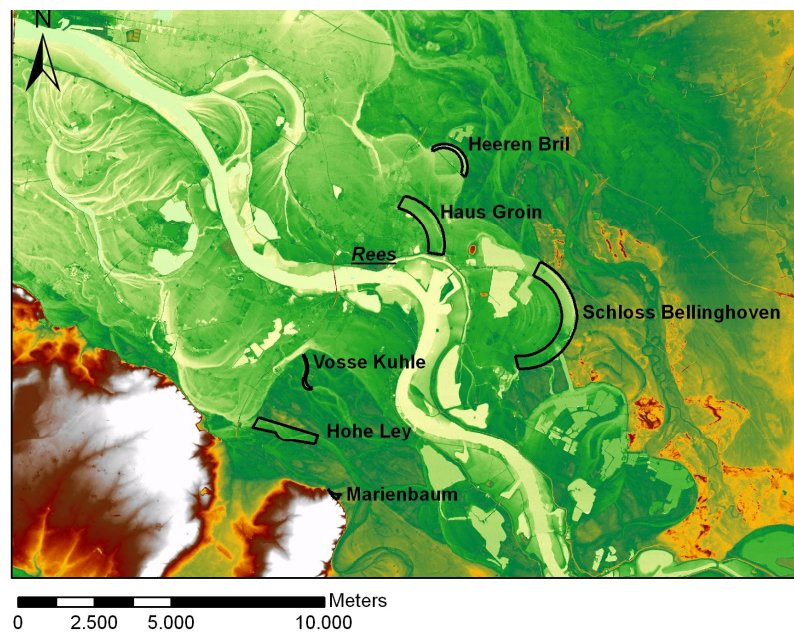
Figure 7: Locations of the residual channels that were sampled with the Boncke piston corer.

In most cases the channels were sampled in their deepest part since this part is most likely to contain the longest and most detailed flood records. In the case of the Schloss Bellinghoven channel sampling the deepest part was not possible because the channel is still partly filled with water. In the case of the Heeren Bril residual channel, the core was taken at an upstream location of the transect over the channel and it is uncertain if the sampling was done at the deepest part of the channel. This location was chosen because preliminary corings done with the Dutch gouge and the Edelman auger had shown that the sediments at this location showed a visually better lamination than the sediments found at the location of the cross section over the channel.

3.3 Loss on ignition

In the laboratory, the cores were opened and photographed. The cores were sampled every 1 or 2 cm, depending on the thickness of the visible layers. Each sample was roughly 1cc in size to be analyzed with the use of the LOI (Loss On Ignition) method. This method was applied in order to find flood signals that were recorded in the channel fill. The size of individual samples varied and was based upon the visible layering and was taken in such a way that one sample represented preferably one layer and did not contain the transition to another layer. This was done to prevent averaging out of the results of the LOI analysis when a sample contains material from both a layer with a high and a layer with a low organic content.

Samples were subsequently weighted, placed in a stove and heated to 105 degrees Celsius for at least 12 hours. During this period the water in the samples evaporated and when the samples were weighted again afterwards, the moisture content was determined. After this procedure the samples were placed in an oven and heated to a temperature of 550 degrees Celsius for a period of 5 hours. These temperatures cause the organic material in the samples to combust (Heiri et al., 2001). By weighting the samples again directly after they exit the oven, the loss in weight due to combustion of organic material was determined in percentage loss.

The result of this procedure was analyzed with change point analysis (CPA). This was done in order to filter out trends caused by for example the distal effect; the further away the river is from the channel fill during the filling process, the more organic the infill will be as less disturbances will bring less and finer grained clastic material into the channel. When a river abruptly changes location an abrupt shift in the LOI over depth will be present (Minderhoud et al., 2010). The CPA removes these trends, which leaves pure spikes (relative weight losses). Next the mean and the standard deviation of the corrected data were computed to filter out background noise. Background noise could for instance be caused by minor flood signals, measurement errors (e.g., when the weight loss percentage is high, the weight of the remaining sample after the LOI procedure is small.

Variations in the outcome of the weighting (caused by for instance air pressure fluctuations) and sample anomalies (e.g., when a sample contains a piece of wood, the percentage weight loss will be relatively large due to the small sample sizes) will have a profound effect). Any peak in the corrected LOI curve that has a deviation from the mean of less than the standard deviation was considered background noise, any deviation larger than the standard deviation was considered a flood signal when the deviation points at a higher clastic content than average.

3.4 Constructing the 1D Hydrological model.

A 1D hydrological model was used to calculate the discharge numbers for water level markers in residual channel fills and on the floodplains. The model consists of a valley cross section that is easily adapted for different time slices with associated floodplain morphology and channel characteristics (Fig. 9). The location of the cross section was chosen in such a way that it covers various terrace levels with channel fills while keeping the effects of narrowing, widening and other effects of the valley on the hydrology in mind (Fig. 8). The cross section was made with the use of a recent DEM (Digital Elevation Model, provided by Landesvermessungsamt Nordrhein-Westfalen, Germany). Over a length of 13,340 m the elevation was sampled from the DEM every 10 meters and if necessary corrected for human influences (constructed dykes, roads and floodplain lowering as a flood mitigating measure or caused by the mining of gravel: Between a distance of 3760 meters and 4190 meters and 5690 and 6300 meters from the start of the transect in the west, the floodplain is currently lowered as a result of gravel mining. This was resolved by locally raising the floodplains in the model by 3 meters to the level of the surrounding floodplains). In the east, the cross section includes the Lange Renne lake. Because of its location and unnatural rectangular shape this lake was assumed to be man made as a flood mitigation measure. It was assumed that such a measure will have been dug in an already existing depression in the landscape to save expensive and time consuming labour. The Lange Renne was thus replaced in the model by a channel with the same dimensions as the residual channel found just to the west of the Lange Renne.
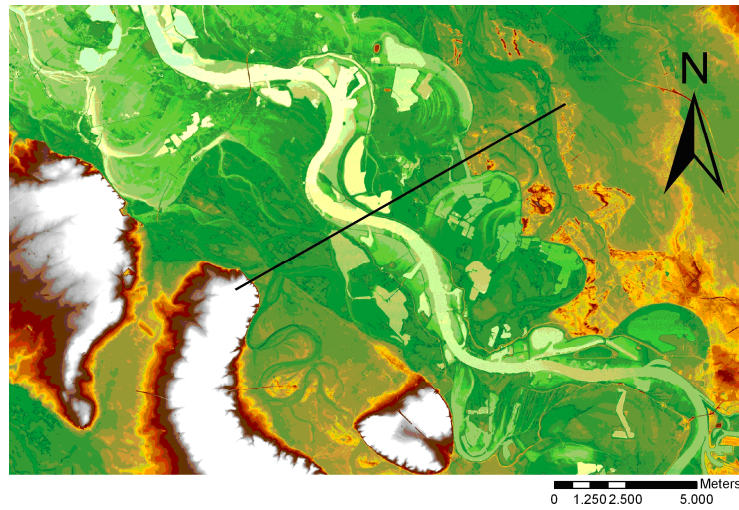
Figure 8: The location of the cross section in the research area. Light green represents the areas with the lowest elevation (~10 m +O.D), red and white the areas with the highest elevation (~70m +O.D.).

