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Spatial distribution of the water exchange through river cross-section – measurements and the numerical model

Maria Grodzka-Łukaszewska*, Zofia Pawlak, Grzegorz Sinicyn

Faculty of Building Services, Hydro and Environmental Engineering,
Warsaw University of Technology, Poland

*Corresponding author's e-mail: maria.lukaszewska@pw.edu.pl

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Abstract: The aspects of surface stability and groundwater exchange recognized by many researchers due to the intensification of agriculture and industry (manifested in, e.g., regulation and dredging of riverbed sediments of rivers) are now widely discussed on the international forum of water policy and management. It is essential to assess the spatial variability of water exchange through the river length and cross sections for the preparation of data and calculation of the groundwater flow model. This article presents research which describes the spatial distribution of the surface water-groundwater interaction within the river cross-section. Two measurement series were carried out to describe its variability. Additionally, a groundwater flow model was developed to simulate and represent the variable nature of water exchange in the hyporheic zone in the river's cross-section. The model was successfully verified by means of measurements of water flux in the hyporheic zone. The precise spatial description of this variability is the first step to determine the possibility of introducing this variable in an accurate manner, within the limits of measurement uncertainties or simulation assumptions, in the construction of mathematical models of groundwater flow.

Introduction

The interaction between surface waters and groundwater can be investigated in two respects: the role of the physical context of this phenomenon and participation in the decision-making processes concerning water management. Surface waters and groundwater are not separate elements of the environment – water exchange of varying intensity occurs between them. The quantitative and qualitative characteristics of such exchange are largely influenced by several physiographic, climatic, and anthropogenic factors (Boano et al. 2014; Jekatierynczuk-Rudczyk 2007; Ward 2016), such as hydraulic conductivity of the sediment layers (Brunke and Gonser 1997; Jekatierynczuk-Rudczyk 2007; Harvey and Gooseff 2015), dynamics of surface flow (Peralta-Maraver et al. 2018), basin ecology (Zieliński and Jekatierynczuk-Rudczyk 2010), channel bends (Boano et al. 2006), river morphology (Schmadel et al. 2017) as well as intensity and duration of rainfall (Siergieiev et. al. 2014).

The threats to the stability of surface and groundwater exchange recognized by many researchers due to the intensification of agriculture and industry (manifested in, e.g., regulation and dredging of riverbed sediments of rivers) are now widely discussed on the international forum of water policy and management (Hendriks et al. 2015). It is now evident that undertaking appropriate activities aimed at protecting and restoring water and coastal ecosystems, dependent on the directions and flux intensities of water exchange in the hyporheic zone, should involve solutions referring to the latest results of hydrological and ecological research (Grygoruk and Acreman 2015). The processes of water exchange between the river and aquifer have a decisive influence on the condition of river ecosystems. Studies on the size and spatial distribution of exchange fluxes in the hyporheic zone have adopted several simplifying assumptions and limitations – in terms of the temporal or spatial variability necessary for model calculations (Anibas et al. 2012). Many hydrogeological situations require accurate evaluation of all three (x, y, z) components, namely: velocity; specific discharge; and flux density. River-aquifer interaction is an example where precise assessment of all three components of groundwater velocity beneath and near the riverbed is essential for correct calculation of flow paths, discriminating between the bottom and bank water exchange.

The range of the hyporheic zone depth is a feature specific to each water reservoir and shows high temporal and spatial variability (Siergieiev et. al. 2014). The depth to the groundwater head – depending on the terrain – is an important factor determining its connection with surface waters

(Jekatierynczuk-Rudczyk 2007; Magliozzi et al. 2018). The seasonal dynamics of water balance parameters also influence periodic changes in the hyporheic zone through the occurrence of periods of high and low water levels (Jekatierynczuk--Rudczyk 2007). Seasonal patterns of water exchange between river and aquifer depend largely on the intensity and duration of precipitation. Directly after precipitation incidents, there are temporary changes in the nature of the river (from draining to infiltrating). Water exchange between groundwater and surface waters results in fluctuations in the water table in rivers (Marciniak et al. 2017).

The numerical solution of the groundwater flow model requires discretization of space, including the discretization of the shape of the river, emphasized in modelling. This discretization is the division of the space of the model into computing cells, where uniformity of variables and parameters of the model are assumed. Considering the need to define rivers in such a discretized model, only the variability of water exchange between the river and aquifer (through the introduction of a bottom sediment parameter) between individual cells of the model can be assumed. Unfortunately, this variability (in the case of the groundwater flow models) is not described in the literature.

The primary objective of this research is describing the spatial distribution of the surface water-groundwater interaction within a Polish river cross-section (the Świder river). The determination of the water exchange variability involved conducting two measurement series. A groundwater flow model was developed to simulate the variable nature of water exchange in the hyporheic zone in the river's cross-section. The model was successfully verified using measurements of water flux in the hyporheic zone.

Material and methods

Groundwater flow model of the river cross-section

There are numerous software tools widely applied to describe hydrogeological conditions and develop groundwater models such as MODFLOW, Visual Modflow and FEFLOW. MODFLOW (McDonald and Harbaugh 1984) is an example of software commonly used to create and calculate regional groundwater flow models and, because of this, it is used mainly for regional-scale conceptualization of the river-aquifer interaction. Like many others, this software numerically solves the groundwater flow equation by means of the block--centered finite difference method and calculates hydraulic heads in grid block centers. Despite its good results in many hydrogeological situations, however, it does not approximate the vertical velocity component with respect to the horizontal components with sufficient accuracy (Nawalany 1993; Zijl and Nawalany 1993; Grodzka-Łukaszewska et al. 2017).

