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Fluvial terrace formation in the northern Upper Rhine Graben during the last 20 000 years as a result of allogenic controls and autogenic evolution

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ABSTRACT

The northern Upper Rhine Graben hosts a well-preserved Late Weichselian and Holocene fluvial terrace sequence. Terraces differ in elevation, morphology, and overbank sediment characteristics. The purpose of this study was to determine the relative importance of allogenic controlling factors versus autogenic evolution on the successive formation of these terraces. For a representative valley segment (the Gernsheim region), results from previous research were integrated with newly obtained borehole data and digitized elevation maps to construct palaeogeographic maps and cross sections. Coarse-grained channel deposits below terrace surfaces were dated using Optically Stimulated Luminescence, and fine-grained abandoned channel fill deposits were dated using pollen stratigraphy and radiocarbon analysis. Initiation of terrace formation was caused by climatic change in the Late Pleniglacial (after ~20 ka), but fluvial response was complex and slow and continued locally until the middle Boreal (~9 ka). Early to Middle Holocene (~6 ka) changes in fluvial style and associated overbank lithofacies are not necessarily controlled by climatic change as was previously proposed. Instead, autogenic processes combined with river reach-specific factors explain the observed terrace development. Continuous incision, autogenic evolution, and high preservation potential provide an alternative explanation for the presence of a terrace sequence in this subsiding area.

1. Introduction

River terraces represent an important component of fluvial archives and are often used to examine fluvial adjustment to natural and human-induced environmental change. The study of fluvial terraces has long served as a core research theme within geomorphology; and the theme maintains much contemporary relevance, being of interest to a broad range of scholars in sedimentology, engineering, hydrology, planning, and archaeology. Terrace formation within the last glacial–interglacial cycle occurs because of allogenic (climate change, human impact, tectonics/base level changes) and autogenic controls (intrinsic behaviour and complex response). On this topic, numerous studies from large and small rivers spanning a range of climatic and physiographic settings have been published, with excellent examples from the Mississippi (e.g., Knox, 1996, 2001, 2006), Danube (e.g., Buch, 1988), Colorado (Blum and Valastro, 1994), Vistula/Weichsel (e.g., Starkel, 2002), some smaller rivers in the UK (e.g., Lewis et al., 2001; Gao et al., 2007), and the Meuse (e.g., Huisink, 1997; Tebbens et al., 2000).

Research has been extensive in the Rhine catchment on Late Weichselian and Holocene terrace formation (amongst many others: Scheer, 1978; Schirmer, 1983; Klostermann, 1992; Schirmer, 1995; Törnqvist, 1998; Berendsen and Stouthamer, 2001; Dambeck and Thiemeyer, 2002; Houben, 2003; Schirmer et al., 2005). A well-preserved terrace sequence along the Rhine trunk valley is found in the northern reaches of the Upper Rhine Graben (Fig. 1). Strikingly, even though the Graben is a subsiding basin, this extensive terrace sequence was formed here. Fetzer et al. (1995), Rosenberger et al. (1996), Dambeck and Thiemeyer (2002), and Dambeck (2005) mapped the Late Weichselian and Holocene surfaces as five terrace levels. Changes in sedimentation, fluvial morphodynamics and environmental conditions during the last 20 ka resulted in specific characteristics (elevation, fluvial style/morphology, overbank sediment characteristics, and soil formation), which can be used to identify each terrace. Nevertheless, which factors control the successive formation of the terraces is not fully known. Authors such as Fetzer et al. (1995), Dambeck and Thiemeyer (2002), Dambeck (2005), and Schirmer et al. (2005) suggested that the terrace sequence formed essentially under the influence of climate change and — later during the Holocene — as a result of human impact. If correct, it implies

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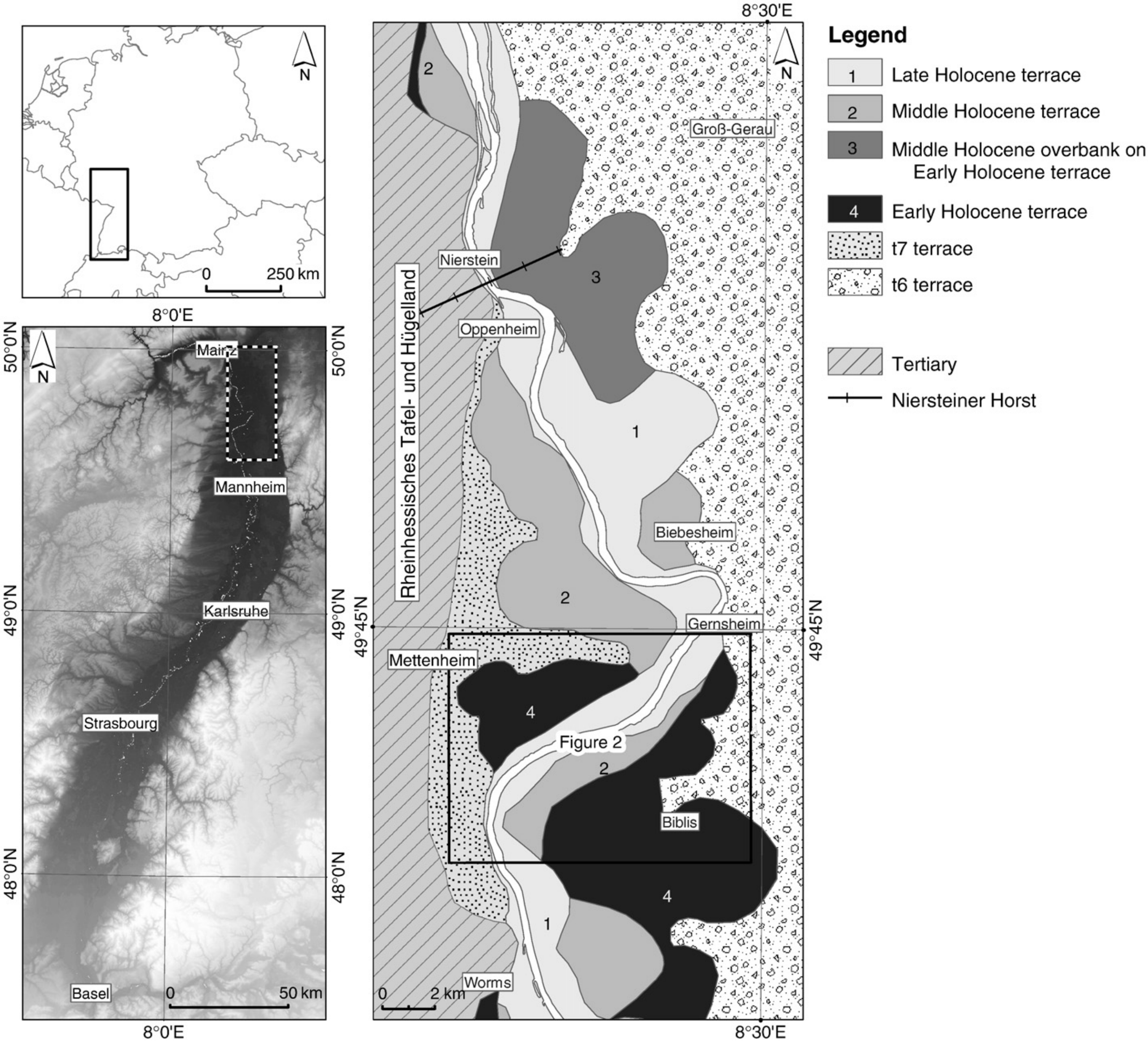

Fig. 1. Setting of northern Upper Rhine Graben (Germany) with Late Weichselian and Holocene terrace levels (Dambeck and Thiemeyer, 2002; Dambeck, 2005).

considerable fluvial sensitivity to climate change because terrace characteristics vary widely, whereas Holocene climate was rather uniform. In addition, although allogenic forcing may be straightforward, the geomorphic response of the fluvial system is likely to be affected by intrinsic behaviour, river stretch characteristics, and catchment characteristics (e.g., size). In the many previous studies in this area, the focus was on allogenic forcing in explaining the terrace sequence, whereby the role of autogenic processes and site-specific characteristics in terrace formation remained unaddressed.

The purpose of this study is to determine and revaluate the importance of allogenic controlling factors versus autogenic evolution during successive formation of the Late Weichselian and Holocene terraces in the northern Upper Rhine Graben. We utilised valley cross sections, high-resolution digital elevation maps, and improved dating of terraces. These data are integrated with results from previous research to describe the palaeogeographic development of the entire northern Upper Rhine Graben in terms of fluvial style and terrace characteristics and the role of allogenic forcing and autogenic evolution.

1.1. Models of terrace formation

Schumm (1977) and Lewin and Macklin (2003) highlighted that terrace sequences can form as a result of intrinsic behaviour of the river system (autogenic evolution). Intrinsic processes are continuously influencing fluvial systems and induce trends and events (threshold crossing in Schumm, 1977) capable of forming valley floor landforms (e.g., terraced surfaces). These mechanisms are active over relatively small spatial and temporal scales (e.g., Hey, 1979a,b; Buch, 1988; Brown, 1991; Houben, 2003), and their pace is strongly controlled by local circumstances. Thus, intrinsic forcing alone is unlikely to result in mappable stratigraphic units that are time-synchronous over considerable distance, i.e., across several river reaches (Blum and Törnqvist, 2000). However, in interaction with allogenic impulses, nonlinear internal system dynamics may cause a complex response of a fluvial system (Schumm, 1973, 1977; Bull, 1991). This means that different river reaches may respond differently to the same instantaneous external forcing, depending on catchment and reach-specific characteristics (e.g., Mol et al., 2000; Taylor et al., 2000; Veldkamp and Tebbens, 2001; the threshold concept of Vandenberghe, 1995). As a result, complex response to an external factor varies spatially; implying correlation of fluvial units resulting from allogenic forcing across large river basins is often difficult or impossible. Only strong allogenic impulses of sufficient magnitude relative to catchment size can be considered capable of overwhelming depositional signatures of intrinsic behaviour and ongoing complex response (e.g., Schumm, 1973; Blum and Törnqvist, 2000). But even then, because morphological adjustment requires time, fluvial geomorphology tends to respond dynamically and lags behind changes in controls (e.g., Vandenberghe, 1995; Törnqvist, 1998; Blum and Törnqvist, 2000; Tebbens et al., 2000; Gibbard and Lewin, 2002; Vandenberghe, 2003). This implies that what is seen as a near-instantaneous response to a certain forcing event may instead be a delayed result from an earlier forcing of larger magnitude (e.g., Busschers, 2008). The extent to which past morphological response remains preserved along a river reach limits the reconstruction of fluvial response to allogenic controls (Lewin and Macklin, 2003). Preservation potential — at millennial timescales — is a result of many local factors, such as incision rates, substrate lithology, lateral channel migration rates and valley morphology. Therefore, local-scale characteristics may decrease the potential for a particular valley reach to represent a viable fluvial archive. Longer valley reaches, however, are more likely to preserve adequate evidence and thus provide greater insight into the controls on terrace formation (Schirmer, 1995). All these considerations complicate attribution of fluvial landforms and terrace levels to both allogenic and autogenic controls.

2. Regional setting

The Rhine River crosses several tectonic basins as it flows from the Alps to the North Sea. The largest tectonic basin is the Upper Rhine Graben, extending 300 km from Basel (Switzerland) to Mainz (Rheinland-Pfalz, Germany). At the northern end of the Graben, the Niersteiner Horst (Peters and Van Balen, 2007; Fig. 1), a relatively stable tectonic block forms a local base level for the river (Dambeck, 2005). During the Quaternary, the Graben was a major sink for the Rhine and local sediments, especially when the northern Upper Rhine Graben (Fig. 1) between Karlsruhe and Mainz subsided, resulting in a relatively complete record of unconsolidated sediments. Maximum absolute subsidence in the entire Upper Rhine Graben was 2 m over the last 10 ka (Peters and Van Balen, 2007). Tectonic deformation along individual intragraben faults does not exceed 0.5 m, and direct influence on the fluvial development in the last 20 ka is considered small or absent (Peters et al., 2005; Peters and Van Balen, 2007). For the study area, this setting resulted in good preservation of fluvial deposits in a relatively wide valley with a low longitudinal gradient.

The Late Weichselian and Holocene terrace sequence is particularly well developed between Biblis and Gernsheim (geological map Gernsheim 1:25 000; Rosenberger et al., 1996), where the total width of the Holocene floodplain is ~10 km and the gradient is only 4 cm/km (Dambeck, 2005). This area (the Gernsheim region, Fig. 2) hosts the most complete terrace sequence within a relatively small area and forms our study area.