The cross section was divided into 5 segments (Fig. 9), coinciding with the vegetation zones described in section 2.4. Each segment was selected on the basis of slope, elevation above the present floodplain, subsoil composition and distance to the river. For each subsection vegetation, palaeotopography and the slope were reconstructed. The vegetation was reconstructed using literature and expert opinions, this was cross checked with pollen analysis of channel fills in the area (Geurts, in prep; Bunnik, pers. comm.). The slope and palaeotopography were reconstructed using literature, the DEM and field data obtained during fieldwork.
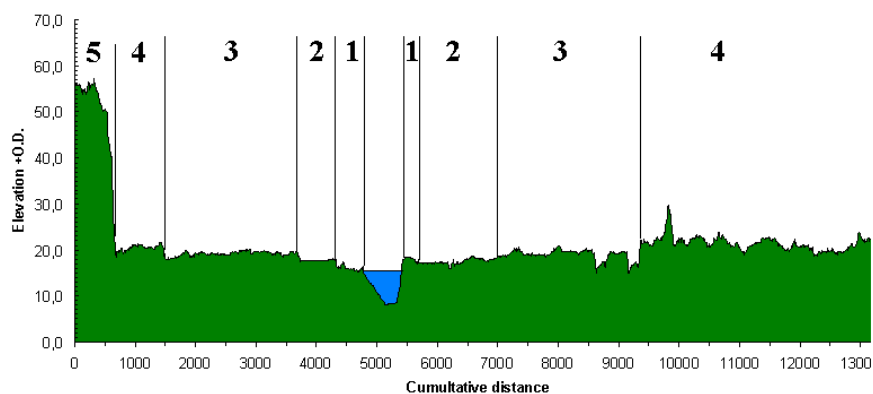

Figure 9: The cross section used in the model. The numbers indicate the vegetation zones as described in section 2.4.

The cross section covers various channels and depressions on several terrace levels. In order to check whether these channels and depressions became inundated for the input water heights in the model a sloped plane with the slope of the river surface was plotted trough the DEM and the specified rise in water level in meters was added to this plane (Fig 10). Using this method it was determined which parts of the cross section get inundated at different water levels. This was a necessity since depressions in the cross section might not be inundated at water levels that exceed the elevation of the bottom of the depression because of blockages upstream and thus were excluded from the calculations.

The cross section of the main channel was based on the dimensions of the various residual channels in the area. These dimensions were obtained by separate coring transects, made over the residual channels of Heeren Bril, Haus Groin and Schloss Bellinghoven to determine their deepest point for LOI analysis and to determine the channel dimensions. For accurate depth estimates of the palaeo channels, the thickness of the top layer of clay is subtracted as it is an overbank deposit that is younger than the river system that formed the residual channel. Not removing this layer would cause an over estimation of the channel dimensions.

The floodplain topography was reconstructed using the incision rate of 1mm/yr as found by Erkens (2008). The parts of the transect used in the model to represent the Holocene floodplains were raised with 0.5 and 0.9 meters (as opposed to the current floodplain elevation) respectively for the Middle and Early Holocene scenarios to compensate for the incision.

The roughness for the channel was obtained from data used by Hesselink et al. (2006). This data is a set of re-evaluated flow measurements by Brunings, taken 1790 and 1792 AD in the River Rhine near the Dutch-German border. While these measurements were taken under normal flow conditions, they are assumed to be the most useable measurements for these research since the river was much more natural at the time of

those measurements than it is now (for instance, no groins were constructed yet at the time of the measurements). It was chosen not to calculate the roughness of palaeo channels using grain sizes obtained with corings, since the used methods for coring did not allow for grain sizes larger than 5 cm's to be sampled. Corings also only offer a pinhole view into the sediments, making it impossible to include the effect of large scale bed forms into the calculation of the roughness. The Chézy roughness for the floodplains was calculated by using equation 7. For the winter scenarios, it was assumed that bare winter vegetation has no influence on the tree density. The Manning surface roughness was determined using the manual for selecting the Manning's *n*, provided by the USGS (Arcement & Schneider, 1989).
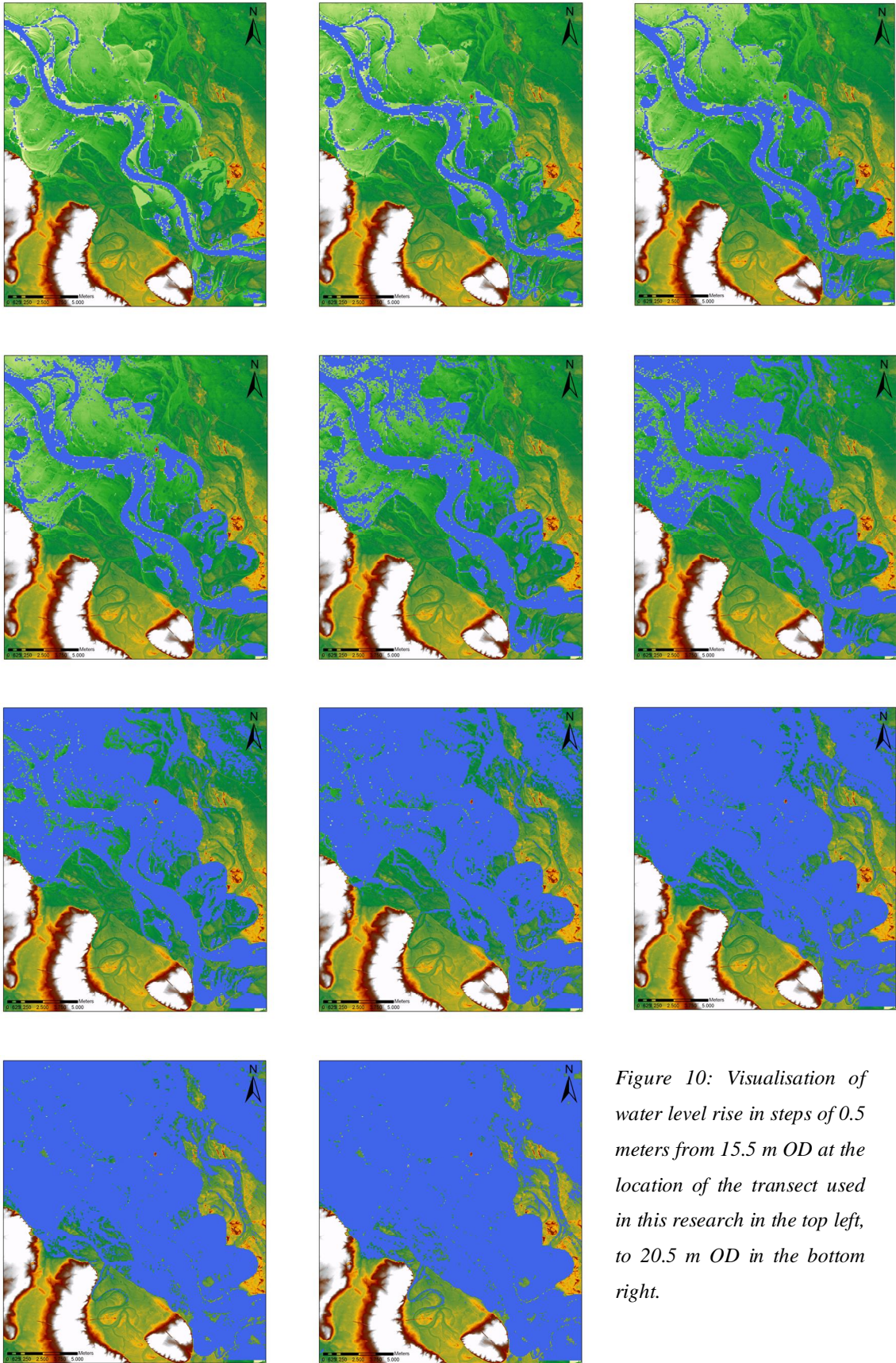
Figure 10: Visualisation of water level rise in steps of 0.5 meters from 15.5 m OD at the location of the transect used in this research in the top left, to 20.5 m OD in the bottom right.