The FEFLOW software designed by DHI-WASY, a German branch of the DHI Group, was applied for creating the model and conducting numerical calculations. This software package works by solving equations of flow, mass, and heat transport in porous and fractured media by a multidimensional finite element method for complex geometric and parametric situations including variable fluid density, variable saturation, free surface, multispecies reaction kinetics, non-isothermal flow, and multidiffusive effects (Diersch 2014; Brunetti et al. 2013). The FEFLOW software has been used in this project, since it presents the opportunity to create a quite precise discretization grid. The 2D spatial division of the model domain into triangular calculation blocks enables precise, flexible and accurate reproduction of complicated elements of the structure to model aquifers (river, boundaries etc.) guaranteeing practical and precise replication of the real shape of the studied area (Leiter 2017). It can be used, for example, to describe the spatial and temporal variability of pollutants in groundwater (Elango et al. 2012), model geothermal processes (Hidayat and Permana 2018), assess the impact of dewatering (Brunetti et al. 2013), assess the impact of climate change on groundwater level (Pandian et al. 2016), or plan strategies for remediation of contaminated groundwater. With regards to calculating the water exchange between the river and the aquifer, the afore-mentioned models MODFLOW and FEFLOW work the same way. Both include the river using a 3rd type boundary condition. The only difference between MODFLOW and FEFLOW that is noteworthy in this article lies in the FEFLOW's ability to accurately reproduce the shape and parameters of the river without losing the time needed for calculations. For this reason, FEFLOW was chosen to conduct the calculations presented here.

The model was calibrated and verified by adjusting the calculated hydraulic head and water exchange between the river and aquifer to the results of two series of measurements. During these measurements, the hydraulic gradient and water flux in the hyporheic zone were measured.

Measurement equipment

One of the first studies concerning equipment for measuring the volume of water exchange between groundwater and surface water began in the 1940s, when a seepage meter was developed. It was used to measure the intensity of water infiltration in a point-based way (Israelsen and Reeve 1944). Further research was based on the determination of water loss from irrigation channels to supply the aquifer by recording changes in the water level occurring in an isolated section of a channel (Robinson and Rohwer 1959; Worstell and Carpenter 1969; Iqbal et al. 2002). It proved to be an expensive and time--consuming method. Throughout the years, the seepage meter has undergone many modifications. The most widely known modification is the device designed by David Robert Lee (Lee 1977). It permits not only measurement of water exchange from surface water to groundwater, but also measurement in the opposite direction. Over the years, the seepage meter has been improved and measurements can now be taken automatically. In the 1980s, a ground infiltrator (Janik et al. 1989) was developed in Poland to determine the intensity of water exchange from surface water to groundwater. Based on this invention, the filtrometer used in this study (Marciniak and Chudziak 2015) was developed.

The determination of the hydraulic conductivity of the bottom sediment in the analyzed cross-section – necessary to develop the groundwater flow model – involved measurement of two parameters, namely infiltration flux density or drainage density, and hydraulic gradient between the aquifer and surface waters. The measurements were done by means of a filtrometer and the gradientmeter developed and made by Prof. Marciniak (Marciniak and Chudziak 2015).

A filtrometer is a device used to measure point infiltration flux density or drainage density. A filter cloche with a known surface is placed in the bottom of the river. The volume of water that flows through it is measured over a specific time period. A gradientmeter is used to determine the hydraulic gradient between the aquifer and surface water. It is composed of two connected tubes, one located in the sediment and the other one in the river. The difference in water levels between the aquifer and river provides the basis for the determination of the hydraulic gradient (Marciniak and Chudziak 2015). The results of measurements undertaken with this equipment may show an uncertainty of 2.7% (Marciniak and Chudziak 2015). The equipment is presented in Figure 1.

Study area

In order to describe the spatial distribution of the surface water-groundwater interaction, a river representative for the region was chosen for the initial study.

The most common type of rivers in Poland (based on the classification of Uniform Water Bodies) are lowland sandy-clayey rivers, which has determined the selection process for the study area. Also, the requirements for the selected cross-section resulted directly from the limitations of the measurement equipment and were as follows: sandy bottom, river depth range 30–100 cm, and width of the river of minimum 10 m (Marciniak and Chudziak 2015).

Taking into account all requirements and limitations, it was decided to undertake the measurements on the section of the Świder river (geographical coordinates: N 52° 8’ 27.622”, E 21° 16’ 11.632”) nearby a town (central Poland, Europe) which, according to the Polish law (Regulation of the Minister of Maritime Economy and Inland Navigation 2019) resulting from the Water Framework Directive, is classified as a sandy-clayey lowland river. Its length is 89.1 km with average slope of 116 cm/km, and a basin covering an area of 1149.8 km^2 (Pietrzak et al. 2018). The average flow is 4.27 m^3/s (IMGW-PIB 2016). The location of the tested cross-section is shown in Figure 2.

Two measurement series were carried out in one selected cross-section on the Świder river. The cross-section’s width was 24 meters. A hydraulic gradient was measured between the aquifer and surface waters at 12 evenly distributed

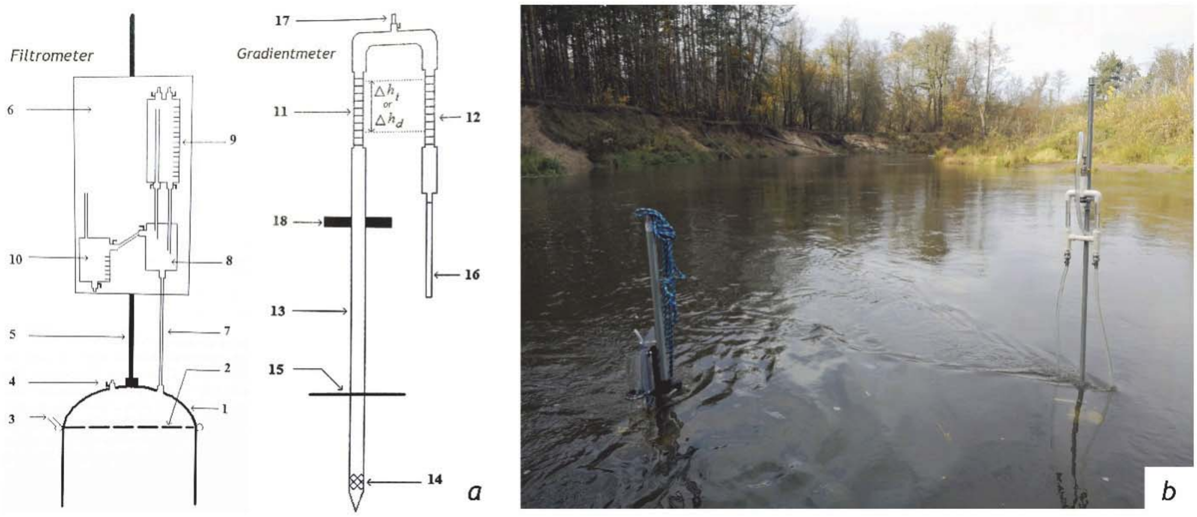

Fig. 1. a) Construction model of filtrometer and gradientmeter (according to Marciniak and Chudziak 2015)