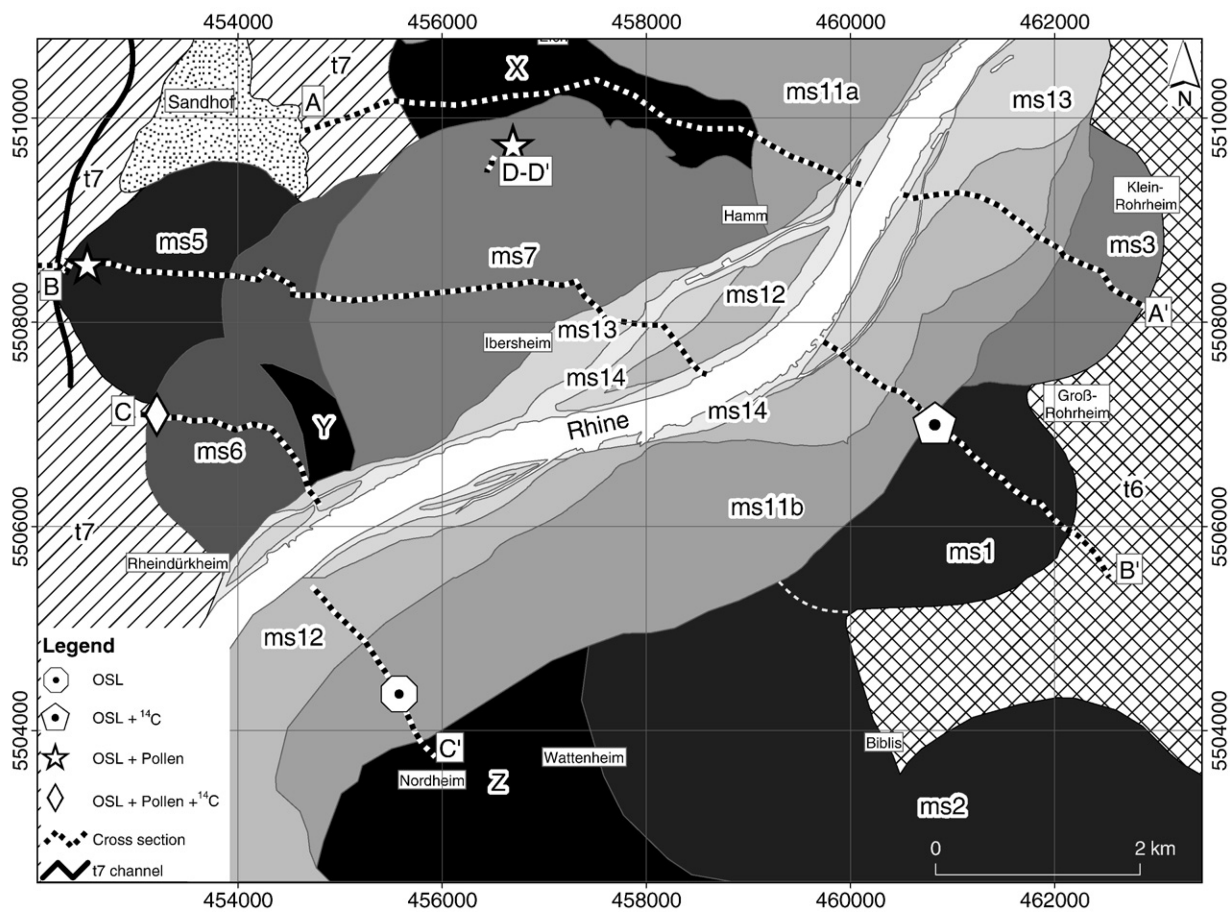

Fig. 2. Area of field study (the Gernsheim region) with Late Weichselian terrace levels (t6, t7), Holocene palaeomeanders (ms 1-14; modified after Rosenberger et al., 1996), location of cross sections, and patches of ambiguous age (X, Y, Z). Dating sites and methods are indicated with different symbols (core PALY1 in ms5, core PALY2 in ms6, core PALY3 in ms7). Location in Fig. 1.

2.1. Late Pleistocene terraces

The palaeogeographic development of the Late Weichselian and Holocene terrace sequence in the northern Upper Rhine Graben has been studied by numerous authors (e.g., Scheer, 1978; Fetzer et al., 1995; Rosenberger et al., 1996; Dambeck and Sabel, 2001; Dambeck and Bos, 2002; Dambeck and Thiemeyer, 2002; Dambeck, 2005). For the Late Weichselian (Late Pleniglacial and Late Glacial), two terrace levels are distinguished: the "Obere Niederterrasse" (or t6 terrace) and the "Untere Niederterrasse" also known as the t7 terrace (cf. Scheer, 1978; Dambeck, 2005). This subdivision of Late Weichselian fluvial deposits is also recognised in other parts of the Rhine catchment, such as along the tributary Main (e.g., Scheer, 1978; Schirmer, 1983) and in the Lower Rhine Embayment (e.g., Klostermann, 1992; Schirmer, 1995; Kasse et al., 2005). In this paper we use the nomenclature of Dambeck and Thiemeyer (2002) and Dambeck (2005). The t6 terrace is present virtually across the full width of the northern Upper Rhine Graben, whereas the t7 terrace level is only present in the western part, embedded in the t6 terrace level (cf. the fill-in-fill type of Schirmer, 1995). The terrace levels are distinguished based on different surface elevation (1–2 m; Rosenberger et al., 1996), geographic position, and soil formation (Dambeck, 2005; Thiemeyer et al., 2005), but not on lithological differences. Scheer (1978) correlated the t6 terrace with a terrace in the Main valley that is older than 20 cal ka BP. The t7 terrace level must then be formed after 20 cal ka BP, but before the Allerød-interstadial, because an undisturbed fall-out layer of Laacher See Tephra (eruption dated 11.06±0.01 ^{14}C ka BP or 13.0–13.2 cal ka BP; Friedrich et al., 1999) has been found in overbank deposits overlying the t7 terrace deposits over considerable distance (at the "Sandhof," Fig. 2; Dambeck and Bos, 2002; Dambeck and Thiemeyer, 2002; Dambeck, 2005; Table 1). Additionally, these overbank deposits were dated to 13.2 (±1.3) ka (Table 1) and a Late Glacial mollusc fauna composition was identified below and above the tephra layer (Dambeck, 2005; Table 1). These ages are supported by data from Scheer (1978), who presented ^{14}C dates of wood and bone found amidst t6 and t7 channel sands (>46.5 and 42.8±1.65 ^{14}C ka BP for t6; 18.3±1.0 and 15.1±0.1 ^{14}C ka BP for the t7; or for the latter two: 21.0 and 18.3 cal ka BP). Dambeck (2005) found no evidence in the northern Upper Rhine Graben for a braided level of Younger Dryas age, in contrast to the distinct terrace found downstream in the Lower Rhine Embayment (Ebing Terrace or Niederterrasse 3; cf. Schirmer, 1995; and Jüngere Niederterrasse; cf. Klostermann, 1992) and the Rhine-Meuse delta (Terrace X; Pons, 1957; Berendsen et al., 1995; Törnqvist, 1998; Berendsen and Stouthamer, 2001).

2.2. Holocene palaeomeander terraces in the Gernsheim region

The Holocene terrace sequence is characterised by large Rhine River abandoned palaeomeanders. Based on crosscutting relationships, sedimentological characteristics, and soil characteristics (Bodenkarte der nördlichen Oberrheinebene 1:50 000; HLfB, 1990), Rosenberger et al. (1996) identified a preserved patchwork of palaeomeanders (Fig. 2). They labelled them in more or less chronological order: "meander system" (ms) 1 through 14, whereby ms1 is the oldest palaeomeander. To avoid any confusion and to allow for direct comparison with previous research, we use this nomenclature in our study when referring to individual palaeomeanders. Fetzer et al. (1995) and Rosenberger et al. (1996) grouped the palaeomeanders into three "terrace levels:" the older, middle, and younger meander system. Dambeck and Thiemeyer (2002) and Dambeck (2005) improved the chronostratigraphic framework and applied it to the entire northern Upper Rhine Graben. In this study, we use their Holocene terrace level classification, grouping the sequence into latest Glacial/Early Holocene (~13–6 ka), Middle Holocene (~6–3 ka) and Late Holocene (~3–0 ka) terraces. Holocene terraces differ based on fluvial style/morphology, overbank sediment characteristics, and elevation. Each terrace is incised into the preceding terrace (Dambeck, 2005); net incision (lowering of channel top level) of successive palaeomeanders is on a decimetre scale.

The Early Holocene terrace hosts well-developed palaeomeanders, which were successively abandoned (Table 1). Meander sinuosity initially increased, but decreased after the middle Boreal (Dambeck and Bos, 2002). By the time of the late Atlantic, the meandering had evolved and produced a Middle Holocene surface (ms11b, Fig. 2). This terrace is characterised by strongly decreased meander sinuosity and more clayey overbank deposits (Dambeck and Thiemeyer, 2002; Dambeck, 2005). In the Gernsheim region, this resulted in almost straight channels (ms11b; Fig. 2), although more to the north in the Upper Rhine Graben large meanders continued to develop (Fig. 1). Deposition of clayey overbank deposits started ~6.7 (±1.1) ka (Dambeck, 2005; HDS-743), nearly synchronous with infilling of the last abandoned channels from the Early Holocene palaeomeanders (6.3 cal ka BP; Table 1). Eutric Vertisols developed in these clayey deposits (Dambeck and Thiemeyer, 2002), giving them their characteristic dark colour and name ("Black Clays"). Deposition of the Black Clay stopped ~2.7 (±0.5) ka at the Subboreal/Subatlantic transition (Table 1). From this time onwards, the river formed the Late Holocene terrace characterised by even smaller meander bends and silty overbank deposits (Dambeck and Thiemeyer, 2002; Dambeck, 2005). In the study area (Fig. 2), the River Rhine transformed into a low-sinuosity single channel river with narrow floodplains, and locally in-channel islands.

3. Materials and methods

To reconstruct the palaeogeographic development of the study area (the Gernsheim region), three detailed cross sections across the entire Late Glacial and Holocene floodplain were cored (Fig. 2). Cross section locations have been selected so that all major palaeomeanders were incorporated. To establish differences in elevation between the successive palaeomeanders, a digital elevation model (DEM) was

Table 1
Abandonment of terraces and palaeomeanders in the study area and dating evidence (locations in Fig. 2)

Palaeomeander	Abandonment (age)	Dating method (sample number)	References[a]
Ms12,13,14	Embanked ~100 a BP	Historical evidence	1
Ms11b	Early Subatlantic (2.7 ka)	OSL dating (HDS-746) not in field study area	1 (Appendix 3)
Ms7	Late Atlantic (~6.0 cal ka BP)	Palynological analyses	1 (Appendix 4–7)
Ms6	Late Boreal (~8.0 cal ka BP)	Palynological analyses	1 (Appendix 4–7)
Ms5	Older than ms6	Crosscutting relationship	3
Ms3	Late Atlantic (6.3 cal ka BP)	^{14}C dating (UtC-7634) and (UtC-9503)	1 (Appendix 2)
Ms1	Middle Boreal (8.6 cal ka BP)	^{14}C dating (UtC-10561) Palynological analyses	1 (Appendix 2, 4–6); 5
Ms2	Early Boreal (9.6 cal ka BP)	^{14}C dating (UtC-7693) and (UtC-9370)	2; 4; 5
Z	Minimum age: 9.8 ka	OSL dating of overbank deposits (HDS-596)	2; 4; 5
T7	Before ~13 cal ka BP	Stratigraphic relationship (underlying Laacher See tephra); OSL dating (HDS-1016); ^{14}C dating	1; 2; 4; 6
T6	Before ~18 cal ka BP	^{14}C dating, correlation with other terraces	1; 6

[a] See references (1 = Dambeck, 2005; 2 = Dambeck and Thiemeyer, 2002; 3 = Rosenberger et al., 1996; 4 = Dambeck and Bos, 2002; 5 = Bos et al., 2008; 6 = Scheer, 1978) for dating techniques and detailed sample information.