The Manning's and Chézy equations (Eq. 1 and 4) were used for calculating the flow velocities for each 10 meter section of the cross section after the water depth, roughness and slopes were determined. The Manning's roughness was estimated using the vegetation reconstruction and the method as given in Arcement & Schneider (1989), the Chézy roughness was calculated based on the vegetation reconstruction alone. The water depth was determined by subtracting the (reconstructed) palaeo surface elevation from the elevation of the water level. The flow velocities were multiplied by the flow area (10 * water depth), giving the discharge for the 10 meter section of the model. The sum of all 10 meter sections is the total discharge for the specified water level elevation. This was done for a variety of scenarios to simulate different conditions; the vegetation roughness in winter and in summer, incision of the Holocene river terrace and different channel dimensions found in the field.

3.5 Model sensitivity

A sensitivity analysis of the model showed that the model is fairly insensitive to changes in floodplain vegetation, causing differences in floodplain roughness. On average the difference in discharge between the summer and winter scenarios are below 5% for both the Manning's and the Chézy calculations. The sensitivity for changes in incision was determined to be around 10% for both the Manning's and the Chézy calculations. The model turned out to be the most sensitive to changes in channel cross-sectional area. The discharges would differ up to 40% between the two channel dimensions used in the Chézy calculations and up to 60% for the calculations using the Manning's equation.

3.6 Model scenarios

Several scenarios were constructed for the model to reproduce different possible conditions during the past flood events. As there was no exact dating performed on some of the water level proxies there were two versions made of every scenario; an Early

Holocene (EH) version (roughly start of the Atlanticum, ~9000 yr. BP.) and a Middle Holocene (MH) version (roughly Subboreal, ~4500 yr. BP). The scenarios were based upon different channel dimensions found in the field: The dimensions of the Haus Groin palaeo channel (HG) and the Schloss Bellinghoven palaeo channel (SB) (Fig.17). Furthermore they were based on different vegetation covers to replicate the difference between bare winter vegetation and rich spring/summer vegetation. The specific roughness for each section/vegetation zone was calculated using the method as described by Arcement & Schneider (1989) (Eq. 3) and the Baptist equation (Eq. 7) and on the different reconstructed water levels since the roughness calculated using the Baptist equation depends on the water depth.

4 Results: Model inputs

4.1 Palaeo stage reconstruction

Loss on ignition analysis on cores taken in the channel fills of the Mariënbaum, Hohe Ley, Vosse Kuhl, Schloss Bellinghoven, Haus Groin and Heren Bril channel revealed flood layers in channel fills that can be used for both flood frequency analysis and as palaeo stage indicators for the water level reached during flood events. For this research the age of the channel fills was determined using pollen analisys (Geurst, in prep; van Munster, in prep; Bunnik, pers. comm.). The channel fills of the Haus Groin, Schloss Bellinghoven and Heren Bril residual channels proved to contain flood layers. For this research no accurate dating techniques were available, these channel fills could thus not be used for a flood frequency analysis at the time of this research. While these fills contain palaeo stage indicators, they were not relevant for this research, focused on extreme flood events, as they would reflect both extreme and less extreme floods without a reliable method of assigning flood magnitudes to the individual flood pulses. The core descriptions and the results of the LOI analysis of these cores can be found in appendix 3. The channel fills of the Mariënbaum and Vosse Kuhl proved to be useful for this research since they do contain palaeo stage indicators of extreme floods and the results of their LOI analysis is described below. The complete core descriptions can also be found in appendix 3.

4.1.1 Mariënbaum residual channel

The Mariënbaum core is located in a depression at the foot of the ice pushed ridge in the west of the research area. The sample core started at a depth of 1 meter below surface and continued to a depth of 3.66 meters below surface, divided over 4 separate sample cores with a hiatus from 1.30 to 1.49 meters. Pollen analisys between 2.15 and 1.80 meters showed this part of the core to be of Preboreal age (roughly 11,500 – 10,500 BP.). The transition from Preboreal to Boreal was around 1.50 – 1.60 meters. A pollen sample around 0.95 meters was determined to be of Late Boreal age (roughly 9500 BP.) (Geurts,

in prep; Bunnik, written communication). LOI analysis of this core (Fig 11) showed a relatively constant organic contents of around 80% for the part of the core that was dated to be of Late Boreal (roughly 10,500 – 9200 BP.) age and younger. This part of the core contained no significant flood signals, it was unlikely that a flood has reached this channel fill since the Late Boreal.
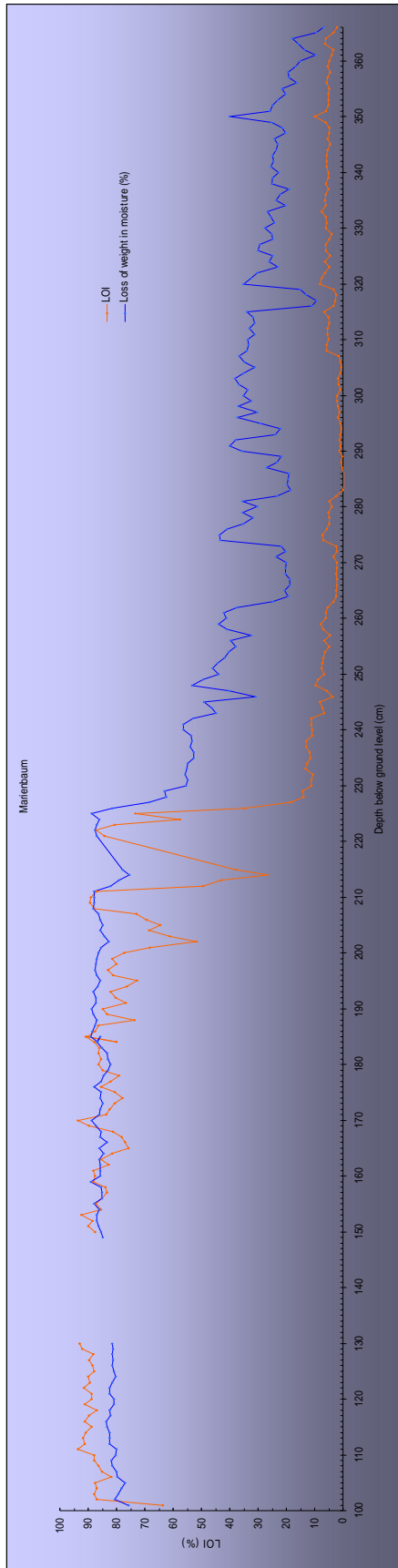
Figure 11: The LOI curve of the Mariënbaum residual channel fill.

