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A hierarchical approach on groundwater-surface water interaction in wetlands along the upper Biebrza River, Poland

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Abstract

Groundwater-surface water exchange studies on natural rivers and wetlands dominated by organic soils are scarce. We present a hierarchical approach to quantitatively investigate and interpret groundwater-surface water interaction in space and time by applying a combination of different field methods including piezometer nests, temperature and seepage measurements. The numerical 1-D heat transport model of STRIVE is used in transient mode to calculate vertical fluxes from thermal profiles measured along the upper Biebrza River, Poland over a period of nine months. The calculated fluxes show no clear spatial pattern of exchange fluxes unless an interpolation of the point estimates on a reach scale is performed. Significance of differences in net exchange rates versus morphological features are investigated with statistical tests. Time series of temperature and hydraulic head of the hyporheic zone are used to estimate the temporal variability of the groundwater-surface water exchange. Seepage meter measurements and slug tests were used for cross validation of modelled fluxes. Results show a strong heterogeneity of the thermal and physical soil properties along the reach, leading to a classification of these parameters for modelling purposes. The groundwater-surface water exchange shows predominantly upward water fluxes, however alternating sections of recharge exist. The exchange fluxes are significantly different dependent on the position of the river in the valley floor and the river morphology where fluxes are more dependent on hydraulic gradients than on river bed conductivity. Sections of higher fluxes are linked to the vicinity of the morainic plateau surrounding the rivers alluvium and to meanders, indicating that a perspective on the fluvio-plain scale is required for interpreting the estimated exchange fluxes. Since the vertical component of the exchange fluxes cannot explain the magnitude of the change in river discharge, a lateral flow component across the alluvial plain has to be responsible. The hierarchical methodology increases the confidence in the estimated exchange fluxes and improves the process understanding, however the accuracy of the measurements and related uncertainties remain challenges for wetland environments.

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1 Introduction

Groundwater-surface water exchange processes take place in the hyporheic zone, the area of saturated sediments beneath and beside streams, rivers and wetlands where groundwater and surface water is actively mixed (Brunke and Gonser, 1997; Boulton et al., 1998; Hayashi and Rosenberry, 2002; Sophocleous, 2002). The processes observed in the hyporheic zone are characterized by significant variability in both time and space (Triska et al., 1993; Constantz, 1998) and by relative strong biogeochemical activity (McClain et al., 2003; Smith, 2005). The complexity and uncertainty surrounding river research and management reflects the need to develop new or more refined tools and methods (Vaughan et al., 2009).

The purpose of this article is to quantify the hyporheic exchange fluxes in space and time for a section of the Biebrza River, Poland. A combination of different methods (Hunt et al., 1996; Weight and Sonderegger, 2001; Kalbus et al., 2006) is applied, including the use of hydraulic gradients, seepage meters and most prominent the thermal method. With this approach we overcome limitations of each individual field method and provide a robust first level investigation for wetland environments. For the understanding of eco-hydrological characteristics of wetlands we need to reliably identify and quantify the relevant groundwater-surface water interaction processes and vice versa. Therefore we hypothesize that the magnitude and variation of fluxes in the hyporheic zone can be examined on a local scale (determined by first order factors like composition of the riverbed, bathymetry and position across the riverbed) and extrapolated to a reach scale. Riverine wetland functioning is however seen as dependent on the groundwater-surface water interaction at the fluvio-plain scale; consequently we assume that fluxes are dependent on second order factors such as e.g. topography, morphology, climate and hydrogeology.

The interaction processes between groundwater and surface water are based on the concept of connectivity, an emerging topic both in hydrological (Bracken and Croke, 2007; Lexartza-Artza and Wainwright, 2009) and ecological sciences (Pringle, 2001;

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Tetzlaff et al., 2007; Boulton et al., 2010). Hydrological connectivity refers to physical linkages of water in different catchment compartments such as rivers and adjacent wetlands (Bracken and Croke, 2007). Connectivity allows the exchange of water, solutes and dissolved matter and as a consequence energy transfer across the riverine landscape (Ward, 1997), determining hydrogeochemical contact times, reaction rates, retention and feedback processes (Fisher et al., 1998; McClain et al., 2003; Buis et al., 2008). Ecological landscape connectivity is defined as a functional relationship among habitat patches owing to the spatial contagion of biotopes and responses of organisms to the structure of the landscape (With et al., 1997). Groundwater-surface water interaction thus constitutes an important link between the river, its wetlands and the surrounding catchment.

The supply of exfiltrating groundwater and the presence of shallow groundwater tables is essential for the maintenance of groundwater dependent wetlands and their habitat connectivity (Succow and Joosten, 2001; Ovaskainen and Hanski, 2004). The vegetation in such environments is often found to depend on the quality, quantity and the pattern of river discharge and groundwater-surface water interaction (Wassen and Joosten, 1996; Batelaan et al., 2003) on a local or reach scale. Virtually all European wetlands are constantly influenced by land use changes, land reclamation, succession processes and habitat fragmentation (Tockner and Stanford, 2002; Hooftman et al., 2003; Smolders et al., 2010) leading to environmental degradation processes like desiccation, acidification or eutrophication (Lamers et al., 2002; Smolders et al., 2006; van Diggelen et al., 2006). Reliable estimates of groundwater flow into a wetland and the understanding of interactions with other system compartments like surface water, soil matrix and organisms play a key role in evaluating the structure of stream systems (Sophocleous, 2002), the sustainability of their wetlands and the conservation of biodiversity (Schot and Winter, 2006).

Various national and international regulations like the European Water Framework Directive (European Commission, 2000) mandate the protection of linked groundwater-surface water systems. To comply with these regulations integrated hydrologic and ecosystem models (National Research Council, 2004; Smith, 2005; Buis et al., 2008) are vital for the development of environmental standards and management schemes for the maintenance, protection and restoration of river catchments. Since the assessment of fluxes across the groundwater-surface water interface is important for the examination of related biogeochemical processes their reliable quantification is an important component of these models.

Groundwater-surface water exchange processes are plagued with heterogeneity and scale problems (Woessner, 2000; Becker et al., 2004; Kalbus et al., 2008); quantification on a local and reach scale is challenging hydrologic sciences since decades. Uncertainties are related to variations of the hydromorphological and physical properties of the riverbed, the riparian zone and the underlying aquifer (Conant, 2004; Fleckenstein et al., 2006; Schornberg et al., 2010). A framework for improved estimation methods for exchange processes is therefore required.

Temperature is a dominant moderator of almost all biological and chemical processes. This makes it an important ecological parameter; however it can also be used as a natural tracer for the detection of groundwater-surface water exchange (Anderson, 2005; Kalbus et al., 2006; Constantz, 2008). The method has proved to be accurate and reliable (Lautz, 2010; Ferguson and Bense, 2011), not least because gathering of thermal data, parameter estimation, establishment of model boundary conditions and calibration are fairly simple (Anibas et al., 2009). Different methodologies have been applied (Anderson, 2005; Kalbus et al., 2006) but most commonly exchange rates have been quantified by inverse modeling of measured temperature profiles (Schmidt et al., 2006; Anibas et al., 2009). Various studies were performed on sites where the riverbed is composed of sand or gravel (Conant, 2004; Anibas et al., 2011); applications on sites dominated by peat soils are not known to the authors. The application of the thermal method represents a point estimate (Becker et al., 2004); the spatial interpolation of distributed sets of these estimates however is described in literature (Schmidt et al., 2007; Anibas et al., 2011).

We gathered river bed temperature profiles in "roaming surveys" (Keery et al., 2007) and continuously monitored river and riverbed temperatures to determine groundwater-surface water interaction by applying the physically based numerical heat transport model STRIVE (STReam RIVer Ecosystem; Buis et al., 2008). The model is used to calculate vertical flux rates in a spatial and temporal distribution by transient thermal simulations. Using GIS, this technique shows exchange patterns and allows the calculation of net mass fluxes across the interface between groundwater and surface water along a river section on a reach scale. Statistical analysis shows influences of geo- and hydromorphology and riverbed heterogeneity on groundwater-surface water exchange on a fluvio-plain scale (Woessner, 2000; Vaughan et al., 2009).

The use of the thermal method however has a limited temporal resolution; methods based on hydraulic head using standpipes, piezometer nests and boreholes (Cey et al., 1998; Baxter et al., 2003) can determine exchange fluxes with a high temporal resolution. We installed a series of piezometers to detect groundwater-surface water interaction by analyzing time series data of hydraulic and temperature gradients. The piezometers also were used for the determination of hydraulic conductivities of the riverbed (Lapham, 1989) by performing slug tests (Fetter, 2001) and by combining head gradients and simulations of the STRIVE model.

Seepage meters finally offer the possibility to measure the exchange flux directly (Lee, 1977), but also show uncertainties in the estimated fluxes related to the technical operation in the field (Murdoch and Kelly, 2003). We used seepage meter for a cross-validation of the results of the thermal and the head based methods.

2 Field site

The study area is situated along the Biebrza River (22°30′–23°60′ E, 53°30′–53°75′ N, Fig. 1) in the Podlaskie Voivodeship, Poland, around 230 km north east of Warsaw. A small part of the catchment area of 7057 km^2 is located in Belarus. The Biebrza River, a right sided tributary of the Narew River, comprises a river reach of 170.6 km

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with an average discharge of 39.2 $m^3 s^{-1}$ at its outlet. The Biebrza River is one of the few natural lowland river systems of its size in Europe (Pałczyński, 1984; Wassen and Joosten, 1996). With an area of 592 km^2, occupying most of the rivers alluvial flood plains, the Biebrza National Park forms a wetland of worldwide significance protected by the United Nations (Ramsar Convention Secretariat, 2008) and by the European Union as a Natura 2000 habitat (European Commission, 1992). The site is the habitat of valuable river marshes and peat lands including highly threatened plant and animal species including the orchid *Liparis loeselii* and the Aquatic Warbler (*Acrocephalus paludicola*) respectively. As a hydro-ecological system the Biebrza River finally serves as a reference area for restorations of managed wetlands (Wassen et al., 2006).

