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Long-term deposition of fine sediments in Vienna's Danube floodplain before and after channelization

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ABSTRACT

Investigations on the deposition processes in current alluvial landscapes are commonly affected by multiple forms of human interventions such as local hydraulic measures and reduced sediment transport caused by upstream reservoirs. The present study focuses on the sedimentation rates and floodplain accretion over 500 yr prior to major river engineering measures. A comprehensive borehole database and research on long-term fluvial dynamics in Vienna's Danube floodplain enabled us to correlate the thickness of fine sediment layers (silt and fine sand) with the morphological age of individual sites before the great regulation program 1870–1875. Five years after the onset of the sedimentation process on top of a gravel bar, the median deposition rate amounted to 18.60 cm yr^{-1}. In the following decades the rate significantly decreased and leveled off after 300 yr with median annual rates between 0.15 and 0.10 cm. Five hundred years after the deposition process had started, the fine sediment layer reached a thickness of 2.64 m, of which half already had been deposited within the first 10 yr.
Stabilization of riverbanks in 1870–1875 significantly boosted the long-term annual sedimentation rates by at least 23–41% (depending on the calculation method), although the volume of the suspended load decreased by 18–45% since around 1880. Assuming equal loads today would hypothetically yield a greater increase of the rates. As opposed to the historical situation featuring intensive lateral erosion, natural levee formation along the protected riverbanks has become a common phenomenon today. The thickness of fine sediment deposition and therefore the long-term "climax level" of the floodplain depends on numerous controlling factors including hydrological regime, sediment volume/size, stream power and riparian vegetation. Human interventions, i.e., bank stabilization, also alter the basic conditions for floodplain accretion, leading to greater sedimentation rates and higher floodplain levels.

1. Introduction

Sediment deposition in floodplains has attracted considerable scientific awareness in recent decades from various points of views. Hydrologically, ongoing overbank sedimentation has progressively reduced the transport and flood retention capacity of the high-water bed along channelized river reaches. This, in turn, has amplified the flood risk in downstream river sections (Middelkoop and Asselman, 1998; Baptist et al., 2004; Hudson et al., 2008; Kiss et al., 2011). From the ecological perspective, regulated river landscapes experience multiple forms of ecological degradation. Because of riverbank stabilization, the absence of lateral channel migration reduces the potential for morphological floodplain "rejuvenation" (Tockner and Stanford, 2002; Díaz-Redondo et al., 2018; Entwistle et al., 2019). Over the long term, the consequences are gradual increase in the elevation of the ground level and increasing depths of the groundwater table in relation to the terrain surface. The results are drier vegetation stands and terrestrialization of floodplain water bodies (Hohensinner et al., 2008; Fujimura et al., 2008; Reckendorfer et al., 2013). Overbank sedimentation also affects the sediment budget of downstream river reaches and the trajectory of delta evolution, as for example addressed for the Danube delta at the Black Sea and the Po delta at the Adriatic Sea (Giosan et al., 2012; McCarney-Castle et al., 2012; Parinello et al., 2021).

Accordingly, sediment deposition in floodplains has become a vital issue in restoration ecology and flood prevention strategies (Hohensinner et al., 2005; Stammel et al., 2011; Geerling et al., 2013; Eberstaller et al., 2018). Recently, the importance of sedimentation processes for humans living in or close to alluvial landscapes has also been

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highlighted in environmental history (Lewin, 2012; Parrinello and Kondolf, 2021; Swayamprakash, 2021). River channelization in most cases also aimed at reclaiming new land to be used for agriculture, forestry or as settlement areas (Haidvogl and Tasser, 2019; Hohensinner et al., 2021a). Studying past human practices of land reclamation in fluvial landscapes informs both about associated modifications of river-floodplain systems and the potential dangers of such practices over the long term (Dotterweich, 2008; Lewin, 2012; Brown et al., 2018).

Sedimentation processes and floodplain formation depend on numerous factors. At the catchment scale, fundamental factors include geological conditions, soil development, vegetation cover and the distance of a specific river reach to upstream sediment sources (Nanson and Croke, 1992; Knighton, 1998; Thonon et al., 2007). At the reach scale, the discharge regime and morphological channel pattern, the available sediment (volume and size), the stream power of the river, and the width of the floodplain play important roles. Locally, inside a floodplain, the distance to active river channels, the floodplain relief, the riparian vegetation, and the type of flooding process are also crucial (Simm and Walling, 1998; Middelkoop and Asselman, 1998; Middelkoop, 2005). Different forms of floodplain inundation such as active overflow (overbank flooding), backwater flooding in one-side connected water bodies or seepage inundation in isolated water bodies are associated with specific sediment fractions and deposition processes (Nanson and Croke, 1992; Asselman and Middelkoop, 1995; Bridge, 2003). Considering the numerous factors contributing to sediment deposition and fluvial erosion, Nanson and Croke (1992) distinguished several types of floodplain formation. Accordingly, “high-energy non-cohesive floodplains” reflect disequilibrium landforms primarily shaped by erosion. “Medium-energy non-cohesive floodplains” are considered to remain in dynamic equilibrium between erosion and deposition over the long term provided that the external hydromorphological framework conditions do not change. Finally, “low-energy cohesive floodplains” are associated with comparably stable single-channel or anastomosing low-energy rivers. In such environments, vertical accretion of fine-grained sediments is the dominant fluvial process (Nanson and Croke, 1992).

Because most larger rivers in the industrialized world have been channelized within the last 150 yr, the problem of progressive sediment aggradation and terrestrialization increasingly threatens riverine biodiversity (Tockner and Stanford, 2002; Booth et al., 2009; Wohl, 2021). One research question raised in this context refers to the “natural” sedimentation processes prior to channelization (Middelkoop, 1997; Lecce and Pavlowsky, 2001; Knox, 2006; Kiss et al., 2011; Klasz et al., 2014). Advances in remote sensing (i.e., Airborne Laser Scanning (ALS)) facilitated the accurate measurement of terrain heights in wooded floodplain areas (Hohenthal et al., 2011). Such methods have therefore been used to estimate sediment rates (i.e., the thickness of newly aggraded sediment layers) after flood events (Notebaert et al., 2009; Stammel et al., 2011; Klasz et al., 2014). Based on several studies (e.g., Simm, 1995; Terry et al., 2002), Middelkoop (2005) reported typical contemporary sedimentation rates between 0.5 and 20 mm yr^{-1}, with up to several-centimeter-thick layers of freshly deposited material after severe floods. In comparison, according to Klasz et al. (2014), average sedimentation rates in the “National Park Donau-Auen” downstream of Vienna amounted annually to approximately 11 mm close to the Danube River over the past 120 yr (when the river was channelized). In floodplain areas remote from the Danube River, annual rates decreased to a mean of 0.3 mm.

Focusing on deposition by flood events in recent decades, however, integrates multiple forms of human interventions involving local river engineering measures and catchment-wide impairments, most of all reduced sediment transport caused by upstream dams and reservoirs (Hesselink, 2002; Benedetti et al., 2007; Hobo et al., 2010; Brown et al., 2018). Studies of the sedimentation rates over several centuries or even millennia are typically based on the analysis of borehole data, making it time-consuming and costly for larger river-landscapes (Middelkoop, 2005; Tanabe et al., 2006; Aalto et al., 2008; Jana and Paul, 2020). Middelkoop (1997) estimated deposition rates in embanked floodplains along the Lower Rhine River based on historical river maps. Accordingly, during the last 150 yr the average rate was approximately 15 mm yr^{-1}. In 350- to 400-yr-old floodplain sections the average deposition rate amounted to only ~5 mm yr^{-1}. Such long-term studies based on historical sources are very rare, leaving considerable gaps in our knowledge about sedimentation rates prior to systematic river regulation. The present study on the floodplain evolution in today's Vienna aims at partially filling this gap by answering the following research questions:

(1) What annual sedimentation rates were typical for the Danube River prior to channelization?
(2) How have sedimentation processes changed as a consequence of channelization?