4.1.2 Vosse Kuhl residual channel

The Vosse Kuhl sample core was taken in a residual channel on terrace 2 (Fig 14). The core started at a depth of 1.37 meters below surface and ended at 3.75 meters below surface, divided over 8 individual sample cores with no overlap between cores. The core had a hiatus between 3.25 and 3.27 meters, a hiatus between 2.40 and 2.44 meters and a hiatus between 1.85 and 1.95 meters. Pollen analysis performed by Geurts (in prep) and Bunnik (written communication 4-2011) showed the bottom of the core to be of Preboreal age (roughly 11,500 – 10,500 BP.) and from roughly 3.70 meters and up to 1.80 meters below surface to be of Atlanticum age (roughly 9200 - 5600 BP.). The transition from the Atlanticum to the early Subboreal was determined to be between 1.80 and 1.50 meters below surface. A sample at 1.40 meters below surface was determined to be of Late Subboreal or Early Subatlanticum age (roughly 3000 BP.). The LOI analysis of the sample core (fig 12) showed that the fill has significant trends in organic matter content. These trends can be caused by for instance climate fluctuations, changes in distance to the active river or the successions inherent to the filling process of a residual channel. The top part of the sampled channel fill, roughly between 1.3 and 1.4 m below surface, showed a rapid decline in the organic matter content. This is due to the clayey layer that covers the channel fill. This clayey layer is an overbank deposit and indicates that since the Late Subboreal or Early Subatlanticum, the Vosse Kuhl residual channel has been inundated at least once, but considering the thickness of the layer, 1.4 m, most likely multiple times. This layer is at present at the surface and is very likely to be disturbed by humans, for instance through plowing. The layer is also located above the ground water table, which made sampling this layer with the Boncke Piston Corer impossible. It was therefore not possible to analyze the organic content of this layer or date it using pollen.
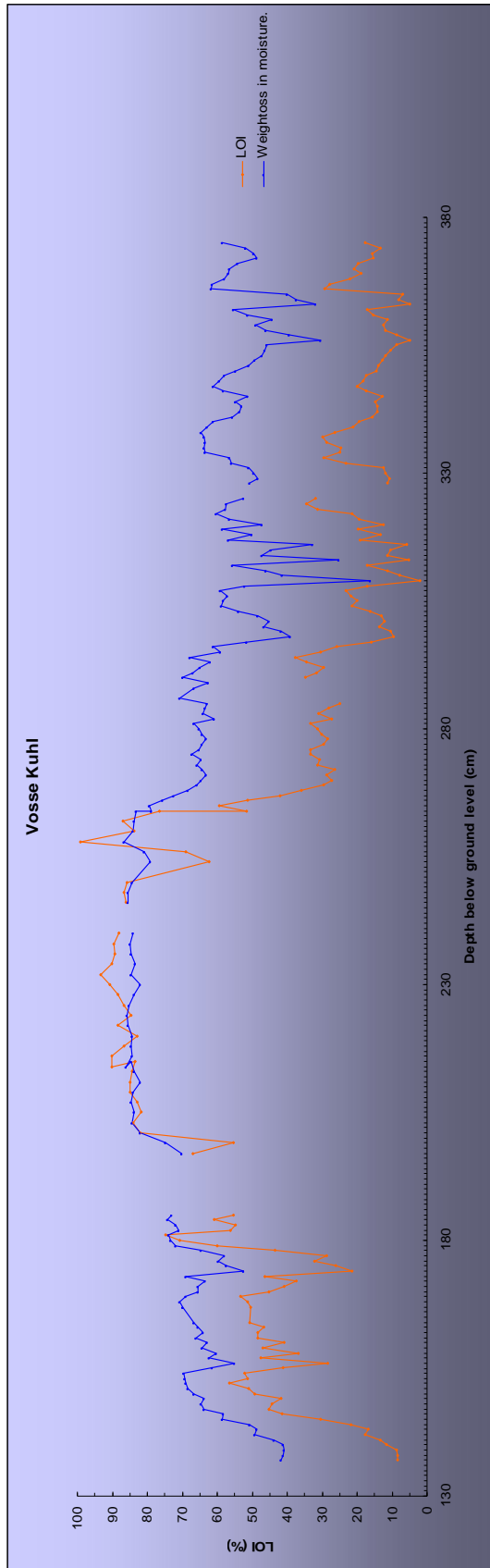
Figure 12: The LOI of the Vosse Kuhl residual channel fill.

CPA analysis (Fig. 13) showed that the Vosse Kuhl sample core contained 8 clear pulses that can be qualified as flood pulses of which the top most pulse around 1.44 meters was of Late Subboreal or Early Subatlanticum age (around 3000 BP.). The depth of 1.44 meters below surface in the residual channel means that this flood reached at least an elevation equal to 18,5 meters +OD at the location of the transect used in the model. These flood pulses can be used as palaeo stage indicators since the Vosse Kuhl residual channel has no sizeable catchment and is expected to only carry water during (extreme) Rhine floods (see fig 10).
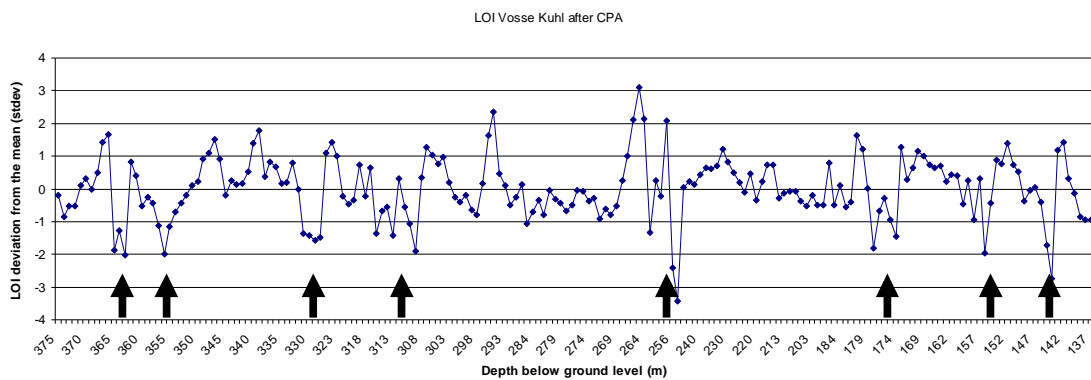
Figure 13: The LOI curve of the Vosse Kuhl core after change point analysis, the arrows indicate the location of flood layers.