1 – shade; 2 – subtend strainer; 3 – perforated pipe with water supply; 4 – vent valve; 5 – bracket; 6 – board; 7 – serpent; 8 – container; 9 – cylinder; 10 – measuring container; 11 – measuring pipe of piezometer; 12 – measuring pipe of surface water; 13 – piezometer; 14 – needlefilter; 15 – subtend roller; 16 – serpent submerged in surface water; 17 – valve; 18 – handle, b) Filtrometer (left) and gradientmeter (right) during measurement

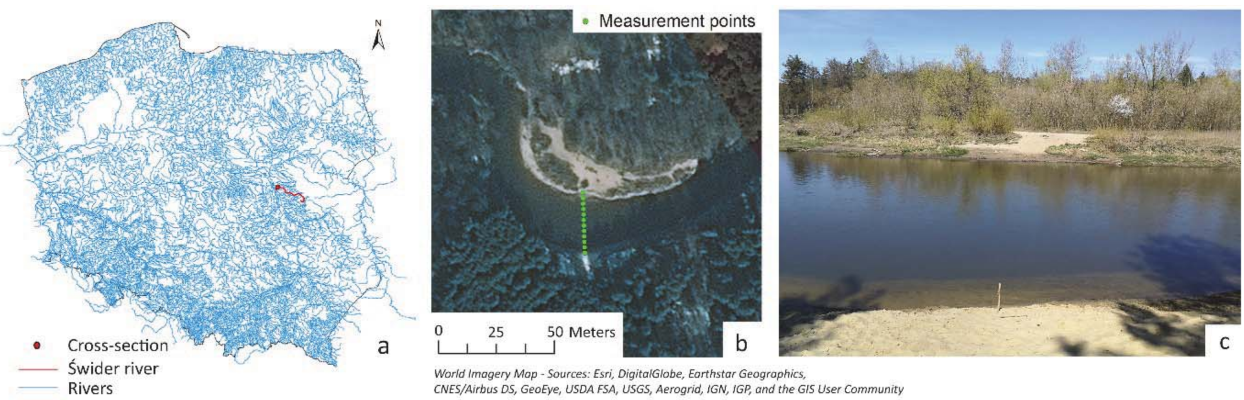

Fig. 2. a) location of the study area in Poland, b) aerial view of the cross-section, c) view of the cross-section (left bank)

measurement points (one measurement every 2 meters) along the cross-section of the river by means of a gradientmeter. The first measurement series was performed on 30 October 2018 after a week of heavy rainfall, and the second one on 10 November 2018 after a week with no major precipitation. The time step between individual measurements using gradientmeter was 10–15 minutes, including the time required to stabilize the results.

The meteorological data of October and November 2018 used in the model are presented in Figure 3. They were obtained from the Polish Institute of Meteorology and Water Management – National Research Institute (IMGW-PIB). The measurement station (named *Warszawa-Bielany*) providing the data is located approximately 26 km north-west of the measurement point.

The water level in the Świder river over the last 10 years (11.2008–11.2018) is presented in Figure 4, with the water level range observed during the measurement series indicated by horizontal lines. The data were acquired from the Polish Institute of Meteorology and Water Management – National Research Institute (IMGW-PIB). These data were presented as the background of the hydrological situation of the presented section of the river. As can be seen, the water level in the river undergoes a certain seasonality throughout the year. Extremely, annual fluctuations have reached a maximum of 1 cm in the last 4 years (2014–2018). The hydrological station (named *Wólka Mlądzka*) providing the data is located approximately 6 km upstream of the measurement point.

The choice of two different meteorological situations allowed for a more differentiated perspective on the studied area. A point measurement of the water flux was also conducted by means of a filtrometer. Bathymetry of the cross-section was determined using a GPS RTK receiver.

Measurement results

The measured difference between water levels in the aquifer and river, together with the calculated hydraulic gradient for both measurement campaigns, are presented in Table 1 and Figure 5.

The measurement results confirm the spatial variability of the water exchange stream in the hyporheic zone of the analyzed river cross-section. Also, the gaining type of the river was determined. However, both measurement series exposed

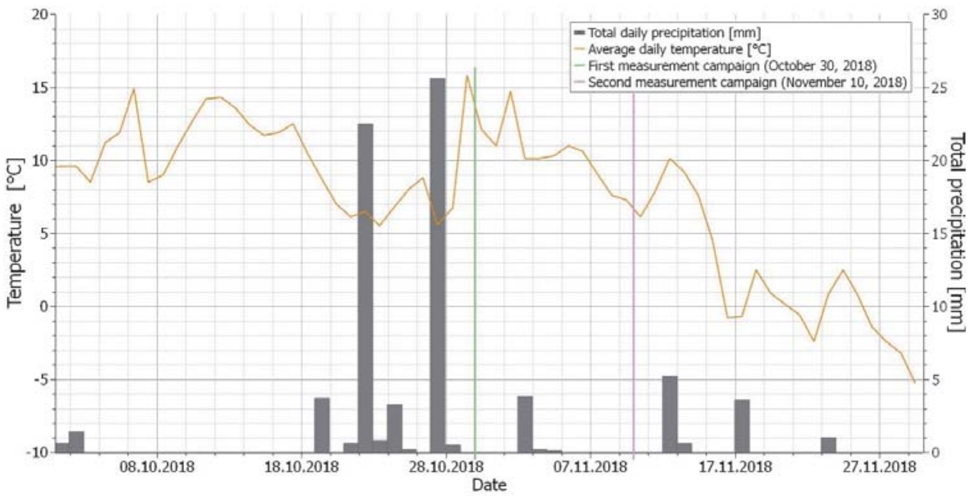

Fig. 3. Average daily temperature and total daily precipitation; October and November 2018