With respect to the first research question, we rely on existing elaborate basic datasets: 10 GIS-based reconstructions of Vienna's Danube

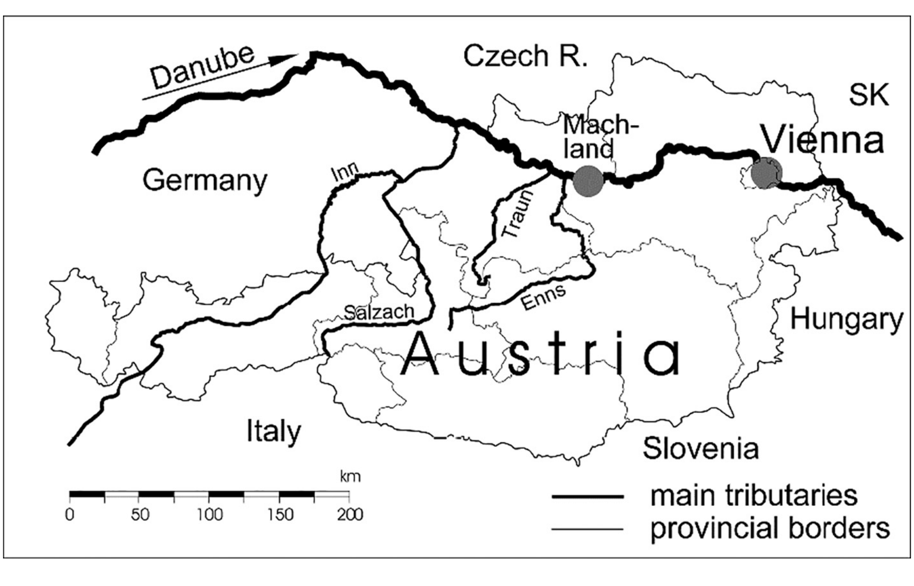

Fig. 1. Location of the study area in Vienna, Austria, with the upstream Machland floodplain and the main Danube tributaries (SK: Slovak Republic).

River landscape between 1529 and 1825, and a large database comprising thousands of geological borehole data (see Section 3). Because Vienna's former floodplain was overbuilt soon after the major Danube regulation program in 1870–1875, the second question will be comparatively discussed combining our results and available studies on the sedimentation processes in the "National Park Donau-Auen" downstream of Vienna.

Our results help to better understand the role of sediment deposition in river sections that have not yet been channelized and to more extensively assess sedimentation processes along regulated rivers in recent decades. This provides a sound data basis for discussing restoration measures in human-modified river-floodplain systems that are currently threatened by progressive aggradation and terrestrialization.

2. Study site

The study site refers to the former Danube floodplain within the city limits of today's Vienna between river-km 1937 and 1919 (Fig. 2). Until the great Danube regulation program in 1870–1875, the currently 18-km-long river reach was paralleled by a large, up to 10-km-wide floodplain. In the present study we focus on a 96.4-km^2-large and maximum 8.6-km-wide part of the floodplain that was hydromorphologically active over the last 1000 yr. While the upper end of that reach is located in the narrower "Wiener Pforte" gap, most of the study site is situated in the large "Marchfeld" plain. Since 1996, the lower end of the study site is part of the "National Park Donau-Auen".

Similar to the Upper Rhine River, the Danube River in Vienna originally showed a complex channel network with several small and larger vegetated islands (Fig. 3). The latter showed similar terrain heights to the adjacent floodplain and, consequently, divided the flow up to bankfull. One or two main navigable main arms structured by large gravel bars were typical. Individual channels showed independent patterns comprising meander, braids or relatively straight courses (Hohensinner et al., 2008). According to the river classification scheme of Nanson and Knighton (1996), in its pre-channelization-state the studied Danube section can be designated as "gravel-dominated, laterally active anabranching river". Based on Nanson and Croke (1992), Vienna's Danube floodplain is best described as a "medium-energy non-cohesive floodplain". Sediments in such floodplains are characterized by gravel, sand and silt. Highly variable flow regimes are only one factor behind the development of such river-floodplain systems. Other common causes and characteristics include additionally intensified

Fig. 2. Study area: Vienna's former Danube floodplain with the Danube River as it was straightened during the great Danube regulation program 1870–1875. Today, most of the past floodplain is urbanized (yellow line: limits of the study area; Donaukanal = Viennese Side Arm until around 1700; orthophoto: Vienna municipal department MA 41 - Stadtvermessung).

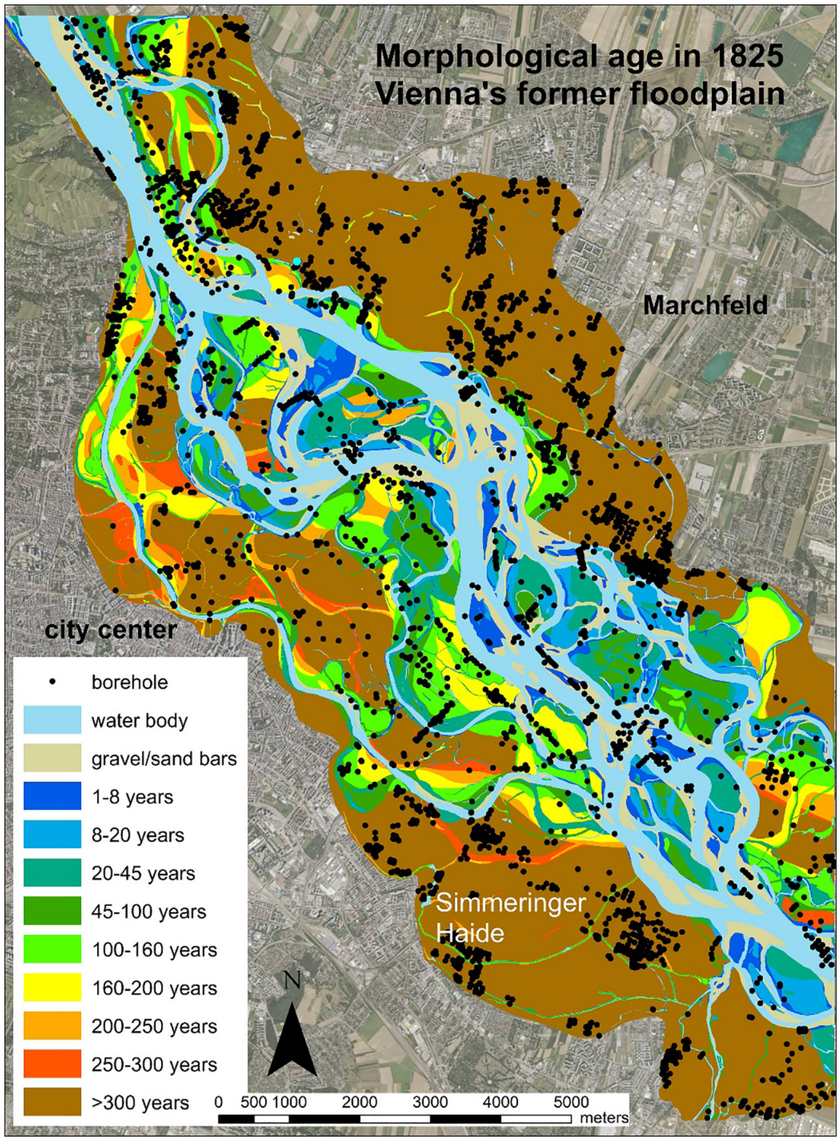

Fig. 3. Morphological site age of Vienna's floodplain in 1825 and locations of boreholes (orthophoto: Vienna municipal department MA 41 - Stadtvermessung).

floods or backwater effects because of ice jams in winter, large woody debris and/or downstream channel constrictions, as well as high loads of coarse bed material, possibly because of larger upstream tributaries (Hohensinner et al., 2008).

Present bankfull discharge amounts to 4800 and 5000 $m^3\ s^{-1}$ (= $Q_{1.3}$-$Q_{1.4}$), which is less than the 1.5-yr return interval discharge usually assumed being equivalent to the bankfull discharge ($Q_{1.5}$ = 5050 $m^3\ s^{-1}$ calculated from annual maximum flood (AMF) approach; based on Klasz, 2020). A hydrodynamic simulation of the Viennese Danube reach in its state in the year 1817 yielded a bankfull discharge between 4700 and 4800 $m^3\ s^{-1}$ (=current $Q_{1.2}$-$Q_{1.3}$ using AMF approach; Hohensinner and Trautwein, 2013). Schmautz et al. (2000) estimated bedload fluxes from upstream Alpine tributaries prior to channelization around 1850 to be approximately 500,000 $m^3\ yr^{-1}$. From that value, because of gradual downstream fining (abrasion, hydraulic sorting), bedload volume progressively dropped while being transported downstream to Vienna. Today, the hypothetical mean annual bedload transport capacity of the channelized Viennese Danube amounts to approximately 340,000 $m^3\ yr^{-1}$ (BMNT, 2018). Though no solid data on former sediment size distributions exist, a somewhat smaller median grain size of the bedload material transported by the pre-channelization Danube is assumed.