4.2 Other palaeo stage indicators

The fieldwork performed for this research showed Holocene clayey and loamy deposits on top of Younger Dryas deposits (Geurts, in prep) on Terrace 2 and on terrace 4 (Fig 14) that are typical overbank deposits for meandering rivers and can thus be used as a palaeo stage indicator for determining the minimum water height reached during Holocene floods.
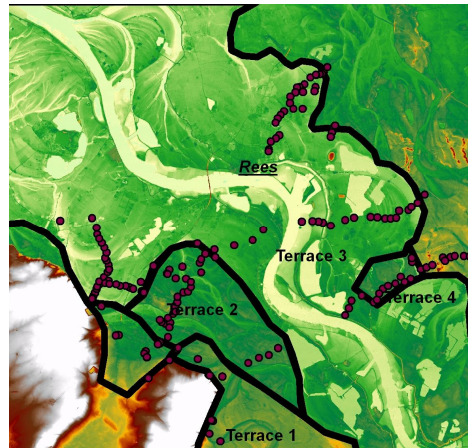
Figure 14: River terraces in the fieldwork area

The highest elevation at which these overbank deposits are found (after correcting for the Holocene river gradient) is 19.4 m. +OD at the location of the transect. Figure 15 shows Holocene point bar ridges on terrace 3 that can also be used as palaeo stage indicators as they are formed below the water surface during high water conditions. Based on their location in the floodplain and on literature (Klostermann, 1989), these point bar ridges are all of Middle to Late Holocene age. Their respective elevations equal to a minimum water level of 18.5 m OD at the location of the transect used for the model. Pointbar ridges on the present day floodplains are excluded from this research. The increased sedimentation rate on the floodplains since the construction of dykes means that the elevations of these ridges can not be used as palaeo stage indicators for floods that occurred before the construction of these dykes because the dykes severely confine the size of the floodplain, resulting in much higher water levels.

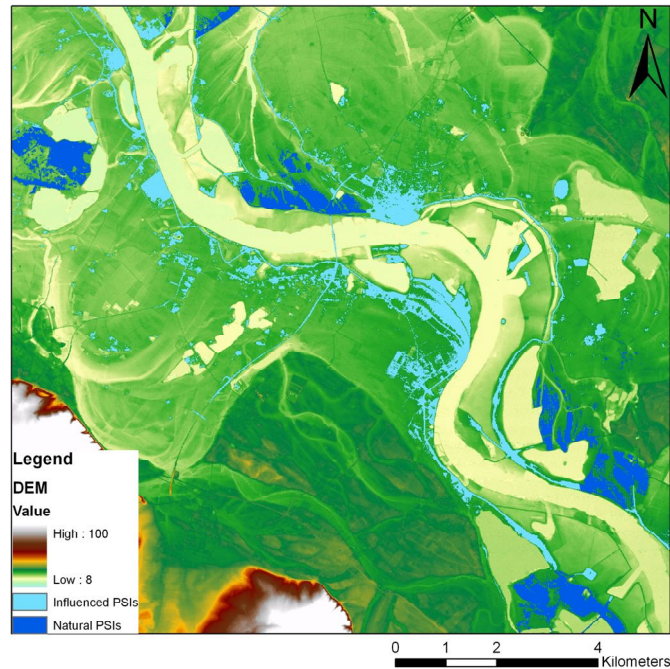*Figure 15: Corrected DEM of the research area. The DEM has been corrected for the slope of the Late Holocene Rhine to indicate the elevation of point bar ridges as opposed to the transect used in the model. Light blue represents palaeo stage indicators that are neglected in this research because they might be influenced by human activities. Dark blue represents relatively undisturbed palaeo stage indicators of which the highest reach an elevation comparable to 18.5 m OD at the location of the transect.*

The youngest flood layer that was identified in the residual channels was the layer at 1.40 m below surface in the Vosse Kuhl residual channel. This layer was dated to be of Late Subboreal or Early Subatlanticum age (Bunnik, written communication). The depth of 1.40 m below surface corresponds with an elevation of 18.5 m OD, similar to the elevation of the point bar ridges. The LOI analysis of the Mariënbaum core, for which water levels must have reached at least 19.7 m OD for the location to be inundated, showed that it is unlikely that this location has been inundated since the start of the Holocene and is therefore considered to be a non-exceedence boundary. Water levels during extreme floods in the Middle Holocene must therefore have reached an elevation of at least 18.5m OD and possibly even 19.4 m OD, but are unlikely to have reached 19.7 m OD (fig 16).
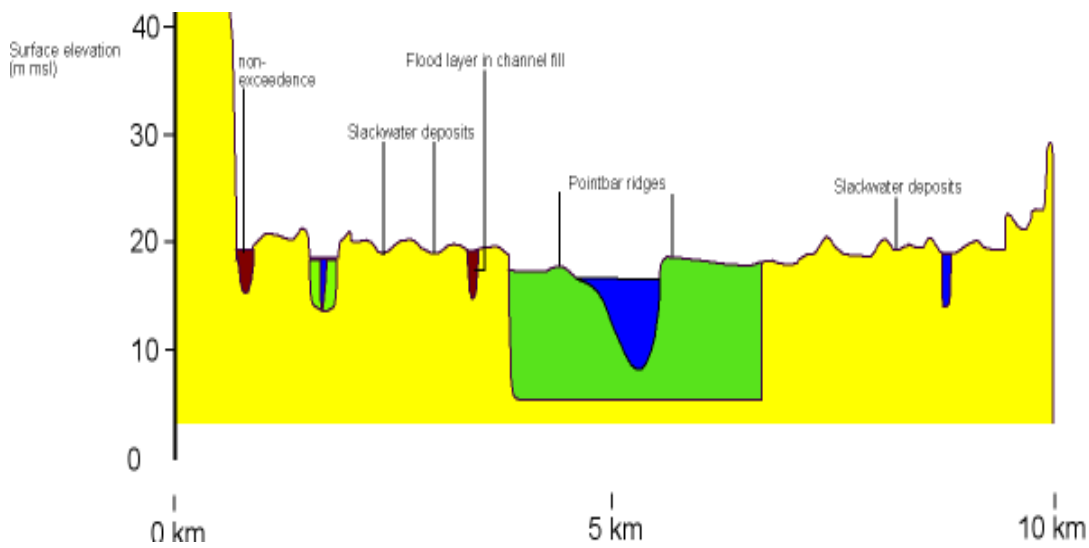
Figure 16: Palaeo stage indicators for the transect used in the model. The Pre-Holocene substrate is depicted in yellow, Holocene channel belts are green, residual channel fills are brown and active channels are marked in blue.