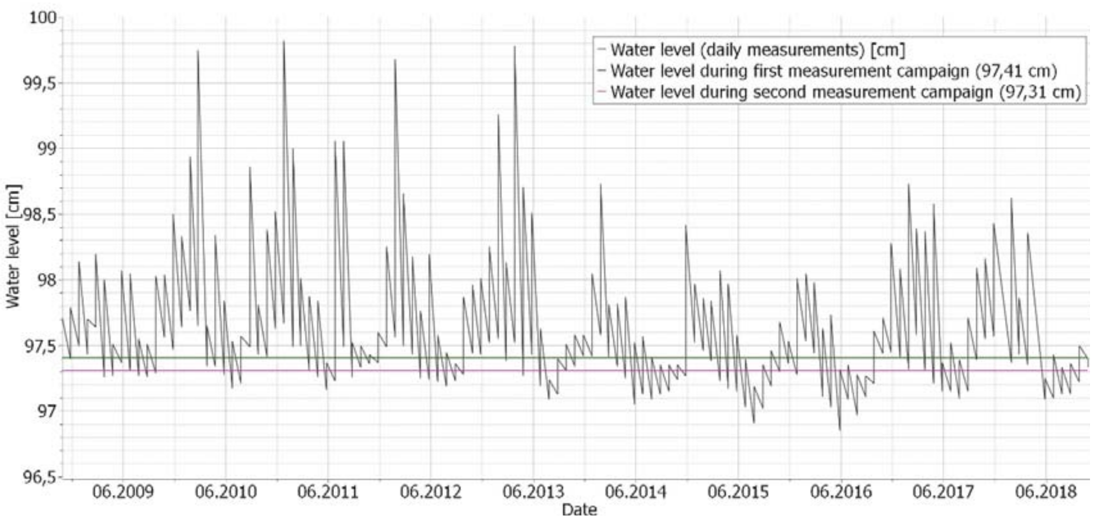

Fig. 4. Water level in the Świder river [cm]; years 2008–2018

the river's losing type at the bed boundaries, suggesting potential variability of the direction of water exchange in the hyporheic zone of the river.

Although both measurements were conducted under different meteorological conditions, the results indicated very similar spatial variability of the hydraulic gradient in the river cross-section. Discrepancies in values measured on different days may be caused by the lack of accurate positioning of the gradientmeter (exactly in the same place). This may indicate a very local impact of the riverbed filtration parameters on the pressure gradient value.

The time interval between both measurement series could have been too short to assess the aforementioned impact of different meteorological conditions. Its assessment may require more measurement series with longer intervals; however it is a promising first step in this scope of research.

The mean value of the measured average filtration intensity was used to calculate hydraulic conductivity. The measurements were carried out on October 30 for three points located in the riverbed at the cross section (due to the limitations of the equipment and depth of sands on the riverbed, it was not possible to make more measurements). The values of filtration intensity measured by means of a filtrometer were 0.9 m^3/d, 0.8 m^3/d, and 0.2 m^3/d, respectively. The lowest significant outlying value was eliminated from the calculations, due to the conclusion that it is too low to be considered accurate.

Table 1. Water level difference between the aquifer and surface water; hydraulic gradient

	October 30, 2018		November 10, 2018	
Measurement point	Difference in the water level [cm]*	Hydraulic gradient [–]	Difference in the water level [cm]*	Hydraulic gradient [–]
1	-4	0.2	-1	0.05
2	6	0.3	4	0.2
3	3	0.15	5	0.25
4	2	0.1	2	0.1
5	10	0.5	8	0.4
6	6	0.3	8	0.4
7	4	0.2	5	0.25
8	2	0.1	5	0.25
9	8	0.4	8	0.4
10	1	0.05	0	0
11	2	0.1	0	0
12	-1	0.05	0	0