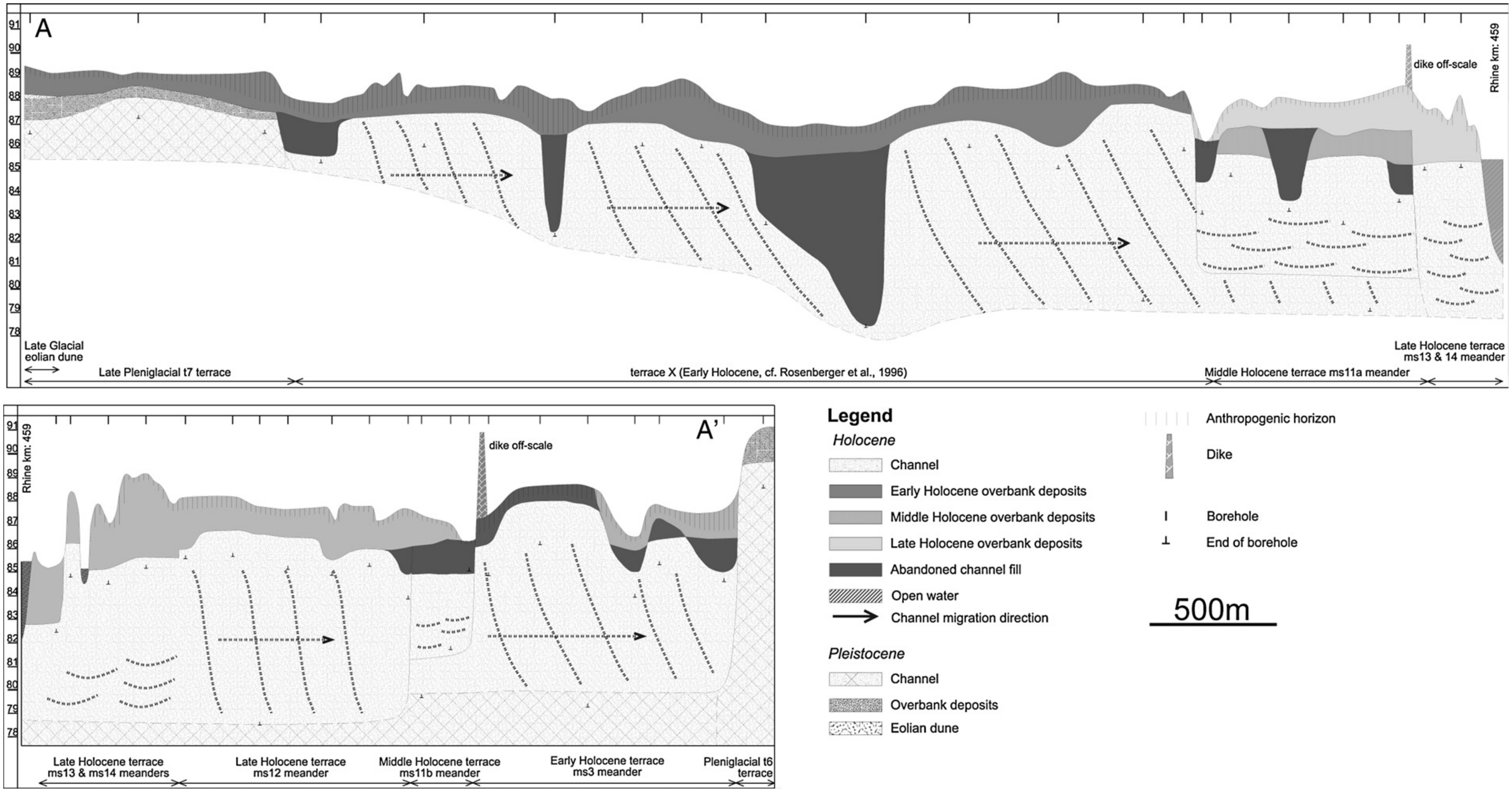

Fig. 3. Cross section A–A′ Eich (Rheinland-Pfalz)–Klein Rohrheim (Hessen), for location see Fig. 2. Channel migration direction (derived from Fig. 2) is indicated for clarification.

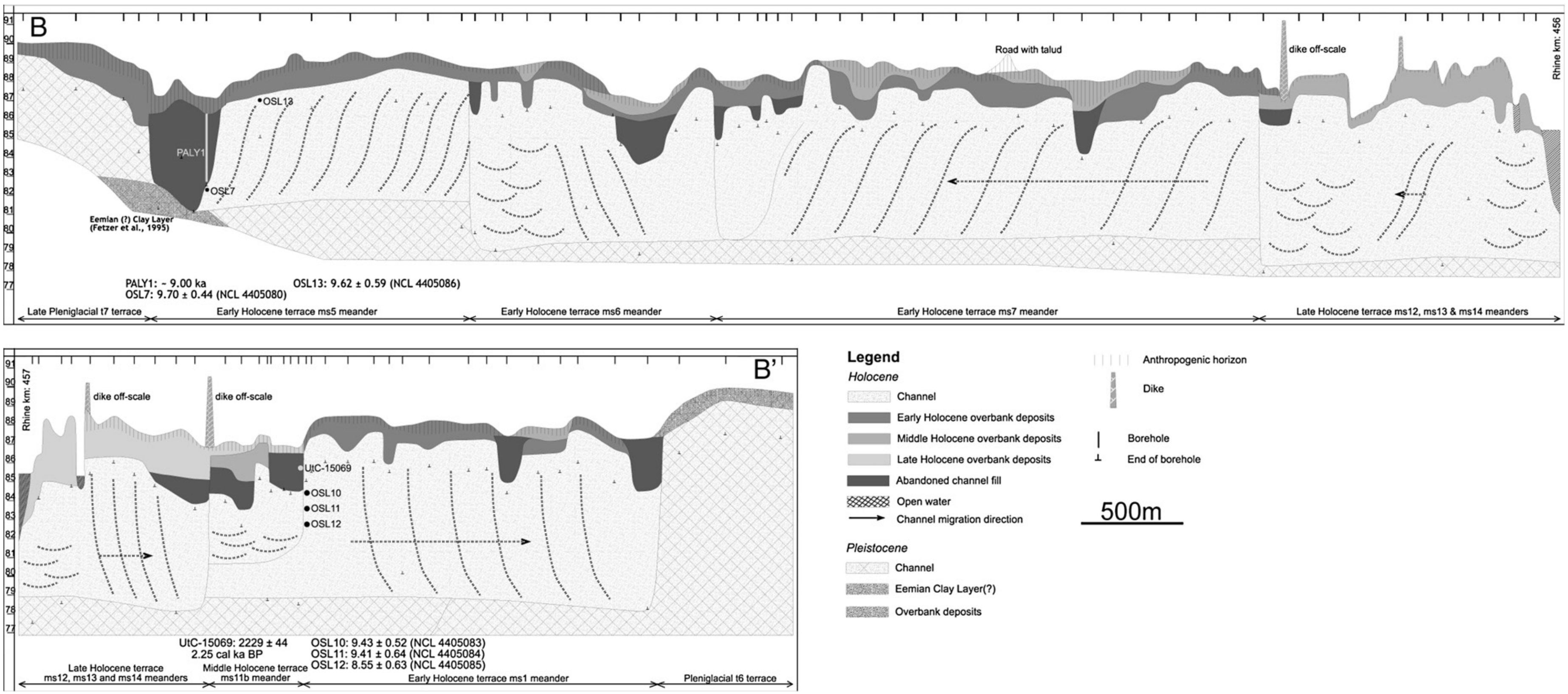

Fig. 4. Cross section B–B′ Mettenheim (Rheinland-Pfalz)–Gross Rohrheim (Hessen), for location see Fig. 2. Channel migration direction (derived from Fig. 2) is indicated for clarification.

constructed. Accumulated results from past research allowed rough age estimation of the deposits, although most palaeomeanders in the Gernsheim region lack direct dating. To test and improve correlations, additional dating evidence for palaeomeanders was collected. At dating sites, detailed cross sections were constructed to find the best suitable sediments and stratigraphic position for dating.

3.1. Field methods

Sediment cores were retrieved with hand-operated drilling equipment (Edelman auger, Dutch gouge, and Van der Staay suction corer; Van de Meene et al., 1979). The cores were logged in the field at 10-cm intervals with regard to texture, median grain size, gravel content (percentage estimated), maximum gravel size, organic (material) content (qualitatively), plant remains, colour, Ca and Fe content (qualitatively), oxidation/reduction, and groundwater level (Berendsen and Stouthamer, 2001). Sediments were classified using the USDA texture classification (USDA, 2005). Soil type and sedimentary facies were interpreted in the field. For the latter, we used a simple facies model based on the "fluviatile series" by Schirmer (1995) and the "alluvial environments" by Macklin and Lewin (2003). The three main characteristic lithogenetic units are (i) in-channel deposits, including gravelly channel lags and sandy channel fill sediments; (ii) overbank deposits consisting of silty natural levee deposits and clayey floodplain deposits; and (iii) abandoned channel fill deposits (i.e., filled oxbow lakes) consisting of silty and clayey deposits, usually rich in organic material. Although the lithology of abandoned channel fills resembles floodplain sediments, their stratigraphic position and thus genesis is very different, resulting in a separate lithogenetic unit.

For radiocarbon dating (2 locations, Fig. 2) and palynological analysis (3 locations, Fig. 2), cores of abandoned channel fills were obtained using a 6-cm diameter gouge. We cored sandy in-channel deposits for Optically Stimulated Luminescence (OSL) dating at five locations (Fig. 2) using two coring methods. Above groundwater, we used the Van der Horst sampling equipment to hammer 30-cm-long plastic tubes into the sandy sediments. Below groundwater, the Van der Staay suction corer was used to retrieve the sand (Wallinga and Van der Staay, 1999). Per palaeomeander, we obtained two to three samples from just below the sampled abandoned channel fill and/or from the adjacent sandy top of the inner bend point bar (Figs. 3–6).

3.2. Pollen and ^{14}C protocols

Cores for pollen analysis and radiocarbon dating (Table 2) were retrieved from the abandoned channels of three palaeomeanders (ms5, ms6, ms7; locations in Fig. 2) at the base of the infill with the greatest depth (PALY1 — Fig. 4; PALY2 — Fig. 5; PALY3 — Fig. 6). Pollen subsamples were prepared following standard pollen preparation methods (Faegri and Iversen, 1989). Pollen assemblages present in basal abandoned channel fills were compared with pollen assemblages from absolutely dated pollen sequences in the northern Upper Rhine Graben and adjacent areas (e.g., Bos, 2001; Dambeck and Bos, 2002; Bos et al., 2008). In this way, the minimum age of infilling and thus the timing of the abandonment of the palaeomeander was reconstructed.

From selected organic levels in core PALY2 and in an abandoned channel of the ms11b palaeomeander, samples were treated with a 5% KOH solution and wet-sieved over a 150-μm mesh, to allow selection of terrestrial macrofossils for accelerator mass spectrometric (AMS) radiocarbon dating. Radiocarbon ages were calibrated using OxCal 3.10 software (Bronk Ramsey, 1995, 2001).

3.3. OSL protocols

In the text below, we provide limited information on the quartz OSL dating methods used and the results obtained. In an electronic supplement (Appendix A) we provide some additional figures and information to complete the OSL information are available. For Quartz OSL dating (e.g., Wallinga, 2002), equivalent dose estimation was carried out on 180–212, 212–250, or 250–300 μm grain sizes, depending on overall coarseness of the sample (Table 3). Samples were treated with HCl and H_2O_2 to remove carbonates and organic material and thereafter with concentrated HF to dissolve feldspars and etch away the outer skin of the quartz grains. Treated samples were sieved again to remove grains that were severely damaged by the HF treatment. The single-aliquot regenerative-dose (SAR) protocol (Murray and Wintle, 2000, 2003) was used for measurement of the equivalent dose. Quartz OSL dating is suitable for these samples as the natural OSL signals are well below the OSL saturation level (Supplementary data, Fig. S-1). Preheat plateau tests on four samples indicated that for some samples the equivalent dose increased with preheat temperatures above 220 °C. We attribute this to thermal transfer from optically less sensitive traps into the main OSL trap. To avoid overestimation of equivalent dose from this phenomenon, we selected a relatively low preheat temperature of 200 °C for 10 s for all natural and regenerative doses. Suitability of the adopted SAR protocol was tested using a dose recovery test (e.g., Wintle and Murray, 2006) on all samples (average given to measured dose ratio: 0.97 ± 0.01, see also Supplementary data, Fig. S-2). Aliquots were only accepted for analysis if they passed recycling and recuperation tests (as suggested by Murray and Wintle, 2003). Absence of feldspar contamination was tested using infrared stimulation. For each sample, 21 to 30 aliquots were accepted for analysis. Because all samples consist of river-transported sediment, incomplete bleaching may occur (e.g., Wallinga, 2002). To avoid bias of results from heterogeneous bleaching, we iteratively rejected single-aliquot equivalent dose results that were more than two standard deviations from the sample mean (Supplementary data, Fig. S-3). Resulting dose distributions show no skewness, which indicates that the OSL signal of the vast majority of grains is completely reset upon deposition and that the mean equivalent dose of this data set can be used for the age calculation. The dose rate was estimated using high-resolution gamma ray spectroscopy, using material surrounding the equivalent dose sample. We assumed water saturation since deposition (20% water by weight) and immediate burial to the present sampling depth for calculating the cosmic ray contribution. All uncertainties in equivalent dose and dose-rate estimation are incorporated in the uncertainty quoted on the final age estimate (1σ).

