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A novel method to evaluate the effect of a stream restoration on the spatial pattern of hydraulic connection of stream and groundwater

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SUMMARY

Stream restoration aims at an enhancement of ecological habitats, an increase of water retention within a landscape and sometimes even at an improvement of biogeochemical functions of lotic ecosystems. For the latter, good exchange between groundwater and stream water is often considered to be of major importance. In this study hydraulic connectivity between river and aquifer was investigated for a four years period, covering the restoration of an old oxbow after the second year. The oxbow became reconnected to the stream and the clogging layer in the oxbow was excavated. We expected increasing hydraulic connectivity between oxbow and aquifer after restoration of the stream, and decreasing hydraulic connectivity for the former shortcut due to increased clogging. To test that hypothesis, the spatial and temporal characteristics of the coupled groundwater-stream water system before and after the restoration were analysed by principal component analyses of time series of groundwater heads and stream water levels. The first component depicted between 53% and 70% of the total variance in the dataset for the different years. It captured the propagation of the pressure signal induced by stream water level fluctuations throughout the adjacent aquifer. Thus it could be used as a measure of hydraulic connectivity between stream and aquifer. During the first year, the impact of stream water level fluctuations decreased with distance from the regulated river (shortcut), whereas the hydraulic connection of the oxbow to the adjacent aquifer was very low. After restoration of the stream we observed a slight but not significant increase of hydraulic connectivity in the oxbow in the second year after restoration, but no change for the former shortcut. There is some evidence that the pattern of hydraulic connectivity at the study site is by far more determined by the natural heterogeneity of hydraulic conductivities of the floodplain sediments and the initial construction of the shortcut rather than by the clogging layer in the oxbow.

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1. Introduction

In the past decades there has been an increasing effort on research and practice according to the restoration of rivers and their floodplains. The main reasons for that are the valuation of river ecosystems as place for species conservation and habitat diversity, recreational and aesthetic purposes, flood protection, enhancing the potential of contaminant deposition and nutrient degradation (Bernhardt et al., 2007; Kondolf et al., 2007; Hester and Gooseff, 2010; Pander and Geist, 2013; Schirmer et al., 2013). This is also reflected in a growing body of legislative directives (Pander and Geist, 2013; Schirmer et al., 2013), e.g. the EU Water Framework Directive demands a good chemical and ecological status of groundwater and surface water (European Commission, 2000). The chemical and ecological status of surface waters is impacted by the adjacent connected aquifer and vice versa. Hence both waters have to be considered when assessing water qualities of either of them. Nevertheless, in river restoration practice the measures most often focus solely on surface waters, whereas the connection of the river and the groundwater below the river bed and the adjacent floodplain is often neglected (Boulton, 2007; Boulton et al., 2010; Hester and Gooseff, 2010).

Previous studies identified the transition zone between stream water and groundwater, the hyporheic zone, as highly relevant for mass exchange, residence time of water and substances in the stream or in the sediment, the chemical and metabolic turnover

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and in general as crucial for water quality (Brunke and Gonser, 1997; Sophocleous, 2002; Boulton, 2007). The spatial extent of the hyporheic zone is mainly determined by two drivers, the hydraulic gradient between the river and the groundwater and the sediment structure (Kasahara et al., 2009), especially the permeability of the stream bed and aquifer sediments (Woessner, 2000; Kalbus et al., 2009). Therefore, clogging of the stream bed, i.e. the sealing of the stream bed with sediments of very low hydraulic conductivity, has been identified as major problem for exchange of surface water and groundwater and the related ecological functions of the hyporheic zone (Sophocleous et al., 1995; Brunke and Gonser, 1997; Sophocleous, 2002).

Fluxes are spatially and temporally heterogeneous due to the spatial heterogeneity of hydraulic conductivity of the sediments and spatial and temporal variability of hydraulic gradients (Woessner, 2000; Malard et al., 2002; Krause et al., 2011; Binley et al., 2013). Different methods are available to estimate fluxes across the interface in a river-groundwater system (Kalbus et al., 2006). Selective approaches, such as vertical temperature profiles (e.g. Schmidt et al., 2006; Anibas et al., 2009), heat pulse sensors (e.g. Lewandowski et al., 2011), hydraulic gradients (e.g. Krause et al., 2012) or seepage meters (e.g. Rosenberry and LaBaugh, 2008) are able to monitor the flux over time for a specific point, but it is not possible to draw conclusions for a whole river section. A method to capture larger areas is distributed temperature sensing (DTS) (e.g. Selker et al., 2006a,b; Krause and Blume, 2013), which is able to detect spots with intense groundwater ex- and infiltration. Another option is to use natural or artificial tracers to determine the degree of interactions (e.g. Négrel et al., 2003; Cox et al., 2007). Beside the different measuring techniques, numerical modelling was often used to examine groundwater-surface water interactions (e.g. Nützmann et al., 2013). One advantage of the latter method is that it does not have to be restricted on the hyporheic zone itself and can include the adjacent floodplain. All methods have in common the large temporal and monetary effort and in most case the restriction to certain areas or certain seasons. Furthermore, most approaches rely on information about hydraulic conductivity or related parameters, which are hard to estimate and result in uncertainties.