* negative values: losing type river, positive values: gaining type river

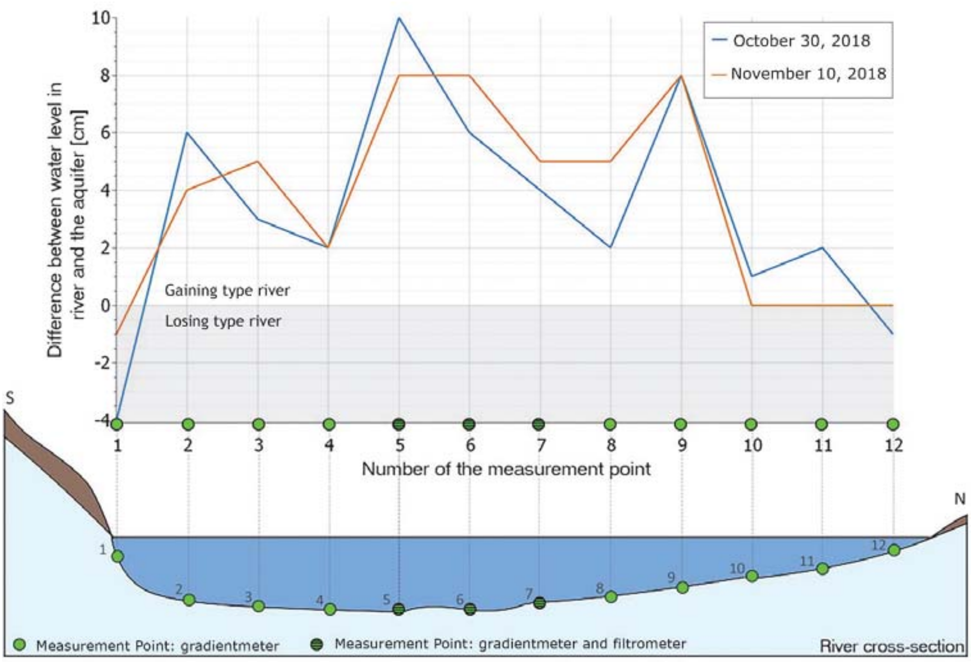

Fig. 5. Measured difference between water level in the river and aquifer

Table 2 presents the hydraulic conductivity determined for individual measurements by means of gradientmeter and filtrometer. Calculations were based on individual results of measurements with use of the filtrometer, taking into account the average result of the gradientmeter measurements. The average of the calculated hydraulic conductivity is approximately 77 m/d, which is within the range accepted for sands.

Model

A two-dimensional steady-state model was developed with a finite element grid consisting of 11885 triangular elements and 6352 nodes. The finite element mesh is presented in Figure 6.

The thickness of the modelled aquifer varies from 9.4 m in the thickest point (left bank) to 3.9 m in the thinnest point (riverbed). The thickness of the sand layer under the riverbed was determined based on a detailed Geological Map of Poland (Baranicka 1976). The elevation of the top of the aquifer layer was determined by GPS RTK measurements. Hydraulic conductivity in the top layer of the modelled riverbed was assumed the same as values calculated from measurement results (Table 2). The remaining areas used the average value obtained during first measurements. At the bottom of the aquifer, a no-flow boundary condition was assumed. On the left and right sides of the model, a head-dependent flux boundary condition was assumed, with a value equal to 1 meter below ground level. In this way, the lateral flow was included in the model. This was due to the lack of indications of the occurrence of the groundwater divide in the area. The river was modelled based on the first type (Dirichlet) boundary condition (the value was assumed based on measurements of the water level in the river).

A spatially variable value of recharge was assumed on the surface of the model. This value depends on the direct precipitation, land cover, terrain slope, type of the surface soil type and land cover characteristic for this area (Duda et al. 2011). The values of the recharge defined in the presented model is presented in the figure below (Fig. 7).

Sensitivity analyses

The sensitivity analysis of the model concerning changes in hydraulic conductivity employed the SENSAN tool (implemented in FEFLOW). A total of 337 simulations were carried out for the sensitivity analysis. During each of them, the value of hydraulic conductivity was increased or decreased in the range from 0 to 50% of its original value for one of 112 individual calculation zones. A large range was used due to significant differences in measurements of hydraulic conductivity (Table 2). Based on the results of these 337 calculations, the difference between the water level in the aquifer and river was calculated. Its variability in subsequent simulations for individual measurement points is shown in Figure 8. The presented graph shows how the difference

Table 2. Determined hydraulic conductivity in the cross-section of the riverbed

Measurement point	Hydraulic conductivity [m/d]	Measurement point	Hydraulic conductivity [m/d]
1	47.22	7	47.22
2	31.48	8	94.43
3	62.96	9	23.61
4	94.43	10	188.87
5	18.89	11	94.43
6	31.48	12	188.87

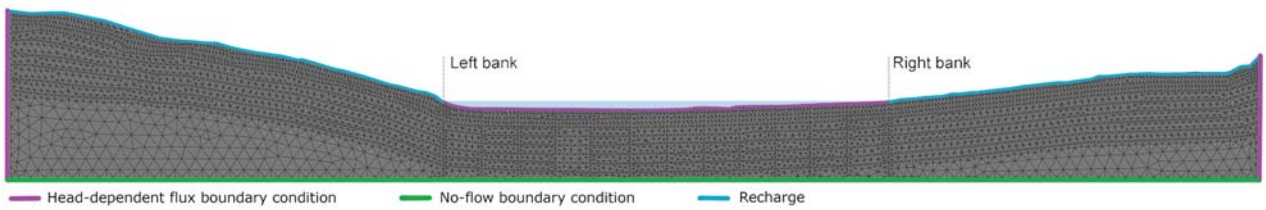

Fig. 6. Cross-section calculated in FEFLOW – finite element mesh with boundary conditions

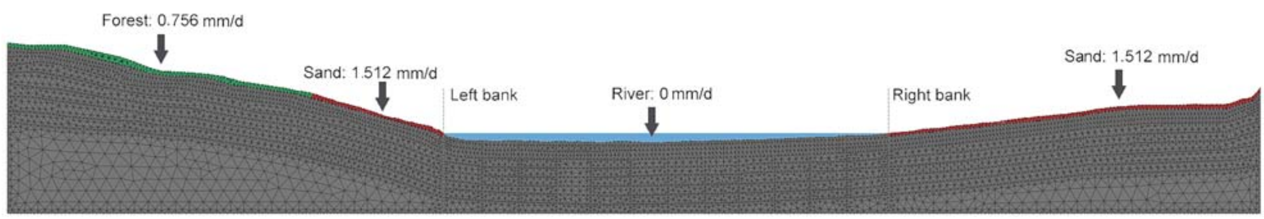

Fig. 7. Spatial distribution of recharge assumed in the model

between water levels at individual measurement points has changed for each simulation in the sensitivity analysis.

Based on the graph analysis, the most considerable changes were observed during simulation No. 227, where the value of hydraulic conductivity in the calculation zone located at the left edge of the modelled riverbed was reduced. Distinct deviations were also observed in simulation No. 115, where the value of hydraulic conductivity was raised. Such substantial changes, up to 1.5 cm, observed in eight of the twelve measuring points, suggest high sensitivity of the model to changes in values of hydraulic conductivity in zones on the left edge of the model. Increasing and decreasing the hydraulic conductivity at the right bank (simulations No. 160 and 273) has no substantial effect on modelling results. The results of measurements closer to the right edge of the model (measurement points 8 and 10–12 in Figure 5) show little sensitivity to changes in hydraulic conductivity, and the calculated values deviate by a maximum of approximately 0.05 cm. The difference in sensitivity to changes in hydraulic conductivity on both sides of the riverbed may be due to higher values of the hydraulic height and hydraulic gradient on the left side of the modelled area, which results in higher values of the specific discharge.