4.3 Channel dimensions

Coring transect over the Heeren Bril, Schloss Bellinghoven and Haus Groin residual channels revealed large differences in the cross-sectional areas of these channels. While all channels are comparable in depth, around 7 meters, the width varies considerably. The Heeren Bril channel is the narrowest channel with a width of roughly 300 meters and a very shallow inner bend. The width of the Schloss Bellinghoven channel is roughly 530 meters, also with (relatively) shallow inner band. The Haus Groin channel has the largest width of roughly 620 meters with a more symmetrical profile (deeper inner bend) (Fig 17).

Channel Dimensions
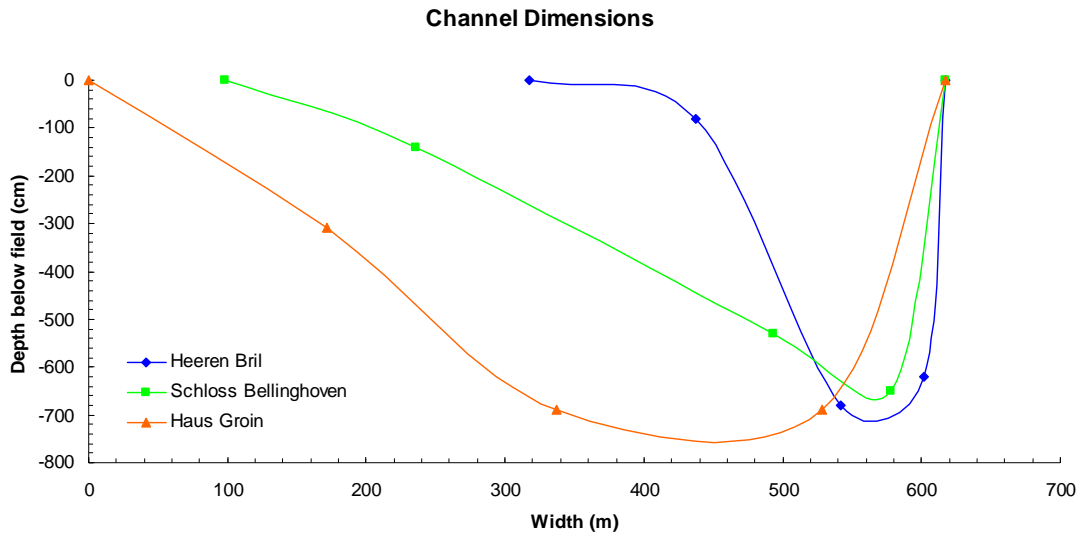
Figure 17: Channel dimensions of the Heeren Bril, Schloss Bellinghoven and Haus Groin residual channels.

4.4 Floodplain topography and roughness values

The slopes for each individual segment of the transect are obtained from research by Erkens (2009) (Table 1).

segment	slope
1	0.00027
2	0.00027
3	0.00027
4	0.00029
5	0.00031

Table 1: Slopes per segment of the transect

For the calculations using the Chézy roughness (Eq 4), the channel roughness was set to be 47 $m^{1/2}s^{-1}$. The Chézy roughness of the floodplain was calculated using vegetation characteristics derived form literature and expert opinions. The tree density was determined to be 0.0275 trees per square meter for vegetation zones 1, 2, 3 and 4, and

0.03 trees per square meter for vegetation zone 5 (section 2.4). The surface friction was set at 42.8 $m^{1/2s}s^{-1}$ (van Velsen, 2003). No roughness was added for undergrowth for the winter scenarios as this would be all but lacking during winter. For the summer scenarios the surface and undergrowth friction was set at 28.8 $m^{1/2}s^{-1}$ for vegetation zones 1, 2 and 3 (Fig 2). For vegetation zone 4 and 5 the surface and undergrowth roughness was set at 32 $m^{1/2}s^{-1}$ (van Velsen, 2003).

For the calculations using the Manning's equation (Eq 1), the roughness for the channel and the floodplains were separately determined using equation 3. For the channel an n_b value of 0.016 $sm^{-1/3}$, an n_1 value of 0.006 $sm^{-1/3}$, an n_2 value of 0.001 $sm^{-1/3}$, an n_3 value of 0 $sm^{-1/3}$ (no obstructions) and an n_4 value of 0 $sm^{-1/3}$ (no vegetation in the active channel) were used. The degree of meandering was minor at the location of the transect, thus the m value was determined to be 1.01 $sm^{-1/3}$. These values were selected using the USGS guide for selecting Manning's roughness coefficients for natural channels and floodplains (Arcement & Schneider, 1989). This resulted in a total roughness value of 0.023 for the channel. The roughness of the floodplains was determined using the same guide. The n values for the floodplains during winter were 0.1 $sm^{-1/3}$ for vegetation zone 1 and 2, 0,08 $sm^{-1/3}$ for vegetation zone 3 and 0.16 $sm^{-1/3}$ for vegetation zone 4 and 5 (Fig 2). For summer conditions, these values were 0.12 $sm^{-1/3}$ for vegetation zones 1 and 2, 0.11 $sm^{-1/3}$ for vegetation zone 3 and 0.18 $sm^{-1/3}$ for vegetation zone 4 and 5 (table 2).

Vegetation zone	Summer roughness ($sm^{-1/3}$)	Winter roughness ($sm^{-1/3}$)
1	0.12	0.1
2	0.12	0.1
3	0.11	0.08
4	0.18	0.16
5	0.18	0.16

Table 2: Manning's surface roughness coefficients for the vegetation zones.

5 Model results and discussion

5.1 Model results

Discharges were calculated to simulate discharges under various conditions (table 3). The highest discharges reconstructed were based upon the dimensions of the Haus Groin residual channel, the location of overbank deposits on terrace 2 and 4 (Fig 14) and a non-exceedence bound (the Mariënbaum residual channel fills which showed no evidence of flooding). These paleo stage indicators were not dated and were therefore simulated for both incision rates representing the Early Holocene and the Middle Holocene. The according discharges using Chézy are 17,000 and 18,700 m^3/s. The lowest discharge reconstructed was found for Early Holocene summer conditions using the Schloss Bellinghoven channel dimensions and the Manning equation. This discharge was 8600 m^3/s. Differences in outcomes between scenarios representing summer and winter vegetation were minor.

