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An improved method for paleoflood reconstruction and flooding phase identification, applied to the Meuse River in the Netherlands

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ABSTRACT

This study investigates Holocene floodplain evolution and flooding phases as experienced in the Lower Meuse catchment, primarily based on grain-size distributions of channel-fill and floodplain deposits in sediment cores. The presence of post-depositional Fe–Mn concretions and resistant organic particulate materials impedes the direct use of grain-size data. By combining end-member modelling results with laboratory observations, we have constructed a Flood Energy Index (FEI), which allows identification of phases of past increased flooding from the grain-size signal. Since concretions and organic-rich sediments regularly occur in floodplain sediments, we emphasize that the quality of a grain-size dataset should be assessed prior to its use for reconstruction of flood events. We suggest that the new approach has potential to become standardized within paleoflood research. The temporal variation of FEI in Meuse sediment cores highlights multi-centennial flooding phases occurring at c. 8500, c. 8000, c. 7600, c. 7000 and c. 5900 cal BP within the fluvio-lacustrine environment (early–middle Holocene). The record of low flood activity in the Subboreal is attributed to a cooler and dryer climate anomaly after the Holocene Climatic Optimum. In the late Holocene, the first flooding phase occurring at c. 2800 cal BP can be linked to the 2.8 ka climate anomaly. During the last two thousand years, the variation of FEI index reveals oscillating flood regimes in the Lower Meuse floodplain. The last three recorded flooding phases most likely coincide with the Roman Period (c. 12 BCE–250 CE), the Medieval Warm Period (c. 950–1400 CE) and the Little Ice Age (c. 1400–850 CE). Despite uncertainty in the age-model, the rapid accumulation rate and amplified flood magnitudes imply increased fluvial instability during the late Holocene, indicating that humans exerted a profound influence on fluvial dynamics in the Meuse.

1. Introduction

Fluvial systems have been influenced by climate change and human activity within a multi-millennial time scale during the Holocene (e.g., Macklin et al., 2012, 2015). A change in the frequency and magnitude of floods is the main direct driver that determines the response of river systems to climatic change (Ward et al., 2008; Macklin et al., 2012; Starkel et al., 2015). Therefore, fluvial systems can act as recorders of climate and human-induced changes and their deposits can thus be used to infer past climatic conditions and human impacts on the fluvial system (e.g., Ely, 1997; De Moor et al., 2008).

Since Baker et al. (1983) introduced the analyses of slackwater deposits from confined bedrock-gorge settings in paleohydrological

research, these deposits have been widely used to reconstruct past floods and flooding regimes, because they can be easily distinguished from other sediment units by applying a series of criteria, such as lithology, grain-size, sediment color, sedimentary texture and structure, and interbedded palaeosols (Ely, 1997; Yang et al., 2000; Huang et al., 2012). However, the use of slackwater deposits in the low-relief landscape that characterizes the lower Meuse floodplain is impractical, because gradually attenuated overbank flow carries sediments away from the river channel more regularly, and the accumulated deposits have adopted a relatively homogeneous structure. This setting makes recognition of clear sequences of flood deposits, dating and assigning a flood stage to reconstruct paleodischarge and assessment of recurrence frequency, a challenging enterprise. Alternative sedimentary archives

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(e.g., gyttja, peat and varved lacustrine sequences) have been used for reconstructing past hydrological events. For instance, well-preserved peat sequences on floodplains and in lakes can be used to study flooding magnitude and frequency (i.e., Ishii et al., 2017). Furthermore, grain-size variation from appropriate sedimentary settings is suited to interpret episodic, high-magnitude floods because the numbers of coarse grains entrained with suspended material increases with flood magnitude, and these coarse sediments can settle out from suspension relatively quickly (Lim et al., 2013; Toonen et al., 2015). However, in a distal floodplain setting the sand input is limited. Therefore, linking the coarse-grained components solely to large floods can lead to misunderstanding because distal floodplain sites do not contain a substantial amount of sand. In addition, the conventional use of grain-size information in paleoflood event reconstructions generally incorporates the entire grain-size distributions (GSDs) without considering the reliability of the dataset (i.e., Toonen et al., 2015). Moreover, iron and manganese (Fe–Mn) concretions in fluvial sediments, which are broadly associated with soil formation and gley phenomena, can be hard to remove/dissolve during sample pretreatment (Schwertmann and Fanning, 1976). To our knowledge, there is nothing reported concerning this issue in published flood reconstructions. The aforementioned issues can however greatly affect the reliability of paleoflood reconstructions in floodplain environments. Hence, caution should be taken when interpreting the grain-size data and linking such information to palaeoflood magnitudes.

In the Netherlands, the instrumental flood record is relatively short. The earliest instrumental discharge record for the Meuse River dates back to 1911, and older historical written flood records start from the 13th century (Buisman, 1995; Glaser and Stangl, 2003). There is no one long-term (millennial scale) flood reconstruction for this fluvial system. As a rain-fed river, the discharge of the Meuse reacts quickly to precipitation changes (Kasse et al., 1995; Ward et al., 2008). Therefore, the information derived from a paleoflood study holds great potential for reflecting the long-term changes of past precipitation.

Here we address the above-mentioned challenges to a study of the paleoflood record for the Lower Meuse. By analyzing the sedimentological properties of a floodplain channel-fill core at 2-cm resolution, we minimize unwanted signals related to the presence of Fe–Mn concretions and organic remains that survived sample preparation. The applicability of the proposed method is evaluated prior to identifying the flooding phases of the Lower Meuse. Given the fact that grain-size data can be biased by post-depositional concretion formation and/or high amounts of organic matter in other sedimentary contexts, we emphasize that the quality of a grain-size dataset should be assessed prior to using it for flooding reconstructions. We suggest that the new method and approach have potential to be standardized in paleoflood research elsewhere. As a building block for future work and a paper that emphasizes on methodology in paleoflood study and its application in the Lower Meuse, we do not detail the human and climatic influences on the Meuse catchment during the Holocene. This synthesized work will be presented in the next step, combined with a new sedimentary record from the same river.