Sensitivity analysis allowed for comprehensive interpretation and detection of places in the model that should be given special attention during calibration and verification. It allowed the designation of zones in which a change in the value of hydraulic conductivity implies a significant change in the model results. Enriching the calibration process with this knowledge allowed for more efficient and effective model analysis. A sensitivity analysis for the hydraulic conductivity helped in improving the manual calibration process, through the identification of areas most vulnerable to value changes.

Calibration results

The model was calibrated simultaneously in terms of two variables:

- gradient differences between the water level in the aquifer and river,
- water exchange (filtration intensity) between the river and aquifer.

Both variables were the result of previously conducted measurement campaigns. The following variables and parameter were calibrated: recharge, head-dependent flux boundary condition and hydraulic conductivity, which was calibrated in 112 individual calculation zones. In the top layer of the modelled riverbed, the applied values of hydraulic conductivity were the same as those determined based on measurements (Table 2). During the calibration, the hydraulic conductivity values of the upper sand layer at the bottom of the river were not changed. Only the values of parameters characterizing the deeper zones were changed. The calibration process was conducted using two calibration methods – manual and automatic (by means of the FEPEST software) and it was applied to the remaining elements. The range of values of the hydraulic conductivity has been adopted according to the information contained in a geological map of this area (Baranicka 1976) and they are comparable with other data from the Świder river valley.

There are no measurements of this parameter located in the immediate vicinity of the discussed cross-section. After calibration, the spatial distribution of hydraulic conductivity was obtained (Fig. 9). The key parameters of the model and their values before and after calibration are presented in Table 3.

The model results (calculated differences between water level in the river and aquifer and calculated filtration intensity) after the calibration process are presented in Figure 10. They reveal a very high correspondence between the measurements and the calibrated model (in terms of water exchange – filtration intensity, and differences between the water levels in the aquifer and river).

The measured water filtration intensity was accurately reproduced in the model, within the measurement uncertainties or simulation assumptions. The difference in water levels in the river and aquifer was reproduced very well on the left bank of the river (measurement points 2–6). On the right bank, the error averaged 2 cm. Lack of accurate calibration of the differences of water levels on the right bank of the river results

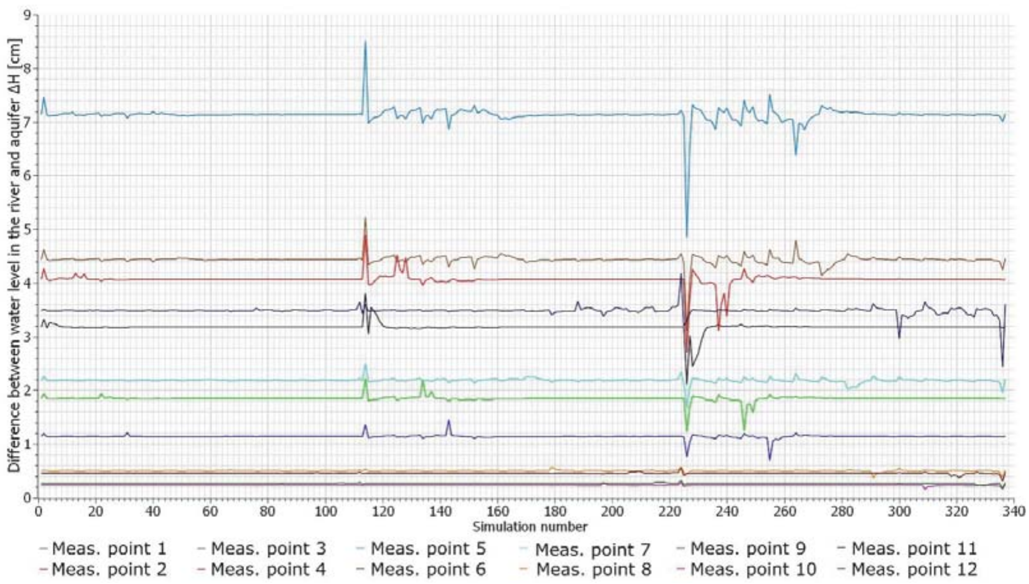

Fig. 8. Sensitivity analyses of the model (for changes in hydraulic conductivity values)

from the simultaneous use of two calibration parameters. An attempt to model better agreement of water levels in the aquifer and river led to a less accurate estimate of the value of water exchange volume (filtration intensity).

Model verification

The evaluation of the usefulness of the model for forecasting involved the verification of results of the second measurement series, conducted on 11th November 2018.

The measurements in November 2018 were preceded by a week of negligible rainfall (the average daily rainfall was 0.59 mm/d vs 7.56 mm/d during the week before the first measurements) which caused the water level in the river to fall. The water level in the aquifer was assumed to have decreased, hence the boundary conditions on both sides of the aquifer were defined as lower than those in the calibrated model. Several simulations were carried out to minimize the difference between the model and the measurement considering both pressure difference measurements and filtration intensity