Geomorphologically the Biebrza Valley is an extensive depression formed during the last glaciations filled with thick deposits of fluvioglacial sands and gravels covered by a variety of organic soils. The Biebrza Valley is divided into three subbasins (Żurek, 1984), the Upper Basin, the Middle Basin, and the Lower Basin, characterized by different hydrological regimes (Byczkowski and Kiciński, 1984) and groundwater-surface water interaction (Okruszko et al., 2006; Chormański et al., 2009).

The Upper Basin, reaching from the springs of the Biebrza River to the village of Sztabin is a 48 km long 1–3 km wide valley (Fig. 2) covering 846 km^2. Topographic elevation of the Biebrza River valley varies between 110 and 130 m a.m.s.l. (above mean sea level), while those of the adjacent morainic plateau and the outwash plain varies between 130 and 180 m (Żurek, 1984). While narrow and showing relative steep gradients close to the spring, the Upper Basin soon becomes a meandering stream reach where the Biebrza River flows through a flat ice-marginal valley. The morainic uplands adjacent to the alluvial plain function as a regional groundwater recharge area and drain the surrounding plateau and the outwash plain towards the river (Pajnowska and Wiencław, 1984). Groundwater, passing through aquitards and hydrogeologic windows is seeping out in the Biebrza valley. The peat lands therefore are mostly groundwater fed, however during spring freshets surface water also infiltrates the alluvia.

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Crossed by ditches of abandoned land reclamation systems, the valley is filled with deposits of varying peat soils of thicknesses of 2 to 5 m. Together with the underlying fluvioglacial gravels and sands the peat layer forms an unconsolidated aquifer. Glacial tills (Pajnowska and Wiencław, 1984; Ber, 2005) however locally separate the sand and gravel layers creating confined aquifers of varying extent resulting in a complex local hydrogeology. The hydrogeological base of the Biebrza catchment consists of Tertiary marls at approximately 0 to −40 m a.s.l.

The Biebrza River catchment is located in the subcontinental/subboreal climate zone with a yearly average temperature of 6.8 °C. The average annual precipitation ranges from 550 to 700 mm yr^{-1}, the evapotranspiration between 460 and 480 mm yr^{-1} (Kossowska-Cezak, 1984). Given the low population density of the area, the current land cover in the morainic uplands consists mainly of arable land and remnants of the natural oak-beech forests. Low lying areas of the catchment are cultivated in an extensive manner as meadows and pastures.

The hydrological regime of the river in the Upper Basin is characterized by a sequence of flood events which are limited in extent by the geomorphologic boundaries of the floodplain. This is the slope crack between valley wall and valley floor indicated by the dashed line in Fig. 3. Floods occur regularly after snowmelt in early spring. The late spring, early summer periods are characterized by low flow whereas summer rain storms occasionally create flood peaks. During the dry periods most of the Biebrza valley is groundwater fed. The spring inundations however are only partly caused by river flooding; groundwater seepage and snowmelt water are present across 80 % of the valley width (Chormański et al., 2011). At the mouth of the Upper Basin at Sztabin the average flow is 4.83 m^3 s^{-1} (Chormański and Batelaan, 2011). At field location No. 4 (Fig. 2) the average discharge during the examined period was with an estimated value of 0.31 m^3 s^{-1} still much lower.

The characteristic low-productive fens are widely abundant (Oświt, 1994; Wassen and Joosten, 1996) but the succession of shrubs and forests is progressing in the alluvial plains (Pałczyński, 1985). Fen-bog transition is stimulated by enhanced infiltration of local precipitation following a subtle lowering of the surface water level of the Biebrza River. Land use changes caused by mechanization and rural exodus also lead to shrub encroachment (Wassen and Joosten, 1996) resulting in increased evapotranspiration.

The examined river stretch is located between the villages Sopoćkowce (Fig. 2, No. 1) upstream and Rogożynek (Fig. 2, No. 4) at the downstream end of the section. We performed most of the presented measurements however between point No. 2 (Stary Rogożyn) and No. 4. The length of the river section is 5670 m, and the average absolute elevation of the water level at No. 2 and 4 is 119.9 and 119.4 m a.s.l. respectively. The average slope of the riverbed was estimated as 0.23 ‰; the river has a width of about 6–8 m and an average depth of 1.1 m along the examined reach. The Biebrza River is free flowing along the entire reach; the river channel is characterized by a rectangular cross-section with steep banks. During low flow in summer the Manning coefficient for this river stretch is about 0.12 (De Doncker et al., 2009). The riverbed is composed of peat of varying consistency; the banks mostly are covered with reed plants.

3 Measurements

3.1 Temperature stick

We established 38 measurement points (Fig. 3), designated as points 200-300 between the villages of Stary Rogożyn (Fig. 3, No. 2) and Nowy Rogożyn (Fig. 3, No. 3) and points 300-400 between Nowy Rogożyn and Rogożynek (Fig. 3, No. 4) to gather temperature profiles of the river bed. Field measurement campaigns of 2 consecutive days were performed by examining points 400–301 the first and 300–200 on the second day. The measurements were executed on 12–13 October, 17–18 November 2007, 5–6 March and 15–16 June 2008 with the so-called T-stick (Fig. 4b) instrument (Anibas et al., 2009, 2011). Additionally, several points were measured on 10 November and 8 December 2007. Using a Topcon GMS-2 GPS receiver with

EGNOS differential correction a relocation of the measurement points was possible with an accuracy of 1 m. The measured temperature profiles consisted of measurements at the groundwater-surface water interface (i.e. 0.0 m) and at 0.1, 0.25 and 0.5 m depth in the riverbed (Fig. 4b). If possible a measurement at the deepest reachable point was taken (in average this was 0.83 m).

3.2 Piezometer nests

At four locations (Figs. 2 and 3) along the river stretch, No. 1, 2, 3 and 4, piezometer nests (Fig. 4a) were installed. At No. 4 two different installations were placed, Fig. 4a and b respectively. Two, three or four piezometer pipes furnished with a filter of 0.15 m were placed at different depth (between 0.15 and 1.20 m) in the riverbed and equipped with temperature (StowAway® TidbiT®, Onset Computer Corporation, Bourne, MA, USA) and/or Diver® temperature and hydraulic head data loggers (Schlumberger Water Services, Delft, The Netherlands) to continuously measure head and thermal gradients in the riverbed. The piezometer nests No. 2, 3 and 4 were furthermore measuring river water levels and temperatures. Raising and falling head slug tests were performed at the piezometer nests No. 2 and 3.

3.3 Seepage meters

Four self-made seepage meters, metal and plastic barrels cut in half of 0.27 and 0.56 m in diameter were pushed into the sediment of the river bed in a zone of around 50 m^2 at Rogożynek (Fig. 3, No. 4). From 16–20 June 2008 nine measurements were performed by collecting during two hours seepage in plastic bags (volume 0.5 l). Pre-filled bags (0.1 l) were used to avoid anomalous short-term influx and to reduce the bag resistance (Murdoch and Kelly, 2003). Average values obtained from all four seepage meters were used.

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4 Methodology

Since the wetlands around the Biebrza River are protected, field methods which are not intrusive or immersive are preferred for the investigation. We applied a set of different field methods to quantify the groundwater-surface water interaction including methods based on hydraulic head, slug tests and seepage meters. The main method applied however is the thermal method (Anderson, 2005; Kalbus et al., 2006; Constantz, 2008).

In the surficial zone of the subsurface the temperature shifts seasonally and diurnally, influenced by the heating and cooling of the land surface. During the summer months the groundwater temperature is generally cooler than stream temperature whereas in winter it is generally the opposite. We assume the groundwater to flow according to hydraulic gradients, hence heat is solely transported by advection and conduction through the system influencing the temperature distribution in the porous media. Nowadays temperature can be measured rapidly as sensors are technically simple, cheap, widely available, and they can be handled easily.

Based on Stallman (1965) and Lapham (1989) the one-dimensional, vertical, anisothermal transport of liquid and heat through homogeneous, porous media is formulated as:

$$\lambda_{\mathrm{e}} \frac{\partial^2 T}{\partial z^2} - v_z\, c_w\, \rho_w \frac{\partial T}{\partial z} = c\, \rho \frac{\partial T}{\partial t} \tag{1}$$

where λ_{e} is the effective thermal conductivity of the soil-water matrix in $\mathrm{J\,s^{-1}\,m^{-1}\,K^{-1}}$, T the temperature at point z at time t in °C, c_w the specific heat capacity of the fluid in $\mathrm{J\,kg^{-1}\,K^{-1}}$, ρ_w the density of the fluid in $\mathrm{kg\,m^{-3}}$, v_z the vertical component of the groundwater velocity in $\mathrm{m\,s^{-1}}$, c the specific heat capacity of the rock-fluid matrix in $\mathrm{J\,kg^{-1}\,K^{-1}}$, and ρ the wet-bulk density in $\mathrm{kg\,m^{-3}}$. The first term of the left hand side of Eq. (1) represents the conductive and the second term the advective part of the heat transport. For convenience we express the vertical groundwater velocity in $\mathrm{mm\,d^{-1}}$. A positive sign stands for water moving from the surface into the hyporheic zone (i.e. groundwater recharge or losing stream reach) and negative sign represents

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water moving from the hyporheic zone into the river (i.e. groundwater discharge or gaining stream reach).