Further discharge calculations were based on a roughly dated flood layer in the Vosse Kuhl residual channel with an elevation of 18.5m OD, which was dated to be around 3000 years age. This flood layer represents a minimum water level that the flood depositioning this layer will have reached in the Middle Holocene and thus represents a minimum discharge. This discharge is determined to be 6000 m^3/s (Manning) and 8000 m^3/s (Chézy) when calculated using the dimensions of the Schloss Bellinghoven residual channel. For the calculations using the dimensions of the Haus Groin channel, the minimum discharges were 12,000 m^3/s (Manning) and 12,900 m^3/s (Chézy).

Model results show a considerable difference in outcomes for the scenarios that were run using Manning's equation and the scenarios run using the Chézy equation, with the former resulting in substantially lower discharges. This is partly the result of roughness values being determined with different methods for both equations and partly the result of the static roughness value for the calculations using the Manning's equation, while the Chézy-based method as described by Baptist takes the reduced drag after vegetation is

submerged into account. The roughness values for the Manning's equation were determined using estimated values, that combined resulted in roughness values for the floodplains and channel. The estimation of components of the final roughness values allows for a stacking of errors. For instance, when one is inclined to estimate higher than average, the sum of all these high estimates for the separate components of the roughness value, will eventually lead to a value that is much too high and vice versa. The roughness values for the Chézy equation were calculated using the formula for submerged vegetation as given by Baptist et al. (2007). Flume tests of this formula showed this method to be highly accurate. While the roughness of the channel was taken from accurate measurements by Brunning in 1792 (Hesselink et al., 2006).

Channel dimension	Time of flood	Water level (m OD)	Season	Manning's discharge	Chézy Discharge
SB	EH	19.7	Winter	8775	12,012
SB	EH	19.7	Summer	8566	11,514
HG	EH	19.7	Winter	14,584	17,104
HG	EH	19.7	Summer	14,388	16,380
SB	MH	19.7	Winter	10,400	13,217
SB	MH	19.7	Summer	9804	12,709
HG	MH	19,7	Winter	16,196	18,678
HG	MH	19,7	Summer	15,959	17,913
SB	MH	18.5	Winter	6480	8260
SB	MH	18.5	Summer	6412	8042
HG	MH	18.5	Winter	12,082	13,238
HG	MH	18.5	Summer	12,014	12,829

Table 3: Model results. SB represents a channel with dimensions of the Schloss Bellinghoven residual channel, HG represents a channel with the dimensions of the Haus Groin residual channel. EH means the model was represented Early Holocene conditions, MH means that the model represented Middle Holocene conditions. All discharges are in m^3/s.

5.2 Discussion of model results

5.2.1. Roughness

Both the Baptist formula and the determination of the Manning's roughness values relied on the same vegetation reconstruction. The Manning's roughness determination with the use of the USGS guide for selecting them Manning's *n* (Arcement & Schneider, 1989) involves estimating roughness of elements that combined result in the total roughness. A systematical under or over estimation will therefore result in a significant over or under estimation of the final value. The Baptist formula uses tree density, stem diameter and height. Flume tests have proven this method to be highly accurate. The added advantage is that the method takes the submergence of vegetation (which reduces drag) into account. It is therefore concluded that the results obtained using the Chézy equation are more reliable for this research, but the Manning's equation has not been excluded in order to be able to compare the results to other studies on the same subject.

5.2.2 Reconstructed discharges

The dimensions of the Schloss Bellinghoven residual channel are considerably smaller than the dimensions of the Haus Groin residual channel (fig 17), leading to much lower discharges in the scenarios using the dimensions of the Schloss Bellinghoven channel. Erkens (2008) shows that channel dimensions, after an initial period at the onset of the Holocene, show little variation. The profile of the Schloss Bellinghoven residual channel suggests that the corings have taken place on the location of a possible sand plug, formed inside the channel at the time of the abandoning of the channel; possibly a new scrollbar. This is evident from the shallow inner side of the residual channel (fig 17). The Haus Groin residual channel shows a more natural profile with deeper inner banks. Furthermore, active channel dimensions are more likely to be underestimated because of the channels being filled in (with multiple layers that can be interpreted as channel lag) during the abandoning process, than they are likely to be overestimated. This, combined with the relatively stable channel dimensions during the Holocene as found by Erkens

(2008) lead to the conclusion that the dimensions of the Haus Groin residual channel (which has been dated to be of Middle Holocene age), are more likely to represent actual channel dimensions during the time of the flood event that formed the flood layer in the Vosse Kuhl channel. The discharge that was calculated using Chézy, the dimensions of the Haus Groin residual channel and the elevation of this flood layer is 12,800 m³/s.

It can therefore be said that the largest Early or Middle Holocene flood that was found in the course of this research had a discharge that is unlikely to have exceeded 18,700 m³/s, but was likely to have reached at least 12,800 m³/s.

5.3 Comparison with recent observed flood records

The discharges in this study were found using a simulation of the river Rhine under natural conditions. The impact of mankind in the last 2000 years has meant that, throughout the whole catchment, large scale deforestation has taken place (e.g. Bos et al., 2003). This increased the runoff and decreased evaporation and infiltration. Artificial meander bend cut offs have reduced the length and sinuosity of the river, basically increasing the slope and decreasing the resistance to flow. The construction of groins and dykes has deepened the river and raised the water level over the deforested floodplain. All this means that under the current conditions a flood wave propagates much faster and that a larger part of the water that is put into the system by snow melt and rainfall will be part of the flood wave during a flood. This means that for reaching peak discharges and water heights as found in this study, nowadays less water in the form of precipitation and snowmelt is needed. Therefore, before results of this study can be implemented in a discharge – reoccurrence comparison for the current day situation or the recent past, the effects of human impacts on the shape and volume of the flood waves needs to be studied. Flood waves reaching a specific discharge and water height in times before human impact will result in much larger flood waves, both in total volume as in peak discharge and water level, in the present day situation (fig 18).
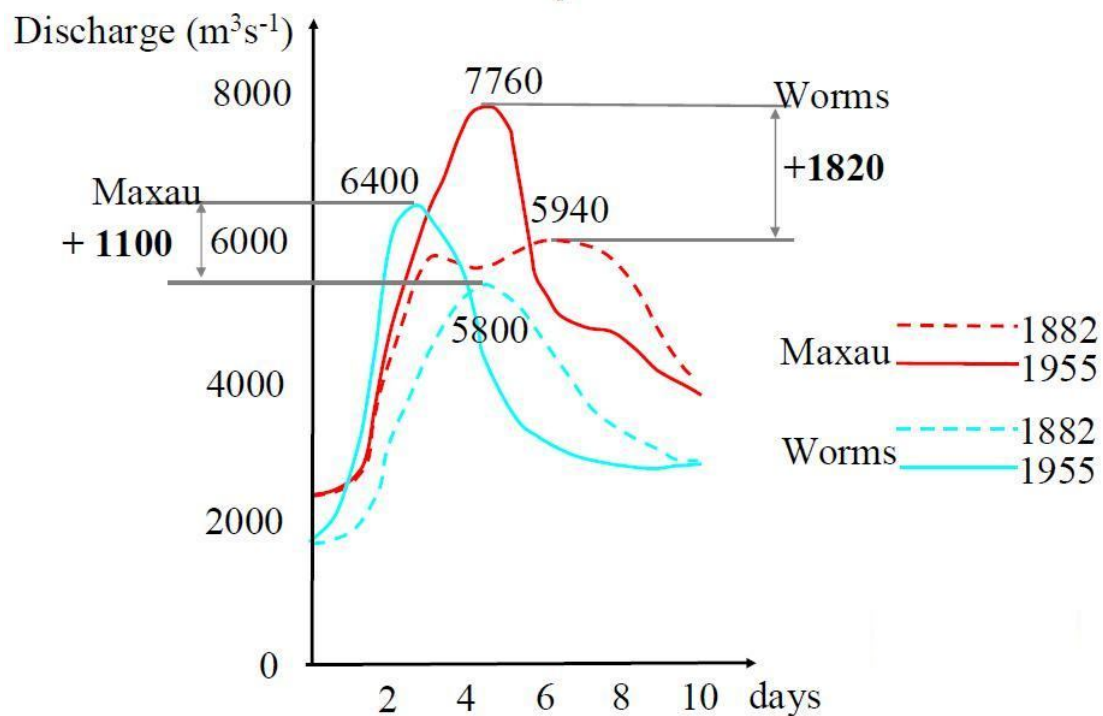
Fig 18: The difference in flood waves caused by different levels of human influence on the environment for two tributaries of the river Rhine. Note that although the peak volumes under the pre-engineering conditions (the situation with a lower human impact on the environment) is similar to those of the continuous lines, the peaks of the latter are considerably higher (after Disse & Engel, 2001)