3.4. Digital elevation model construction

A digital elevation model (DEM; Fig. 7) was constructed to more accurately establish differences in elevation between palaeomeanders. The Deutsche Grundkarten scale 1:5000 (LVRP, 1966–1979; HLV, 1977–1990) contain elevation data either as contour lines with 0.50- or 0.25-m interval or as point data that are accurate on the decimetre scale. By combining both types of data and using the ANUDEM interpolation algorithm (Hutchinson, 1989), a DEM with a minimum vertical resolution of 0.50 m was created. The obtained DEM (grid cell size 2×2 m) was validated using a random selection of original, initially not used, data. This yielded an averaged squared correlation coefficient (r^2) of 0.92.

4. Results

4.1. Description of cross sections

Three cross sections (A–A′; B–B′; C–C′) are shown in Figs. 3–5. Most borings penetrate the fine clastic sediments at the surface and terminate in the underlying sands. Deeper cores (up to 10 m) usually end in gravelly deposits. The lithological sequence in the cross sections is very uniform. Usually, the sands are overlain by 1- to 2-m-thick heterogeneous successions of fine-grained deposits (silt and clay).

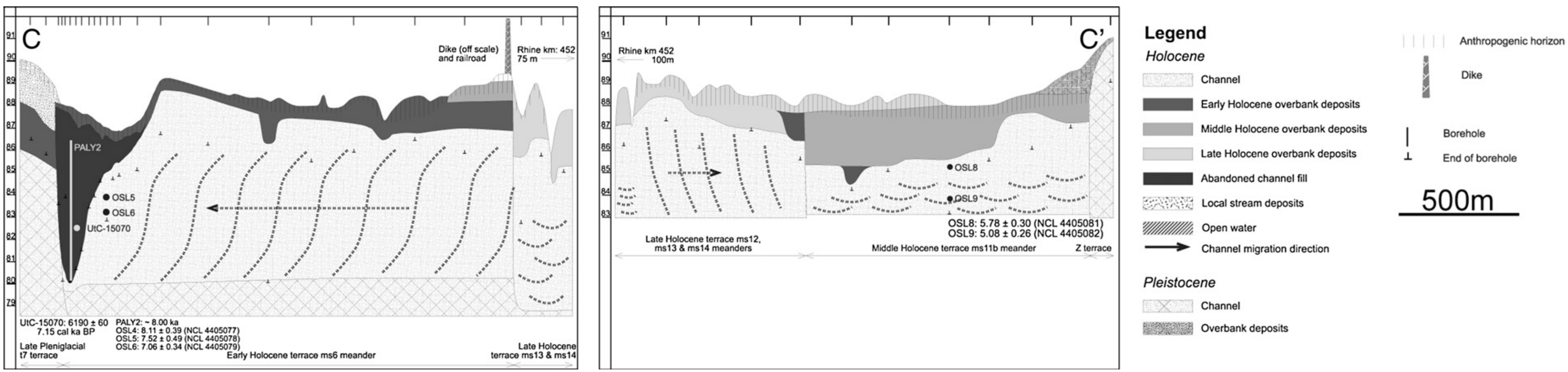

Fig. 5. Cross section C–C′ Rheindürkheim (Rheinland-Pfalz)–Nordheim (Hessen), for location see Fig. 2. Channel migration direction (derived from Fig. 2) is indicated for clarification.

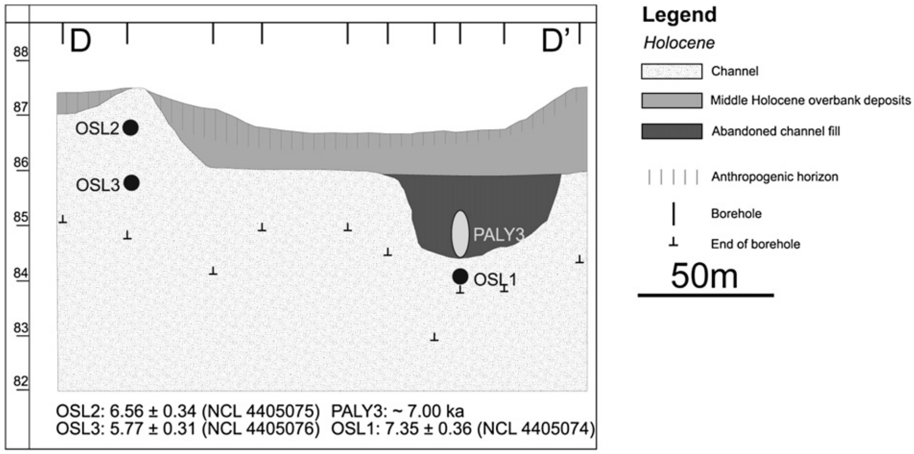

Fig. 6. Cross section D–D′—detailed cross section across the abandoned channel of palaeomeander ms7.

Sands range in grain size from 150 to 850 µm and often fine to coarse gravel can be found at the base. Sorting ranges from very poorly to well sorted. Most sandy deposits are moderately well sorted and, in general, sorting improves toward the top. In some cores, a large-scale fining-upward sequence is present, whereas in others several smaller fining-upward sequences were recognised. The degree to which the fining upward is developed differs from core to core. Sandy basal deposits are interpreted as bedload deposited as in-channel bars and mid-channel fills.

Coarse and gravelly deposits at the base of channel deposits are interpreted as channel lag deposits and indicate maximum channel depth. Channel lag deposits are typically moderately well sorted gravelly sands with a coarse sand matrix (420–1400 µm), whereas underlying terrace deposits generally consist of poorly sorted, medium coarse sand (210–420 µm). Because most channel lag deposits consist of reworked older terrace sediment, distinguishing channel lag deposits from underlying terrace deposits is sometimes hard. By identifying channel lags in 25 cores, we were able to estimate the depth (5–6 m) of all Holocene palaeomeanders (Figs. 3–5). This depth appears remarkably uniform in all cross sections, which may seem unusual for river channels where depth varies along curved channel segments (e.g., Bridge, 2003). This is related to the position of

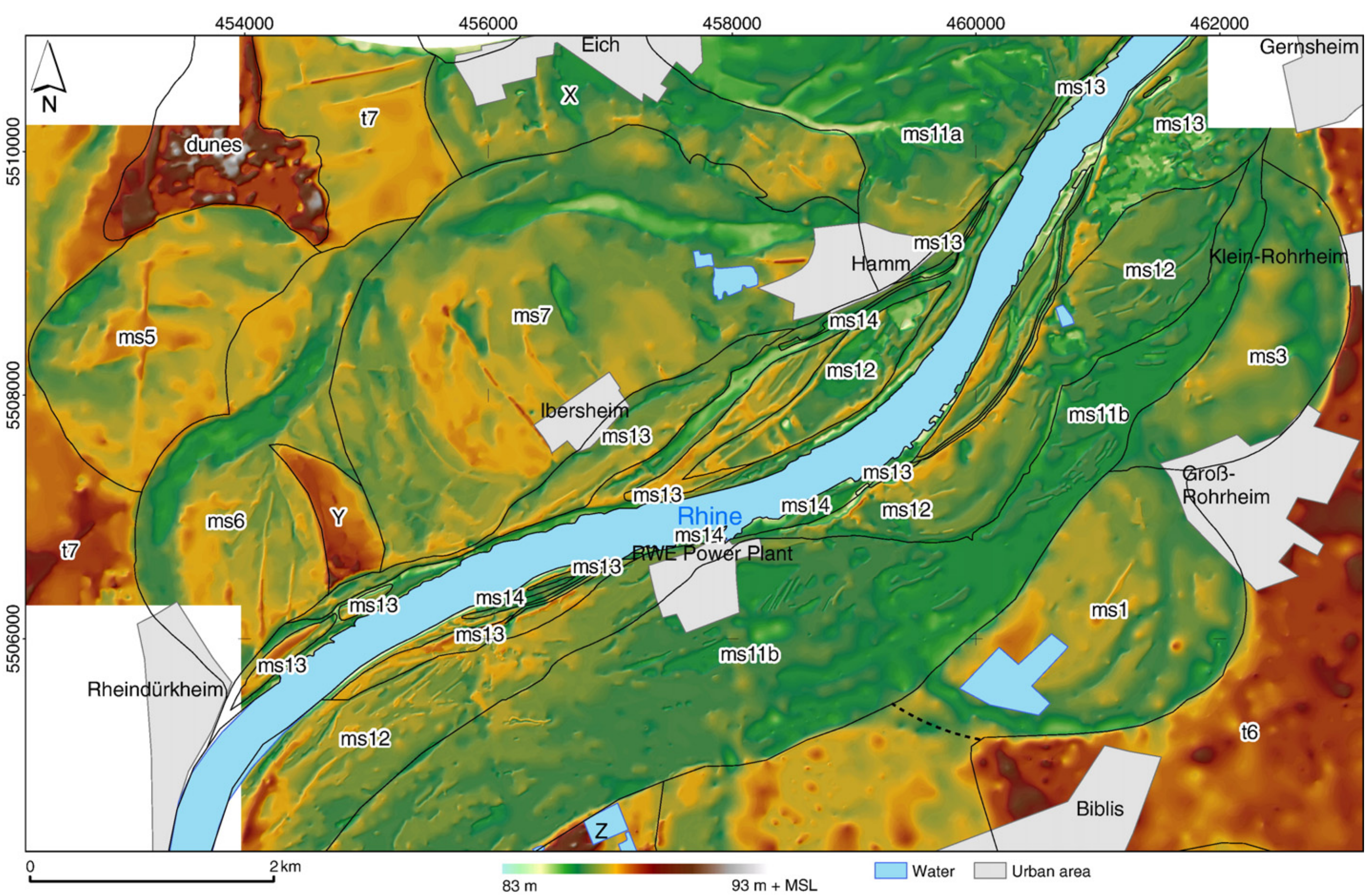

Fig. 7. Digital elevation model of the study area (terrace and palaeomeander classification from Rosenberger et al., 1996).

our cross sections that all cross the palaeomeanders at their central axes, where depths of the channels are similar during single meander formation.

Within the sandy channel deposits, channel-shaped features of considerable size occur (100–300 m wide), mostly filled with silty and sandy clay loam. These features are interpreted as abandoned channel fills based on their stratigraphic position and morphology.

The composition of the overlying finer layers varies throughout the cross sections. In some locations they are clayey (clay, clay loam, sandy clay loam), elsewhere they contain more silt (silty clay loam, silt loam, loam) or sand (sandy loam). A fining-upward sequence in these layers, for example a sequence from sandy loam/loamy sand to loam to clay loam, is present at most locations. The silty and clayey deposits are interpreted as natural levee and overbank deposits because these finer sediments are considered to be the result of suspended sediment deposition in floodplains in slow-flowing or stagnant water outside the channel. The transition between predominantly suspension-settled and underlying mostly bedload-carried deposits in boreholes is typically gradual. In the cross sections, the boundary between overbank and channel deposits was drawn where texture changes from dominantly silt/clay to dominantly sand.

4.2. Interpretation of the cross sections

Using crosscutting relationships between palaeomeanders and overbank criteria (texture) of Dambeck and Thiemeyer (2002) and Dambeck (2005) enabled grouping of all deposits within their terrace framework. For the Holocene deposits, we presume that the time lag between the deposition of channel deposits and the onset of overlying overbank deposition was minimal, so that the sedimentation should be attributed to the same system (see also Schirmer, 1995). This presumption is based on the idea of simultaneous changes in morphology of the palaeomeanders and pedogenic characteristics of Holocene overbank deposits (Dambeck and Thiemeyer, 2002). However, for the Late Weichselian terrace levels (t6, t7) this is less clear, as discussed below.

The highest terrace (91 m +MSL) at the easternmost part is interpreted as the Pleniglacial t6 terrace (Figs. 1 and 2). On top of the t6 channel sands, a layer of sandy loam is present. This layer may represent overbank deposits associated with the t6 terrace or be deposited by the channels of the Late Pleniglacial t7 terrace level. Flooding of the Pleniglacial t6 terrace level during the Holocene likely did not occur frequently or result in substantial deposition. This finding is based on the deposits being sandier than subsequent deposits, the large elevation difference of 2 m between the Pleniglacial t6 terrace and adjacent Holocene, and strong soil formation in overbank deposits (Dambeck, 2005). In support of this interpretation is the fact that during an extreme flood event in 1882/1883 A.D. flood waters only impacted the Late Holocene, Middle Holocene, and the lowest Early Holocene terraces (Dambeck, 2005).

The high-elevated surface in the westernmost part of the cross sections is interpreted as the Late Pleniglacial t7 terrace, based on elevation and stratigraphic position. In the west, cross section A–A′ (Figs. 2 and 3) ends at a Younger Dryas eolian dune (Sandhof; Dambeck, 2005), that overlies Allerød overbank deposits. Farther to the east, cross section A–A′ (Fig. 3) crosses the t7 terrace that grades into a lower elevated surface with large abandoned channel fills. On top of the Late Pleniglacial t7 channel deposits, overbank deposits are present, largely related to younger channels. Soil formation indicates that overbank deposition ended before the Middle Holocene at the very latest (Dambeck, 2005). In cross sections B–B′ and C–C′, these overbank sediments are abnormally thick, probably because of sediment supply from small alluvial fans coming from adjacent hillslopes that form the Graben rim in the west.