The thermal method is an indirect method; the data measured in the field must be processed with a heat transport model in order to derive quantitative estimates of the flow velocity or flux. We apply inverse thermal modeling in which the calculation of vertical groundwater fluxes is achieved by solving Eq. (1) with transient boundary conditions. A vertical 1-D heat transport model STRIVE (Buis et al., 2008; Anibas et al., 2011) is used, based on the ecosystem modeling platform FEMME (Soetaert et al., 2002). The heat transport module obtains a best model fit by changing the value of the vertical groundwater velocity (v_z) and minimizing the difference between the measured and simulated temperature distributions by user defined internal integration and fitting routines (Soetaert et al., 2002).

4.1 Thermal model

The STRIVE model for the Biebrza River is descritizised as a vertical, one-dimensional, heterogeneous, saturated soil column of 5.0 m length and composed of 100 layers. The spacing of the model layers follows a sinusoidal function, providing layer thicknesses of 0.001 m at the upper and lower boundary to reduce discritization errors close to groundwater-surface water interface, while the thickness of the layers is increasing towards the center of the model domain to 0.08 m.

A continuously measured surface water temperature data set (i.e. solid line in Fig. 5) forms the upper boundary of the model domain. The lower boundary is defined as a constant temperature at 5.0 m depth (i.e. dashed-dotted line), where it is assumed that no significant changes in temperature occur over time (Anibas et al., 2009). One temperature profile measured with the T-stick, indicated by crosses, is used to initialize the model (i.e. the profiles of 10 or 11 October 2007), whereas the other three, at some points four T-stick measurements, are used to fit the modelled temperature distributions. With STRIVE's VODE (variable-coefficient ordinary differential equation) numerical integration routine two hourly output values were created. The dotted lines

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in Fig. 5 show simulated temperatures of the respective depths of the T-stick measurements for the best model fit. The result of the simulation represents an integration of the flux over the given simulation period.

The described model setup was also applied with temperature time series data from piezometers. The higher amount of data points used to fit the model allows reducing the simulation period and thus a higher temporal resolution.

A time series of surface temperature measurement was used for each of the 38 measurement points to serve as upper model boundary. In STRIVE this boundary and the uppermost measurement point (i.e. at 0.0 m, Figs. 4b and 5) should have identical values. Since there were only three different time series of surface water temperatures available for the entire river section the time series were linearly interpolated to fit with the corresponding T-stick measurement.

According to pedological information derived from soil maps (Banaszuk, 2000) and information collected from drillings vertically heterogeneous soil profiles were assigned for the model. At Rogożyn Nowy (Fig. 3, No. 3) a stratigraphy consisting of peat up to 2.2 m depth with a sandy soil layer below (Fig. 4a) was found and used for all but one measurement point. At Rogożynek (Fig. 2, No. 4) information from installed piezometer nests was available. A soil column consisting of a sandy surficial layer of 0.35 m thickness followed by a peat layer till 2.20 m depth and a sandy layer up to 5.00 m depth was defined.

According to Stonestrom and Constantz (2003) and Anderson (2005) λ_e for saturated sands ranges usually between 1.4 and 2.2 $\mathrm{J\,s^{-1}\,m^{-1}\,K^{-1}}$, whereas for peat λ_e varies between 0.4 and 0.7 $\mathrm{J\,s^{-1}\,m^{-1}\,K^{-1}}$. As λ_e varies less than an order of magnitude for all soils found in river beds, λ_e values usually can be taken from literature. This is a significant advantage compared to methods based on Darcy's law as λ_e is equivalent K_v, the hydraulic conductivity, which varies over several orders of magnitudes (Chen, 2000). In order to handle the heterogeneities along the examined field location three representative sets of physical-thermal parameters (Table 1) were estimated and assigned to the 38 measurement points. The thermal characteristics were determined

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based on physical properties of peat of the Biebrza River valley (Churski and Szuniewicz, 1994; Gnatowski et al., 2010), as well as literature including Farouki (1986), Peters-Lidard et al. (1998), Schwärzel et al. (2002) and Côté and Konrad (2005). Alternatively, uniform sets of parameters were used for simulations.

4.2 Conceptual model

In Fig. 6 we present a conceptual figure for the estimation and understanding of groundwater-surface water interaction. Figure 6a shows the field measurements representing point estimates valid on a "local scale" not bigger than 1–10 m along and across the river reach. First order factors including riverbed bathymetry, the composition of the riverbed and the position of the measurement across the river influence the vertical fluxes, indicated by arrows of various sizes. Spatially distributed point estimates along and across a river reach can be interpolated leading to Fig. 6b, where on a "reach scale" spatial relationships and net exchange rates between the hyporheic zone, the river bed and river can be examined. The spatial distribution of exchange fluxes and their patterns are indicated in grey scales. For a meandering river reach converging flow lines cause higher exchange fluxes at the convex banks of meanders and relative low exchange fluxes in the concave banks since the flow lines are diverging. To interpret the derived flux pattern it is thus necessary to investigate the system in a wider context, the "sub-basin" or "fluvio-plain" scale (Fig. 6c), since the quantity of exchange fluxes is also dependent on second order factors like topography, climate, hydrogeology, hydromorphology and vegetation. These features are indicated in Fig. 6 by meanders, the changing distance between the river and he slope crack, the different hydrogeological layers and their spatial abundance with respect to the river course and changing vegetation patterns across the alluvium and the morainic plateau respectively. The slope crack is defined as the break of the slope between valley wall and valley floor The combination of all these features influences the groundwater-surface water interaction via the riverbed as in the center of the alluvium for example it is expected that the groundwater discharge is lower then in the vicinity of the slope crack

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(van Loon et al., 2009). A better understanding of the hydro-ecological functioning is finally achieved when the determined results are used to update and further improve the investigative and modeling efforts of the system.

5 Results and discussion

5.1 Heterogeneity of the riverbed

Figure 5 is representative for the temperatures measured and simulated along the examined reach of the Biebrza River. Exemplarily the temperature profiles measured between 12 October 2007 and 16 June 2008 of two measurement points, 308 (Fig. 5a) and 302 (Fig. 5b) are shown together with corresponding surface water temperature time series. Since the temperatures are clearly different between the two points, it is obvious that strong temperature variations exist between the 38 T-stick measurements along the examined river reach. On 15–16 June 2008 for example maximal temperature differences of 7.86 °C and 6.14 °C were detected for the measurements at 0.0 m and 0.5 m depth respectively. Minimal spatial differences were measured on 5–6 March 2008 with values of 4.17 °C at 0.0 m depths and 2.62 °C at 0.5 m. Measured thermal gradients between 0.0 and 0.5 m had a range of 0.06 °C to 5.77 °C, whereas the measurement campaign of 12–13 October 2007 showed the lowest gradients with an average of 1.45 °C. The campaign of 17–18 November 2007 indicated the maximum gradients with depths (i.e. 3.37 °C). The minima and maxima of the measured surface water temperature time series within the simulation period were 3.75 and 16.53 °C for point No. 2 and –0.14 and 21.25 °C for point No. 4 respectively. The average surface water temperature of the winter season of 2007–2008 was 4.41 and 1.99 °C for No. 2 and No. 4 respectively.

Alongside with the thermal pattern the T-stick measurements revealed strong spatial variations in the physical consistency of the peat soil in the riverbed. Therefore we assume that the variation in temperature indicates, beside differences in groundwater

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fluxes, spatial heterogeneity of soil physical properties between the measurement points. A classification was achieved by a simple manual and visual examination point for point, repeated at every measurement campaign. Especially two peat types were distinguished, one showing a soft, loose structure whereas the other is fairly compact. Together with the underlying sand they are designated as “soil profile I” and “soil profile II” respectively. At point 400, where the river approaches the morainic upland, the riverbed becomes sandy and a more heterogeneous stratigraphy is present, e.g. “soil profile III”. This peat was assigned with similar physical values as soil profile I. A few measured profiles showed intermediate characteristics of soil profiles I and II; they were either classified according to one which best fit the model or average values of simulated fluxes were used for further analysis. For the different peat soils parameter sets were assumed as summarized in Table 1.

Soil profile I is characterized by a dark, black colour and a muddy consistency. Often no clear interface between surface water and riverbed is present; the region up to around 0.10–0.15 m depth the peat behaves like a suspension with a gradually decreasing porosity. Since the interface is not well determined it is difficult to define the absolute position of the temperature measurement. The pedological map of the Biebrza National Park indicates that this soil is predominantly composed of “reed peat” (Banaszuk, 2000). Temperature measurements of soil profile I indicate highly dampened temperatures with depth and flat thermal gradients as can be seen from measurement point 308 (Fig. 5a). In contrast to experiences with sandy soils, where a similar thermal pattern indicates high discharge fluxes, peat soils must be assigned with low thermal conductivity, high heat capacity and porosity values to get an acceptable model fit. Consequently by applying the respective parameter values of Table 1 these locations eventually show quite low flux estimates.