Another method which directly estimates the spatial distribution of the hydraulic properties of the sediments is hydraulic tomography (e.g. Yeh and Liu, 2000; Zhu and Yeh, 2005). In a network of spatially distributed wells the response to an artificial pressure signal induced by a pump at one well is recorded at all other wells. The procedure is repeated by sequentially circulating the pump through the other wells. With packers each well can be segregated in different depth intervals and by circulating the pump through the depth intervals at all wells the depth integrated estimation of hydraulic properties can be enhanced to a 3D-tomography (Yeh and Liu, 2000; Cardiff and Barrash, 2011). With an inverse model the spatial distribution of the hydraulic properties of the aquifer is estimated from the interplay of all the observed hydraulic head series. Up to now most of the non-numerical hydraulic tomography studies aimed to map the small scale variability of the hydraulic properties of the sediments on the lab to plot scale and used artificial pressure pulses (Yeh et al., 2009; Cardiff and Barrash, 2011). Recently there were attempts to extend hydraulic tomography to the groundwater basin scale and to use natural pressure signals, e.g. river stage fluctuations as signal (Yeh et al., 2009).

Similarly we use in the present study river stage fluctuations as natural pressure signals and study their propagation in the aquifer before and after a stream restoration measure to study groundwater-stream water interactions. Time series of hydraulic head reflect effects of different causes, like river stage fluctuations, groundwater recharge, precipitation, evapotranspiration, measurement errors, etc. (Yeh et al., 2009). In contrast to the aforementioned approaches we decomposed the hydraulic head series into independent components using a principal component analysis in order to disentangling the different effects.

The study was conducted at a section of the river Spree and its floodplain in the east of Berlin. Here an island is formed by an artificial stream channel (shortcut) and an oxbow. As restoration measure the shortcut was detached from the river at its upstream end, the former oxbow was reconnected to the stream and its clogging layer was excavated. The site was equipped with 15 groundwater observation wells, 2 river stages and 2 hyporheic wells, where data loggers measured every hour two years before and after the restoration. Please note that in our study we do not focus on the small scale heterogeneity of the hydraulic properties of the sediments as it would be important e.g. for estimations of the flowpaths of contaminants or the study of biogeochemical processes in the hyporheic zone. Instead we investigated the effect of the removal of the clogging layer and the change of the river course on the hydraulic connection of stream and groundwater.

Our analysis is based only on hydraulic head data and does not require any additional information. Please note that therefore our analysis is restricted to the transmission of pressure waves. We use the term "hydraulic connectivity" in contrast to the broader concept of "hydrologic connectivity" which is defined by Pringle (2001) as "water-mediated transfer of matter, energy, and/or organisms within or between elements of the hydrologic cycle" to account for that. The presented approach does not allow direct conclusions on related mass fluxes, flowpaths and water exchange rates (Lewandowski et al., 2009; Page et al., 2012). Instead the permeability for pressure signals is a necessary prerequisite for the exchange of mass fluxes. Thus, the hydraulic connectivity between the observation wells can be used as proxy for the relative differences in effective hydraulic conductivity of the floodplain sediments between the observation wells. With this integrative measure the problem of measuring the small scale variability of hydraulic conductivity in the floodplain is avoided.

To that end, we followed the approach presented by Lewandowski et al. (2009) for a time period where the river section was not restored and applied a principal component analysis on time series of groundwater heads and stream water levels. Based on the findings of Lewandowski et al. (2009) we hypothesized that (1) due to the restoration the hydraulic connectivity between the oxbow and the nearby groundwater will increase and that (2) in the shortcut the river bed will be clogged due to the reduced stream velocity, resulting in decreasing hydraulic connectivity between the shortcut and the adjacent groundwater.

2. Methods

2.1. Study Site

The Freienbrink site is situated in the floodplain of the lowland river Spree about 30 km east of the centre of Berlin (52°22′06″N, 13°48′25″E). The discharge of the river Spree is regulated by the Weir Grosse Tränke located 10 km upstream and varies usually between 5 and 20 $m^3\,s^{-1}$ (Nützmann et al., 2013). At the site, a straight, artificial channel (shortcut) and an old meander (oxbow) form an artificial island. The shortcut was constructed in the 1960s to increase the flow velocity within the river and lower the surrounding groundwater table for agricultural purposes. In the first two years of the monitoring period, the shortcut served as the main stream channel and the oxbow was nearly completely blocked at its upstream end with a dam (Fig. 1). Some pipes inside the dam that connected the oxbow to the main stream were blocked with fine sediments. Therefore, the flow velocity in the