The cross sections traverse nine different Holocene palaeomeanders (Fig. 2), with every younger palaeomeander consecutively lower in elevation. The Early Holocene set of palaeomeanders east of the River Rhine consists of ms1 (cross section B–B′, Fig. 4) and ms3 (cross section A–A′, Fig. 3); and west of the River Rhine: ms5 (cross section B–B′, Fig. 4), ms6 (cross section B–B′ and C–C′, Figs. 4 and 5), and ms7 (cross section B–B′, Fig. 4). Palaeomeanders ms1, ms3, ms6, and ms7 all are of similar channel depth, ~7 m. The maximum depth of the ms5 palaeomeander is 5 m, which is shallow compared to other Early Holocene palaeomeanders (such as adjacent ms6 and ms7). The abandoned channel of the ms5 meander is incised into a layer of cohesive loamy clay (Fig. 4). In general, all Early Holocene channel sediments are overlain by loamy overbank deposits. The overbank deposits of ms7 — the youngest and lowermost Early Holocene palaeomeander — are commonly overlain by characteristic blackish, clayey Middle Holocene overbank deposits (Black Clays). This is also the case in the lowest morphological units (e.g., abandoned channels) of palaeomeanders ms1, ms3, and ms6.

Middle Holocene deposits are present in all three cross sections at a lower elevation than the Early Holocene deposits. The elevation difference between the top of the Middle Holocene point bar ms11b and the Early Holocene point bars is ~1.5 m. In the cross sections A–A′ (Fig. 3) and B–B′ (Fig. 4), two channel lags on top of each other were found; the upper one is presumed to belong to the Middle Holocene point bar. The lower channel lag correlates in depth with the channel lag of the Early Holocene palaeomeanders ms1 and ms3 and is, therefore, probably a remnant of these palaeomeanders. Possibly, the channels of ms11b were shallower (~5 m) than those of Early Holocene age. Channel deposits of ms11b are covered by clayey Middle Holocene overbank deposits (Black Clays) that are, because of the low elevation, commonly overlain by Late Holocene overbank deposits. Late Holocene point bars of palaeomeanders ms12, ms13, and ms14 occur in all cross sections adjacent to the present day River Rhine. The top of the Late Holocene channel deposits, which are at the same elevation as the top of the Middle Holocene point bar (cross section A–A′, Fig. 3 and B–B′, Fig. 4) if not slightly higher (cross section C–C′, Fig. 5). Channel depth (on average ~7 m) is greater than that of the Middle Holocene channels, as is evident from the occurrence of deeper channel lags (Figs. 3 and 4). The thickness of the silty Late Holocene overbank deposits is considerably larger than the thickness of Early and Middle Holocene overbank sediments in all cross sections.

4.3. New dating of individual palaeomeanders

In addition to dating results of Dambeck and Thiemeyer (2002) and Dambeck (2005), the palaeomeanders were dated using the following techniques:

(i) Crosscutting relationships in planview (maps, geomorphology, DEM) and cross sections were used as relative dating evidence. An important observation from the DEM (Fig. 7) is that the Middle Holocene palaeomeander (ms11b) has the lowest elevation and forms the deepest floodplains.

(ii) The timing of the abandonment of the Early Holocene palaeomeanders ms5, ms6, and ms7 was estimated by applying pollen analyses on their abandoned channel fills. Fig. 2 shows the locations of the analysed pollen sequences, as summarized in Table 2. In core PALY1 (Fig. 8A), the pollen assemblage at the base of the infill with abundant deciduous trees is dominated by *Corylus* (Hazel). Bos et al. (2008) dated the maximum distribution of Hazel in this area to the Boreal and the occurrence of *Tilia* (Lime) indicates that this pollen assemblage resembles the middle Boreal (~9.0 cal ka BP). The onset of the infill of core PALY2 (Fig. 8B) may be dated to the early Atlantic (~8.0 cal ka BP) given the high values of *Quercus* (oak) pollen throughout the entire diagram and the absence of human

Table 2
Abandonment age estimates from palynological analyses and radiocarbon dating for Early and Middle Holocene palaeomeanders in the Gernsheim region

Sample name	Pollen diagram	Palaeomeander (Fig. 2)	Stratigraphic position	Depth (m)	Elevation (m +MSL)	Estimated abandonment (cal ka BP)
PALY1	Fig. 8A	ms5	Abandoned channel fill (Fig. 4)	2.10–5.00	87.1	~9.00
PALY2	Fig. 8B	ms6	Abandoned channel fill (Fig. 5)	3.50–8.00	87.8	~8.00
PALY3	Fig. 8C	ms7	Abandoned channel fill (Fig. 6)	1.10–2.65	86.8	~7.00
Sample name	Analysed material	Palaeomeander	Stratigraphic position	Depth (m)	Elevation (m +MSL)	Estimated abandonment (cal ka BP)
UtC-15070	Terrestrial seeds	ms6	Abandoned channel fill (Fig. 5)	5.37	87.8	7.15[a]
UtC-15069	Charcoal	ms11b	Abandoned channel fill (Fig. 4)	1.5	86.8	2.25[b]

Radiocarbon dating was performed at the R.J. van de Graaff laboratory, Utrecht University.
[a] Radiocarbon age: 6190±60 ^{14}C.
[b] Radiocarbon age: 2229±44 ^{14}C.

influence in the lower part of the infill. This interpretation is further confirmed by the radiocarbon date from this core (see below). The pollen diagram from location PALY3 (Fig. 8C) shows high pollen values of *Quercus* and subsequent increase of anthropogenic indicators (*Plantago lanceolata*, *Artemisia,* Cerealia*)*. This pollen assemblage correlates to biozone Atlantic II (sensu Kalis et al., 2003; see dating below), indicating that infilling started in the middle Atlantic (~7.0 cal ka BP).

(iii) Two abandoned channel cores were additionally dated using radiocarbon dating of organic material. In the PALY2 core, we sampled calcareous gyttja (organic mud) halfway through the core (sample number UtC-15070) at a depth of 537 cm at the transition from biozones Atlantic I to Atlantic II (Fig. 8B). An assemblage of seeds and fruits of riparian herbs (*Phragmites australis*; *Typha* sp.; *Scirpus lacustris*; *Carex* sect. *Acutae*) was selected. The age of abandonment in this palaeomeander is dated at 6190±60 ^{14}C or 7.2 cal ka BP (Table 2). Biozone Atlantic II is characterised by higher values of nonarboreal pollen and Poaceae, which both indicate more open forest vegetation. Kalis et al. (2003) date this transition at ~7.5 cal ka BP, which is in good correspondence with our dating results. The abandonment age from the radiocarbon sample is somewhat younger than indicated by the pollen analyses — probably because the radiocarbon sample is located 2.5 m higher in the abandoned channel fill sequence. Therefore, we consider this age to be a minimum age of abandonment of the ms6 meander. The second sample for radiocarbon dating (sample number UtC-15069), consisting of charcoal, was retrieved from a abandoned channel fill of the ms11b terrace level (Fig. 4) and yielded 2229±44 ^{14}C or 2.3 cal ka BP (Table 2).

(iv) We dated five palaeomeanders (locations in Fig. 2) using OSL (dating results and stratigraphic position in Table 3). All OSL dates fit the palynological analyses (Table 2) and seem to be approximately correct, even though dates in sequence (OSL2/3; OSL4/5/6; OSL8/9; OSL10/11/12) are often not in the correct stratigraphic order. Dose rate measurements for two samples (Table 3; OSL3, OSL6) showed unexpected high values, possibly from sampling of a finer matrix above and/or below the sample used for equivalent dose estimation. As a consequence of the high dose rate measurements, the OSL ages for these two samples are likely underestimated, and they are rejected.

Dating results refine the chronology of Dambeck and Thiemeyer (2002) and Dambeck (2005) (Fig. 9). Although single dates on meander cut offs and channel abandonment imply that changes are sudden, it may take several decades before a channel is abandoned and oxbow-lake fill in starts. In-channel OSL dates (Table 3) indicate the last moment of meander formation (bar deposition), whereas dates from the base of the abandoned channel fill (radiocarbon, pollen analyses) mark the onset of abandoned stage.

Abandonment of the Early Holocene palaeomeander ms5 is dated slightly younger than previously estimated, as it was abandoned in the early/middle Boreal (~9.5 ka; OSL7, OSL13, PALY1). Palaeomeander ms1 was abandoned after ~9.4 ka (dating of point bar sands; OSL10, OSL11, OSL12) but before ~8.6 cal ka BP (dating of abandoned channel infill; Dambeck, 2005; Table 1). Palaeomeander ms1 was abandoned a little later than Early Holocene palaeomeander ms5, which was by then already succeeded by palaeomeander ms6 (Fig. 9). The results of pollen analysis, radiocarbon dating and OSL dating of palaeomeander ms6 indicate abandonment at the onset of the Atlantic (~8.0 ka; OSL4, OSL5, UtC-15070, PALY2), and for the Early Holocene palaeomeander ms7 during the middle Atlantic, ~7.0 ka (OSL1, PALY3) or perhaps a little later (~6.6 ka; OSL 2). Dambeck (2005) suggested that this palaeomeander was connected to palaeomeander ms3, which was abandoned ~6.3 cal ka BP or earlier (Table 1). Possibly both palaeomeanders were cut off simultaneously between 7.0 and 6.3 ka (Fig. 9). Dating of channel sands (~5.78 ka; OSL8 and ~5.08 ka; OSL9) of Middle Holocene palaeomeander ms11b confirms that this palaeomeander was still active in the second half of the Subboreal (Fig. 9). This is further confirmed by dating of an abandoned channel fill of the ms11b (UtC-15069). Although this sample is located high in the fill sequence (1 m above the ms11b channels sands), it gives a minimum age of 2.2 cal ka BP for abandonment.

5. Palaeogeographical development

We integrated all available data into a palaeogeographical reconstruction (Fig. 10) for the Gernsheim region. The locations of the channels in Fig. 10 are based on the geological map (Rosenberger et al., 1996), as far as they are preserved. To position the eroded channels on the maps for a given time interval, we interpolated between preserved meander fragments, assuming constant width of the channels. The trends in fluvial style and terrace level as mapped in the Gernsheim region are used to describe the development of the entire northern Upper Rhine Graben, where this diverges from the general trend this will be indicated in the text.

During the Pleniglacial (Fig. 10A), the River Rhine was a braided river with minor overbank deposition and very sandy and gravelly channel deposits in the northern Upper Rhine Graben. Its deposits are preserved as the t6 terrace level that is present at the eastern margin of the study area (Figs. 2 and 10B). This terrace formed across the entire width of the northern Upper Rhine Graben, indicating large-scale reworking and perhaps aggradation.