In contradiction to soil profile I, profile II is characterized by a stable, compact consistence of the river bed with a clear interface between the riverbed and the surface water. According to the pedological map (Banaszuk, 2000) this soil type is associated with “moss-sedge peat” or “alder swamp peat”. The temperature variations over time

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and depth are, compared to soil profile I, much stronger (Fig. 5b), indicating different fluxes and different soil properties. For soil profile II a higher thermal conductivity and a lower heat capacity have obviously to be applied. The final values (Table 1) were established by manual calibration runs of the transient thermal model in STRIVE leading in general to higher flux estimates than for soil profile I. This difference is underlined by statistical tests (Kolmogorov-Smirnow and Mann-Whitney U tests ($N = 38$; Level of significance $p = 0.05$). This correlation is not observed in case uniform thermal and physical properties are assumed for all measurement points, which leads to rather uniform flux estimates along the reach. This indicates that the estimation and classification of thermal properties of peat soils on a local scale is important to be able to correctly observe and interpret spatial relationships on the reach and fluvio-plain scale.

The stratigraphy of the riverbed influences the estimated fluxes when the soil parameters change relatively close to the groundwater-surface water interface. Test runs with STRIVE showed that the influence of the sand layer at a depth of 2.2 m below the peat is limited since no measurements have been performed at these depth and the exchange of thermal energy at this depth is relatively low.

Viewed in the broader context (i.e. fluvio-plain scale; Fig. 3) the fluvial plain in the upper part of the section has a constant width of around 367 m. From point 208 till point 303 it is widening up to a width of 777 m. Between point 304 and 310 the alluvium is steeply narrowing again and the width remains around 289 m until the lower end of the section. With differing distances from the slope crack the two soil profiles indicate a lateral heterogeneity in pedology. Soil profile I is in average farther away from the right slope crack of the valley and is found closer at the left side, whereas for soil profile II this is the opposite. The pedological map (Banaszuk, 2000) also shows different soil composition in the center of the floodplain and towards the left side of the alluvium. This finding however could not be supported with the Kolmogorov-Smirnow and a Mann-Whitney U tests ($N = 34$, $p = 0.05$). Slug tests performed at piezometers across the right side of the alluvium (Fig. 3) indicate a decrease in horizontal hydraulic conductivity K_h between the slope crack and the river course. The values decrease slightly from

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$0.65\,\mathrm{m\,d^{-1}}$ to $0.58\,\mathrm{m\,d^{-1}}$ across the alluvium towards the river, whereas in the riverbed the value drops to $0.10\,\mathrm{m\,d^{-1}}$.

5.2 Spatial variation

Conant (2004) and Anibas et al. (2011) suggest that strong temperature differences of spatially distributed measurement points indicate a heterogeneous pattern of groundwater-surface water exchange. Figure 7 shows the results of the STRIVE simulations using the parameters of Table 1 as bar graph for the 38 measurement points. The colours of the bars indicate the different soil profiles I, II and III in blue, red and green respectively similar to the dots in Fig. 3. The point estimates of the exchange fluxes at these locations are integrated over the examination period of 9 month and hence can be seen as average fluxes over the given time. The maximum flux is observed at point 210 with an exfiltration of $-37\,\mathrm{mm\,d^{-1}}$ and the minimum exfiltration rate is $-6.3\,\mathrm{mm\,d^{-1}}$ at point 311. Point 215 shows an infiltration rate of $4.8\,\mathrm{mm\,d^{-1}}$. Point 209 and 320 belong to soil profile I and show high discharge values, whereas points 205 and 300 (i.e. soil profile II) show relatively low discharge values. This is in contradiction to the general significant difference found between the exchange values of type I and II soils, and indicates that the spatial exchange pattern is not only depending on the composition of the riverbed. An average Root Mean Square Error (RMSE) between the measured and the simulated temperatures of 0.44 °C has been obtained. The RMSE for soil profile I is with 0.39 °C lower than of soil profile II with 0.50 °C.

A thermal steady state analysis as presented by Anibas et al. (2011) using the measurements of 5–6 March 2008 shows according to Spearman rank R and Gamma tests ($N=38$; $p=0.05$) a significant correlation between the spatial pattern of the steady-state and the transient simulations.

Transient heat transport simulations applying uniform physical parameter for the 38 measurement points showed however, a rather uniform distribution of fluxes along the river course. Furthermore, the model fit, obtained for this analysis, was all but

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satisfactory (e.g. RMSE values of more than 1.5 °C where common). From Fig. 8 it is obvious that the flux estimates along the river are fairly heterogeneous, at first instance no clear patterns are visible. We used a multilog radial bases function (Surfer 8.04, Golden Software, 2003) for the spatial interpolation of the locally determined point estimates (Fig. 6), where the parameter R^2 was set as 1800. An anisotropy ratio of 2.5 and an angle of 20° were used to represent the orientation of the river course. Figure 8b shows the range of the interpolated flux estimates with a maximum exfiltration of $-29.4\,\mathrm{mm\,d^{-1}}$ and a maximum infiltration of $4.7\,\mathrm{mm\,d^{-1}}$. An average flux of $-10.4\,\mathrm{mm\,d^{-1}}$ was calculated leading to a (vertical) net exchange rate of $-5.44\,\mathrm{l\,s^{-1}}$ along the entire river section of 5670 m. Given an average discharge of $0.31\,\mathrm{m^3\,s^{-1}}$ this is 1.8 % of the average surface water discharge at point No. 4. Compared with a study of the Belgian Aa River (Anibas et al., 2011), a significantly bigger stream, the Biebzra gains 0.32 % per km, which is considerably less than the Aa with 0.42 % per km. Both are low land rivers, but the different hydrogeology, peat versus sandy riverbed at the Aa, determines strongly these differences.

Notice that some parts of the river course show infiltrating and others discharging conditions. The section between the measurement points 301–211 is slightly infiltrating whereas just further upstream, between the points 210–206, the highest exfiltration rates are estimated. 91.4 % of the interpolated river surface shows exfiltrating characteristics, whereas 8.6 % indicates recharge. The exfiltrating zone shows a net flux of $-5.54\,\mathrm{l\,s^{-1}}$ whereas for the infiltrating zone a flux of $0.10\,\mathrm{l\dot{s}^{-1}}$ is estimated. The section halfway along the examined river section and locations relatively far from the slope crack show the lowest flux values. Since the total river length from the source to measurement point No. 4 is just about 15 km (Fig. 2) the estimated net exchange fluxes alone cannot explain the amount of surface water discharge measured in the river. A lateral flow component has to be responsible for this difference, which confirms the hypotheses from van Loon et al. (2009).

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Figure 8a presents measurements of surface water discharges at 7 positions along the river section between No. 2 and 4 (De Doncker et al., 2009). The results indicate significant changes in discharge along the section and a discharge at the downstream point (i.e. No. 4), which is in the same range as at the upstream point (i.e. No. 2). The obtained fluxes from the thermal analysis are too low to explain the variations in discharge. However, the interpolated net flux (Fig. 8b) shows a comparable spatial trend along the river; especially the strong discharge zone between points 210–206, the slight recharge zone between points 301–211 and the increasing discharge between 319 and 400 are reproduced. Again, in line with van Loon et al. (2009) a lateral contribution of groundwater flow to the river can be accounted for the differences. Because of the growth of macrophytes estimates of surface water discharge are however difficult to perform and their results may also have a considerable error band.

Statistical tests have been performed on the reach scale using the flux estimates of Fig. 7. Since the dataset is not normally distributed (supported by Lilliefors and Shapiro-Wilk tests, $N = 38$), non parametric statistical tests have been applied. Although the population size N is relatively small compared to other works like Anibas et al. (2011), some relationships between the magnitude of vertical flux values and morphologic features can be examined.

A significant correlation (Spearman Rank Order Correlations, Gamma correlations and Kendall Tau Correlations tests, $N = 38$; $p = 0.05$) is found for the flux rates of the measurement points versus the distance of each point to the slope crack of the morainic plateau (indicated as dashed lines in Figs. 3 and 8b). Along the right side of the river section high fluxes correlate with short distances, whereas such a correlation for the left side of the river is not significant until p is increased to 0.10. Higher fluxes are detected closer to slope crack where predominantly soil type II is abundant (i.e. the right side of the alluvium) indicating decreasing flux rates across the flood plane. van Loon et al. (2009) suggest the occurrence of groundwater discharge at the slope crack between valley wall and floor and a shallow permeable zone within the alluvium, which allow shallow lateral flow towards the river. Results from a groundwater model of van Loon et al. (2009) confirm these high groundwater discharges along the interface, which supports the relationship between flux measurement (river) location in the alluvium and flux quantity.

We classified the flux estimates according to their position along the river reach to investigate the relationship between the morphologic features and the calculated fluxes. In general the morphology of the river consists of straight sections and meanders. Measurement points located on the convex edges of meanders (e.g. cut banks) and where the river flow is straight, parallel or perpendicular to the general orientation of the river valley, are grouped and examined with a Mann-Whitney U test ($N = 38$, $p = 0.05$). The test indicated that fluxes on the edges of meanders are significantly higher than at other morphological positions. By adopting the Mann-Whitney statistical test ($N = 23$; $p = 0.05$) differences between other features, like sections of parallel and perpendicular flow with respect to the general flow direction could not be revealed. The high fluxes on the outer edges of meanders can be explained by the combined effect of the (usual shorter) distance to the slope crack between morainic upland and alluvial plain and the convergence of groundwater flow lines towards these points. In general points closer to the left side of the alluvium show in general low fluxes, which can be an indication that the groundwater discharge from the right side of the alluvium is stronger than from the left side, caused probably by differing soil and/or hydrogeologic composition since soil type I and II are indicated closer to the left and the right side of the alluvium respectively. Since the reach scale hydromorphology only can explain partly the flux differences a hierarchical approach is necessary to understand the remaining variability in fluxes. These points out that the fluvio-plain scale where second order factors are taken into account is inevitable to interpret the results gained from the reach scale.