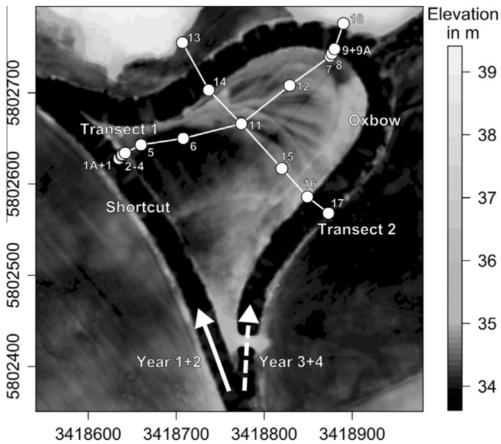

Fig. 1. Elevation map based on a LIDAR-Scan from 3rd December 2009 with 1 m gridsize and 0.3 m resolution for altitude in projection ETRS89 UTM Zone 33. First transect of groundwater observation wells in west-northeast direction and second transect in northwest-southeast direction. At both ends of the first transect there was a water level gauge situated in the stream (1A + 9A). The filled arrow marks the main stream flow in the first two years of the study, the dashed arrow the main stream flow in the third and fourth year. The flow through the other reach was blocked in both situations.

oxbow was almost zero and as a consequence an organic silt layer had developed in the oxbow with thicknesses varying spatially between zero and more than 1 m (Nützmann and Lewandowski, 2009). The organic silt layer had a hydraulic conductivity k_f between 10^{-6} and 10^{-5} ms^{-1} and an effective porosity n_e of 0.5. Compared to the surrounding aquifer with k_f of 10^{-4} to $5 * 10^{-5}$ and n_e of 0.15–0.2 (Nützmann et al., 2013), the organic layer can be attributed as a clogging layer.

At the end of the second year of measurements, in October/November 2008, the clogging mud in the oxbow was excavated with a suction dredger. The oxbow was reconnected as main stream channel on the 27th November 2008. After the reconnection of the oxbow, a dam was built at the upstream end of the shortcut and closed on the 9th December 2008. Hence, there was a switch in the river course from the shortcut to the oxbow (Fig. 1). In the subsequent two years of the monitoring period the former shortcut was blocked with the dam at the upstream end and the former oxbow was reactivated. Only at high water levels the dam was overflown.

The river intersects an unconfined aquifer of about 20 m thickness. The floodplain consists mainly of medium to fine grained sandy sediments of glacial and fluvio-glacial origin (Lewandowski et al., 2009). In the northern part the floodplain adjoins a steep hillslope to a 5 m higher plateau (Lewandowski et al., 2009). The topography of the floodplain is the result of morphological work of the river (Fig. 1). The western part of the floodplain is characterized by a depression with fine grained sand and silt, where the groundwater level repeatedly exceeded the surface (Fig. 1). Starting from the middle of the floodplain, a ridges-swales structure with about 0.4 m difference in elevation established due to the meandering oxbow. Here, coarser sediments with intermediate organic layers are present (Pöschke et al., 2014).

Most of the time groundwater is exfiltrating into the stream. A mean groundwater exfiltration rate of 233 $L\,m^{-2}\,d^{-1}$ with a groundwater flow velocity between 10^{-7} and 10^{-6} ms^{-1} was estimated by Nützmann et al. (2013). Maximum lateral infiltration of river water into the aquifer is less than 4 m (Lewandowski et al., 2009). Velocity of pressure wave propagation (celerity) from the stream into the aquifer was found to be about 1550 md^{-1}, thereby three to four orders of magnitude higher than the velocity of groundwater mass flux (Lewandowski et al., 2009). For a general elaboration on velocity of pressure waves (celerity) vs. velocity of water particles (mass fluxes) in hydrology, see McDonnell and Beven (2014).

2.2. Water level measurements

Groundwater heads at 15 groundwater observation wells, hydraulic head at 2 hyporheic wells and stream water levels at 2 river stages were measured at hourly intervals with data loggers (Aquatronic, Kirchheim/Teck, Germany, ±1 mm) along two transects. Most of the groundwater wells had filter screens of 2 m length and the upper end of the filter screen at approximately 25–70 cm below the ground, with the exception of observation well No. 11 with 9 m filter length. The first transect consisted of 10 shallow groundwater wells crossing the island in southwest-northeast direction from the shortcut to the oxbow (Fig. 1). Measurements were conducted from 1st October 2006 to 1st October of 2010. Additionally, at both ends of the first transect surface water level was measured in the oxbow and in the shortcut (gauges no. 1A and 9A). Adjacent to the stream water gauges two hyporheic wells no. 1 and 9 were installed that screened at 0.5 to 1.5 m below the riverbed at gauge no. 1A and between 1.5 and 2.5 m at gauge no. 9A (Lewandowski et al., 2009). The second transect consisted of five additional groundwater observation wells that crossed the island in northwest-southeast direction and the first transect at observation well no. 11 (Fig. 1) and measured groundwater level from 12th November 2007 to 1st October of 2010.