The onset of formation of the successive Late Pleniglacial t7 terrace level (Fig. 10B) must be dated after ~20 ka (Dambeck, 2005; for dating evidence see Section 2.1). This level has an ~1.5-m lower elevation than the t6 level, implying that incision took place after 20 ka. The strong reduction of width was accompanied with a change in fluvial style. The t6 terrace level is considered to be fully braided, whereas the t7 level reveals one remnant of a winding channel in the westernmost part of the study area (Fig. 2), which can be traced over a distance of 15 km (Dambeck, 2005). Because the width and depth (<3 m) of the filled channel are small, we presume multiple semisinuous channels to have been active at locations where Holocene channels were active

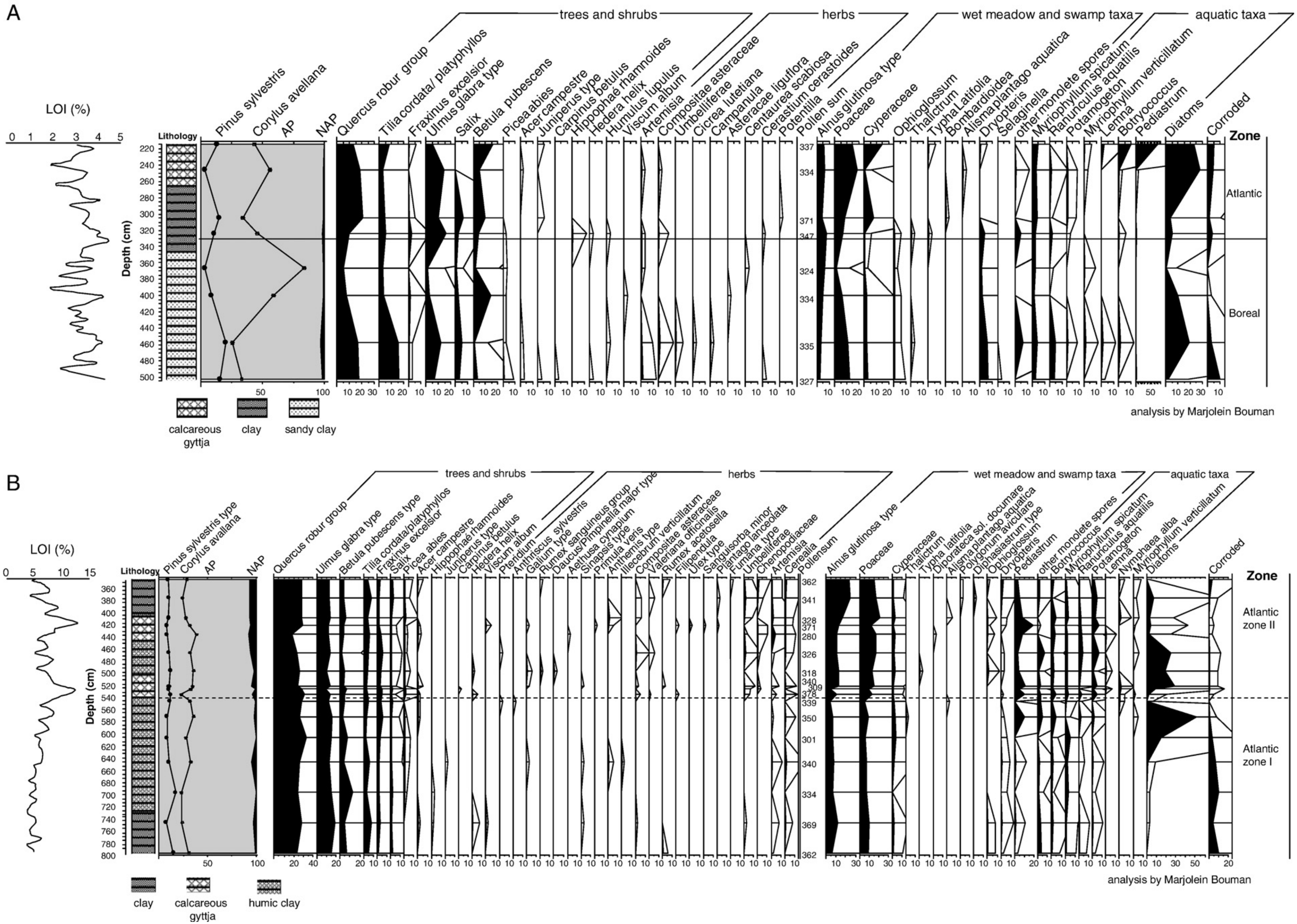

Fig. 8. Pollen diagram Sandhof, abandoned channel fill of palaeomeander ms5 (A); pollen diagram Mückenhäuser Hof, abandoned channel fill of palaeomeander ms6 (B); pollen diagram Eich, abandoned channel fill of palaeomeander ms7 (C).

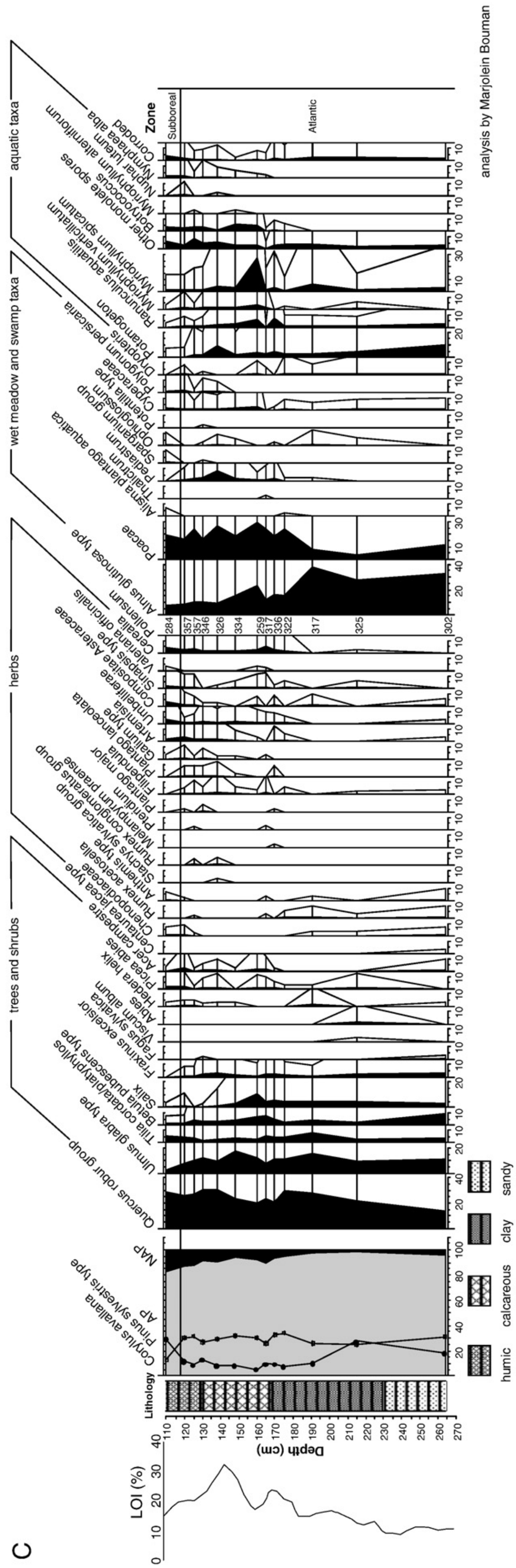

Fig. 8 (*continued*).

Table 3
OSL dates for Early and Middle Holocene palaeomeanders in the Gernsheim region

Sample name	NCL code	Age (ka)	Palaeomeander (Fig. 2)	Stratigraphic position	Surface elevation (m +MSL)	Averaged depth below surface (m)	Used grain size range (μm)	Radionuclide concentration (Bq kg^{-1})			Total dose rate (Gy/ka)	Equivalent dose (Gy)
								^{238}U	^{232}Th	^{40}K		
OSL1	4405074	7.35±0.36	ms7	Abandoned channel floor (Fig. 6)	86.8	3.08	180–212	12.04±0.21	12.77±0.46	483±21	1.726±0.069	12.7±0.3
OSL2	4405075	6.56±0.34	ms7	Mid-channel bar (Fig. 6)	87.6	1.28	250–300	11.98±0.19	10.93±0.33	489±19	1.694±0.066	11.1±0.4
OSL3	4405076	5.77±0.31	ms7	Mid-channel bar (Fig. 6)	87.6	2.52	250–300	16.51±0.16	16.89±0.37	685±4	2.287±0.073	13.2±0.6
OSL4	4405077	8.11±0.39	ms6	Mid-channel bar (Fig. 5)	87.0	2.38	212–250	13.54±0.21	13.25±0.37	522±6	1.857±0.061	15.0±0.5
OSL5	4405078	7.52±0.49	ms6	Mid-channel bar (Fig. 5)	87.0	2.41	250–300	15.06±0.23	12.48±0.45	416±20	1.581±0.065	11.9±0.6
OSL6	4405079	7.06±0.34	ms6	Mid-channel bar (Fig. 5)	87.0	2.76	250–300	13.96±0.14	14.12±0.32	608±3	2.072±0.067	14.6±0.5
OSL7	4405080	9.70±0.44	ms5	Abandoned channel floor (Fig. 4)	87.1	3.35	212–250	14.56±0.14	14.26±0.30	424±3	1.644±0.055	15.9±0.5
OSL8	4405081	5.78±0.30	ms11b	Point bar (Fig. 5)	88.0	3.60	212–250	15.76±0.22	13.92±0.37	496±22	1.797±0.072	10.4±0.3
OSL9	4405082	5.08±0.26	ms11b	Point bar (Fig. 5)	88.0	4.25	180–212	13.55±0.21	11.78±0.38	505±6	1.769±0.060	9.0±0.4
OSL10	4405083	9.43±0.52	ms1	Point bar (Fig. 4)	86.8	3.44	212–250	12.33±0.15	12.44±0.31	480±4	1.689±0.056	15.9±0.7
OSL11	4405084	9.41±0.64	ms1	Point bar (Fig. 4)	86.8	4.12	250–300	11.58±0.18	10.52±0.33	392±18	1.397±0.060	13.2±0.7
OSL12	4405085	8.55±0.63	ms1	Point bar (Fig. 4)	86.8	4.80	212–250	12.48±0.20	11.66±0.45	446±18	1.565±0.063	13.4±0.8
OSL13	4405086	9.62±0.59	ms5	Terminal point bar (Fig. 4)	88.5	1.65	212–250	20.91±0.25	19.22±0.59	372±18	1.696±0.064	16.3±0.8

OSL dating was carried out at the Netherlands Centre for Luminescence Dating at Delft University of Technology.

later. The width reduction, winding channel, and the net removal of sediment (net incision) during formation of the t7 terrace level indicate Rhine discharge concentrated in fewer channels relative to the t6 braided stage. Taking into account clear indications for meandering in the next younger levels, this allows classifying the t7 level as a transitional system from braided to meandering. Although braided rivers can also be characterised by overbank deposition (e.g., Brierley and Hickin, 1992), the more-transitioned Late Pleniglacial t7 river system is likely the source of the overbank deposits on the Pleniglacial t6 terrace level.

Terrace levels from the Allerød-interstadial (13 ka) to the Preboreal (~11 ka) are remarkably poorly preserved in the northern Upper Rhine Graben (Fig. 10C). Dambeck (2005) suggested that the surface hosting the villages Nordheim and Wattenheim (indicated with a "Z" in Fig. 2) dates from this period. Crosscutting relationships make this surface older than the Preboreal/Boreal meander to the east (Fig. 2), but it is on average lower in elevation (and therefore younger) than the Late Pleniglacial t7 terrace level (Fig. 7). Based on the elevation data from the DEM (Fig. 7) and crosscutting relationships, the surface just east of the "Sandhof-dune" (indicated with an "X" in Figs. 2, 3, 7) and a part of the point bar belonging to palaeomeander ms6 (area indicated with a "Y" in Fig. 7) may have been coeval. The preserved patches are too small to reveal the associated fluvial style, but Dambeck (2005) described thin and very sandy Younger Dryas overbank deposits. Therefore, we hypothesize that this level developed under transitional conditions from braided to meandering — similar to the t7 terrace level — and was not fully braided.

The next younger level is palaeomeander ms5 (Fig. 2), which is part of the Early Holocene meandering system (Fig. 10D). It developed mainly during the Preboreal. Dating results (OSL7, OSL13, PALY1 and OSL10, OSL11, OSL12; Tables 2 and 3) suggest that the ms5 palaeomeander was coeval with the Early Holocene ms1 palaeomeander at the eastern part of the Early Holocene floodplain. It seems, however, unlikely that these two palaeomeanders were directly connected to each other. This is based on the connection of the ms1 meander to the ms2 palaeomeander more to the south (Fig. 2),

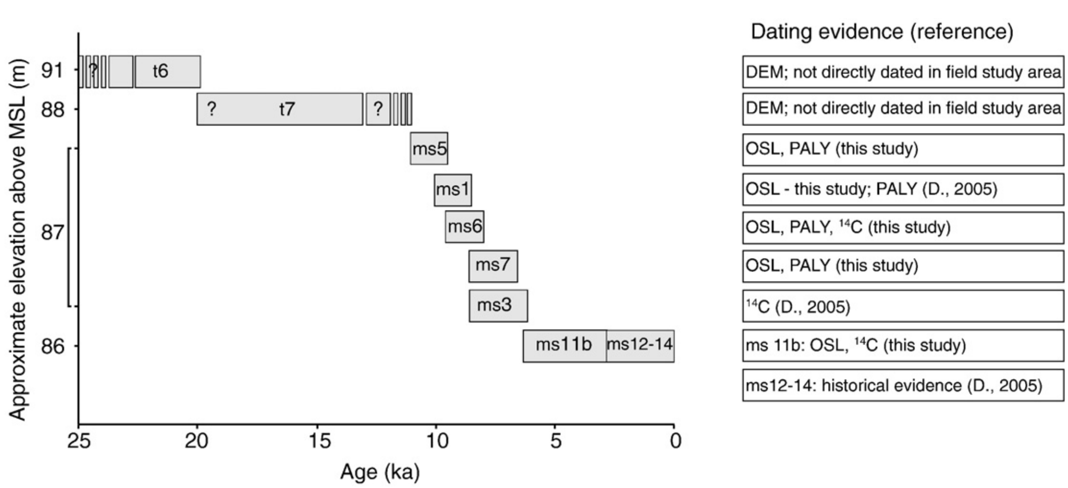

Fig. 9. Period of activity and approximate elevation of Late Weichselian terrace and Holocene palaeomeanders in study area with dating evidence (OSL=optically stimulated luminescence; PALY=palynological analysis; ^{14}C=radiocarbon dating; DEM=digital elevation model) and references (D., 2005=Dambeck, 2005). Elevation values are estimated from the top of channel sands in the cross sections and are averaged per terrace (group of palaeomeanders). Note that the y-axis is not on scale and that numbering of palaeomeanders is not chronological.