Between 1878 and 1884, suspended load (silt) made up between 5.4 and 6.5 million t yr^{-1} in Vienna (Penck, 1891). The numerous reservoirs and dams built upstream of Vienna since the 1950s largely prevented bedload transport and reduced the suspended load to an average 3.9 million t yr^{-1} between 1957 and 1965 (Gruber, 1969). For the period 1982–1995, Nachtnebel et al. (1998) reported annual suspended loads between 3.0 and 3.5 million t. Newer daily records from the gauge Hainburg-Straßenbrücke approximately 40 km downstream of Vienna yield a mean volume of suspended load of annually 4.87 million t between 2008 and 2017 (4.16 million t without the 300-yr flood in 2013; unpublished data, https://ehyd.gv.at/).

Historical records reveal information on the vertical structuring of Danube floodplains in Austria. Accordingly, prior to channelization, the mean limit between unvegetated gravel/sand areas and sites colonized by perennial vegetation referred to summer mean water level (SMW = 1.7 m above mean annual low water and 0.3–0.4 m above mean annual water level, respectively; Hohensinner et al., 2004, 2008). Vegetated sites above SMW were primarily formed by silt (deposited suspended load) and fine sand (Fig. 4).

The first hydraulic constructions are documented from the 1540s onwards (Thiel, 1904). They were carried out at the upper end of the study area close to the bifurcation of the "Viennese Side Arm" (today's

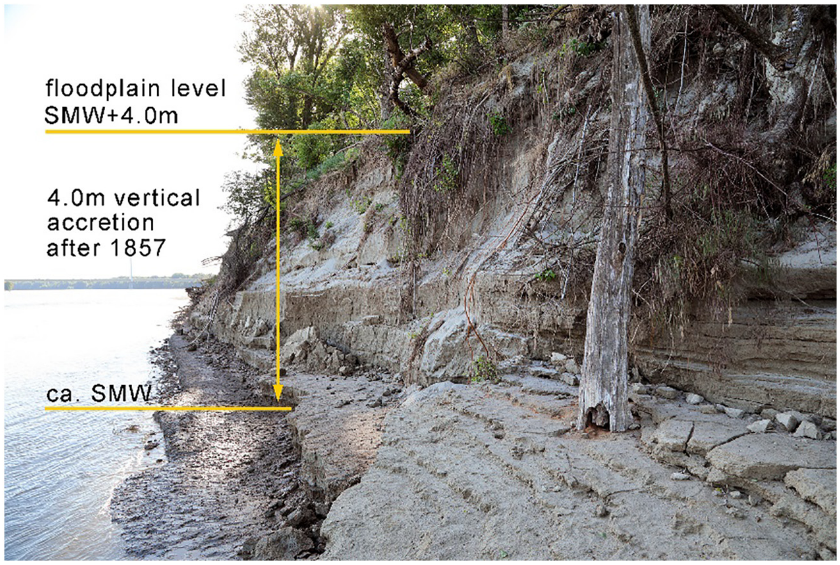

Fig. 4. Bank erosion following restoration measures in the "National Park Donau-Auen" reveals the typical layer sequence of alluvial soils close to Vienna: Sandy gravel form the basis that reaches up to approximately summer mean water level (SMW). In the shown location the former gravel bar has been topped by a 4-m-high layer of fine sediments (silt, fine sand) after 1857 (photo: Christian Baumgartner 2017, National Park Donau-Auen GmbH).

"Donaukanal") and the main channel of the Danube. Following a phase of amplified flood activity possibly caused by climate change in the Alpine Danube catchment (so-called "Grindelwald fluctuation"), hydraulic works were intensified from the 1560s onwards (Hohensinner et al., 2013a). Since then, over centuries, the main hydraulic constructions focused on improving the flow conditions for navigation in the "Viennese Side Arm". The next phase of adverse climate conditions starting around 1768 gave rise to the first systematically planned flood protection dikes. However, soon after their construction they were destroyed or overflowed by severe floods and remained largely useless. Until around 1825, except for the "Viennese Side Arm", hydraulic constructions were implemented only locally. In the following decades until 1870, most of the riverbanks of the major Danube channels had been gradually protected and stabilized (Hohensinner et al., 2013a). Finally, during the great Danube regulation program 1870–1875 a new main channel was excavated, several side arms were filled up and a new flood protection scheme was implemented. Since then, Vienna's former floodplain is largely separated from the Danube River (Hohensinner and Schmid, 2016).

3. Material and methods

The present study is based on two major datasets:

(a) In the framework of the research project "Environmental History of the Viennese Danube" (ENVIEDAN) and of other studies, Vienna's fluvial landscape was reconstructed based on more than one thousand historical sources, archaeological findings and geological information (Winiwarter et al., 2013). The reconstruction was conducted using the "regressive-iterative GIS method" developed by Hohensinner et al. (2013b), yielding 14 GIS maps of the river landscape at 14 points in time between 1529 and 2010 (Lager, 2012). Ten of the GIS maps that were selected for the present study show the Viennese Danube prior to major hydraulic constructions (1529, 1570, 1632, 1663, 1704, 1726, 1780, 1805, 1817 and 1825; see Digital Atlas of the City of Vienna: https://www.wien.gv.at/kulturportal/public/grafik.aspx?bookmark=S8OZRTIo2kU-aMtRGF6AuRhwpUhiWNwI-b). In another project ("Enough wood for city and river? Vienna's wood resources in dynamic Danube floodplains") the ten GIS maps were combined to calculate the potential minimum and maximum age of the floodplain terrain in 1825 (Hohensinner et al., 2016). Based on these values the potential mean age of each site was derived. The resulting GIS dataset comprises the morphological age of 9735 sites referring to the time before major hydraulic constructions (Fig. 3). In order to consider deposition of material during floods after 1825, additional GIS reconstructions (1849 and 1875) were used. Moreover, each floodplain site was checked regarding whether it might have been flooded between 1825 and 1875 when the new flood protection system was finally realized. For that, a database on historical hydraulic measures at the Viennese Danube and a second one on historical floods were consulted (Hohensinner and Hahmann, 2020; Hohensinner, 2020). The final dataset comprises the morphological age of the individual floodplain sites corrected for the last possible inundation and, consequently, deposition of sediments.

(b) In the course of a long-term hydro-geological research project conducted by "Wiener Gewässer Management Gesellschaft mbH" by order of Vienna's municipal department "MA 45 – Water Management", the data of 65,000 boreholes were analyzed and used to generate a hydrological 3D layer model of Vienna's near surface geological basement (Pfleiderer et al., 2019; Grupe and Payer, 2020, in prep.; based on raw data provided by MA 29 – Bridge Construction and Foundation Engineering). In that data base, 17,600 of the boreholes are located in Vienna's alluvial zone (Grupe et al., 2021). Of these, 7200 sites that do not feature human landfills were selected. At landfill sites it remains unclear whether the surface of the former floodplain terrain was artificially lowered before the landfill. In a next step, all sites located outside of the study area – most of them north of the study area in the "Marchfeld" plain – were discarded. Finally, 3120 sites (borehole data) with mean morphological ages between 7 and 463 yr were used for the present study (Figs. 3 and 5). To analyse the historical deposition rates of fine sediments that aggraded on the basal (sandy) gravel-bed material, the thickness of silt, fine sand and humus layers was identified for each site.

The last work step comprised correlating the morphological age and the thickness of the fine sediment layer at the selected sites (Section 4.3). This yielded calculated mean annual sedimentation rates that were dependent on the respective morphological age.

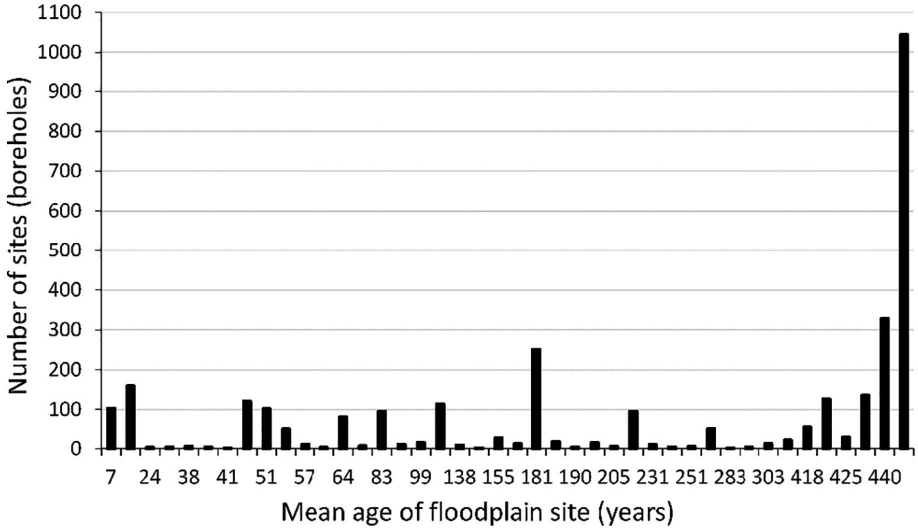

Fig. 5. Distribution of the morphological site age of the boreholes used for this study.