2.3. Pre-processing of the data

Short data gaps of up to seven hours, which resulted from the biweekly water quality sampling and maintenance of the pressure transducers and loggers, were interpolated using natural splines (less than 0.5% of the readings per year at all wells). Apparent offsets between subsequent periods that could have been due to inexact reinstallation of the devices were adjusted with the offset of the linear extrapolation before and after the corruption. Lengthening of the cable of the devices in the wells were identified by comparison with biweekly manual readings of water levels and groundwater heads and corrected with simple linear regression.

Longer gaps in the data or periods with distorted data were interpolated with the best multiple linear regression model (package 'leaps' in R (R Core Team, 2013)). Two abrupt shifts in the water level of the oxbow compared to the water level in the shortcut and the nearby groundwater wells at 27th November 2008 12:00 and 9th December 2008 15:00 were attributed to the reconnection of the oxbow and the later decoupling of the shortcut with the new dam and hence were not corrected. At maximum, correction comprised 3026 consecutive readings, that is, 34.5% of the readings of the fourth year at groundwater well no. 5. Second most affected was well no. 11 with 538 consecutive readings (6.1%) in the third and 421 consecutive readings (4.8%) in the second year. At another ten events correction was up to 250 consecutive readings or 2. 9% of the readings per year.

2.4. Principal component analysis

Time series of groundwater and river water level show a very close linear correlation ($r = 0.98$), which indicates a good hydraulic connectivity between both water bodies. Nevertheless, there is no perfect correlation. Aside from measurement noise and possible artefacts, these differences presumably have to be ascribed to different factors that affect different sites to different degrees. Among these, pressure signals induced by river water level fluctuations as well as groundwater recharge, depending on vegetation and the soil properties of the overlying vadose zone, are considered to play a major role (Lewandowski et al., 2009). Thus every water level series can be regarded to be the result of different superimposing effects. Our analysis aimed at extracting the effect of river water level fluctuations from these mixed signals by following the approach presented by Lewandowski et al. (2009).

Firstly, the deviation from the mean of groundwater heads and river water level was calculated for each time step. Afterwards, each of these residual time series was normalized to zero mean and unit variance to ensure equal weighting. Then, a principal component analysis (PCA) was applied to the prepared data. PCA performs an eigenvalue decomposition of a data matrix, yielding a series of independent components. We considered these components as depicting different drivers of hydraulic head fluctuations.

In order to analyse for long-term shifts PCA was performed for single hydrologic years separately (October throughout September yielding 8760 readings at each observation well and 8784 readings in the second year, respectively). For the first transect four years of water level measurements were available since 2006, whereas measurements at the second transect started one year later. To enable comparison between the two transects, a joint analysis with data from both transects was performed for the second, the third and the fourth year. Since the data acquisition at the second transect started at the 24. November 2007 (7476 readings at each observation well), the second year does not cover the full hydrological year 2007/2008. The documentation of the changes in the system in annual resolution is a compromise between comparability with the preceding study of Lewandowski et al. (2009), precision of the effect of the restoration we wanted to measure and comparability of the different PCAs among each other. Furthermore, the PCA was performed for each quartile of the hydrologic year to account for inter-annual variability.

Loadings on a component are the expression of a component at the different sites. They were calculated as Pearson correlation coefficients of the z-normalized residuals of the water level series and the values (scores) of the component (compare Lewandowski et al., 2009). The stability of the loadings in each observation year was estimated with the mean of the loadings of the 4 quartiles of the hydrologic year and their corresponding confidence intervals. The areal expression of the first component was estimated as thin plate splines – a kriging variant implemented in the package 'fields' in R (R Core Team, 2013) – based on the mean of loadings of the 4 quartiles of the third observation year, as it was the year with the lowest inter-annual variability.

All the statistics and calculations were done with the free software package R, version 3.0.1 (R Core Team, 2013).

3. Results

3.1. Characteristics of water level dynamics and floods

The mean amplitude of groundwater level fluctuations was approximately 0.9 m for the first three years of the monitoring and approximately 1.3 m in the last year of the study. The groundwater observation wells, except the two hyporheic wells, have a mean distance of groundwater level to the surface of approximately 60 cm with a standard deviation of approximately 20 to 25 cm. The mean difference in water level from one hour to the next hour was between two to three mm. The mean difference in water level among the groundwater observation wells on the island was in the magnitude of a few cm. The mean groundwater level increased from the first year to the second year by approximately 15 cm and from the third year to the fourth year by approximately 13 cm. The increase in mean groundwater level and amplitude in the last year is due to several flood events: one week of flooding in December 2009, three weeks in January 2010 and 1 month of flooding from the mid of August to the mid of September 2010. For more details including graphs of the water level series characteristics please see Lewandowski et al.'s (2009) examination of the first year of the observation period.