2. Setting

The Meuse originates from northeastern France and flows northward through Belgium, passing through the uplifted Ardennes. After initially continuing northward in the Netherlands, the flow direction changes to the west before entry into the low-relief Rhine–Meuse delta plain and debouchment into the North Sea. The study site, Well–Aijen (referred to as WA hereafter), is located in the Limburg Province of the southern Netherlands, and about 45 km upstream of the Rhine–Meuse delta apex. Tectonically, this site lies on the Venlo Block of the Roer Valley Rift System (Fig. 1a). The relatively subsiding Venlo Block is separated from its adjacent uplifting Peel Block by the Tegelen Fault Zone (Van Balen et al., 2005). Geological studies indicate that average

Fig. 1. (a) Location map of the study area and major fault zones in the southern Netherlands. The research site (red box) is located on the subsiding Venlo Block which is separated from the southwestern Peel Block by the Tegelen Fault Zone (Van Balen et al., 2005). (b) Digital elevation map (<https://ahn.arcgisonline.nl/ahnviewer/>) shows the Well–Aijen Holocene floodplain elevation. The light-yellow area depicts the distribution of the Holocene floodplain. The dashed lines outline the abandoned channels. East and west of the Meuse Late Glacial terraces, aeolian dunes and cover sands are present (Kasse et al., 1995; Woolderink et al., 2018). The insert shows core WA11 and core WA16 within the WA paleochannel; note that the two cores are aligned parallel to the river flow. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

subsidence rates in the Roer Valley Graben and the Peel Block are ~88 mm/ka and ~46 mm/ka respectively during the Quaternary (Houtgast and Van Balen, 2000). The subsidence rate for the Venlo Block is possibly even lower, and there is no evidence indicating

tectonic movement during the Holocene.

In the Roer Valley Graben the Meuse has a broad Holocene floodplain (~3.5 km wide) due to its highly sinuous channel. On the Peel Block the Holocene floodplain is very narrow (~500 m wide) along the relatively straight, incised channel. On the Venlo Block, where WA is situated, the floodplain gradually widens towards the Rhine–Meuse delta. The Holocene floodplain at WA is confined by Lateglacial terraces and superimposed inland dunes (Fig. 1b). Directly to the east, aeolian inland dunes derived from braided river deposits were formed during the Younger Dryas (Kasse et al., 1995). During the transition from the Lateglacial to the Holocene, the Meuse changed from a braided to a meandering system in response to climate change (Kasse et al., 1995; Huisink, 1999; Tebbens et al., 1999) and, as a consequence, many braided channels were abandoned (Fig. 1b). The study area has been extensively investigated by means of paleogeographic mapping, establishing regional pollen records and constructing frameworks for prehistoric human activity (e.g., Zuidhoff and Van der Velde, 2008; Bos et al., 2014; Bos and Zuidhoff, 2015; Hoek et al., 2017; Isarin et al., 2017; Rensink, 2017; Woolderink et al., 2018). In 2011, an abandoned paleochannel at WA was studied using hand-drilling and a trial trench. Archeological finds indicated that this trench revealed a sediment record that covers the last ~8000 years (Bos and Zuidhoff, 2015). Thus, the WA site could be ideal for tracing past fluvial activity. According to the Lateglacial-to-Holocene terrace stratigraphy of the Meuse (e.g., Hoek et al., 2017; Woolderink et al., 2018), it follows that the WA channel was active during the Preboreal and Boreal (early Holocene). Radiocarbon dating together with a palynological study confirm its abandonment in the late Boreal (Bos and Zuidhoff, 2015). Since then, the abandoned channel received sediments when bankfull discharge was exceeded during flooding. After the abandonment, the main channel of the Meuse was positioned roughly at its present location (Kasse et al., 1995), and no apparent migration has been recognized. Our observations support this. First, Mesolithic and Neolithic archeological sites were preserved in the area between the WA paleochannel and the present river (Bos et al., 2014; Bos and Zuidhoff, 2015), implying there was no eastward migration in the past. Second, the existence of paleochannel relicts between the studied paleochannel and the Lateglacial terraces precludes the westward erosion after the channel abandonment (Fig. 1b).

3. Material and methods

3.1. Coring and scanning

In 2011, core WA11 was drilled using a Begemann coring device into the deepest part of the WA paleochannel to produce a record for a paleoecological study (Bos and Zuidhoff, 2015). In 2016, a second core (WA16) was retrieved from the same paleochannel, 14 m NW of the core WA11 site. The two cores are aligned parallel to the present river flow direction (Fig. 1b). Both sediment cores reached the coarse-grained channel bed and have respective lengths of 6.29 m (WA11) and 6.26 m (WA16). The top 30 cm of core WA16 was discarded because this interval represents soil that is heavily disturbed by modern farming activities.

Core WA16 was recovered using PVC tubes with an inside diameter of 66 mm. It was transported to TNO Geological Survey of the Netherlands, where it was split, described, photographed (registered as core B52E3302) and sampled. One core-half was used for sampling, the other half was scanned with a GEOTEK Multi-Sensor Core Logger at VU Amsterdam at 0.5 cm resolution to obtain a record of magnetic susceptibility and gamma-ray bulk density. Before scanning, the surface of the core halves was cleaned using a plastic blade. The magnetic susceptibility was measured using a Bartington MS2E point sensor and the bulk density record was corrected for variations in sediment thickness according to the Geotek calibration protocols (<http://www.geotek.co.uk/support/downloads/>).

3.2. Grain-size analysis and end-member modelling

We obtained 292 samples for grain-size analyses, continuously sampled with boundaries placed at 2 cm intervals. The method described in Konert and Vandenberghe (1997) was used to pretreat the samples for grain-size analysis. Samples from the upper 340 cm clastic sequence were successively pretreated with 10 ml 30% H₂O₂ and 5 ml 10% HCl to remove organic matter and calcium carbonates respectively. Additional H₂O₂ was added to the organic-rich samples from the lower organo-clastic part of the core (below 340 cm) to remove excess organic matter. Grain-size distributions (GSDs) ranging from 0.1 to 2000 µm were measured with a Sympatec HELOS KR laser-diffraction instrument at the VU Amsterdam.