4. Results

4.1. Morphological age of floodplain terrain

An overview of the distribution of the age of the floodplain terrain in the study site prior to major hydraulic measures in 1825 is provided in Fig. 3. It reveals the part of the contemporary floodplain that was subject to erosion and deposition processes within the last 500 yr, or 300 yr related to 1825 (all except brown areas in Fig. 3). The northern part of the study area in the "Marchfeld" plain as well as the southeastern part ("Simmeringer Haide") proved to be the most stable over the long term. The situation is similar for several islands close to the historical city center. After 1825, until the final damming of the Viennese floodplain in 1875, local flood prevention measures prohibited further inundation and sediment deposition. Taking these local measures into account yields slightly different site ages than reflected by Fig. 3. The temporal distribution of the 3120 sites (boreholes) analyzed is presented in Fig. 5. In total, 42 different mean site ages can be distinguished, ranging between 7 and 463 yr. Most of the boreholes are located in very old parts of the floodplain (compare Fig. 3).

4.2. Borehole data

The distribution of the thickness of the fine sediment layers consisting of humus, silt and fine sand for each age class is shown in Fig. 6 (omitting age classes < 5 values). The median thickness averaged

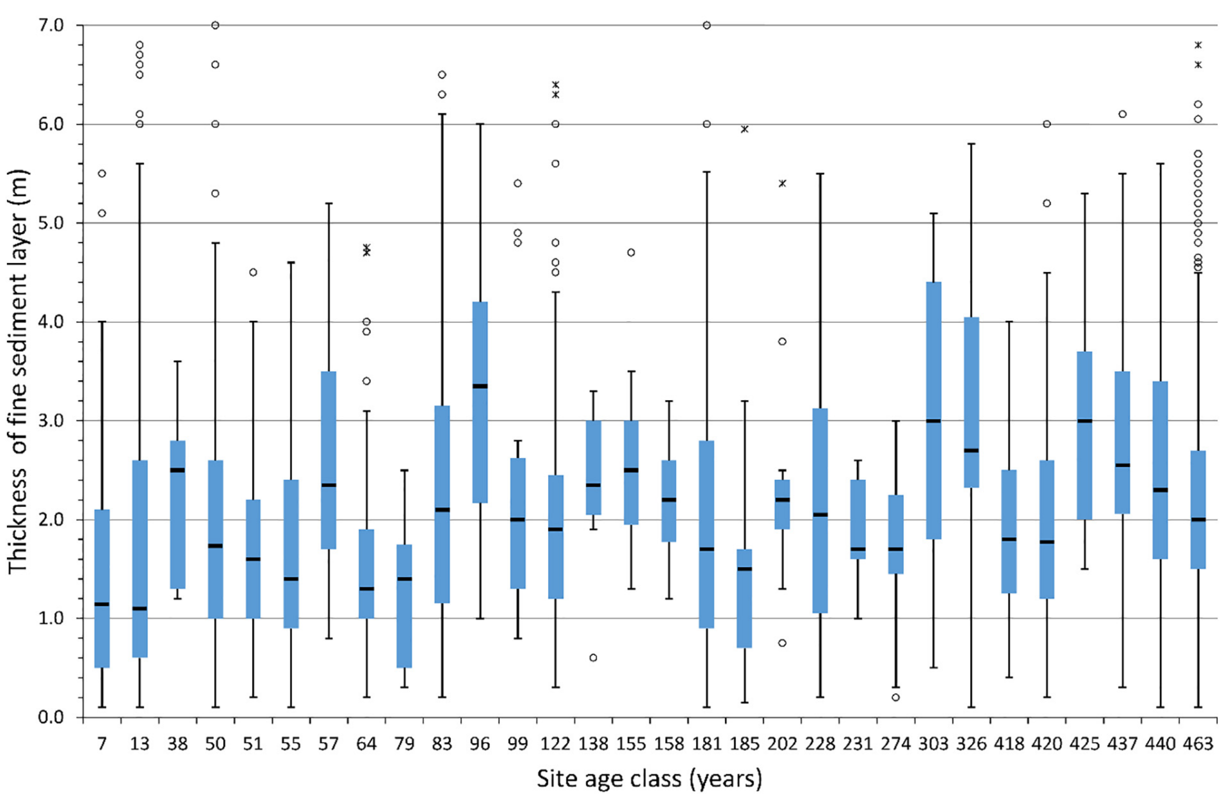

Fig. 6. Distribution of the thickness of the fine sediment layer (m) per site age class (years; classes with values <5 are not shown).

over all boreholes is 2.0 m, and minimum and maximum values are 0.1 and 8.5 m, respectively. As assumed, the youngest sites show significantly thinner fine sediment layers than older ones. Nonetheless, layer thickness is highly variable between ages of 38 and 463 yr. Thus, Pearson's correlation ($r = 0.141$; $p < 0.0005$) indicates only a very weak relationship.

4.3. Sedimentation rates prior to channelization

A simple calculation of annual sedimentation rates by dividing layer thickness through site age (years) generally yields rates that are too high. This type of calculation merely provides a mean rate interpolated over the entire respective time span/site age (see Table 2, column (a)). To solve this issue, we derived a potential function based on the median values shown in Fig. 6 for layer thickness per age class (see black function curve in Fig. 7). This function was used to calculate median layer thicknesses for characteristic site ages with 5, 10, 25 yr, etc. (Table 1, middle column). Because the derived function failed to figure the youngest age classes, the thickness values for the age classes 5 and 10 yr were calculated using a smaller dataset including solely the younger age classes (Table 1, right column). The two function curves based on the original median values and on the corrected values (synthesized function) are shown in Fig. 7. The next step involved calculating the differences (=increase) in layer thickness between successive age classes. Finally, the derived differences were standardized for years, resulting in annual sedimentation rates (cm yr^{-1}) between individual points in time. Many boreholes are spatially clustered within areas characterized by the same site age, which could statistically bias the results (Fig. 3). Calculating the sedimentation rates based on the median thickness of each site age class (Fig. 6) helps to mitigate this statistical effect.

The results presented in Table 2 illustrate the decreasing sedimentation rates over time. Five years after the deposition process on a particular gravel bar has started, annual sedimentation rate would amount to 20.58 cm based on a simple temporal interpolation (see (a) in Table 2). The incremental cumulative approach (considering the layer thickness changes between individual points in time), however, yields 18.60 cm yr^{-1}, an approximately 10% lower value. Five hundred years after the deposition process had started on gravel, the interpolation yields only 0.40 cm yr^{-1}. According to the cumulative approach, in comparison, annual sedimentation even dropped to 0.10 cm yr^{-1} after 500 yr (see (c) in Table 2). The difference in sedimentation rates using different calculation methods is displayed in Fig. 8. A simple linear interpolation by dividing the layer thickness through the number of years would smooth the curve of annual sedimentation rates (black dotted curve based on the blue median values). In contrast, incremental cumulative calculation of sedimentation rates between two subsequent points in time yields a more pronounced curve as a function of time

Table 1
Median thickness of fine sediment layers calculated for characteristic site ages: (a) based on measured values using the potential function shown in Fig. 7; and (b) related values corrected for the younger age classes (right column).

Site age	Median thickness of layer (m)	
(yrs)	(a) Measured	(b) Corrected
0	0.00	0.00
5	1.27	0.93
10	1.42	1.31
25	1.64	1.64
50	1.83	1.83
100	2.04	2.04
150	2.18	2.18
200	2.28	2.28
300	2.43	2.43
400	2.54	2.54
500	2.64	2.64

Table 2
Sedimentation rates (cm yr^{-1}) and median thickness of fine sediment layer calculated for characteristic points in time: (a) interpolated by dividing layer thickness through the site age based on all borehole data; (b) interpolation using median sedimentation rates per site age class (compare (b) in Fig. 8); and (c) related values calculated incrementally based on the differences in layer thickness between individual points in time and corrected for the younger age classes (cumulative approach; compare (c) in Fig. 8). 25%- and 75%-values refer to the lower and upper quartile, respectively.