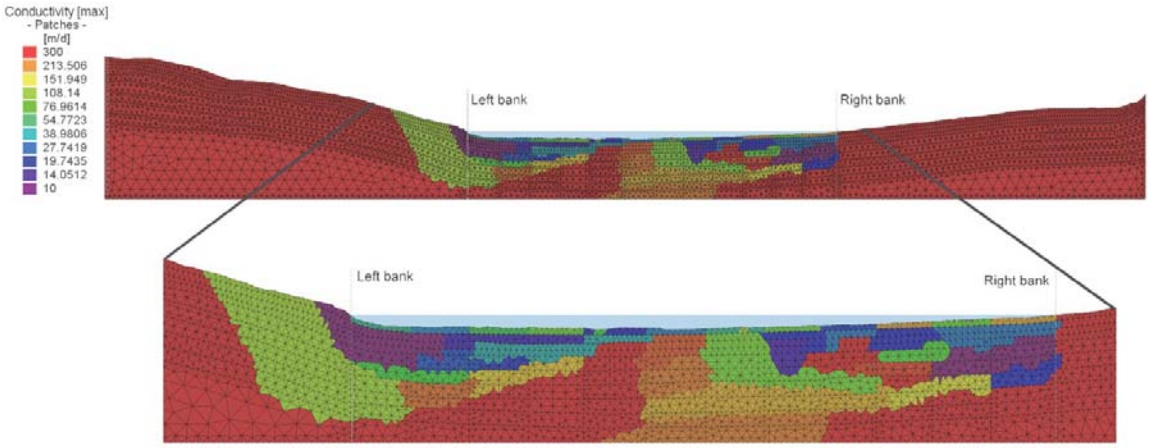

Fig. 9. Spatial distribution of hydraulic conductivity after calibration

Table 3. Model parameters before and after calibration

Recharge values [l/d]			Hydraulic head (head-dependent flux boundary condition) [m]			Hydraulic conductivity [m/d]	
Land cover	Before calibration	After calibration	Location	Before calibration	After calibration	Before calibration	After calibration
Forest	1,51	0,756	Left bank	96	96,2	18–188 Table 2	10–300 spatial distribution on figure 9
Sand	1,51	1,51	River	93	93,2		
River	1,51	10–4	Right bank	92,5	92,5		

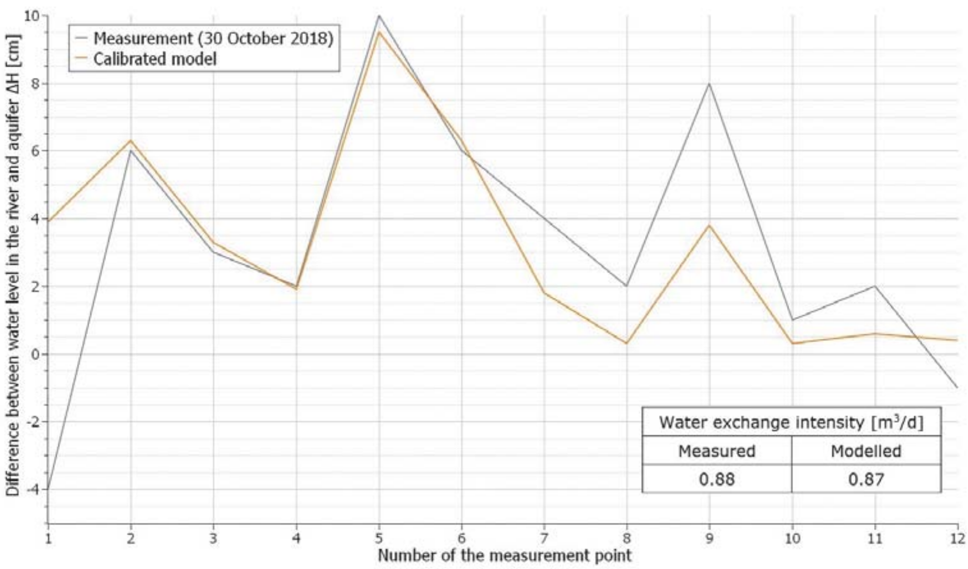

Fig. 10. Model results after calibration

between the river and aquifer. During this process, the hydraulic head for both sides of the model was changed. The hydraulic conductivity remained unchanged.

The verification of the model showed satisfactory results concerning differences in water heads (Fig. 11). Like in the case of the calibrated model, the differences are well reproduced on the left bank of the river (measurement points 2–6). In the middle (points 6–9), the error is higher, reaching approximately 4 cm. The error results in imperfect calibration at these points for the results of the water head differences. It should be noticed that the model calibration process was based not only on the pressure difference measurements but also on agreement of the measured water exchange (filtration intensity) between the river and aquifer. Thus, the differences between the model results and the measurements obtained during the verification process are acceptable. The model verification showed good reproduction of the actual results for the measurement points on the right bank (points 10–12).

Discussion

It is essential to assess the variability of the water exchange through the river length and through the cross sections for the preparation of data and calculation of the groundwater flow models. In the case of mathematical modelling of groundwater flow, for models at a regional scale, the interaction between the river and aquifer is usually introduced into the model using a head – dependent flux boundary condition. Many studies confirm differences in the intensity of water exchange between the river and aquifer with consideration of different points along the length of the river (e.g. Boano et al. 2006; Revelli et al. 2008). These differences can also be seen over a longer period of time or/and after hydrological events like flood (Kasperek et al. 2012). The literature, however, provides no studies or recommendations regarding the proposed averaged distances with different water exchange flux between the river and aquifer. The considerations in the literature regarding the possibility of averaging the width of the river and dividing a computing cell defined as a river into a larger number of computing layers (Brunner et al. 2010). The model resolution, i.e., grid size and length of a river reach that needs to be evaluated to get a representative average amount of hyporheic exchange for a study site, is not known. Because this variability has not been sufficiently documented in the literature so far, the results obtained in this research will help to understand this scientific issue better. After this initial step of the study, further research will include not only other cross-sections, but also the identification of groundwater-surface water interactions along the length of the river. Thus, it will be possible to determine the length of the river on which a pattern of spatial variability of the water exchange stream in the hyporheic zone may be taken as an average, providing for a sufficiently good representation of reality.

Nowadays, there are not enough common measuring methods which will allow for direct quantitative assessment of the filtration flux. The measuring devices used in this research were also successfully used and tested in the Parsęta River, Poland (Chudziak 2013). The results of the measurements of the specific discharge and the hydraulic conductivity of sediments have shown the variability of measured values along the river (Chudziak 2013). Anibas et al. (2015) have 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 in a section of the Biebrza River in Poland. All the studies presented so far have been focused on representing the variability along the river rather than in the cross section.