After pretreatment, we observed that 43 samples in the interval 40–210 cm still contained sand-sized undissolved concretions, and several samples from the lower 300 cm still contained resistant particulate organic remains. In order to understand their physicochemical properties, concretions from six samples were analyzed with a scanning electron microscope (JCM-6000Plus) at the VU Amsterdam. The results show that concretions from the same and from different depths can have different chemical compositions (supplementary material, Fig. S3). It was suspected that these post-depositional concretions and organic remains would affect the ‘true’ GSDs. Hence, another six samples were taken from both the clastic and organo-clastic part of core WA16 and pretreated with different methods to evaluate the reproducibility of the GSDs. These testing methods involve the use of ultrasonic homogenizer and additional reagents (supplementary material, Table S1). The tests show that none of the methods could completely dissolve the concretions.

We deployed end-member (EM) modelling to decompose the GSDs into a set of end members (Prins and Weltje, 1999; Weltje and Prins, 2003). This approach has been designed to offer the simplest explanation for the observed variations in GSDs in terms of grain-size EM. This EM modelling approach has been successfully applied for extracting representative paleoclimate indicators and detecting mixing patterns of sediments supplied from multiple sources (e.g., Prins et al., 2007; Shang et al., 2016). Also, the EM modelling has been applied to study transport processes and fluvial dynamics (Toonen et al., 2015), thus demonstrating that the use of EMs is geo-genetically meaningful and not merely a descriptive methodology. Here, we used the AnalySize package (Paterson and Heslop, 2015; Van Hateren et al., 2018) to obtain EMs and to determine their proportional contributions in each grain-size sample. The coefficients of determination (r^2) were calculated to assess the minimum number of EMs (cf. Prins and Weltje, 1999). The coefficients of determination represent the proportions of the variance of each grain-size class that can be reproduced by the approximated data. This proportion is equal to the squared correlation coefficient (r^2) of the input variables and their approximated values (Prins and Weltje, 1999; Prins et al., 2000).

3.3. Thermo gravimetric analysis

A Leco TGA701 analyzer at VU Amsterdam was used to determine organic matter and calcium carbonate content (Heiri et al., 2001). To achieve greater contact area and better equilibrium conditions, the samples were ground into powder after being dried at 50 °C overnight. Each analysis started with the removal of residual moisture by purging the closed environment during heating to 105 °C. The second heating step was performed in a pure oxygen atmosphere, the temperature being ramped up to 330 °C, thus combusting the labile organic matter. At the third step, under air atmosphere, the reaction temperature rose to 550 °C to combust the resistant organic matter. The last heating step was performed in a CO₂ atmosphere, by heating up the sample to 1000 °C, and different types of carbonate were dissociated. In this manner, the total organic matter (TOM, sum of labile and resistant organic matter) and the carbonate contents were obtained.

Table 1

Radiocarbon dated samples from core WA11 (Bos and Zuidhoff, 2015) and WA16 (this study). Depths of dates from core WA11 were transferred to correlated depths in WA16 (Section 1 in supplementary material).

Lab code	Depth in WA11 (cm)	Depth in WA16 (cm)	^{14}C age (yr BP)	Calibrated ages (median)	Calibrated ages 2σ age ranges	Material dated
SUERC-44817	344.5	347.7	4363 $\pm$ 35	4926	5039–4851	wood
GrM-13857		448	6090 $\pm$ 20	6954	7006–6891	wood pieces
SUERC-44816	450.0	451	6111 $\pm$ 35	6987	7157–6893	wood
GrM-13859		458	6140 $\pm$ 20	7048	7157–6954	wood pieces
SUERC-44818	523.5	523	7069 $\pm$ 35	7896	7966–7836	leaves
GrM-13860		552	7205 $\pm$ 20	8004	8044–7964	leaves/bud scales
SUERC-44822	586.5	584.7	7848 $\pm$ 35	8625	8755–8548	leaves
GrM-13858*		623	9095 $\pm$ 25	10,239	10,267–10,208	seeds/wood*
SUERC-823*	629	626	8845 $\pm$ 35	9947	10,156–9744	seeds*

* marked deepest samples might be from reworked materials.

3.4. Geochronology

3.4.1. AMS ^{14}C dating

Four AMS ^{14}C dates were obtained from the lower organo-clastic part of core WA11 by Bos and Zuidhoff (2015) (Table 1). We visually compared the two cores as the first step to understand the sedimentary lithofacies. The correlation of the cores shows that the sedimentary lithofacies and their marker horizons in the two cores are essentially the same (supplementary material, Fig. S1). To constrain the chronology for core WA16 further, four terrestrial plant macrofossil samples were picked from it and ^{14}C dated at the University of Groningen (Table 1). An age–depth model was constructed for the organo-clastic part (units 2 and 3) and the basal clastic part of unit 4 by using Bayesian statistics (Blaauw and Christen, 2011) using the IntCal13 calibration curve (Reimer et al., 2013).

3.4.2. Pollen zone subdivision and additional age control methods

No identifiable organic matter for ^{14}C dating was present in the upper 340-cm clastic part of core WA16. Therefore, the established Holocene regional biostratigraphy, pollen assemblages of core WA11 (Bos and Zuidhoff, 2015), and a tuned age model based on sedimentary proxies for accumulation rates were used for chronology (cf. Minderhoud et al., 2016). The pollen assemblage zones and archaeological finds have allowed subdivision of the sedimentary strata according to main cultural and climatic zones (supplementary material, Fig. S5). Specifically, paleobotanical interpretations reveal that the condensed intervals of 334–314 cm and 314–274 cm correspond with the Bronze Age (c. 4000–2750 cal BP) and Iron Age (c. 2750–2000 cal BP), respectively. The shallowest sample in which pollen investigation was possible originates from 274 cm and is indicative of the inception of the Roman Period (c. 2000 a BP). Variable textural and grain-size variation of the upper clastic sequence indicate a dynamic sedimentary environment. Therefore, we link the granulometry-based parameters (FEI index, see 4.4) with accumulation rate to establish a tuned age model for the interval 314–70 cm (Section 4 in the supplementary material). This method is modified after Minderhoud et al. (2016), who linked loss-on-ignition with accumulation rate. The modelling method described there can also be performed on other ‘continuously sampled’ sedimentary proxies for vertical aggradation, for example grain-size or geochemistry data (Minderhoud et al., 2016).