5.3 Temporal variation

STRIVE also can simulate changes in groundwater and surface water exchange with some temporal resolution. Analysis of Dujardin et al. (2011) showed that transient simulations with a period of one week are feasible using the presented model set up

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if sufficient (i.e. continuously measured) data is available to fit the model. Figure 9a shows thermal data of piezometers No. 4 for the period 3 March to 20 June 2008. In Fig. 9b flux results based on hydraulic gradient data are compared with transient simulations of weekly duration from STRIVE. The global trend of groundwater-surface water interaction is well reproduced by the heat transport model. The model however fails to reproduce sharp peaks of exchange flows. Since the measurement accuracy of the used thermal sensors is less than 0.3 °C a sufficiently high temperature gradient and time is needed to detect temperature changes with depth to get a reliable flux estimate over the given simulation period. This and initialization errors limit the temporal resolution of STRIVE to 1–2 weeks. The thermal model integrates the exchange fluxes over a vertical domain of 5.0 m assuming a constant flux rate in depth, however in reality a vertical heterogeneity in flux rates is possible (Chou, 2009). Since the hydraulic head data covers a vertical domain of not more than 0.6 m absolute differences in flux rates as well as sensitivity of both methods to changing flow conditions are likely.

Flux estimates with a higher temporal resolution however can be generated by connecting the heat transport model with hydraulic gradient data from the piezometer nests. Values for vertical hydraulic conductivity K_v were estimated for periods with stable hydraulic gradients by calculating respective flux rates with STRIVE using transient simulation and by applying Darcy's law (Lapham, 1989). Using data from piezometer nest No. 2 a K_v of $0.22\,\mathrm{m\,d^{-1}}$ was estimated. K_v of piezometer nests No. 3 and 4 are $0.81\,\mathrm{m\,d^{-1}}$ and $0.05\,\mathrm{m\,d^{-1}}$ respectively (Table 2). Table 2 also shows the estimates of the horizontal conductivity K_h derived from falling and rising head slug tests in the respective piezometer nests No. 2 and 3. The anisotropy K_h/K_v ranges from 0.9 at No. 2 to 8.1 at No. 3, which is despite its range in agreement with literature values (Chen, 2000).

The estimated K_v values were then applied on time series data of hydraulic gradients measured in the piezometers to calculate hourly values of exchange flux. Figure 10 shows the results of the analysis between 13 September 2007 and 20 June 2008. For Piezometer nests No. 2 a continuous dataset is available showing an average infiltration

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of 4.8 mm d^{-1}. Piezometer No. 3 shows an average exfiltration of −25.6 mm d^{-1} in the period of 13 September 2007 and 8 December 2007, whereas No. 4 show a respective value of −78.9 mm d^{-1} during 4 March 2008 and 20 June 2008. The respective analysis of the points 200, 300 and 400 show exchange fluxes of a lower magnitude (Fig. 7) but of a comparable distribution, low infiltration, exfiltration and strong exfiltration, respectively. The highest determined fluxes at the location showing the lowest K_v indicates that the quantity of the fluxes along the reach is primarily determined by differing hydraulic head gradients rather than by differences in hydraulic conductivity. Point No. 4 shows a high temporal variability of exchange fluxes. The point is located close the slope crack and has relatively high fluxes, highlighting the influence of the exfiltration zone at the interface (van Loon et al., 2009). Exfiltration is dominating; long periods of relative stable flow conditions are interrupted by peaks of river discharge where the magnitude of the exchange fluxes alters rapidly and can adverse the flow direction from exfiltrating to infiltrating conditions. Infiltration rates of 5.8 mm d^{-1} where calculated, while during exfiltrating conditions flux values reach −104.3 mm d^{-1} at piezometer No. 4.

Piezometer No. 3 is located far from the slope crack in the middle of the alluvial plain and shows in comparison to No. 4 lower values of exchange fluxes and less fluctuation; minima and maxima of −3.0 and −49.7 mm d^{-1} were determined. Piezometer nest No. 2, with respectivee values of 32.1 and −6.8 mm d^{-1}, shows compared to piezometer nests No. 3 and 4 predominantly an infiltration of surface water into the hyporheic zone. Piezometer nest No. 2, in comparison with No. 4 is located farther away from the slope crack, the valley floor is wider explaining the differences in exchange fluxes and the peat resembles soil profile I. Piezometers however are difficult to place and to maintain, especially when they are placed directly in the riverbed. The retrieval of correct head gradient data is challenging in comparison with temperature measurements. A STRIVE simulation of thermal data from piezometer No. 1, where unfortunately no useful head data sets could be retrieved shows that the flux there is within the range of the results of the other piezometers (Table 2).

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5.4 Validation

By calculating K_v using STRIVE and head gradients the two methods were not applied independent from each other. Therefore we performed seepage meter measurements in Rogożynek (Fig. 3, No. 4) to have an independent validation of our previously presented results. By calculating K_v using STRIVE and head gradients the two methods were not applied independent from each other. Therefore we performed seepage meter measurements in Rogożynek (Fig. 3, No. 4) to have an independent validation of our previously presented results. The magnitude of the seepage meter fluxes follow the temporal pattern of the time series of fluxes based on head gradients measurements. For the examined period between 16–20 June 2008 the head based fluxes however yielded a higher flux of $-36\,\mathrm{mm\,d^{-1}}$, whereas the seepage meters indicated an average flux of $-14\,\mathrm{mm\,d^{-1}}$. A transient thermal simulation of temperature profiles collected close to the placed seepage meters using STRIVE resulted in a flux of $-10\,\mathrm{mm\,d^{-1}}$.

6 Conclusions

A hierarchical approach to quantitatively investigate and interpret groundwater-surface water interaction in space and time was presented by applying a combination of different field methods along a section of the Upper Biebrza River in Poland. Temperature profiles taken with a T-stick instrument and time series data of hydraulic head and temperature gradients measured in piezometer nests allowed the detection of "hot spots" and "hot moments" (McClain et al., 2003) of groundwater-surface water exchange in the hyporheic zone, whereas slug tests and seepage meter measurements are used for cross validation of the model results. With the combination of different field methods the limitations of each single method can be overcome; this increases the credibility of the obtained results.

Thermal modeling using STRIVE was performed using spatially distributed T-stick measurements and time series data from piezometer nests; an acceptable agreement

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between them was found. The thermal method also showed a good agreement with head based flux estimates and are validated with a number of seepage meter measurements. A vertical heterogeneity of these fluxes is observed but possible (Chou, 2009), since in contradiction to the thermal model the head and seepage meter measurements are performed close to the groundwater-surface water interface, hence a perfect agreement between these methods is unlikely.

In an area dominated by wetlands and peat soils the range of measured temperatures is high even during a single measurement campaign. In opposition to what is valid for sandy river beds we conclude that large variations in measured temperatures and strong thermal gradients are not only explained by differing fluxes but also by changing physical properties of the river bed. These differences are controlled also by varying soil properties (i.e. first order factors) resulting in a scattered pattern of estimated flux rates at a local scale (Figs. 6a–7). The soil properties however are not only heterogeneous with depth; they also vary along the river course and across the flood plain. This finding results in the necessity for a classification of model parameter sets for different measurement points and might be a reason for limiting application of thermal analysis in systems like the Biebrza River. We addressed this phenomenon by defining stratified soil profiles and introduced different sets of physical properties (Table 1). This is a simplification taking into account only the most evident differences found in the field, likely we do not cover complete physical and hydrological heterogeneity of the entire river section.

In comparison to a sandy riverbed the thermal conductivity of peat is low and the heat capacity high, thus the damping of the thermal signal with depth is strong. Significant changes in temperature occur therefore close to the interface, temperature measurements at shallow depths are therefore preferred. Since the exact position of the interface between riverbed and surface water is often not easy to define the techniques of collecting field data in peat environments should be subject of future improvement. An additional challenge for field investigations is the fact that protected wetlands obviously are not easily accessible and may underlie also legal or environmental restrictions that can hamper scientific investigation. An investigation of the soil parameters including laboratory tests, field measurements with improved spatial resolution and detailed assessment of the stratigraphy of the river bed are likely to further improve the model output. A classification of the parameters as described will however be always necessary.

The introduction of interpolated temperature time series defining the upper boundary of the transient thermal model of the T-stick measurements may lead to biases. Moreover the real exchange fluxes are not constant over time (Fig. 10) influencing the thermal pattern in the riverbed and consequently may influence the flux estimates integrated over long periods of time, especially when only a few profiles are available to fit the model (Fig. 5). Finally the estimated fluxes are in general fairly low, which can magnify the influence of other uncertainties of data on the simulated fluxes. We therefore regard the accuracy of the modelled results in the Biebrza River lower than of comparable works on sandy riverbeds or other more homogeneous environments (Anibas et al., 2009, 2011; Dujardin et al., 2011). The presented study is a first level investigation of the exchange processes in the area; by using the chosen parameter sets the calculated RMSE values of around 0.45 °C have an acceptable magnitude. Furthermore the STRIVE model is one dimensional vertical and hence cannot give any insight on lateral or longitudinal flow vectors within the riverbed (Fairley and Nicholson, 2005).