3.2. Principal component analysis

For the data set of the first transect the first principal component depicted 70% of the total variance in the first year and 70%, 65% and 63% in the subsequent years (Fig. 2). Please note, that the range of loadings on the first component describes the *relative* differences in the correlations of the first component with the original water level series at the observation wells because the PCA was applied on the z-normalized residuals of the water level series. Hence, we refer to loadings close to one as high loadings and loadings close to minus one as low loadings.

At the stream water gauges no. 1A in the shortcut and no. 9A in the oxbow loadings were very high on the first component throughout the observation period with the exception of lower loadings for no. 9A in the third year (Fig. 3). At the groundwater observation wells next to the shortcut (no. 2, 3, 4) loadings were very high on the first component and exhibited approximately the same loadings compared to the gauges in the stream (Fig. 3). The hyporheic well no. 1 below the stream bed depicted substantial lower loadings than observation wells no. 2 to 4 in the second year, and slightly lower loadings in the first and fourth year (Fig. 3). Only in the third year loadings at observation well no. 2 were slightly lower than loadings at hyporheic well no. 1.

Groundwater wells next to the oxbow showed very low loadings with slightly increasing loadings on the last meter to the oxbow from observation well no. 7 to 9 (Fig. 3). Only in the fourth year loadings were substantially higher at observation wells no. 8 and 9 than in the years before. Observation well no. 12 was on the same level as the gauges next to the oxbow. Observation wells no. 5, 6 and 11 in the middle part of the island showed decreasing loadings along the first transect from the shortcut to the oxbow. The loadings in this middle part showed the highest variability between the years and also within their quartiles. From the first to the second year all loadings in the middle part increased and decreased thereafter until the fourth year.

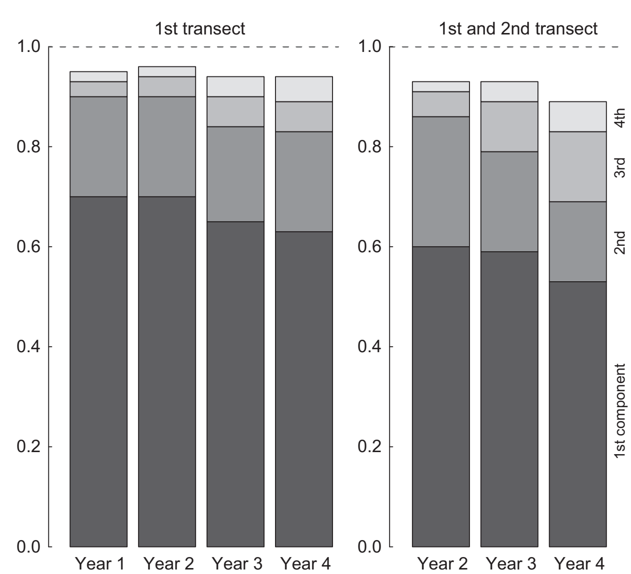

Fig. 2. Ratio of overall variance explained by the first four principal components of the PCA of the data set of the first transect (left) and of the PCA of the joint data set from both transects (right).

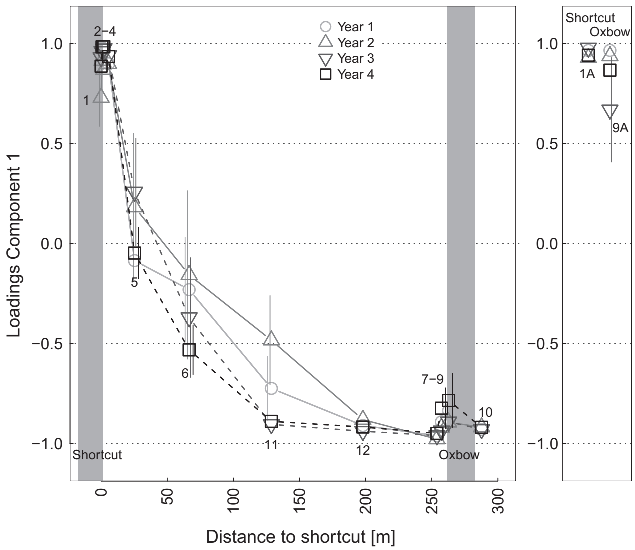

Fig. 3. Loadings of the groundwater observation wells of the first transect on the first principal component vs. distance to shortcut (left) and of the stream water gauges (right). Mean of loadings of the 4 quartiles of the hydrologic year (Oct.–Sept.) and their corresponding confidence intervals. Only confidence intervals >0.2 are shown. On the left panel the grey bars indicate the position of the stream.