X-ray fluorescence analysis (XRF) was performed on a black layer at ~70 cm depth by using the Pananalytical MagiX Pro XRF at the VU Amsterdam. The XRF results (supplementary material, Fig. S2) confirm deposition of this black layer during the period of coal mining further upstream along the Meuse in the middle 19th century (Stam, 2002).

4. Results

4.1. Description of the sedimentary units

Six lithostratigraphic units were distinguished in core WA16, based on the sediment characteristics such as color, sedimentary structure, grain-size composition, organic and carbonate content, core density and magnetic susceptibility. Generally, the base of core WA16 consists of coarse sands and gravels (unit 1). They are overlain by organic-rich deposits (units 2 and 3) and clastic deposits (units 4–6) (Fig. 2a, b). GSDs obtained from unit 1 are strongly bimodal with modes centered around 10 and 400 μm (Fig. 3). In units 2–6, the GSDs have dominant modes at 8–20 μm and occasionally a secondary mode above ~100 μm (Fig. 3). Table 2 summarizes the granulometric information for each unit.

Unit 1 (626–610 cm) starts from the base of the core (Fig. 2b). This unit has a dark grey color and a rapid fining-upward trend (Fig. 4a). At the very base it consists of well-rounded gravels and coarse sands. The core density is highest at the base and decreases gradually upwards (Table 2; Fig. 2f).

Unit 2 (610–450 cm) exhibits dark-brown and black colors. Light-grey clastic laminations with millimeter-scale thickness were observed (Fig. 4c, d). The median grain-size curve shows minor variations around 8.1 μm (Table 2). This unit contains a large portion of organic matter (Table 2; Fig. 4b–d), as reflected by the increased total organic matter (TOM) values and low carbonate content (Fig. 2d, e). The core density is rather low in this unit, varying between 0.24 and 1.38 g cm^{-3} . Magnetic susceptibility is very low throughout this unit (Fig. 2g).

Unit 3 (450–330 cm) consists of clastic sediments mixed with organic matter and has a color and structure similar to unit 2. Macroscopic charcoal particles are present (Fig. 4e, f), their frequency and size showing a decreasing trend towards the top of the unit. There is no clear change in median grain size, but the organic-matter content is significantly lower than in unit 2 (Fig. 2d). Calcium-carbonate content co-varies with TOM values. Core density gradually increases throughout this unit and magnetic susceptibility is slightly higher compared to unit 2 (Fig. 2f, g).

Unit 4 (330–165 cm) has a gradual contact with unit 3 and is characterized by homogenous yellowish clay (Fig. 2a). In this unit, oxidized rusty spots occur sporadically (Fig. 4g, h). Notably, at 220–200 cm, the sediment color changes from yellow to brown, whereas most of the sediment properties show only minor variations. The median grain size shows a series of peaks at ~250 cm (Fig. 2c). TOM reaches its lowest level with an average of 5.9%. Core density and magnetic susceptibility increase whereas carbonate content decreases slightly (Fig. 2e–g).

Unit 5 (165–70 cm) is distinguished from unit 4 by its brown silty clay with loose structure and coarser grain size (Fig. 4i). The median grain size varies between 8.4 and 16.5 μm (Fig. 2c). TOM values (around 8%) and calcium-carbonate content (1.5–1.7%) remain at a

Fig. 2. Sedimentary characteristics of core WA16. (a) photograph of the core; the asterisks show the calibrated ^{14}C dating positions from core WA11 (red, Bos and Zuidhoff, 2015) and WA16 (blue, this study); (b) shows the unit subdivision with bars along the right side marking the observed clastic flooding layers (see 4.1, thickness is not on scale); unit 1 represents channel bed sediments that are composed of gravels and sands; units 2–3 are the organo-clastic infilling of the abandoned channel; units 4–6 are flooding sediments mainly consisting of clays and silts; (c) median grain-size; (d) total organic matter (TOM) content; (e) calcium carbonate content; (f) core density; (g) magnetic susceptibility. Horizontal dashed lines mark the unit boundaries. Note that the basal part of the core (Unit 1) has not been analyzed by TGA (d, e) and core scanning. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

low level (Fig. 2d, e). At about 100 cm, the median grain size, core density and magnetic susceptibility increase significantly, whereas carbonate content decreases slightly (Fig. 2).

Unit 6 (70–30 cm) consists of silty clay sediments and is separated from unit 5 by an 8 cm thick black layer at 70–62 cm (Fig. 2a). The median grain size varies at $\sim 9\ \mu\text{m}$. TOM and calcium-carbonate content reach up to 18% and 2.5%, respectively (Fig. 2d, e). Core density decreases to $0.85\ \text{g cm}^{-3}$ (Fig. 2f). The black layer was also observed in core WA11. Starting from the bottom of this layer, magnetic susceptibility increases sharply to the highest recorded value (80 SI units) at the top of this layer, and then reduces rapidly to about 30 (SI units) (Fig. 2g). The geochemical analysis revealed enriched Zn and Pb concentrations in this layer (Fig. S2).

The organo-clastic interval of core WA16 (units 2–3) was examined to identify clastic flooding layers through a variety of sedimentological criteria that are commonly used in paleoflood hydrology (Benito et al., 2003; Benito and Thorndycraft, 2005): (i) color change; (ii) silt and clay deposits intercalated between organic-rich deposits; (iii) horizontal clay laminations. These flooding layers appear as single layers as well as clustered laminations in the sediments, and the thickness varies between 0.1 and 0.6 cm (Fig. 4c–f). The black bars in Fig. 2b and Fig. 4c–f show the position of the clastic flooding layers (thickness is not to scale).

4.2. Age–depth model

The AMS ^{14}C ages were calibrated to calendar years before present (1950 CE). The calibrated ages in units 2 and 3 do not deviate from a steady increasing age trend with depth (Table 1, Fig. 2, Fig. 5), suggesting that the sediment accumulation took place progressively and

without reworking in the organo-clastic sequence. The tuned age–depth relation for the clastic sequence is presented in Fig. 5, and the detailed information is provided in the spreadsheet of the supplementary material. Average accumulation rates were derived for units 1–3, and for the overall clastic sequence (Fig. 5).