Hydromorphology at reach scale seems to play a key role for the hyporheic exchange (Cardenas, 2008; Boano et al., 2009). By interpolating the scattered point estimates on a reach scale (Figs. 6b and 8) the role of morphologic features and net exchange rates can be studied. Statistical tests show a significant dependence of the exchange flux to the distance of the measurement point to the slope crack (Fig. 8b) and the influence of meanders on the groundwater-surface water exchange. These effects are stronger for the right side of the fluvial plain especially visible at the upper and lower end of the measured section where the river approaches the morainic plateau (Fig. 8b). We regard consequently the right bank of the river stretch responsible for the greater

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share of groundwater discharge into the river. This is due to different soil types (or a different hydrogeology in general) dominating the left and the right side of the alluvium. Hydrological models (van Loon et al., 2009) of the area suggest surficial lateral flows across the fluvial plain. We regard them as being important for the rivers hydrology since a pure vertical flow component cannot explain the magnitude of river discharge. Discharges measured along the reach by De Doncker et al. (2009) show a stronger variation as these calculated by the spatial interpolation which may also indicate the influence of a lateral flow component. In general however the surface water discharge measurements of De Doncker et al. (2009) indicate a comparable spatial distribution of net groundwater discharges with the spatial interpolation. The investigated Biebrza River section shows a net infiltration along nine percent of the reach. Most of the time exfiltration and infiltration exist side by side; the infiltration however is fairly small.

Differences in flux are thus caused by differences in pedology together with morphology and topography. The convex side of the meanders shows higher fluxes because these parts of the river are closer to the morainic plateau, have a higher conductive underground and groundwater flow lines are converging at this point. The groundwater-surface water exchange thus could have an influence on the formation of meanders and on the soil properties and the conductivity of river bed. The morphology of the river therefore will not just influence the quantity of fluxes; they themselves influence the geomorphology.

Transient thermal simulations can be performed over different time scales with STRIVE. Periods of months or seasons are feasible, assuming relative constant groundwater fluxes; time scales of less than a week are however problematic because of initialization errors and limited data points to fit the model. If compared with head measurements results from piezometer nest No. 4 perform well. Short flux peaks however cannot be reproduced with the thermal model.

Since it is often difficult to get reliable values for the vertical hydraulic conductivity K_v of riverbeds, especially for heterogeneous peat soils, STRIVE is capable to connect information of hydraulic gradients with modelled exchange fluxes for its estimation. We

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found K_v values varying over one magnitude along the reach; the variation in flux is therefore related to differences in head gradients rather than conductivity changes.

These results, their heterogeneity and complexity underline the importance to select the appropriate scale for monitoring and interpretation (Vaughan et al., 2009) of the exchange processes and their determining factors. We therefore suggest a hierarchical approach to interpret and understand the determined groundwater-surface water exchange fluxes (Fig. 6). Point estimates of the exchange fluxes are representative on a local scale, where the first order factors (e.g. the composition of the riverbed, riverbed bathymetry apparent surface water and groundwater temperatures, elevation and position across the riverbed) have to be taken into account. The variability along the river course however cannot be explained by the first order factors alone. Spatial patterns become visible when the results are analyzed on a reach scale. There, “hot spots” of high or low exchange fluxes and zones of ex- and infiltration and relations between the exchange fluxes and morphologic and topographic features can be identified. To understand the underlying mechanisms of interaction, however an even wider scope, the fluvio-plain or sub catchment scale (determined by the second order factors like topography, morphology, climate, vegetation and hydrogeology) is necessary. It is thus indispensable to interpret fluxes determined on a local scale via thermal modeling in a wide perspective. Head gradients for example are related to the topographic and morphologic features determined on the fluvio-plain scale. The groundwater-surface water exchange pattern however might underlie specific temporal and spatial variations at each of the discussed scales, local, reach and fluvio plain.

The quantitative information of groundwater-surface water interaction or simply measured temperatures can be used to improve the parameterization, calibration procedure and therefore the accuracy of modelled hydrological or ecological transport, retention and reaction processes for the Biebrza River and its wetlands. This will not just improve the understanding of the hydro-ecologic functioning of the site but further establish the Biebrza National Park as reference area of worldwide significance (Chormański et al., 2009; Dabrowska-Zielinska et al., 2009). A better understanding of the interaction processes between the river and its adjacent wetlands and the hyporheic zone of this particular ecosystem helps to develop unerring procedures for its management and conservation; practices which can than be transferred to other locations were protection or restoration efforts are needed, planned or already established.

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Table 1. Physical and thermal properties of the soil profiles defined for the Upper Biebrza catchment.

Soil*	Porosity Φ	Specific heat capacity c in $J\,kg^{-1}\,K^{-1}$	Density ρ in $kg\,m^{-3}$	Thermal conductivity λ_e in $J\,s^{-1}\,m^{-1}\,K^{-1}$	Description
Peat	0.95	3900	1100	0.4	Soil profile I and III
Peat	0.80	3300	1300	0.7	Soil profile II
Sand	0.42	1300	2000	1.8	Soil profile I, II and III

* completely saturated

Properties of the liquid phase (e.g. water): c_w, ρ_w and λ_e are $4180\,J\,kg^{-1}\,K^{-1}$, $1000\,kg\,m^{-3}$ and $0.6\,J\,s^{-1}\,m^{-1}\,K^{-1}$ respectively.

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Table 2. Estimated vertical hydraulic conductivity K_v using thermal and hydraulic head gradient data.

Piezometer nest No.	begin[1]	end	Vertical flux[2] v_z in mm d^{-1}	Vertical hydraulic gradient[3] φ in cm	Vertical hydraulic conductivity K_v in m d^{-1}	Horizontal hydraulic conductivity[4] K_h in m d^{-1}	
1	6 Aug 2007	28 Aug 2007	−24.5	–	–	–	–
2	18 Jan 2008	4 Mar 2008	3.5	3	0.22	0.26	1.2
2	8 Feb 2008	22 May 2008	−6.5	2	0.22	0.26	1.2
3	6 Aug 2007	28 Aug 2007	−36.2	−4	0.81	0.10	0.1
3	9 Mar 2008	28 Apr 2008	−20.4	−3	0.81	0.10	0.1
4A	6 Aug 2007	28 Aug 2007	−21.9	−36	0.05	–	–
4B	6 Aug 2007	28 Aug 2007	−38.4	−36	0.05	–	–
4B	14 Apr 2008	15 Jun 2008	−29.8	−29	0.05	–	–

[1] simulation period
[2] using transient STRIVE simulations
[3] from piezometer nests
[4] from falling and rising head slug tests

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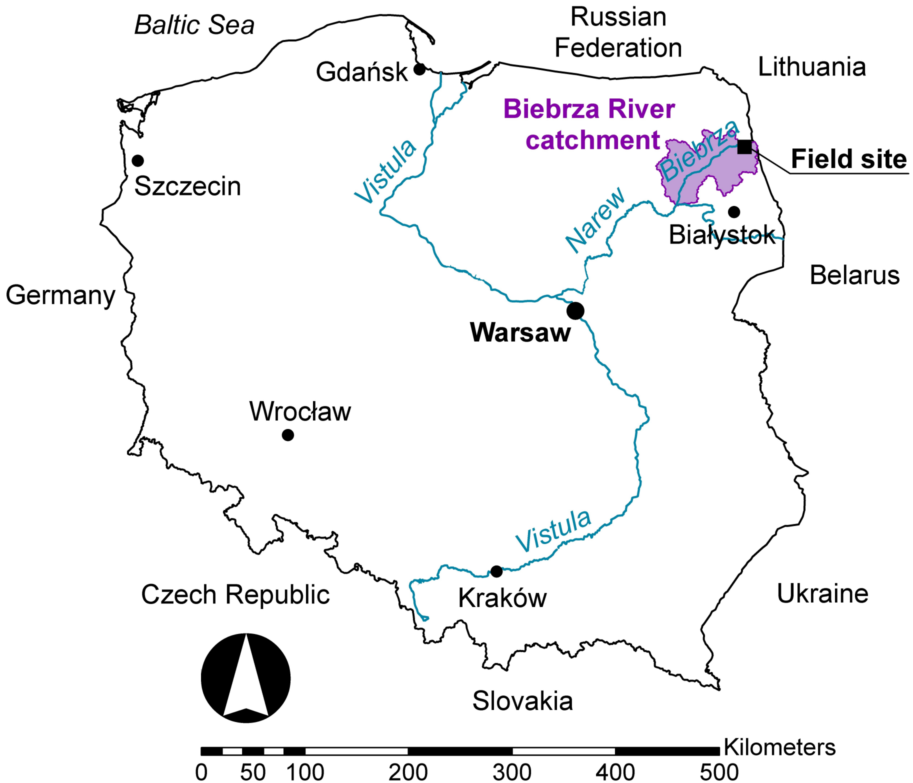

Fig. 1. Location of the Biebrza River catchment in Poland.