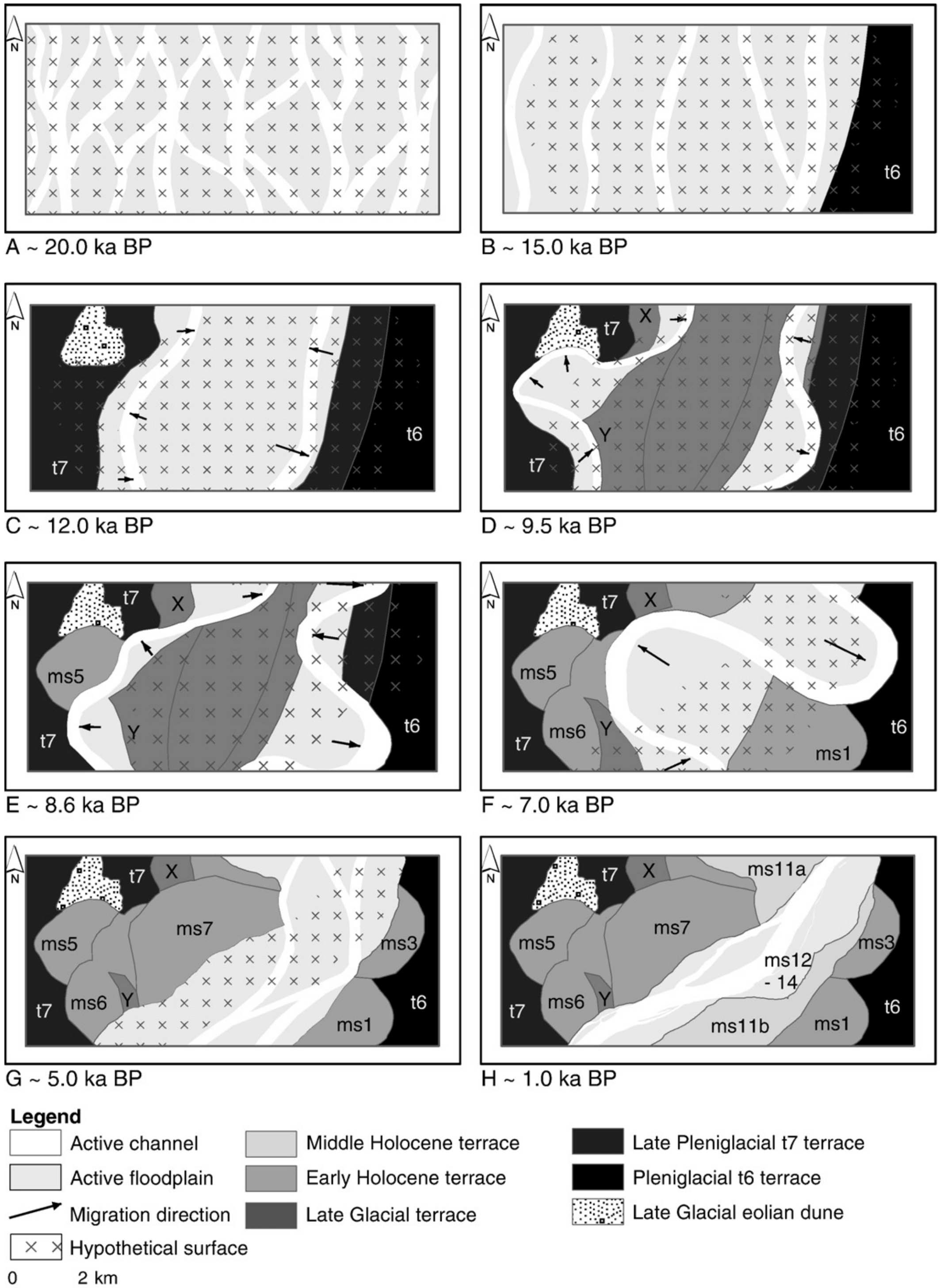

Fig. 10. Eight time slices showing reconstructed palaeogeographic development of the field study area. Figures integrate new results (this paper) with results from Rosenberger et al. (1996) and Dambeck (2005).

evidenced by dating results (Tables 1 and 2). Furthermore, channel depths differ between ms1 (~7 m) and ms5 (~5 m). The two channels may have been separated from each other by remnants of an older terrace level (e.g., Y and Z in Fig. 2). Probably, discharge was not evenly distributed over the two channels, which may explain the shallowness of palaeomeander ms5 (Fig. 4).

At ~9.0 ka (Table 2), the ms5 meander was cut off by the channel that formed the Holocene ms6 meander (Fig. 10E). The two-channel situation continued after the abandonment of ms5, when the meanders ms6 and ms1 coexisted (between ~9.5 and ~8.6 ka). With the abandonment of meander ms6 (~8.0 ka), flow became concentrated in a single channel belt (ms3, ms7, ms11-14).

After a cut off that abandoned the ms1 palaeomeander, the Early Holocene ms7 and ms3 meanders developed (Fig. 10F). The ms7 and ms3 meanders were cut off nearly simultaneously in the middle Atlantic between 7.0 and 6.3 ka (Tables 1–3).

The resulting Middle Holocene level is — in the Gernsheim region — geomorphologically characterised by the absence of large winding meanders, but several poorly developed palaeomeanders are recognised (Fig. 10G). The preserved parts of the Middle Holocene ms11b system consist of low-sinuosity channels of relatively small width and shallow depth (<5 m) — very different from the preceding and following palaeomeanders. Incision continued during formation of ms11b, and the top of the channel deposits is ~1 m lower than the top of the Early Holocene channel deposits. The characteristic Middle Holocene overbank deposits (Black Clays) are found on top of the ms11b channel sands, but also on the lower parts (abandoned channels, youngest palaeomeanders) of the Early Holocene surfaces.

After 2.7 ka (Table 1), river flow concentrated in a single low-sinuosity channel (Fig. 10H). Net incision between the Middle and Late Holocene was apparently small or even absent, which resulted in Late Holocene overbank deposits commonly covering Middle Holocene surfaces. Late Holocene overbank sediments are generally thicker (>1 m) and contain much more silt than older overbank sediments.

6. Discussion

From the results the following observations are highlighted:

(i) Every single younger terrace level formed during the last ~20 ka is lower in elevation, implying that incision of the terrace sequence starts after ~20 ka and continues until the Late Holocene terrace level (which seems not to be incised anymore);
(ii) preservation of fluvial deposits from the Allerød-interstadial and Younger Dryas-stadial is poor;
(iii) locally, two parallel meandering channels were active during the Early Holocene;
(iv) a local trend of decreasing meander sinuosity from the late Atlantic onward exists, resulting (in the Gernsheim region) in a shallow multichannel system during the Middle Holocene.

We evaluate the potential importance of autogenic processes and allogenic controls on the formation of the terrace sequence. While doing so, the study area is compared with other stretches of the River Rhine and other river catchments in Europe.

6.1. Complex response to glacial–interglacial climate change

The climatically driven transition from braided to meandering fluvial style in the northern Upper Rhine Graben started during the Late Pleniglacial, after ~20 ka, and is associated with the onset of incision and formation of the t7 level within the t6 level. During the Late Pleniglacial and Late Glacial, alpine glaciers retreated significantly (Ivy-Ochs et al., 2004) and formed large alpine lakes (e.g., Bodensee/Lake Constance), which acted as sediment sinks (Hinderer, 2001). This reduced sediment delivery to the Upper Rhine Graben and transformed the River Rhine from a transport-limited to a supply-limited river (terminology cf. Houben, 2003). In the Upper Rhine Graben, this forcing in sediment supply owing to properties of the upstream catchment, superimposed on climatic warming, is likely to have been the trigger for the onset of incision and transition from braided to meandering. Seemingly, this trigger was much more important for *initiation* of terrace formation than the climatic change during the Late Glacial–Holocene transition. Leigh et al. (2004) found a braided-meandering transition that started before the onset of the Late Glacial (~17 ka BP) in unglaciated catchments in the southeastern United States. This additionally underlines that the braided-meandering transition in fluvial style is not exclusively related to glacial–postglacial transitions such as the onset of the Late Glacial (~14.5 ka BP).

During the Late Glacial (and probably also in the Younger Dryas), a fluvial system of transitional fluvial style continued to be present in the northern Upper Rhine Graben. Subsequently a clearly meandering system emerged, of which the oldest meanders must have formed within the Preboreal. However, in parts of the northern Upper Rhine Graben (e.g., the Gernsheim region), initially two meandering channels existed at positions inherited from a preceding multichannel transitional system. These channels coexisted for some hundreds of years until the middle Boreal. The complete transition from a fully braided to a meandering system was slow.

Recognition of braided-transitional stages developing to single channel meandering systems at the last glacial interglacial boundary is widely recognised across northern Europe (Vandenberghe, 1995), and commonly during the Younger Dryas fluvial style temporarily reversed (e.g., in the Lower Meuse valley) (Huisink, 1997; Tebbens et al., 2000). Not all fluvial systems show response to each considerable climatic oscillation, which can be explained by geomorphic threshold concepts and lagged climate–vegetation interaction (Vandenberghe, 1995). Slow and stagewise transition from fully braided via slightly curving/multichannel to multichannel meandering and single channel meandering was also reported by Vandenberghe (1995) for the Late Glacial River Warta (Poland). In the United Kingdom, Brown et al. (1994) and Gao et al. (2007) found Early Holocene multichannel system deposits, suggesting that in these smaller river systems the transition from braided to meandering (in their case associated with the Younger Dryas–Holocene climatic transition) took considerable time. For the northern Upper Rhine Graben, this adjustment can be seen primarily as a slow-but-steady complex response to initial climate change during the Late Pleniglacial.

6.2. Nonpreservation of Allerød and/or Younger Dryas terrace levels

Smaller climatic variations during the Late Glacial probably delayed completion of the braided-meandering transition. In the northern Upper Rhine Graben, these were apparently too insignificant to cause geomorphic adjustment. Clearly developed meandering Allerød and/or braided Younger Dryas terrace levels are described for areas farther downstream along the River Rhine (Pons, 1957; Klostermann, 1992; Berendsen et al., 1995; Schirmer, 1995; Törnqvist, 1998; Berendsen and Stouthamer, 2001), in tributaries of the River Rhine (e.g., Andres et al., 2001), and in many other European River systems (e.g., Brown et al., 1994; Huisink, 1997; Starkel, 2002; Gao et al., 2007). In contrast, the northern Upper Rhine Graben is characterised by terraces that developed under transitional conditions from braided to meandering, which are poorly preserved.

One possibility is that these terraces formed as gradual continuations of the preceding system and may never have formed as distinct levels. This implies a lack of geomorphic response of the system, making geomorphological recognition of terraces difficult. An allogenic explanation for the lack of a distinctively formed Younger Dryas terrace may be that changes in vegetation cover in the Rhine drainage basin upstream of the study area were too minimal to change discharge regimes significantly — and thus not changed the mode of ongoing complex response (see Section 6.1). Indeed, the persistent presence of pine forest cover over the southern half of the Rhine catchment throughout the Younger Dryas (Dambeck and Bos, 2002; Bos et al., 2008) is in contrast to the downstream, northerly parts that are almost completely deforested (e.g., Rheinische Schiefergebirge; Litt and Stebich, 1999). Andres et al. (2001) pointed out that only the highest elevated (>400 m above MSL) tributary valleys in the southern half of the Rhine catchment were deforest during the Younger Dryas but that, nevertheless, the upper and lower courses of the valleys showed a transition to a braided system. Apparently, the larger trunk River Rhine is much more buffered to these climatic changes. An additional possible intrinsic explanation is that discharge of the River Rhine in the northern Upper Rhine Graben includes a relative large volume of Alpine meltwater. This could have provided a relatively large base flow, especially during the Late Pleniglacial and successive Bølling and Allerød-interstadials when Alpine glaciers were shrinking in volume (Ivy-Ochs et al., 2004). Along its downstream reaches, where pluvial/nival-dominated tributaries join the River Rhine, the contribution of Alpine discharge to discharge regime is much smaller. The large base flow in the northern Upper Rhine Graben may have facilitated the development of the transitional t7 system, thereby contributing to the slow response of the system. A subsequent decreased base flow as a result of readvancing glaciers in the Alps during the Younger Dryas may provide a supplemental explanation for the absence of a fully braided terrace level. Peak discharge may have been further buffered by the existence of large glacial lakes in the alpine part of the catchment during the Younger Dryas (see Section 6.1).