4.3. End-member modelling analysis

The median coefficient of determination (r^2) reveals that a four-EM model explains $> 85\%$ of the total variance of the GSDs (Fig. 6a). Using a five-EM model or six-EM model does not significantly improve the performance, while a three-EM model performs very poorly ($r^2 < 0.7$) in the range from 53 to 220 μm (Fig. 6b). Although the median r^2 values of the four-EM model decrease from 0.8 to 0.6 in the grain-size interval 105–200 μm (Fig. 6b), reflecting a slightly reduced explanatory power in this range, we consider the four-EM model to be the most suitable mixing model for the WA16 grain-size dataset. Fig. 6c shows the grain-size distributions of the four EMs, with the finest labelled EM1 and the coarsest labelled EM4. EM1 has a mode at 7 μm . The modes of EM2 and EM3 are 20 μm and 31 μm , respectively. The secondary mode of EM3 at $> 100\ \mu\text{m}$ (Fig. 6c) is associated with the decreased coefficient of determination within that range (Fig. 6b). The coarsest EM4 has a dominant mode at 450 μm and a secondary mode at 10 μm .

Fig. 6d denotes the relative contribution of the four EMs in relation to their stratigraphic position. EM1 and EM2 are dominant throughout the core. Their total contribution is about 85% in units 2–6 (Table 2). EM1 varies around 50% at 610–270 cm, from 270 to 230 cm its contribution is very low (average of 20%), at 230–165 cm it again reaches to 50%, but from 165 cm its contribution declines again ($< 38\%$). The average abundance of EM2 at 610–560 cm is 35%, but it decreases

Fig. 3. Grain-size distributions (GSDs) for each unit. Unit 1 shows bimodal grain-size distributions with modes at 10 and 400 μm . The GSDs in the upper five units show that the deposits generally consist of fine-grained particles dominated by a mode at 8–20 μm . Note that some of the GSDs in units 2 and 5 have irregular tails at grain sizes > 100 μm .

gradually to 470 cm. EM2 then increases to an average of 43% in units 3 and 4. In the middle of unit 4, it shows a series of spikes (Fig. 6d). In units 5 and 6, EM2 accounts for an average of 65%. The average contributions of EM3 in unit 2 and units 5–6 are 24% and 15%, respectively. In units 3–4, the average proportion of EM3 is 8%, with variation within a narrow range. EM4 is dominantly present in the core base but almost absent in units 2–6 (Fig. 6d).

5. A new method for paleoflood reconstruction: The Flood Energy Index (FEI)

To date, linking the coarse fraction of grain-size distributions (e.g., the P90, the percentage of the sand fraction, or a coarse EM score) to paleoflood magnitudes is becoming a relatively widely applied approach in paleohydrological research (e.g., Knox, 2006; Toonen et al., 2015; Munoz et al., 2018). For several reasons, we consider this method

suboptimal. For example, the lack of substantial sand content in a sediment record does not necessarily mean that no flood events occurred in the past. Minor floods with a limited inundation area may not pass a uniformly set critical P90 value or minimum EM score, and appear unrecorded in otherwise continuous series. More importantly, if (part of) the grain-size record is ‘contaminated’, the (problematic) GSDs should be corrected by filtering out the noise. Specifically, in units 2 and 5, the coarse tails of the GSDs (> 100 μm) exhibit anomalous shapes (Fig. 3). One reason for this could be that this grain-size range includes contamination from sand-sized Fe–Mn concretions in unit 5 and particulate organic remains that survived from sample pretreatment in unit 2 (e.g., Toonen et al., 2015). We have therefore developed a Flood Energy Index (FEI) that normalizes the raw grain-size parameters by incorporating the flood-sensitive coarse populations and excluding the contaminated grain-size portions. This index highlights the increased floods that occurred within a certain period by quantifying

Table 2
Sediment characteristics of each unit from core WA16.

Unit	Sediments type	Median (μm)	Depth range (cm)	Sand (vol %)	TOM (wt%)	Carbonate (wt%)	Density (g cm^{-3})	P _{EM1}	P _{EM2}	P _{EM3}	P _{EM4}
6	Brown silty clay	8.1–8.5 (8.2)	70–30	2.11	8.9–15.1 (10.2)	1.9–2.5 (2.0)	0.06–1.90 (1.40)	33.3	45.3	9.9	0
5	Brown silty clay	8.4–16.5 (10.6)	165–70	2.25	3.8–9.8 (5.0)	1.4–2.0 (1.6)	1.07–2.24 (1.71)	15.9	69.5	14.5	0.1
4	Yellowish clay	5.4–19.6 (7.6)	330–165	1.53	4.2–11.4 (5.9)	1.5–2.3 (1.9)	1.11–1.99 (1.63)	48.4	45.7	5.8	0.1
3	Clastic sediments mixed with organic matter	6.3–9.7 (7.5)	450–330	3.36	9.3–25.8 (17.5)	2.1–3.1 (2.5)	0.71–1.72 (1.25)	50.6	39.4	9.8	0.3
2	Organic dominant infills	6.0–19.2 (8.1)	610–450	7.31	15.2–69.4 (38.0)	1.1–3.4 (2.3)	0.24–1.38 (0.88)	48.6	25.0	24.2	2.2
1	Gravels and sands	9.2–335.8 (91.6)	626–610	45.65	4.7–26.4 (12.4)	1.2–2.7 (2.1)	1.05–2.05 (1.38)	12.9	10.4	10.8	65.9

Numbers in parentheses are mean values.

Fig. 4. Photographs of the units. (a): unit 1 shows a fining-upward trend, the base contains gravels and coarse sands while the top consists of silts; (b)–(d): well-preserved wood pieces, closely-spaced laminae and clastic flood deposits are highlighted with white lines in unit 2; (e) and (f): charcoal spots (white arrows) and flood layers (white lines) in unit 3; (g) and (h): homogeneous yellowish sediments and rusty spots in unit 4; (i): brown clays and silts with rusty spots in unit 5; (j): unit 6 with a black layer containing high concentrations of Zn and Pb. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Fig. 5. Bayesian age-model with 2σ error below 314 cm and tuned age-model (dash-dotted line) above 314 cm. Two calibrated ^{14}C ages in unit 1 were excluded due to the uncertainty of erosion (see 4.2 and 6.1). The triangles which represent time boundaries are derived from archeological evidence (BA = Bronze Age, IA = Iron Age, RP = Roman Period). The dot at 70 cm reflects peak coal mining activity upstream along the Meuse in the middle of 19th century (Stam, 2002). Average accumulation rates were derived for units 1–3, and for the overall clastic sequence. See lithological explanations for core units 1–6 in Fig. 2b.

the relative contribution of coarser grains in flood deposits. This means the FEI prepares channel-fill grain-size records for interpretation in terms of flooding phases (cf. Macklin et al., 2005), rather than for single event-magnitude quantifications. What to include and exclude in the coarse grain-class range is dependent on the GSDs, the EM-modelling results and laboratory observations.