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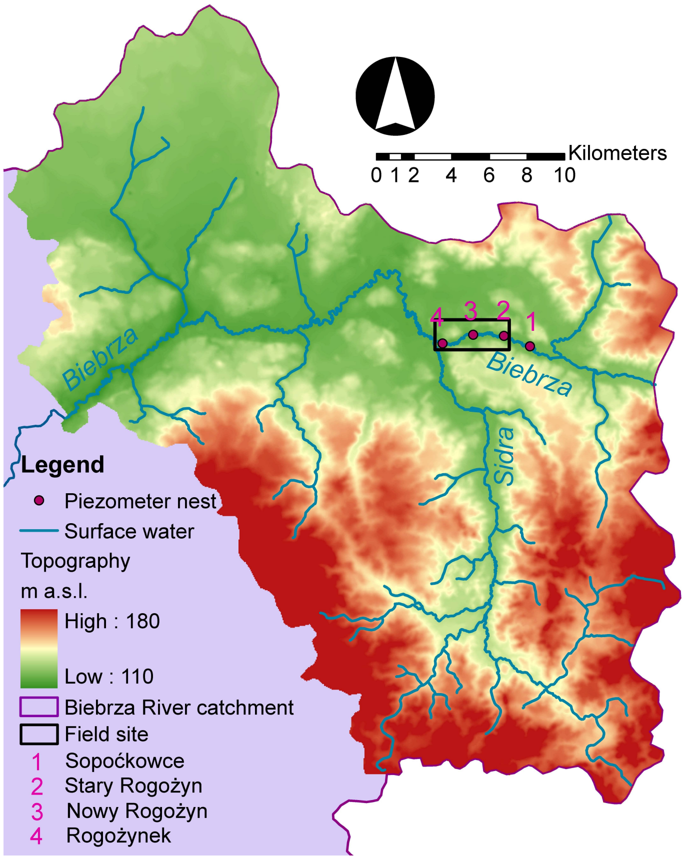

Fig. 2. Digital elevation model of the Upper Basin of the Biebrza River. The dots indicate the locations of the piezometer nests. The black box indicates the river section where the T-stick measurements have been performed.

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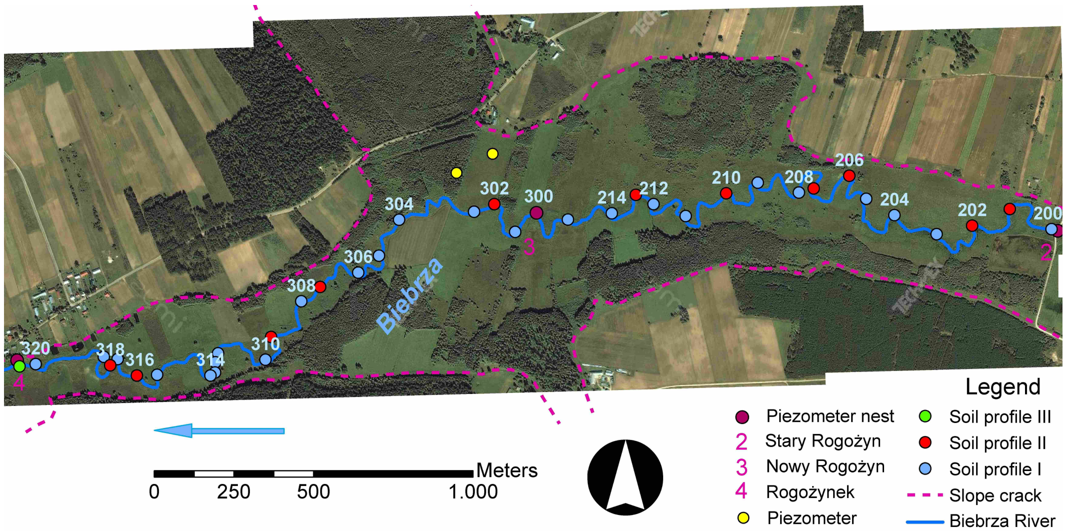

Fig. 3. Location of the 38 points of the T-Stick measurements along the Biebrza River. The purple dots indicate the location of piezometer nests. The dashed line indicates the maximum extent of the alluvium or floodplain (i.e. the slope crack between valley wall and valley floor); a tributary is entering the alluvium from the south, in the north the alluvium extents into a paleochannel of the Biebrza River. On the right side of the alluvial plain two piezometer nests are indicated. Orthophotomap source: http://www.zumi.pl.

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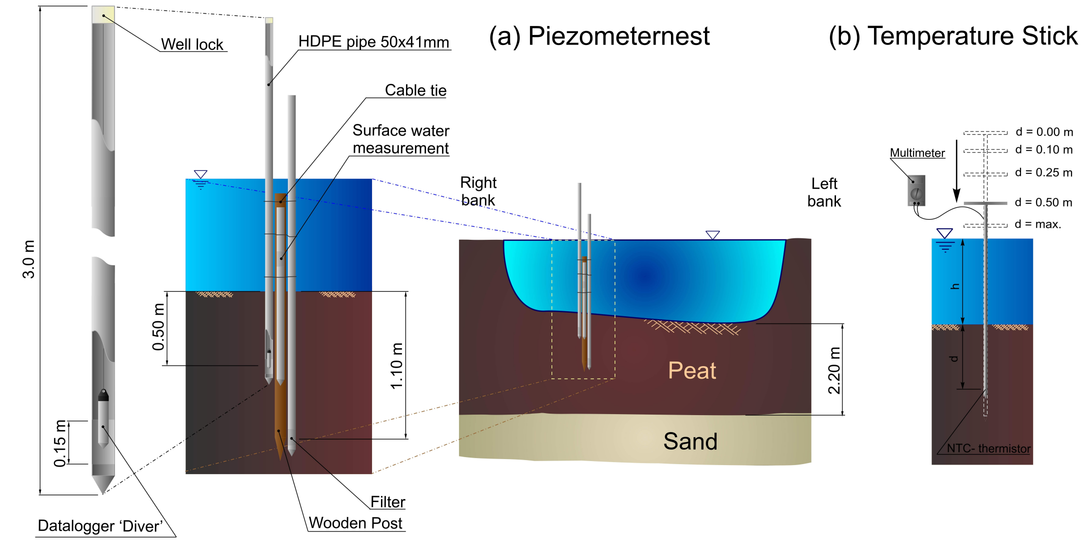

Fig. 4. **(a)** Setup for measuring temperature profiles and hydraulic head in the Biebrza River with piezometer nests equipped with data loggers, as example piezometer nest No. 2. **(b)** Scheme for measuring of temperature profiles in the riverbed with the T-stick instrument.

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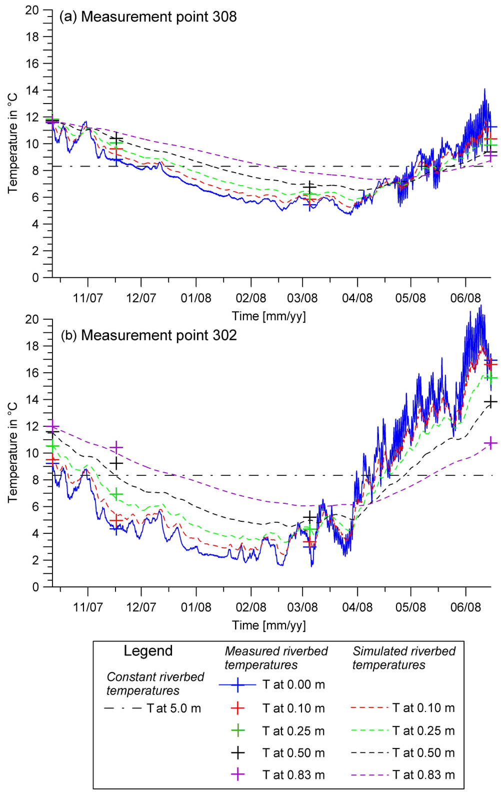

Fig. 5. Setup of the transient STRIVE model with results from measurement point 308 **(a)** and 302 **(b)**. The riverbed temperature at 0.0 m depth serves as upper model boundary, a constant temperature at 5.0 m depth (the dashed-dotted line) as lower boundary. The crosses indicate the measurements with the T-stick instrument, whereas the dotted lines indicate the simulated temperatures at the respective points for the best model fit. **(a)** Soil type I, flux = −6.5 mm d^{-1}, RMSE = 0.41 °C; **(b)** Soil type II, flux = −26.2 mm d^{-1}, RMSE = 0.46 °C.

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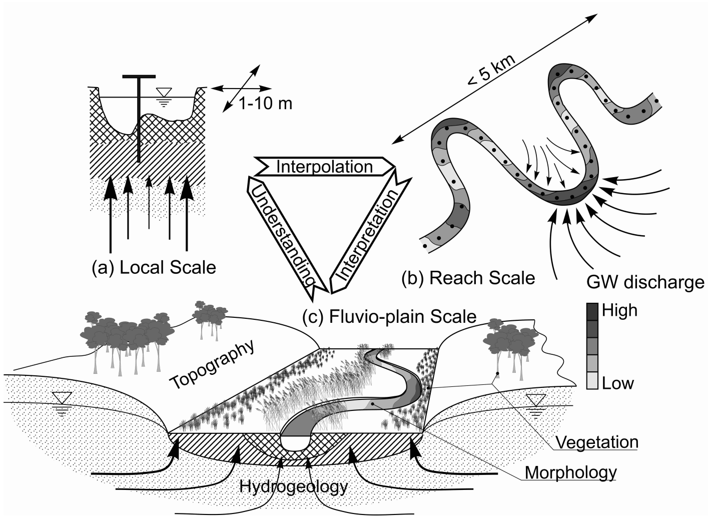

Fig. 6. Concept for the estimation and understanding of groundwater-surface water interaction. **(a)** Setup of field measurements representing point estimates on a local scale. Bathymetry, composition of the riverbed and position of the measurement across the river influence the vertical fluxes (e.g. first order factors). **(b)** Spatial interpolation reveals distributed exchange patterns on a reach scale indicated in grey scales. Hydromorphological features like converging and diverging flow lines cause different exchange fluxes at convex and concave sides of meanders respectively. **(c)** Interpretation of the groundwater-surface water interaction system is possible when the system is analyzed at sub-basin or fluvio-plain scale. The quantity of exchange fluxes is dependent on morphology, topography, climate, vegetation and hydrogeology (e.g. second order factors). The morainic plateau outside the fluvial plain is composed of heterogeneous loamy sand deposits. The flat alluvial valleys are filled with different types of organic soils determined by the location of the river and the associated vegetation. While the morainic plateau is dominated by agriculture and forest, the alluvial plain shows reed vegetation in the center of the valley and sedges closer to the slope crack. The improved understanding of the hydrological system is used to update and improve future investigation and modeling efforts **(a)**.