For the joint data set (water level data from both transects) the first principal component depicted 60% in the second year and 59% and 53% in the third and fourth year (Fig. 2). The main pattern in the first transect with high loadings next to the shortcut, and decreasing loadings to the east was similar to that of the separate analysis based on the first transect. The loadings of the two variants were correlating in the three common years with an r^2 of at least 0.97.

In the northern part of the second transect groundwater well no. 13 was loading constantly low and groundwater well no. 14 constantly loading high on the first principal component over the entire observation period (Fig. 4). In the southern part of the second transect groundwater wells no. 11, 15, 16 and 17 were jointly shifting from slightly positive loadings in the second year towards negative loadings in the third year and stayed on this level in the fourth year (Fig. 4). Inter-annual variability was prominent in the second year for wells no. 11, 15, 16 and 17 and in the last year for wells no. 13, 16 and 17 (Fig. 4). There was no regular seasonal pattern along both transects. The overall spatial pattern remained relative stable throughout the observation period (Figs. 3–5).

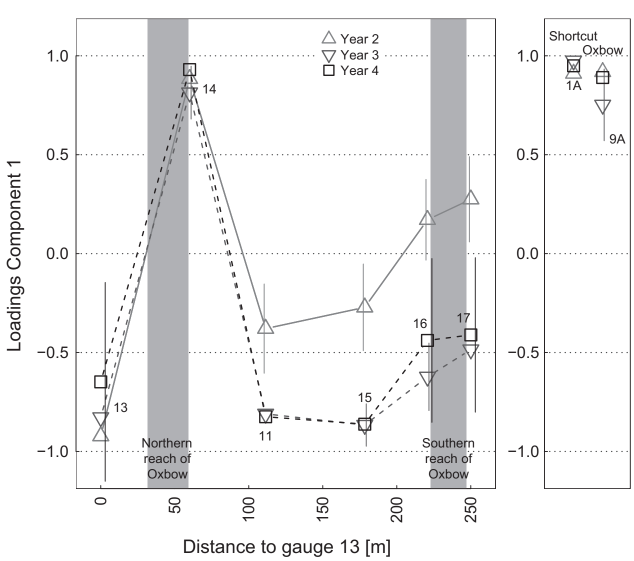

Fig. 4. Loadings of the groundwater observation wells of the first principal component of the second transect vs. distance to the northern end of transect two at observation well 13 (left) and of the stream water gauges (right). Mean of loadings of the 4 quartiles of the hydrologic year (Oct.–Sept.) and their corresponding confidence intervals. Only confidence intervals >0.2 are shown. On the left panel the grey bars indicate the position of the stream.

4. Discussion

4.1. Interpretation of the first component

In a more detailed analysis of the first year of the dataset, Lewandowski et al. (2009) identified the first component as dampening and delay of the fluctuations of the water level in the river. We follow their interpretation as loadings of the stream water level gauges exhibit loadings close to one in all of our analyses (Figs. 3 and 4). Thus the first component explains almost 100% of the deviation of the stream gauges from the mean behaviour of all observation wells. This implies that processes that impact the groundwater head in the wells could hardly have any additional effect on the stream water level gauges. Thus, the loadings on the first component can be used as a quantitative measure for the hydraulic connectivity between river and groundwater.

High positive loadings on the first component imply that the respective time series of the groundwater head deviates from the spatial mean in the same way as the stream water level gauges. That allows the conclusion that groundwater head at the respective site is strongly affected by the stream water level fluctuations. In contrast, high negative loadings point to a weak impact *relative* to the other observation wells, and zero loadings to intermediate effects.

Our prior assumption was that the observed water level dynamics at the single sites are a mixture of different superimposing effects. In fact the first component explains only 53–70% of the spatial variance, indicating that other factors have substantial effects on the observed groundwater heads as well. To investigate in particular the hydraulic connection of the river and the groundwater by analysing the original time series, e.g. using cross-correlation, would therefore not have been satisfactory.

4.2. Spatial patterns of hydraulic connectivity

The oxbow and the shortcut have been connected at their downstream end throughout the entire study period and exhibited the same water level. The slightly lower loadings at hyporheic well no. 1 compared to observation wells no. 2 to 4 on the western end of the first transect are interpreted as slightly lower hydraulic conductivity of the river bed compared to the river bank. Along the first transect (Fig. 1) the impact of river level fluctuations decreased with distance from the stream up to well no. 11 and remained approximately stable for the eastern half of the transect (Fig. 3). As the loadings of the groundwater wells in the eastern part of the island are substantial lower than the loadings of the stream wells and the loadings of the groundwater wells in the western part of the island, we conclude that groundwater wells in the eastern part of the first transect are hydraulically disconnected from the oxbow relative to their hydraulic connection to the shortcut (Fig. 5). Only groundwater wells no. 8 and 9 adjacent or even underneath the oxbow are slightly more influenced by the water level fluctuations in the oxbow (Fig. 3). The same pattern was already found by Lewandowski et al. (2009) who suggested that the slight increase is either due to the heterogeneity of the aquifer or due to a very low - but not zero - hydraulic connectivity of the clogging layer.