In this study, we chose grain sizes between 31 and 105 μm as the flood indicator. The upper limit of 105 μm was selected due to its

proximity to 100 μm and ready accessibility in raw grain-size datasets. The coarse EM3 requires a higher flow-energy to be transported than the finer EM2 and therefore can be linked with peak discharge during floods, although the direct use of EM3 is inappropriate due to the ‘contaminated’ part at its coarse end (Fig. 6b–c). The mode of EM2 (20 μm) corresponds with the modes of the bulk GSDs (8–20 μm) in units 2–6 (Fig. 3). The finer EMs (EM1 and EM2) represent suspended and wash load populations, for which it is assumed that limited differentiation occurs between different flood magnitudes, as these fine particles are also in suspension during normal flow. Here, we chose the mode of coarse EM3 as lower boundary for reconstruction of the FEI as it can better represent the coarse component of floods. Therefore, we define the FEI as:

$$FEI = \left(\frac{V_{31-105}}{V_{0.12-105}} \right) \times 100$$

Where the V_{31-105} and $V_{0.12-105}$ represent the volume fraction with a grain-size range of 31–105 μm and 0.12–105 μm , respectively. By doing this, the FEI excludes the grain-size signals in the fraction above 105 μm . EM4 is not used due to its absence in most part of unit 2–6. FEI variation against core depth is shown in Fig. 6e. We used a locally weighted polynomial smoothing function to prevent undue emphasis being placed on single data points. In order to assess the reliability of FEI, we conducted a series of statistical analyses comparing the FEI with other paleoflood proxies, including median, mean, 90th percentile (P90) and the mean of sand (> 63 μm) fraction (MS).

6. Discussion

6.1. Holocene floodplain evolution

The Well-Aijen paleochannel contains a fairly complete continuous depositional record of Meuse flood deposits from ~8600 cal BP to modern times. This makes the study site very suitable for studying depositional changes and fluvial dynamics in the Lower Meuse catchment, as continuous records are extremely rare in such a dynamic fluvial system (Bos and Zuidhoff, 2015). The reason that the WA record is relative complete is that organic-rich sequences are very much stable sedimentary environments that offer the opportunity to derive continuous records that can be precisely dated (Goslin et al., 2018). Our chronology for the organo-clastic part of the sequence supports this (Fig. 5).

The distinct GSDs of unit 1 and unit 2 reflect a rapid change in sedimentary environment (Fig. 4a–b); which implies that the channel was abandoned over a short timescale. Vivianite was observed in the channel-bed deposits (unit 1), indicating a reducing environment

Fig. 6. The end-member modelling results and the FEI variation against core depth. (a) Median and mean r^2 on all grain-size classes as a function of the number of end-members. A four-end-member model is sufficient to explain > 85% of the total variance; (b) r^2 for each size class for end-member models with two to six end-members; (c) The modeled end-members (EMs) according to the four-end-member model (4EMM); (d) Proportional contribution of the EMs of the 4EMM; (e) FEI variation against the core depth based on the 4EMM. Horizontal dashed lines mark the unit boundaries. Shaded areas in (b) and (c) indicate the decreased r^2 at grain-size class larger than 100 μm (see 4.3).

during or after the deposition. After that, the abandoned WA paleochannel changed into a fluvio-lacustrine environment. At the very base of unit 1 ^{14}C results produced relatively old basal dates (Table 1) that imply a rather low accumulation rate (0.2 mm a^{-1}) for this unit (Fig. 5). Also, an exceptionally coarse sand bed (at 615 cm, Fig. 2c, 5) is interbedded within relatively finer-grained deposits. We suspect this coarse bed was formed during a short-lived flood event, which might have caused a hiatus or reworking of sediment. Therefore, we focused our interpretation on the part of the channel fill above 610 cm, which covers the time interval since ~ 8600 cal BP. This underlines that a risk may exist for ^{14}C dating when using organics buried in active channel bed deposits (Chiverrell et al., 2009; Howard et al., 2009).

The overall fine-grained deposits in core WA16 (Fig. 3) reflect that the flood water transporting capacity was limited at the core site. Hence, the well-preserved leaf remains and fresh wood fragments can be considered as originating from local vegetation. Even though the archeological excavation indicates that humans had already appeared at Well-Aijen during the Mesolithic and Neolithic (Bos et al., 2014; Bos and Zuidhoff, 2015), the impact of prehistoric people on the landscape and on river behavior was probably limited (Kalis et al., 2003). This is also evidenced by the high percentage of tree-pollen species and low proportion of agricultural pollen species (i.e., *Cerealia*) (supplementary material, Fig. S5). The moderate sedimentation rate in the organo-clastic part of the sequence (Fig. 5) resulted from the local

morphodynamics in a natural depositional setting rather than representing general floodplain accumulation rate (Toonen et al., 2012). From unit 2 to 3, the accumulation rate is nearly halved and the median grain size shows a minor decrease. This can be attributed to the disconnection from the later-formed active river channel and the gradual infilling of the abandoned channel (Toonen et al., 2012). The anti-phase variations of TOM (increasing trend) and core density (decreasing trend) in unit 2 (Fig. 2) indicate that the organic-matter input was dominant over the clastic-material input. This feature is different in unit 3, in which the reduced TOM and elevated core density are indicative of the dominant clastic sediment input. The transition from unit 3 to unit 4 features sediment-composition and color change, indicating the transition from an abandoned channel into a floodplain that only received sediment input during floods. If this change was caused by a depositional hiatus, we would expect sharp parameter changes at ~ 330 cm. However, neither the grain size in core WA16 nor the pollen data in WA11 show an apparent shift (Fig. S5, Fig. 2c, e-g). Thus, we think this transition reflects a continuous channel infilling rather than a hiatus.