When the results of this study, a minimum discharge estimate of 12,800 m^3/s and a highest estimate of 18,700 m^3/s, are compared to the present day standard of an estimated 16,000 m^3/s for a 1/1250 yr flood, it is clear that these estimates are roughly within the same range, even when it is taken into account that no reoccurrence times were determined for the extreme floods in this study. Further research into the reoccurrence times of these floods and on how to translate palaeo discharges to modern day discharges is required for a better comparison of the discharges in this study to the 1/1250 year standard.

6 Conclusions and recommendations

The objective of this study was to reconstruct peak discharges of extreme palaeo floods of the River Rhine near Rees, Germany. Using palaeo geographical reconstructions, vegetation reconstructions and sedimentary archives, it proved possible to estimate the values of the parameters required to reconstruct these discharges. In order to reconstruct these discharges, the following information was needed:

(1) Slackwater deposits in channel fills and on terraces provided excellent palaeo stage indicators. In this study it was shown that during the Holocene, the maximum water level did not exceed 19.7 m OD, while it did reach at least 19.4 m OD. These palaeo stage indicators, and thus the moment of depositioning, were not dated. A roughly dated palaeo stage indicator in a residual channel fill resulted in a minimum water level of 18.5 m OD during the Late Subboreal or Early Subatlanticum. This water level elevation is further supported by the height of Middle to Late Holocene floodplain morphology that also reaches 18.5m OD.

(2) Previous studies and the fieldwork undertaken for this study showed that a palaeo reconstruction of the landscape during the Holocene was possible. Incision rates from previous studies provided the correct elevation of the river bed and floodplain, while the corings provided the necessary information on the slopes and ages of the terraces and the location and diameter of the channel. Where needed, a pollen analysis provided information about the ages of channel fills.

(3) Lack of an fully natural riparian forest in Northwestern Europe forced this study to rely on literature and expert opinions to reconstruct the palaeo vegetation on the floodplain. Using this method, tree densities, heights, stem diameters and rejuvenation rates for both summer and winter conditions were obtained. This information was used to estimate the surface roughness of the floodplain for the calculations of the floodplain roughness using the method as described by Baptist (2007). Sensitivity analyzes of these roughness values showed that the effect of changes in vegetation on the outcomes of discharge calculations is small (less

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than 5% difference between the outcomes for winter and summer vegetation conditions).

(4) While it was possible to obtain palaeo stages from flood layers in laminated channel fills, it proved impossible in this study to link specific flood layers in channel fills to flood layers in other channel fills. Combined with a lack of accurate dating of channel fills, it was therefore impossible to obtain reliable information on the recurrence times of floods that were recorded in the fills. More detailed dating of these layers might make this possible in the future. However, pollen dating enabled the rough dating of some flood layers, which led to the conclusion that during the Late Subboreal or Early Subatlanticum, water levels must have reached at least 18.5m OD at the research location.

This information was used to simulate scenarios with different water level data for the Early and Middle Holocene. These simulations resulted in a highest possible discharge during the Early and Middle Holocene of 18,700 m^3/s. They also resulted in a minimum discharge of 12,800 m^3/s for a flood that occurred during the Late Subboreal or Early Subatlanticum (roughly 3000 years BP.). It can thus be concluded that the highest, dated, flood event that was found in sedimentary records in the research area near Rees (Germany) had a minimum discharge of 13,200 m^3/s, but will not have exceeded 18,700 m^3/s. These discharges took place in a natural river and will have to be translated to discharges that would happen in the heavily human influenced river of the present for the same meteorological conditions. For this translation, a separate research is necessary.

Comparison of these discharges to the extrapolated data of observed recent floods shows no discrepancy between the extrapolated discharges and the discharges found in this research, meaning that this study shows that the estimated discharge of 16,000 m^3/s for a 1/1250 yr flood is realisitic. The discharge of 23,800 m^3/s as found by Herget & Meurs (2010) for a flood event near the city of Cologne (Germany) in 1342 AD is considerably higher than the discharges found in this study. While this might be partly due to the different conditions, this study focused on a natural river, while Herget & Meurs focused on a river that was heavily influenced by mankind, this discharge exceeds the discharge

estimates gained from the extrapolated observed flood records. It is therefore concluded that the discharge found in their study is too high.

This study has resulted in a range of discharges for flood events of the river Rhine during the Early and Middle Holocene. These discharges have occurred under natural circumstances; present day response to similar rainfall events that cause these flood events may be different. It is therefore imperative that further research is performed on the differences between flood wave volume, height and propagation in the present day Rhine and the natural palaeo Rhine. Furthermore, flood signals found in the channel fills in this and other researches will need to be accurately dated. When this dating is accurate, it will be possible to link flood layers in channel fills with a lower elevation to flood layers in channel fills with a higher elevation. Flood layers that ware found on lower elevations, but not on higher elevations are unlikely to have exceeded the threshold discharge needed for the flooding of the higher located residual channels. This information combined can lead to a palaeo recurrence times reconstruction for floods, which in turn can be used to improve the design discharge estimates that are currently used by policy makers. This knowledge is needed to guarantee the safety and economic welfare of the parts of the Netherlands that is at risk of river flooding.

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