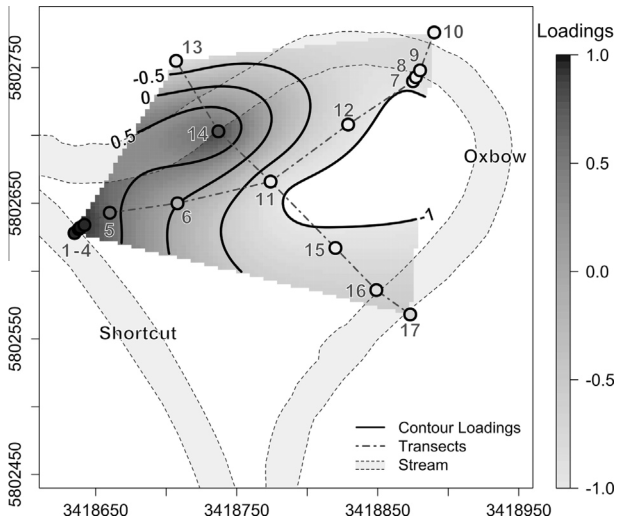

Fig. 5. Spatial interpolation of loadings of the first component based on the mean of the loadings of the 4 quartiles of the third observation year (Oct.–Sept.). Projection is ETRS89 UTM Zone 33. Values < −1 are artefacts and excluded.

Results from the second transect show high connectivity with the oxbow only for well no. 14 in its northern reach, which is also not affected by the stream channel restoration, since the pattern stays stable over the four years (Fig. 4). This specifies the hypothesis of areas of relatively better hydraulic connection in the north-western part of the island (Fig. 5). Observation wells no. 16 and 17 close to the southern reach of the oxbow show higher hydraulic connectivity with the stream than the observation wells no. 11 and 15 in the middle of the island and also than the observation wells in the north-eastern part of the first transect. This specifies the area of lowest hydraulic connectivity in the north-eastern part of the island, which is the area where the inner bank of the meander is and was located. Since the flow velocity is lowest at this position during the formation of the meander, material of finer grain sizes has been accumulated in that area. This explanation is also supported by investigations of Pöschke et al. (2014), who conducted a ground penetrating radar survey to characterize the sediment composition. They could show that the north-eastern part of the island is characterized by finer grain sizes close to the meander (well no. 7) in comparison to the western part, where the sediment adjacent to the channel bed is much coarser.

The low hydraulic connectivity of the observation wells no. 10 and 13 in the hillslope of the adjoining plateau is not due to the hydraulic gradient induced by the topographic gradient, because the first component captures the pressure-wave signals of the river.

4.3. Temporal variation of hydraulic connectivity

The analysis was performed for single years separately in order to capture changes of connectivity. Loadings of the stream gauges were constantly close to one as it is mandatory for the interpretation of the first component as influence of river level fluctuations on groundwater levels. The lower loading at stream gauge no. 9A in the third year (Fig. 3) could be ascribed to the reconnection of the oxbow and the construction of the new dam in the shortcut.

Loadings of observation wells close to the Spree river at the western end of the first transect, and of groundwater well no. 14 close to the oxbow in its northern reach were close to one in all years of our study, pointing to a constant high connectivity between river and aquifer. Clogging of the stream bed in the shortcut due to the reduced stream velocity was not detected during the two years after the restoration. We assume that the first clogging sediments that settled throughout the first year have been removed due to the intense floods in the fourth year. Nonetheless we expect clogging of the shortcut in successive years.

In the central and south-easterly part of the island wells no. 5, 6, 11, 15, 16 and 17 exhibit substantial shifts between single years, although the general spatial pattern remains approximately the same (Figs. 3 and 4). There the loadings are highest for the second year and smallest in the fourth year, whereas those of the third year are similar to those of the fourth year. We did not find an unequivocal explanation for that temporal shift. Such systematic shift can hardly be explained by artefacts of the measurements. In addition, a corresponding change of the properties of the aquifer can be excluded. All of these wells are located close to the same former river bed, although this had not been intended. The associated interbedding of different substrates such as gravel, silt, former clogging layers and peat might still affect groundwater flowpaths and lead to a closely coupled behaviour (Nützmann and Lewandowski, 2009).