In the Iron Age (unit 4, Fig. 5), the significant increase of agricultural pollen indicators (Fig. S5) indicates that humans have exerted profound influence on the landscape since then. At the same time, the river Meuse began to raise and widen its natural levees, the distal edge of which expanded across the study side (Bos and Zuidhoff, 2015). Although uncertainty exists in the age-depth model, and the

sedimentation rate could have fluctuated due to the migrating nature of the fluvial system (Toonen, 2015), the onset of Iron Age deposition (~ 2.7 ka BP) at ~ 314 cm and the polluted black layer at ~ 70 cm together constrain the overall accumulation rate to 0.9 mm a^{-1} for that interval (Fig. 5), which is significantly higher than the accumulation rate in the ^{14}C dated organo-clastic part of the sequence. Moreover, assuming the existence of a hiatus at ~ 220 cm (where the lithofacies changed) would give an even higher accumulation rate ($> 0.9 \text{ mm a}^{-1}$). Prolonged human influence in the catchment (i.e., soil erosion upstream) and floodplain from the Iron Age, Roman Period and Medieval Warm Period (De Moor et al., 2008) is considered to account for overall higher sedimentation rates at our site and in the wider floodplain of the lower Meuse (Pierik et al., 2017).

6.2. Evaluation of the FEI

We suggest that two conditions need to be fulfilled in order to use FEI appropriately in paleohydrological contexts. First, tectonic movement should have been minimal because this can change the river morphology and the connectivity between sediment traps and river channel. Second, channel migration should be limited. This is because large-scale migration can produce different index values even when floods have similar magnitudes. As stated before, these two prerequisites are satisfied in this study. In order to assess the performance of the FEI, results were statistically compared with direct grain-size descriptors (median, mean, P90 and MS) that have been used as proxies in flood magnitude reconstruction (Benedetti, 2003; Arnaud, 2005; Czymzik et al., 2013). Here, three subsets of grain-size data from unit 4, unit 2 and unit 5 were examined (Fig. 7). The grain-size composition in unit 4 represents a ‘clean’ dataset because there are neither concretions nor excess organic matter, which could cause biases in the dataset. Based on the EM-modelling results and sedimentological observations, the datasets from unit 2 and unit 5 were contaminated by concretions and/or organic remains. In these units, we expect that using the direct descriptors is inappropriate and FEI should outperform the other proxies.

In unit 4, the statistical analysis shows that there are good linear relationships ($0.77 < r^2 < 0.97$) between FEI and the other proxies (Fig. 7a). This means, when using a ‘clean’ grain-size dataset, all these parameters (FEI, median, mean, P90 and MS) are able to represent the coarse-grained population which links with paleofloods. In contrast, the correlations are poor in organic-rich unit 2 (Fig. 7b), with more scattered data points, as in the r^2 varies between 0.56 and 0.83. This indicates that either the FEI or the other paleoflood proxies have failed to work as paleoflood indicator(s). This is likely due to the incorporation

of organic remains in the grain-size composition (i.e., Toonen et al., 2015), and our laboratory observations support this. Therefore, we argue that a more reasonable approach in reconstructing past floods is to adopt a reliable grain-size composition rather than incorporating the problematic portion. Certainly, some GSDs may not have been contaminated but, when the problematic GSDs cannot be distinguished from the overall dataset, rejection of the problematic proportion is more logical. This principle should also be applied to unit 5, in which there are quasi-linear correlations ($0.83 < r^2 < 0.96$) between FEI and other proxies (Fig. 7c) regardless of the observed concretions. The correlations plausibly show that all these parameters work well. This is because the post-depositional Fe–Mn concretions were treated as ‘real particles’ by the laser-diffraction instrument during grain-size analysis. The scanning electron microscope images show that these fragile concretions have irregular shapes with sizes larger than $100 \mu\text{m}$ (Fig. S3). To what extent these concretions have affected the grain-size dataset is of great importance in FEI construction. In this case, the reason to exclude the grain-size range larger than $105 \mu\text{m}$ is that the particle size of the concretions is an order of magnitude larger than the mode of GSDs in units 2–6 (Fig. 3). Thus these concretions have undoubtedly enlarged the values of P90, median, mean and MS. Incorporating these problematic components can result in overestimation of the paleoflood magnitudes in paleodischarge reconstructions, as the concretions are much coarser than the sediment particles.

The formation of these concretions is associated with soil processes, including gleying phenomena. As these post-depositional processes are ubiquitous worldwide, including in monsoon and equatorial zones, we propose that grain-size datasets should be evaluated before being used in paleoflood reconstructions. The FEI method proposed here provides an approach to remove the effect of the post-depositional processes. Therefore, we think the selective use of grain-size datasets should be considered when the regular laboratory treatment is unable to produce a robust dataset due to the existence of Fe–Mn concretions and residual particulate organics. Based on the above, we think the method of constructing a FEI has potential to be popularized when encountering similar challenges elsewhere. By applying this method, the effect of contaminated portions in grain-size datasets can be minimized.

6.3. Holocene paleoflooding phases

FEI variation through the Holocene is illustrated in Fig. 8. The FEI record starts at 8.6 cal ka BP because of the uncertainty of the basal ^{14}C dates. The higher FEI highlights the flooding phases that are reconstructed from core WA16. The visually determined clastic flooding layers were plotted against the age axis with a scale showing the

Fig. 7. Statistical analysis of FEI with other paleoflood proxies: P90 = 90th percentile, Median = median grain-size, Mean = mean grain-size, MS = mean of sand fraction. Unit 4 represents sediments with a ‘clean’ dataset without Fe–Mn concretions and organic matter; while part of the samples from units 2 and 5 contain minor admixtures of organic remains and Fe–Mn concretions, after conventional sample pretreatment, respectively. The results show that there are good linear relationships between FEI and other proxies in units 4 and unit 5, and poor correlations in unit 2.