An alternative explanation is more site-specific: terraces may have formed, but their preservation is poorer than later fluvial deposits, because of the subsequent major change in fluvial style. It is recognised that different types of fluvial style have different preservation potential (e.g., Lewin and Macklin, 2003). The change in fluvial style and channel pattern at the Late Glacial–Holocene transition affected the preservation potential of prior fluvial units. Extensive lateral migration (sufficient space and the low gradient) of the meandering channels in Early Holocene may have counteracted the reduction of floodplain width by incision. In that case, the result is poor preservation of Late Glacial deposits and extensive preservation of Early Holocene deposits.

6.3. Human influence on terrace formation

In the study area, incision continued throughout the Late Glacial and most of the Holocene. Since ~2.7 ka, however, net floodplain lowering stalled. Because incision depends on the sediment load/discharge ratio, two explanations are possible: (i) the river had fully adapted its morphology to Holocene discharge and a longitudinal equilibrium profile was established; or (ii) new allogenic forcing markedly changed the sediment delivery. Although the first explanation may be true, it could not be tested with the data discussed in this study. Arguments for changed sediment supply toward the fluvial system come from archaeological sites and hillslope erosion studies (see below). Such an increase would change the natural sediment-limited system to a system carrying more suspended sediment, forcing deposition. The marked change of the overbank lithofacies toward silty overbank sediments associated to the Late Holocene time frame supports this explanation. The relative coarsening of the overbank deposits is most likely the result of human-induced soil erosion in the loess areas upstream (as found elsewhere, e.g., Knox, 2006). Large-scale deforestation during the Bronze Age, Iron Age, and Roman times in the hinterland (e.g., Friedmann, 2000; Singer, 2004; Lechner, 2005; Bos et al., 2008) resulted in significant changes in sediment availability. Initially, this led to increased deposition on hillslopes as colluvium (e.g., Lang and Hönscheidt, 1999; Lang, 2003) and in smaller catchments upstream as alluvium, which is quantified by several studies (e.g., Lang and Nolte, 1999; Houben, 2003; Mäckel et al., 2003; Rommens et al., 2006; De Moor, 2007; Seidel and Mäckel, 2007). In the following Iron Age, human-induced sedimentation is also recognised in the trunk valley, for example in the southern Upper Rhine Graben (Mäckel and Friedmann, 1999; Mäckel et al., 2003; Seidel and Mäckel, 2007). The time lag between human-induced sedimentation in upstream and downstream areas may be caused by filling of small-scale sediment sinks underway (e.g., Lang et al., 2003). Seemingly, fluvial response to human influence in terms of terrace formation and overbank sedimentation is less complex and at least quicker than the response to Late Glacial climate changes, although in the Late Holocene the system merely changed its suspended load composition. Dambeck and Thiemeyer (2002), Dambeck (2005), and Bos et al. (2008) concluded that increasing human impact during the Subboreal and especially Subatlantic was superimposed on climate change and resulted in a change in the fluvial morphodynamics (flow concentration and locally decreasing meander sinuosity, possibly by increased bank resistance — see further discussion below). We support this viewpoint, but given the considerable change in overbank composition, it seems unlikely that the subtle cooling and increase in precipitation at the Subboreal–Subatlantic transition (e.g., Van Geel et al., 1996) is equally important as the strong and widespread Bronze and Iron Age human land cultivation. Human impact on fluvial systems during the last millennia, including increased overbank sedimentation and decreased channel sinuosity, was also recognised along the Rhine trunk river downstream (Klostermann, 1992; Schirmer, 1995; Schirmer et al., 2005) and in many other central European river systems (e.g., Vandenberghe and De Smedt, 1979; Schirmer, 1983; Szumanski, 1983; Schirmer, 1995; Kalicki and Plit, 2003; Macaire et al., 2006).

6.4. Terrace formation as a result of intrinsic response

During the Late Atlantic (~6.5 ka), the river straightened with much smaller meanders and multiple shallow “secondary” channels, although in adjacent areas fluvial style remained unchanged with ongoing development of large meanders (Dambeck, 2005). Farther downstream, in the Lower Rhine Embayment, meander sinuosity also varies during the Holocene, but the trend and timing is different (e.g., Schirmer, 1995). This implies that intra-Holocene changes in meander sinuosity and/or fluvial style in the northern Upper Rhine Graben did not occur on a drainage basinwide scale, but rather on a local scale. If climate is considered to control these changes in fluvial style (Schirmer, 1995; Dambeck and Thiemeyer, 2002; Schirmer et al., 2005), this can only be a satisfying explanation in combination with complex response, i.e., when different reaches respond differently to the same external factor. Climatic variations during the Holocene are of smaller magnitude compared to the Late Weichselian. Because the River Rhine showed no distinct geomorphic response to Late Glacial climate changes, crossing of geomorphic thresholds by intra-Holocene climate changes are unlikely. Another possible allogenic forcing is human impact, which is for the first time visible in the late Atlantic pollen spectra (e.g., Kalis et al., 2003; Bos et al., 2008). However, Neolithic deforestation and agriculture were still practised on a limited scale, and it was not until the Bronze and Iron Ages (from ~4.0 ka onward) that human cultivation became significant (e.g., Dix et al., 2005; see also Section 6.3). This also excludes human impact as a direct control for Middle Holocene terrace level formation. As an alternative for presumed allogenic forcing, we here explore the possibility that Middle Holocene changes in fluvial style and terrace formation are caused by intrinsic response to local factors. Because of the presence of a local base level, the valley slope of the northern Upper Rhine Graben is the lowest of the Upper Rhine catchment (Dambeck, 2005). When the River Rhine was fully meandering during the Boreal, the gradient decreased even farther. Consequently, the stream power of the River Rhine was low, which hampered further incision and attainment of a gradient-equilibrium. As a result of randomly occurring meander cutoffs, locally gradient (and consequently, stream power) increased again. This resulted in localised incision, each time thereby forming a new “terrace”. For example, in the Gernsheim region the gradient of the River Rhine locally almost doubled (from 0.08 to 0.13 m/km) by meander cutoff at the end of the Early Holocene. Local meander cutoff (an autogenic process) can result in terrace formation and local changes in fluvial style. Subsequently, it requires more energy for an incised and straightened stretch of the river to become meandering again, as more material (lateral higher positioned older terraces) has to be removed. Moreover the incised fluvial system is reworking older floodplains, characterised by fine deposits, which have a higher bank resistance (cf. Schumm, 1977; Brown et al., 1994). At a specific moment in time (after sufficient incision), the very low gradient and stream power may have hindered formation of new large-scale meanders by the incised Middle and Late Holocene system, which resulted in smaller bends instead. We suggest that this specific moment in time may represent a geomorphic threshold in the sense of Schumm (1973). This implies that intrinsic response of the river to site-specific factors (low gradient, incised valley) also provides a satisfactory explanation for the local trend of decreasing meander sinuosity from the late Atlantic onward. Increased suspended sediment delivery during the Late Holocene (an allogenic forcing, see Section 6.3) is likely to have caused a further increase of bank resistance and thereby amplified the existing trend. However, important to realize is that the nature of the response can be spatially limited because of localised site-specific factors. This is supported by locally large meanders continuing to develop in the Middle and Late Holocene, perhaps because local differences in subsurface or gradient.

An argument often used in favour of allogenic-controlled changes in fluvial style at the end of the Atlantic is the apparently simultaneous distinct change in texture of the overbank deposition (Black Clays). For example, Dambeck (2005) suggested that a large-scale allogenic forcing — stabilising climatic conditions — may explain the increase in organic material and the fining of the available sediment. However, the required climate and associated vegetation cover (dense deciduous forest cover) were already present from the early Atlantic onward, and later climate and vegetation changes during the Atlantic were very small (e.g., Kalis et al., 2003; Bos et al., 2008). Furthermore, across the Rhine catchment, formation of the Black Clays did not occur synchronously and, in places, did not occur at all (for example no Black Clays formed in the Lower Rhine Embayment, Klostermann, 1992; Schirmer, 1995; personal observations). Because deposition of the Black Clays is associated to the Middle Holocene terrace level with a characteristic fluvial style, we suggest it to be a result of specific facies conditions. The River Rhine changed locally to a shallow, multichannel plain with frequent flooding over larger areas. The dense Atlantic deciduous forests (e.g., Bos et al., 2008; Fig. 8B) in the low-gradient floodplain hampered flow of water during floods and yielded a large amount of biomass (Dambeck, 2005). We suggest that the Black Clays are not reflecting a new climate- or human-induced source of sediment (compare Dambeck and Thiemeyer, 2002). Instead, it had been present as suspended load in the River Rhine during earlier millennia already, but it was not until the specific facies conditions came into existence that the Black Clay could be deposited.

6.5. Terrace sequence formation in the northern Upper Rhine Graben

Even though the Graben is a subsiding basin, an extensive terrace sequence was formed. In order to form terrace levels, the existence of an incision potential is the first important factor. Here, incision potential exists because the transport-limited glacial River Rhine used the Graben as a sediment sink during the last phase of the Pleniglacial (especially where the Neckar fan enters the northern Upper Rhine Graben; Busschers, 2008, chapter 5). Initial major climatic warming and the formation of the Swiss lakes changed the system to supply limited, transforming the Graben into a sediment source and initiating incision (c.f., Bull, 1991). This incision was uninterrupted despite Late Glacial and intra-Holocene climatic fluctuations (as discussed in Sections 6.1 and 6.2). Ongoing incision promoted the preservation of older deposits (i.e., the process of terrace formation). The specific tectonic situation of the Graben tempered the incision rate and provided enough space for lateral channel migration, resulting in a fairly high preservation potential. Even when climate (and thus discharge and sediment supply) would remain constant, a river system with large preservation potential would be flanked by patches of floodplain of different ages. This implies that the presence of the entire terrace sequence can be explained by three conditions (ongoing incision; an inherited setting that offers preservation potential; sufficient time having passed since incision set on). Because under nonallogenic forcing conditions (e.g., constant climate) incision is still influenced by (local) nonlinear autogenic river behaviour (see Section 1), terraces may sometimes have different morphological characteristics. As a result, the final appearance of the terrace sequence in the northern Upper Rhine Graben is to a large extent the product of intrinsic behaviour and complex response (which are both dependent of site-specific factors). We thus suggest for our study area that climate change was important as an initial trigger, but did not necessarily influence individual terrace formation afterwards. This interpretation differs from that of Schirmer (1995) and Schirmer et al. (2005), who interpret Holocene terraces along the Rhine trunk valley (including the northern Upper Rhine Graben) to be a direct result of repeated changes in climate initially and human impact later in the Holocene. The terrace levels form an event-like sequence, which is easily misinterpreted to be a result of specific events. In practice, terrace level characteristics form the deterministic criteria to group a continuous process into separate levels in time and space, which is inevitable for mapping purposes. We suggest that subsequent coupling terrace levels directly to external factors may result in erroneous interpretations, even if fluvial style/morphology changes between successive levels.

7. Conclusions

In the northern Upper Rhine Graben, five terrace levels were formed that each differ in elevation, morphology, and overbank sediment characteristics. The highest level is a braid plain that was abandoned in the Late Pleniglacial (t6); the next lower level (t7) marks Late Pleniglacial to Late Glacial activity. The intra-Late–Glacial climatic changes did not result in a distinct geomorphic response and/or are poorly preserved. The meandering Holocene Rhine produced multiple palaeomeanders at increasingly lower elevation that group into three terrace levels.

From our study we conclude that

(i) climatic warming during the Late Pleniglacial triggered the onset of incision and the transition toward a meandering system. Strong indications have been found that the Late Glacial terrace level was formed by a slightly curving (low sinuosity) multichannel transitional system, but no indication of a Younger Dryas braid plain. At the onset of the Holocene, the system became fully meandering. Locally, two meandering streams — inherited from the multichannel transitional system — were active until the middle Boreal. This indicates that climate change was the most important factor controlling fluvial development during the Late Weichselian, although response was complex and slow.

(ii) Terrace formation continued during the Holocene, forming three successive terrace levels. Intrinsic response to the strongly decreased gradient and local tectonic background situation provides an alternative explanation for Middle Holocene terrace formation. Because intra-Holocene climate changes are small, autogenic fluvial evolution became dominant.

(iii) During the Subatlantic, incision of the river system ceased and overbank sediments became siltier. This is explained as a result of human impact in the contributing drainage basin.

(iv) The terrace sequence is explained by a complex interplay of both allogenic and autogenic controlling factors and site-specific characteristics such as preservation potential and tectonic background situation. Interpreting and correlating terraces over larger distances as a result of a single allogenic factor must be done with care.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.geomorph.2008.07.021.

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