Wells no. 8 and 9 close to the north-easterly reach of the oxbow exhibit almost no shift in time. Loadings are close to minus one for the first three years, indicating low hydraulic contact with the oxbow. This suggests that the clogging layer is not the only reason for the lower hydraulic connectivity of the hyporheic well no. 9 below the oxbow compared to the hyporheic well no. 1 below the shortcut. Instead we suggest that the effective hydraulic conductivity of the sediments around the shortcut in the north western part of the island is higher than of the sediments around the oxbow (Fig. 5). This can be attributed to the construction of the shortcut. The shortcut itself and the surroundings are situated in sandy sediments, which were dislocated artificially. In contrast, the sediments around the oxbow are characterized by inclined layers of sand and organic material of the recent and former point bars. On the other hand the loadings at well no. 14 are close to one for the whole observation period of the second transect, suggesting a strong impact of natural spatial heterogeneity of effective hydraulic conductivity of the floodplain sediments (Fig 4).

Only in the fourth year loadings at wells no. 8 and 9 are slightly higher (but not significantly higher), pointing to increased, although still fairly low connectivity (Fig. 3). Likewise a minor increase of loadings between the third and fourth year is observed at wells no. 13, 16 and 17 close to the oxbow (Fig. 4). This is consistent with our prior assumptions of increasing connectivity due to stream channel restoration. However, this is more a little piece of evidence rather than a proof and the development has to be checked in the following years. Anyhow, the effect is much weaker than assumed.

5. Conclusions

The data driven PCA approach used in this study is easy to apply and requires only time series of hydraulic heads in the aquifer and in the stream. It splits up the water table dynamics into independent components which represent different drivers of the hydraulic head dynamics in the system and the spatial expression of those drivers can be analysed. In the present study, the first component captures the propagation of river stage fluctuations into the aquifer and can therefore be used to describe the hydraulic connectivity of groundwater and stream water. Areas of relative low and high hydraulic connectivity could be identified, although it is not possible to quantify mass fluxes and water exchange rates. Despite the small differences in water table of a few cm along the groundwater transects and intense periods of flooding in the fourth year, it is possible to derive throughout the observation period spatial and temporal consistent patterns of one distinct hydraulic process, namely the propagation of the river water table fluctuations, on the hectare-scale (Figs. 3–5). One benefit of the presented approach is that hydraulic head devices are long established and often already permanent part of hydrologic monitoring networks. Once installed the operating effort is very low. The hydraulic head data are usually measured with relative high temporal resolution of minutes to hours and stored with data loggers.

The presented approach can be used to optimize the monitoring network and monitor further developments of the groundwater-stream water system (Page et al., 2012), as in our example after a stream channel restoration. It can also provide a framework for further modelling of groundwater-stream water interactions (Lewandowski et al., 2009), help to identify the hydrological active functional properties on the landscape scale (Lischeid et al., 2010), regionalize the different hydrological contributions to the aquifer dynamics (Longuevergne et al., 2007), or be used as explorative data analysis tool to develop hypothesis such as the spatial patterns of areas of relative low and high effective hydraulic conductivity of floodplain sediment as done in the present study.

Our results demonstrate that even two years after the restoration hydraulic connectivity of stream water and the groundwater next to the new main stream channel (oxbow) had not reached the level of connectivity next to the old artificial stream channel (shortcut). We identified three factors to explain this finding. That is (1) that the change in stream velocity and sedimentation rate in the oxbow and in the shortcut due to the switch of the main stream channel might need more time to effectively change the stream bed permeability in such a scale that it alters the spatial pattern of hydraulic connectivity in the floodplain. The two other options are that hydraulic connectivity around the shortcut is in the long run higher than around the oxbow due to the spatial heterogeneity of the sediments in the floodplain, whether the heterogeneity is (2) natural or (3) induced by the digging of the artificial shortcut directly into the sandy floodplain sediments. To distinguish between this three factors and their relative contribution to the observed spatial pattern of hydraulic connectivity further analysis of the spatial distribution of the sediments around the stream channels and their hydraulic conductivities, as well as the further monitoring of the development of hydraulic connectivity after the restoration is necessary.

With this study the importance of comprehensive monitoring of restoration measures even several years after the restoration and the explicit identification of specific properties of the system to be restored is underpinned (Kondolf, 1995). Option one of our suggested set of influencing factors exemplifies that restoration of connectivity of groundwater and surface water is a process that might last for several years after the initial restoration of the surface water, while on the surface the restoration might appear already successfully completed. The proposed differences in effective hydraulic conductivity of the alluvial sediments in different parts of the floodplain (option two and three) highlight the importance of pre-studies that identify the specific natural conditions before the restorations (Kondolf and Micheli, 1995; Woolsey et al., 2007; Pander and Geist, 2013). Focusing only on the hydrogeomorphology of the surface waters or individual target species in restoration practice might be insufficient to restore the functional properties of the river ecosystem (Kondolf et al., 2006; Woolsey et al., 2007). If groundwater-stream water interactions are understood as substantial features of the system to be restored, this has to be taken into account in the restoration and monitoring practice, as well as in the evaluation of restorations.

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