Years after begin of sedimentation process	Sedimentation rates (cm yr^{-1})					Median thickness of layer (m)
	(a) Interpolation	(b) Interpolation	(c) Incremental cumulative			
	All boreholes	Median values	25%	**Median (50%)**	75%	
5	20.58	20.40	8.60	**18.60**	30.60	0.93
10	11.39	11.80	6.60	**7.51**	7.80	1.31
25	5.21	5.73	1.99	**2.21**	2.28	1.64
50	2.89	3.32	0.69	**0.76**	0.83	1.83
100	1.60	1.92	0.40	**0.43**	0.45	2.04
150	1.13	1.39	0.26	**0.27**	0.28	2.18
200	0.88	1.11	0.19	**0.20**	0.21	2.28
300	0.63	0.81	0.15	**0.15**	0.15	2.43
400	0.49	0.64	0.11	**0.11**	0.11	2.54
500	0.40	0.54	0.09	**0.10**	0.10	2.64

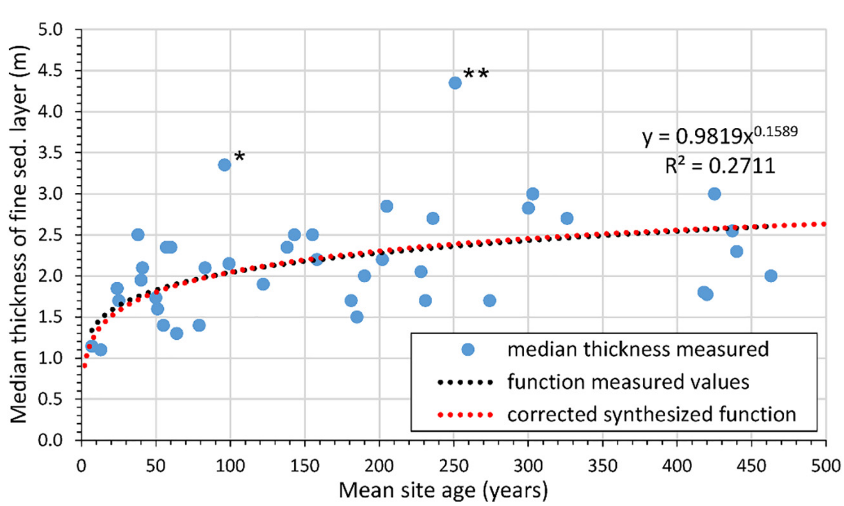

Fig. 7. Potential functions based on the original median values for layer thickness (black dotted curve based on the blue median values) and on the corrected values (synthesized function; red dotted curve). The indicated function refers to the original values (* median refers to several boreholes directly behind a flood protection dike constructed in the 1770s–1780s; ** median is based on only four borehole data).

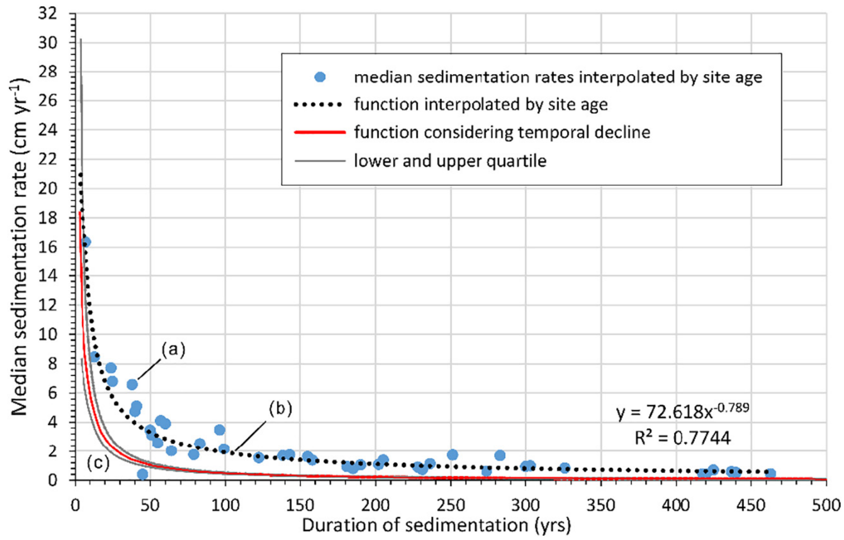

Fig. 8. Comparison between different methods for estimating annual sedimentation rates (cm yr^{-1}): (a) blue dots: median rates calculated for each site age class; (b) black-dotted curve based on median rates: sedimentation function derived by dividing the layer thickness by site age (interpolation over time, compare (b) in Table 2); and (c) red curve: sedimentation function considering temporal decline of deposition (compare (c) in Table 2). The indicated function refers to the black-dotted curve.

(i.e., duration of the sedimentation process; see red curve). The lower and upper quartiles indicated in Fig. 8 and Table 2 show that the potential range between the quartiles is much greater in shorter time spans. The longer the observed time period, the more the variation levels out.

5. Discussion

5.1. What annual sedimentation rates were typical for the Danube River prior to channelization?

Both the cumulative and interpolated sedimentation rates show a distinct decrease, with higher rates at the beginning of the deposition process on top of a gravel bar and lower rates the longer a site exists. The two calculation methods highlight a basic problem in the interpretation of deposition processes: The rates appear to accelerate the shorter the time periods that are considered. This effect is a consequence of "time-averaging" of geomorphological processes across different time scales (Gardner et al., 1987; Benedetti et al., 2007). The incremental cumulative approach used in the present study thus provides more accurate results than an interpolation by dividing layer thickness by site age (deposition period). Because many other studies present interpolated sedimentation rates, we also include them here for comparison.

Starting with 18.6 cm yr^{-1} in the fifth year, the rate dropped to 2.2 yr^{-1} 25 yr after the onset of sedimentation (according to the interpolation method the rates dropped from 20.40–20.58 to 5.2–5.7 cm yr^{-1}; Table 2 and Fig. 8). This means that the fine sediment deposits accumulated to a median height of 93 cm within the first five years. Another five years later (i.e., after 10 yr), the respective fine sediment layer had already attained a median thickness of 131 cm (Fig. 9). Similarly, Wolman and Leopold (1957) also stated a rapid increase in floodplain elevation along US streams in the first 10 yr. Applying a quantitative model of floodplain growth, Moody and Troutman (2000) presented similar results for several US rivers.

Strictly speaking, the observed decrease of deposition rates is not a function of time. Rather, it depends on the progressively aggrading sediments that result in increasing layer thickness and, consequently, higher elevation of the floodplain terrain. The higher the terrain surface

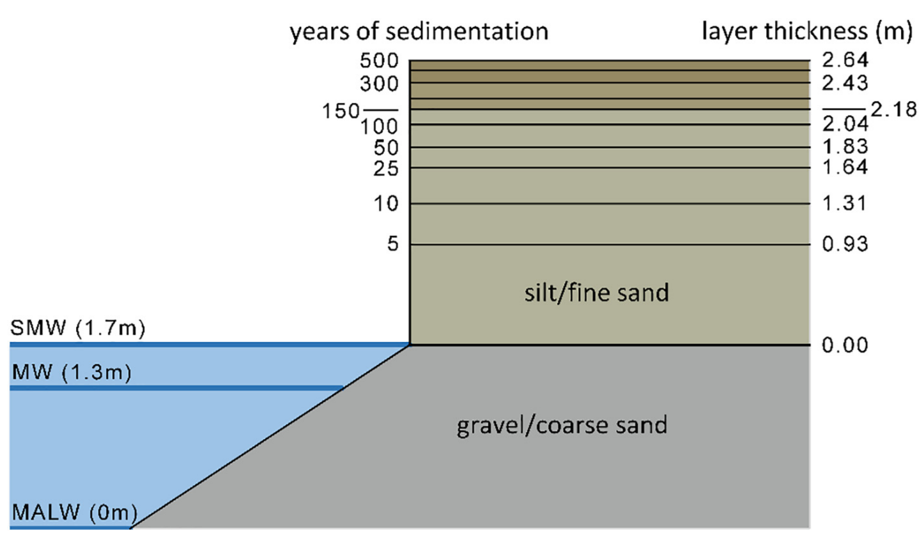

Fig. 9. Schematic stratigraphic cross section of Vienna's floodplain prior to channelization based on calculated median sedimentation rates (water levels: MALW = mean annual low water, MW = mean water, SMW = summer mean water).

level is, the less frequently it will be inundated during floods, and the potential for sediment deposition gradually diminishes. Finally, after e.g., 500 yr, only extreme floods may reach the floodplain level. Thus, sedimentation processes, i.e., floodplain accretion, and flood regime (frequency, magnitude, duration) are closely interrelated. The general result is that floodplain levels are adapted to the river/reach-specific flow conditions (Wolman and Leopold, 1957; Moody and Troutman, 2000). Besides the hydrological regime, additional significant factors influencing floodplain evolution are the available sediment supply (volume and size), the type of riparian vegetation as well as the energy environment (stream power Ω as a function of discharge Q and channel slope S) manifested, for example, as lateral erosion (Nanson and Croke, 1992; Knighton, 1998; Thonon et al., 2007).