After the analysis and interpretation of the fieldwork and modelling studies, the main conclusion is that the research on the spatial variability of water exchange between the river and the aquifer should be investigated further. As natural objects, rivers are strongly connected with the other components of the environment, varying highly along its reaches. The cross-section presented here is a first step in the field of research into their heterogeneity. The measurement results confirmed the occurrence of spatial variability of the water exchange stream in the hyporheic zone of the analyzed river cross-section. Analysis of the measurements results revealed the

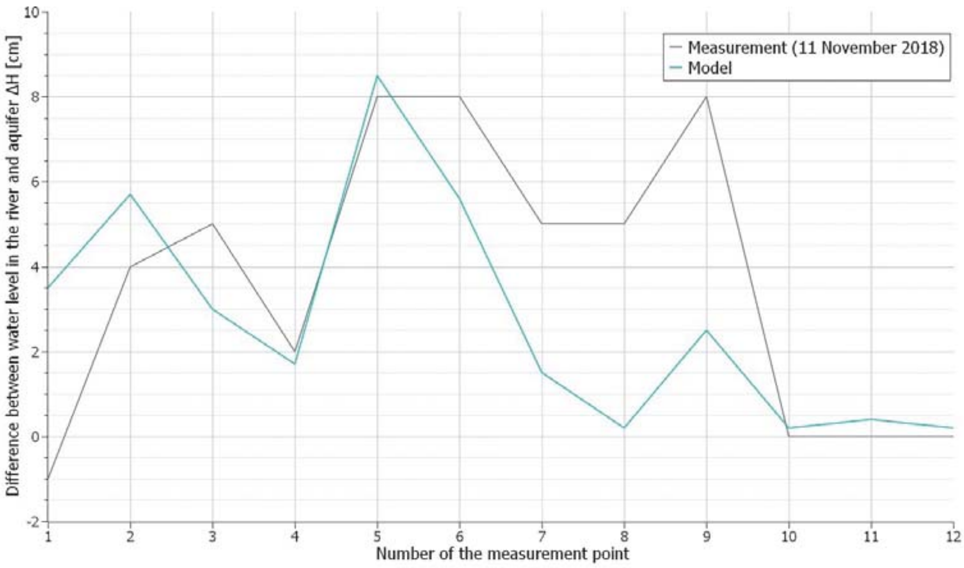

Fig. 11. Water level difference between the aquifer and surface water, model verification (11 November 2018)

heterogeneity in the water exchange in the hyporheic zone of the river in its cross-section.

Although both measurement series were conducted under different meteorological conditions, their results showed similar spatial variability of the hydraulic gradient in the river cross-section. This correlation was also confirmed by the modelling of groundwater flow, permitting for the reproduction of the repeatable spatial variability of the hydraulic gradient in the examined cross-section. Obtaining such agreement indicates that the process of water exchange between the aquifer and the river depends mainly on the spatial distribution of hydraulic conductivity of the aquifer under the river bottom.

The characteristics of the section of the river, in particular the change of spatial and temporal water exchange between the river and the aquifer, may have great consequences in considering the impact of this exchange on the other environmental components of the environment. However, the observed heterogeneity may be caused by the local characteristics of the differential pressure of the river and may also be an effect of heterogeneity of sediments within the channel (Anibas 2015), so it should be further examined.

Conclusions

The main goal of the presented studies was to characterize the spatial distribution of the surface water-groundwater interaction within the river cross-section. It has been achieved by through conducting two series of measurement campaigns and developing a groundwater flow model. The model was verified using data from both measurement series. A satisfactory correlation was determined between the model results and the observations, which indicates the possible potential of the studied subject to contribute towards improving the process of modelling water exchange in the hyporheic zone. The use of simultaneous measurements of gradient differences between the water level in the aquifer and river and water exchange (filtration intensity) enabled determining the value of hydraulic conductivities in the upper sand layer at the river bottom to be determined. On the other hand, the use of this information in the river-aquifer water flow model allowed for better identification of hydraulic conductivity throughout the entire cross-section (in deeper zones).

The Authors described the spatial distribution of the surface water-groundwater interaction within the river cross-section located in Poland (the Świder river). The study has been conducted using the measurements (two series) and modelled using a 2D numerical model. Both measurement series, however, exposed the river's losing type at the bed boundaries. This suggests potential variability of the direction of water exchange in the hyporheic zone of the river, which would need to be further examined in the next phases of the study. The presented study was an initial step in determining this pattern of spatial variability of the water exchange stream in the hyporheic zone.

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Przestrzenny rozkład wymiany wody w przekroju rzeki – pomiary i model numeryczny

Streszczenie: Problem opisu interakcji wód powierzchniowych i podziemnych jest podejmowany przez wielu badaczy, między innymi ze względu na intensyfikację rolnictwa i przemysłu przejawiające się m.in. regulacją i pogłębianiem osadów rzecznych rzek. Problemy te są obecnie szeroko dyskutowane na międzynarodowym forum polityki i gospodarki wodnej. Do przygotowania danych i obliczenia modelu przepływu wód podziemnych niezbędna jest ocena zmienności wymiany wody na długości rzeki i jej przekrojach. W artykule przedstawiono badania opisujące przestrzenną zmienność interakcji wód powierzchniowych i podziemnych w przekroju rzeki. Aby opisać tą zmienność, przeprowadzono dwie kampanie pomiarowe. Dodatkowo opracowano model przepływu wód podziemnych w celu przedstawienia zmiennego charakteru wymiany wody w strefie hyporeicznej w przekroju rzeki. Model został pomyślnie zweryfikowany za pomocą pomiarów przepływu wody w strefie hyporeicznej. Dokładny przestrzenny opis zmienności przestrzennej intensywności wymiany wody pomiędzy rzeką a warstwą wodonośną jest pierwszym krokiem do określenia możliwości dokładnego wprowadzenia tej zmiennej do budowy modeli matematycznych przepływu wód podziemnych.