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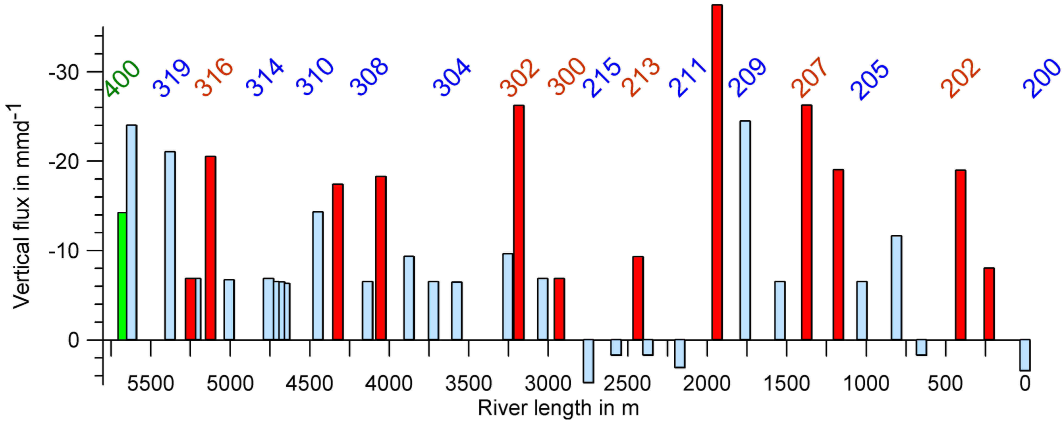

Fig. 7. Results of transient simulations per measurement point with STRIVE. The bars indicate the integrated fluxes per measurement point between 11 October 2007 and 17 June 2008; the colours of the bars indicate soil profiles I, II and III respectively in blue, red and green.

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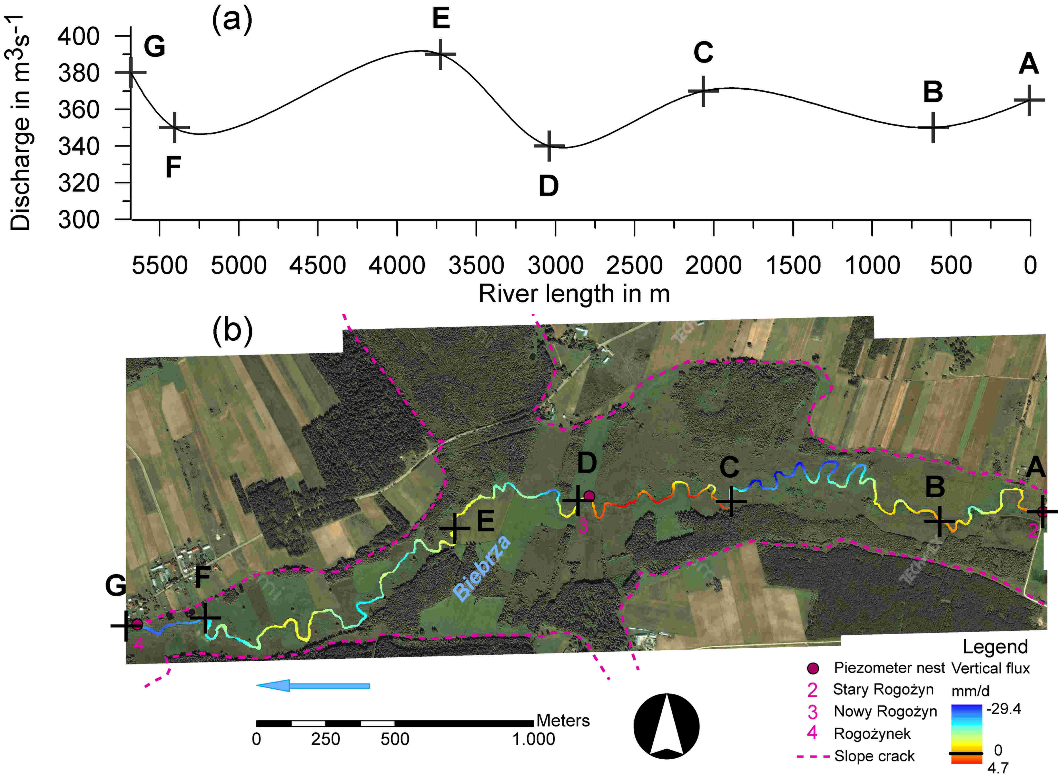

Fig. 8. (a) Surface water discharge measurements A–G along the river section according to De Doncker et al. (2009). **(b)** Spatial interpolation of the point estimates of the transient simulation on a reach scale indicated as coloured band. The location of the surface water measurements A-G are indicated by crosses. Orthophotomap source: http://www.zumi.pl.

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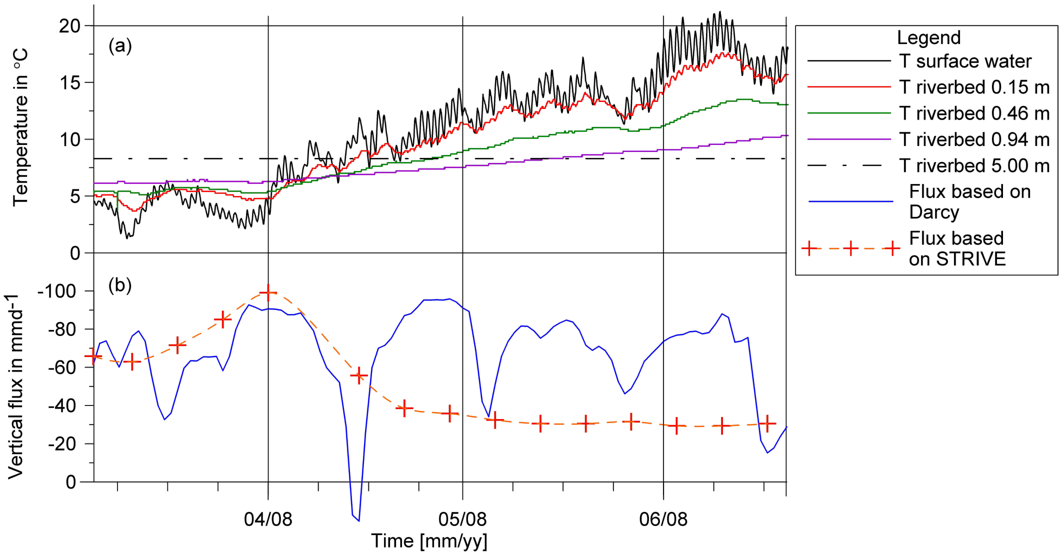

Fig. 9. **(a)** Surface water temperature and measured groundwater temperatures at different depths in piezometer nests No. 4. **(b)** Corresponding estimated fluxes using weekly transient thermal simulations with STRIVE as well as daily averaged fluxes based on Darcy calculations.

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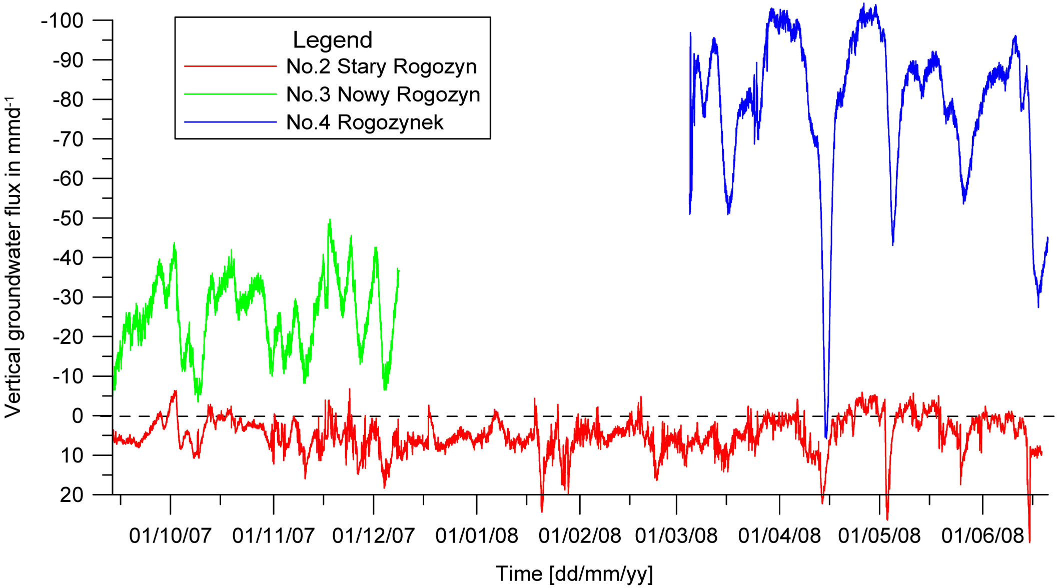

Fig. 10. Temporal distribution of the surface water-groundwater interaction in the riverbed of the Biebrza River based on time series data of hydraulic head and hydraulic conductivity values derived with STRIVE.

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