Because of different forms of channel adjustments and floodplain inundation (active overflow, backwater flooding, seepage inundation), alluvial landscapes feature a great variety of depositional processes. Besides overbank vertical-deposition, also lateral point-bar accretion, braid-channel accretion in wider channel profiles and abandoned channel accretion are typical – each associated with specific sediment fractions (Nanson and Croke, 1992; Asselman and Middelkoop, 1995; Bridge, 2003). Accordingly, sedimentation rates and the thickness of fine sediment layers are strongly linked to the specific conditions of a deposition site within the fluvial system (Wolman and Leopold, 1957; Hudson et al., 2008). Vienna's anabranching Danube landscape also originated from several sedimentation processes. Within-channel deposits up to the bankfull stage, such as lateral point-bar and mid-

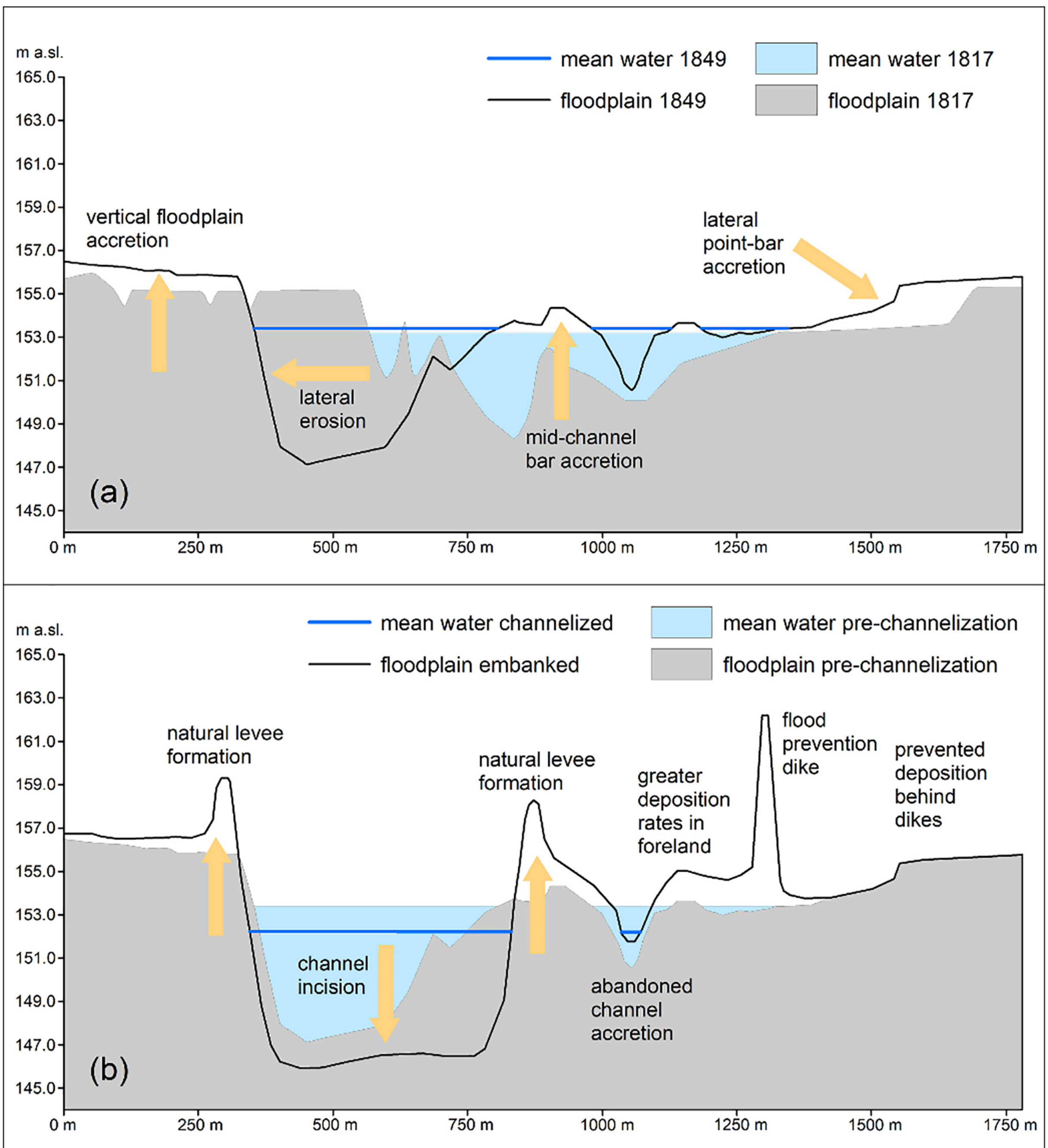

Fig. 10. Channel dynamics and sedimentation processes under natural conditions (a), and after channelization and embankment (b). Prior to channelization, lateral erosion was counteracted by lateral point-bar and mid-channel bar accretion. After channelization, deposition processes prevail (schematic illustration based on historical surveys).

channel-bar accretion on top of gravel/sand bars, were typical phenomena (Fig. 10a). The latter process led to the formation of new vegetated islands in larger river arms. Outside the active river channels, areal overbank deposition and infilling of abandoned channels predominated. The data elaborated so far, however, do not allow a concrete quantification of the diverse deposition volumes. Different types of sediments, in turn, are crucial for the emergence of adapted plant communities. For example, riparian vegetation that colonizes lateral point-bar accretions shows trajectories diverging from those in abandoned, terrestrialized backwaters (Müller, 1995; Drescher and Egger, 2013).

The different factors controlling local deposition processes make it difficult to compare sediment rates between different river systems. Annual rates between 2.0 and 0.05 cm reported by Middelkoop (2005) based on various studies (e.g., Simm, 1995; Terry et al., 2002) were typical for the unregulated Danube River at sites with morphological ages between 25 and 500 yr (Table 2). Reported maximum sedimentation rates of several centimeters during severe floods would refer to approximately 10-yr-old floodplain sites in Vienna before major regulation works. Comparing long-term deposition rates in the area of the Lower Rhine River with those of the present study reveals additional insights. During the last 150 yr (prior to 1997) the average rates in the Rhine floodplains amounted to approximately 1.5 cm yr^{-1} (Middelkoop, 1997). At the historical Danube River, the same rate is evident for time periods between 100 and 150 yr after the onset of sedimentation based on analogously averaged (interpolated) values (compare (a) and (b) in Table 2). Considering the cumulative approach, however, after 100 to 150 yr, the median deposition rates fell to only 0.43 to 0.27 cm yr^{-1} (see (c) in Table 2). A similar situation is evident for 350- to 400-yr-old sections of that Rhine floodplain, where averaged (interpolated) rates were 0.5 cm yr^{-1} (Middelkoop, 1997). In the present study that same rate is evident for 400- to 500-yr-old sections of the Danube floodplain, depending on the calculation method (see (a) and (b) in Table 2). In comparing the data from these Rhine and Danube floodplains, note that the former was already significantly human altered by flood prevention dikes more than 150 yr ago. Our data from Vienna, in contrast, refer to floodplain sections that were largely untouched by large-scale hydraulic interventions (except for local embankments, smaller or already breached dikes).

Despite such apparent similarities, the stratigraphic conditions discussed here strongly depend on the individual river (reach) type and flow regime. Moreover, they are based on a stationary perspective, i.e., stable hydromorphological conditions as long as the external controlling factors do not change. Alluvial landscapes, however, in most cases integrate several sections that emerged under varying climatic and hydrological conditions (Owens and Walling, 2002; Asselman et al., 2003; Macklin et al., 2010). Prior to channelization, Vienna's floodplain, for example, featured sections from at least five periods characterized by differing flood regimes and associated fluvial dynamics (Hohensinner et al., 2013a). This means that a clear answer to the first research question must be formulated in the context of the given climatic conditions and forms of catchment-wide and local human interventions (e.g., historical land use/cover changes; Hohensinner et al., 2021b).