Fig. 8. The FEI variation through the Holocene. Grey line and dots are raw data; thick black line represents the smoothed curve to prevent undue emphasis being placed on single data points. Arrows above the curve indicate the increased flooding phases according to our sediment analyses. Bars along the age axis represent the thickness of visually determined fine-clastic flood laminae in core WA16. The FEI record starts at 8.6 cal ka BP due to the uncertain ages at the base of core WA16. The top part of the figure indicates the cultural and climatic periods: IA = Iron Age, RP = Roman Period, DA = Dark Ages, MWP = Medieval Warm Period, LIA = Little Ice Age, HCO = Holocene Climatic Optimum.

thickness (Fig. 8). We think the closely-spaced thin layers at around 440, 460, 500 and 520 cm (Fig. 2a) were formed during a series of short flood events. Therefore, these thin layers are integrated as single floods. These observed flooding layers correspond well with the increased flooding phases derived from FEI variation (Fig. 8). In the Mesolithic (11.6–6.8 cal ka BP) and Neolithic (6.8–4.0 cal ka BP), five multi-centennial flooding phases occurred; at c. 8500, c. 8000, c. 7600, c. 7000 and c. 5900 cal BP. The pollen data (i.e., occurrence of *Cerealia*, *Anthoceros punctatus*, *Rumex* and *Plantago*) reveal that humans began to alter the landscape by grazing and land reclamation at this stage (Bos and Zuidhoff, 2015). However, human activity in this period was still low and thus unlikely to have exacerbated the magnitude and frequency of floods.

In the time-period from 5 to 3 cal ka BP, the FEI is characterized by low values with small fluctuations, indicating the flood activity was low (Fig. 8). Meanwhile, the regional agricultural pollen species also declined (Fig. S5). This period followed the termination of the Holocene Climatic Optimum (9–5 ka BP), which featured relatively wet and warm climate conditions (Wanner et al., 2008). Intermittent glacial advances have been recorded in the Northern Hemisphere after the Holocene Climatic Optimum (Wanner et al., 2008 and references therein). The dryer and cooler climate phase during the Subboreal (5–2.5 cal ka BP) may have led to a low discharge regime in the Meuse, suppressing flooding phases and sedimentation rates. The role of human activity in changing the flooding regime is unclear.

In the late Holocene, the first high flooding phase occurred at c. 2800 cal BP (Fig. 8). This period corresponds with the early Iron Age. This phase may have resulted from the 2.8 ka climate anomaly, which was characterized by cooler and wetter climate conditions in Europe and North America (Van Geel et al., 1996, 1998). During the last two thousand years, the FEI index reveals an oscillating flooding magnitude that will have echoed through in an irregular sedimentation rate (Fig. 5 and Fig. 8). Based on the tuned age-model for the late Holocene deposition, the FEI record tentatively reveals that the last three flooding phase occurred at c. 2000, c. 700 and c. 300 cal BP (Fig. 8). The first of these coincides with the Roman Period (c. 12 BCE – 250 CE), and this identification agrees with stratigraphic observation, as the FEI values were derived from grain-size information at ~274 cm in core WA16, which corresponds with the onset of early Roman Period deposition (Bos and Zuidhoff, 2015). The latter two phases most likely coincide with the Medieval Warm Period (c. 950–1400 CE) and the Little Ice Age (c. 1400–1850 CE). During the Dark Ages (c. 200–800 CE), flooding was less severe. Previous studies support these interpretations. First, the wet

and warm summers that occurred during periods of Roman and medieval prosperity (Büntgen et al., 2011) may have contributed to intense floods. Second, widespread deforestation and population expansion in the Roman Period and the Medieval Warm Period stimulated soil erosion and further increased the volume of sediment input to the Meuse (Berendsen and Stouthamer, 2000; De Moor et al., 2008; Ward et al., 2008, 2009).

Notably, the FEI is higher and with larger amplitude in the late Holocene, indicating greater the flood intensity. The more fluctuating flooding regime and high accumulation rates (Fig. 5) indicate rather unstable hydroclimatic conditions in the late Holocene. However, to provide more accurate dating and resolve flooding phases for the late Holocene, it would be preferable to sample a paleochannel fill of younger age instead of a middle Holocene channel fill and a late Holocene floodplain overburden. Nevertheless, the FEI index shows considerable potential to resolve flooding phases both in subaqueous organic facies as well as in floodplain deposits, enabling cross-correlation between channel-fill and floodplain sites in lowland rivers such as the Meuse.

7. Conclusion

The Well-Aijen paleochannel provides a record of phasing in paleoflood activity that spans a large part of the Holocene. The lower organo-clastic part of the sequence (units 2–3) represents a fluvio-lacustrine swamp environment, which formed after channel abandonment in the early Holocene. Natural depositional processes dominated over limited human interference. The upper clastic part of the sequence (units 4–6) formed since the Bronze Age, and was affected by enhanced human interference, which accelerated sediment accumulation rates.

This study indicates that post-depositional Fe–Mn concretions and organic residuals affect the quality of grain-size records, and further influence EM modelling results and data interpretation in paleohydrological studies. Combining laboratory observations with EM modelling results, we argue that the problematic portions of grain-size distributions should be rejected. By excluding the ‘contaminated’ portions of our grain-size data, we constructed a Flood Energy Index (FEI) to highlight flooding phases through the Holocene for the Lower Meuse. We think this method is capable of minimizing unwanted signal contamination and has potential to be a standard method when similar challenges are encountered elsewhere. We suggest that the quality of a grain-size dataset should be critically assessed prior to using it for flooding reconstruction.

Multi-centennial periods with increased floods were identified at c. 8500, c. 8000, c. 7600, c. 7000 and c. 5900 cal BP within the fluvio-lacustrine part of the meuse sequence (early–middle Holocene). The low sedimentation rate and low flooding activity in the Subboreal is attributed to a cooler and dryer climate anomaly after the Holocene Climatic Optimum. In the late Holocene, the first flooding phase at c. 2800 cal BP can be linked to the 2.8 ka climate anomaly. The last three inferred flooding phases occurred at c. 2000, c. 700 and c. 300 cal BP and coincide with the Roman Period, Medieval Warm Period and Little Ice Age, respectively. Rapid sediment accumulation rates and more severe floods reveal increased fluvial instability in the late Holocene, also indicating that humans had begun to have a profound influence on fluvial dynamics.

Data availability

Data related to this article can be requested from corresponding author.

Declaration of conflicting interests

The author(s) declared no potential conflicts of interest with respect to the research, authorship, and/or publication of this article.

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Appendix A. Supplementary data

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