5.2. Influence of riparian vegetation and vertical floodplain configuration

The hydraulic roughness of the riverbanks and in particular bank vegetation also substantially affect sedimentation rates and, accordingly, the celerity of floodplain accretion (Leopold and Wolman, 1957; Simm and Walling, 1998; Corenblit et al., 2007). Historical sources on the Danube River in Austria reveal that the lower limit of woody pioneer vegetation (mostly various willow species and German Tamarisk (*Myricaria germanica*)) was approximately equal with the summer mean water level (SMW) during the vegetation period (SMW = approximately, the mean water level + 0.3/0.4 m; Hohensinner et al., 2004, 2016). This vegetation limit roughly also coincided with the transition from (sandy) gravel deposits (bedload) to fine sediments in form of silt and fine sand. This accordance is by no means arbitrary. Rather, the type of sediment and the pioneer vegetation are intrinsically interconnected in a feedback loop (James et al., 2002; Gurnell et al., 2006; Västilä and Järvelä, 2017). In the case of the Danube River, (sandy) gravel bars that reach SMW feature adequate conditions for the germination of pioneer species (groundwater depth, inundation frequency and duration). The growing vegetation increases the hydraulic roughness of the surface of the bar. Consequently, at flows higher than SMW, flow velocity and shear stress are reduced, favoring the deposition of sediment with smaller grain sizes (also known as “comb effect”; Lavaine et al., 2015). The newly deposited material supports the growth of new riparian plants that, in turn, once more increase hydraulic roughness and boost sedimentation rates. This interplay between inundation, vegetation and sedimentation is recognized as a main process for (initial) floodplain accretion or emergence of vegetated islands (Piégay, 1997; Corenblit et al., 2007, 2009; Gurnell et al., 2012). Once the deposited material has reached a higher level where inundation frequency decreases, sedimentation rates and the influence of vegetation on this process also decrease. Fig. 7 and Table 1 illustrate that these feedback processes gradually level out and would theoretically cease at a certain elevation of the floodplain terrain above the water level, i.e., SMW.

In the case of the Viennese Danube this “climax level” (in the sense of a relatively stable quasi-equilibrium level that developed over a longer time period without disturbance/erosion; compare “typical floodplain” according to Wolman and Leopold, 1957) would be reached after roughly 500 yr, with a final thickness of the fine sediment layer of 2.64 m – in other words 2.64 m above SMW (Fig. 9). In comparison, August Höchsmann, a hydraulic engineer who was entrusted with regulation plans, estimated the maximum thickness of the fine sediment layer in the Machland floodplain 160 km upstream of Vienna at 2.5 to 2.8 m, which coincides well with the findings of the present study (see location of the Machland in Fig. 1; Höchsmann, 1848; Hohensinner et al., 2004). Considering that SMW was 1.7 m above zero point of gauge (approximately, the mean annual low water, MALW; Hohensinner et al., 2008), the climax floodplain level would theoretically reach an elevation of 4.34 m above MALW (or 3.04 m above mean water level) prior to major hydraulic measures around 1825. Based on historical topographical surveys, Hohensinner et al. (2016) and Hohensinner and Jungwirth (2016) reported floodplain terrain levels between 2.7 and 6.7 m, with a weighted average of 3.8 m above MALW. The latter value is 0.54 m lower than the presented hypothetical climax level (4.34 m above MALW). Note, however, that the mean elevation of 3.8 m also includes larger areas of younger and thus less aggraded and lower sites of the floodplain. As stated in Section 4.2, based on Fig. 6, the median thickness of the fine sediment layer of all analyzed boreholes is 2.0 m. Adding this value to the elevation of SMW (1.7 m above MALW) yields a median elevation of the floodplain terrain of 3.7 m above MALW, virtually identical to the mean weighted elevation of 3.8 m presented in the former studies.

5.3. Channel dynamics and lateral erosion

The available historical records show no evidence of natural levee formation along the Austrian Danube River. This is potentially because of poor accuracy in historical surveys. Nonetheless, even the topographical survey of the Viennese Danube floodplain in 1849 with a vertical resolution of 38 cm does not indicate fluvial structures that could be interpreted as such (Streffleur and Drobny, 1849). Levee formation as a result of amplified deposition of silt and sand along riverbanks is a phenomenon typically associated with low-energy systems such as meandering or anastomosing rivers (Nanson and Croke, 1992; Makaske, 2001; Klasz et al., 2014). Historically, such levees are documented for the meandering Danube section in southern Hungary, for example (Iványi et al., 2012). After floods, they functioned as a kind of natural polder-system, which retained the water in the floodplains over weeks or even months. To control the floodplain inundation and the inflow to floodplain water bodies to improve fisheries, small drainage canals were excavated through the natural levees (so-called "fok-system"; Guti, 2001). Such low-energy rivers generally feature comparably stable channels, and cohesive sediments predominate over coarser material such as gravel (Nanson and Knighton, 1996).

In contrast, the alluvial sections of the Austrian Danube River were characterized by intensive lateral erosion and channel avulsions. Median lateral erosion along all active river arms in Vienna's floodplain amounted to 2.4 m yr^{-1} in the periods 1805–1817 and 1817–1825, with a maximum of 33 m yr^{-1} (measured at the broadest site of each eroded bank section, $n = 657$). In such a "laterally active anabranching river" (according to Nanson and Knighton, 1996), most shorelines along the active channels shifted faster than vertical accretion of sediments and, thus, natural levee formation could proceed. Mixed sediment loads consisting of gravel, sand and silt were typical (see Section 2). According to the floodplain classification scheme of Nanson and Croke (1992), the former Viennese river landscape refers to a "medium-energy non-cohesive floodplain". Based on historical surveys and hydrodynamic flow simulations (Hohensinner and Trautwein, 2013), the specific stream power of the historical Danube (1817–1831) ranged between 14 and 19 W m^{-2} at bankfull flow in Vienna and in the downstream present-day national park. Accordingly, in respect of specific stream power Vienna's floodplain would have referred to the subtype B3, i.e., a "meandering river with lateral migration floodplain" (10–60 W m^{-2}). Nonetheless, based on the descriptions of floodplain types provided by Nanson and Croke (1992), Vienna's past alluvial landscape clearly showed fluvial forms and processes as specified for "wandering gravel-bed river floodplains" (type B2, 30–200 W m^{-2}). This discrepancy can potentially be explained by channel controlling factors that are not mentioned in the here cited classification of fluvial landscapes. In the case of the Viennese Danube, almost annually occurring (severe) ice jam floods amplified fluvial dynamics and fostered channel avulsions (Hohensinner et al., 2013a). Moreover, ongoing tectonic subsidence at the lower end of the study area might have affected morphological turnover processes (Lüthgens et al., 2017).

Between 1663 and 1825, on average annually 0.42% (~25 ha) of the floodplain terrain was eroded and new terrain aggraded to the exact same extent (Hohensinner et al., 2016). Over the long term the system remained in dynamic equilibrium or "quasi-equilibrium" (Leopold and Maddock, 1953; Wolman and Leopold, 1957) as hypothesized for such types of floodplains by Nanson and Croke (1992; compare Fig. 10a). Examining shorter time periods (1780–1805, 1805–1817, 1817–1825), annually between 0.40 and 0.56% (24–34 ha) of the Viennese floodplain were eroded (Hohensinner et al., 2016). According to the percent erosion rates, theoretically, the entire floodplain would have been renewed within 179 and 250 yr ("floodplain turnover time"; Wohl, 2015). In reality, however, over the long term, the channels often re-occupied former abandoned locations of the floodplain and did not erode all of the older alluvial terrain. In comparison, Aalto et al. (2008) estimated a much longer hypothetical time span for reworking of the entire Strickland River floodplain in Papua New Guinea (~920 yr). The floodplain width there is similar to Vienna's alluvial zone (~10 km), but the Strickland River is a lowland meandering river with less fluvial dynamics and tenfold more suspended load than the Austrian Danube. Wohl (2015) states that floodplain turnover times typically increase downstream because floodplain width generally increases faster than channel width. Documented turnover times range from decades in very small river-floodplain systems to 300–500 yr for larger rivers in Washington, USA, with catchment sizes between 1100 and 1200 km^2 (O'Connor et al., 2003; for comparison: Danube catchment size at Vienna: 102,000 km^2). In Vienna, bank erosion primarily affected morphologically younger terrain closer to the main river arms (compare Fig. 3). Thus, the terrain eroded between 1817 and 1825 showed a median age of only 20 yr compared to that of the entire floodplain with approximately 300 yr.

Three-dimensional reconstructions of the Danube River in the Machland floodplain upstream of Vienna revealed that the net erosion/aggradation significantly fluctuated prior to channelization (see location in Fig. 1). Between 1812 and 1817, because of intensive avulsion processes of the main river arms, annually 126,000 m^3 of gravel and fine sediments were eroded per km^2 of floodplain terrain. At the same time, an annual volume of 109,000 m^3 km^{-2} was deposited resulting in an annual net erosion of 17,000 m^3 km^{-2} (based on Hohensinner et al., 2014). The surplus of the eroded and, thus, downstream-released material in the entire 10.25-km-long Machland floodplain amounted to 400,000 m^3 yr^{-1}. Of this, according to the stratigraphic floodplain model presented in Hohensinner and Jungwirth (2016), hypothetically 31% were formed by fine sediments and 69% by bedload (gravel and coarse sand). This means that the 1812–1817 Danube avulsion in the Machland alone was equivalent to approximately 80% of today's hypothetical annual bedload transport capacity of the Viennese Danube (~340,000 m^3 yr^{-1}, see Section 2). In contrast, the released fine sediments made up only ~4% of the total suspended load (mean load between 1878 and 1884: 5.95 million t). Most probably, the entire volume of the released material was not directly transported 160 km downstream to Vienna. Instead, it was partly – or most likely largely – deposited in the next downstream floodplain and later remobilized. For the subsequent period between 1817 and 1821 that was hydromorphologically less turbulent (reflecting a stabilization phase after the intensive fluvial dynamics), our model yielded an annual net deposition of 4400 m^3 km^{-2} in the Machland floodplain.

5.4. How have sedimentation processes changed as a consequence of channelization?

Intensive fluvial dynamics have probably wiped out initially aggraded natural levees along the historically anabranching channels of the Danube River and perpetually reworked and renewed the floodplain features. Today, because of channelization in the nineteenth century, the alluvial reaches of the Upper Danube River are stabilized. Specific stream power at bankfull flow significantly increased from originally 14–19 W m^{-2} to currently 56 W m^{-2} because of channel narrowing and straightening. Without bank protection, today's much higher stream power would promote lateral erosion and channel widening. This was the case in the "National Park Donau-Auen" downstream of Vienna, where the riprap was locally removed (Fig. 4). Here, a 5 to 20 m broad floodplain strip was eroded at the inner (convex) bank within only one year (Klasz et al., 2008). Because of the absent channel/shoreline migration, along regulated river reaches vertical floodplain accretion and natural levee formation are commonly observed (Middelkoop, 2005; Hudson et al., 2008). In the national park, Klasz et al. (2014) reported natural levees with heights between 0.3 and 3.7 m. Since the channelization of that Danube section from 1880/90 until 2009/10 (~120 years), the average annual sedimentation rates amounted to 1.1 cm close to the river and 0.03 cm in more remote floodplain sections (Klasz et al., 2014). According to our model for the near-natural Viennese floodplain, an average (interpolated) rate of 1.1 cm yr^{-1} was typical for 155- to 203-yr-old floodplain terrain, depending on the calculation method (see (a) and (b) in Table 2). Similarly, a rate of 0.03 cm yr^{-1} is valid for the much more than 500-yr-old floodplain sections under pre-channelization conditions. Based on the incremental cumulative approach, a rate of 1.1 cm yr^{-1} was calculated for much younger floodplain sections. In this case such deposition rates already occurred between 25 and 50 yr after the onset of the sedimentation process (see (c) in Table 2).

The comparison of the averaged (interpolated) values shows that sedimentation close to the river proceeded between 23 and 41% faster (depending on the calculation method (a) or (b) in Table 2) along the channelized Danube River than under near-natural historical conditions. Though comparable cumulative sedimentation rates are not available for the channelized Danube River, it can be assumed that the difference is somewhat greater using the cumulative calculation method to estimate the rates. A concrete comparison of the deposition rates in remote floodplain areas of the channelized river (0.03 cm yr^{-1}) with the historical situation is difficult because of the approximative character of the function (curve) after a time period of 400–500 yr (Fig. 8).

In addition to the local or sectional hydraulic interventions, also the upstream sediment supply, i.e., suspended load, is critical. Climate change and human land cover modification in the upstream catchment significantly influence the volume and type of the available sediment at a certain river reach (Walling, 1995; Xu, 2003; Gell et al., 2009; Giosan et al., 2012; Hohensinner et al., 2021b). Moreover, the construction of bedload barriers in headwaters and weirs for hydropower plants not only prevents bedload transport but also significantly reduces fine sediments (suspended load). As for the Danube River in Vienna, the suspended load was reduced by between 18 and 45% since the 1880s (based on Penck, 1891; Nachtnebel et al., 1998; unpublished data, https://ehyd.gv.at/). Though fine sediments are still deposited in floodplains during floods, bank erosion has almost ceased because of bank protection. Consequently, lateral channel erosion cannot balance the deposited material in the floodplains (Fig. 10b). This contributes to the sediment deficit in the river (Klasz et al., 2014). The estimated overbank deposition rate of about 416,000 m^3 yr^{-1} for the entire "National Park Donau-Auen" is equivalent to between 18 and 20% of the mean annual suspended yield (Klasz et al., 2014). In this respect, the significant reduction of the suspended load since the 1880s partly counteracts ongoing floodplain deposition. If the volume of transported fine sediments would still be the same as in the late nineteenth century, then the current deposition rates would be hypothetically greater and the sedimentation process would be significantly faster than estimated by Klasz et al. (2014). Thus, the above-described difference to the pre-channelization state would be more pronounced.

6. Conclusion

Vienna's comprehensive borehole database and research on long-term fluvial dynamics provide information on the sedimentation processes in the past Danube floodplain. Specifically, the sedimentation rates over the last 500 yr can be estimated by correlating the thickness of fine sediment layers (silt and fine sand) on top of the gravel-dominated riverbed and gravel bars with the morphological age of individual floodplain sites prior to major river engineering measures (channelization, flood prevention). Five years after the beginning of the sedimentation process at a particular gravel bar, the median deposition rate was 18.60 cm yr^{-1}. In the following decades the rate significantly decreased and leveled off after 300 yr with median annual rates between 0.15 and 0.10 cm. Approximately 500 yr after the onset of the deposition process, the fine sediment layer reached a thickness of 2.64 m, of which half was already deposited within the first 10 yr (Fig. 9). On the very long term, we expect a dynamic equilibrium between floodplain sedimentation and lateral erosion/channel migration (Fig. 10a).

Channelization and stabilization of riverbanks significantly boosted floodplain sedimentation – most strongly in areas close to the river (Fig. 10b). Natural levees along the riverbanks arose soon after river regulation. They reached heights of up to about 3.7 m within 120 yr. Levee formation was probably not a typical phenomenon in the Austrian Danube floodplains prior to channelization. Intensive fluvial dynamics, i.e., lateral bank erosion, prevented the evolution of distinct natural levees. As a consequence of channelization the sedimentation rate increased by 23 to 41% (depending on the calculation method) compared to historical near-natural conditions. Hypothetically, the rate increase was significantly higher when taking the historical volume of suspended load into account. The load transported by the Viennese Danube River decreased by between 18 and 45% (depending on the compared time period) in the twentieth century as a consequence of the numerous bedload barriers in Alpine headwaters and reservoirs along the Upper Danube River and its tributaries.

The data indicate that a median "climax level" of the deposited material is reached after approximately 500 yr when the sedimentation rates and the accretion of the fine sediment layer almost ceased. This quasi-equilibrium "climax level" is typical for individual river reaches and depends on numerous controls including the hydrological regime, volume and type of sediment supply, channel slope, floodplain width and riparian vegetation. It may develop under stationary hydromorphological conditions as long as the external controlling factors (climate, land use in the catchment, etc.) do not change. Over longer time frames, larger floodplains mostly integrate several parts that aggraded under different climatic and human-altered conditions. In the floodplain, each phase of changed environmental conditions is reflected by different sedimentation rates and potential "climax levels" of the aggraded material. Moreover, the phase of floodplain accretion (initial phase, mid- or final phase) in which a certain alluvial site is currently in is relevant. Because the controlling factors may change within centuries or even decades, parts of a floodplain may not reach the final "climax level". Instead, the deposition process is altered, resulting in a new, lower or higher "climax level". In this respect, riverbank stabilization is also a controlling factor that significantly influences sedimentation rates and the potential evolution of floodplain soils. In the absence of erosion processes, the floodplain terrain piles up until a level is reached that can be hardly overflowed even during larger flood events.

The diverse nature of rivers and alluvial landscapes makes it difficult to compare individual deposition processes. This contribution presents first basic data on the long-term sedimentation processes and floodplain accretion. It provides a sound basis for interpreting floodplain deposition in current, significantly human-modified river landscapes.

Data policy statement

Basic data are partly subject to third party restrictions. Selected data are available on request.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found in the online version, at https://doi.org/10.1016/j.geomorph.2021.108038. This data include the Google map of the most important areas described in this article.

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