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Combining flux estimation techniques to improve characterization of groundwater–surface-water interaction in the Zenne River, Belgium

J. Dujardin · C. Anibas · J. Bronders · P. Jamin · K. Hamonts · W. Dejonghe · S. Brouyère · O. Batelaan

Abstract The management of urban rivers which drain contaminated groundwater is suffering from high uncertainties regarding reliable quantification of groundwater fluxes. Independent techniques are combined for estimating these fluxes towards the Zenne River, Belgium. Measured hydraulic gradients, temperature gradients in conjunction with a 1D-heat and fluid transport model, direct flux measurement with the finite volume point dilution method (FVPDM), and a numerical groundwater flow model are applied, to estimate vertical and horizontal groundwater fluxes and groundwater–surface-water interaction. Hydraulic gradient analysis, the temperature-based method, and the groundwater flow model yielded average vertical fluxes of –61, –45 and –40 mm/d, respectively. The negative sign indicates upward flow to the river. Changes in exchange fluxes are sensitive to precipitation but the river remained gaining during the examined period. The FVPDM, compared to the groundwater flow model, results in two very high estimates of the horizontal Darcy fluxes (2,600 and 500 mm/d), depending on the depth of application. The obtained results allow an evaluation of the temporal and spatial variability of estimated fluxes, thereby helping to curtail possible consequences of pollution of the Zenne River as final receptor, and contribute to the setup of a suitable remediation plan for the contaminated study site.

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J. Dujardin (✉) · C. Anibas · O. Batelaan
Department of Hydrology and Hydraulic Engineering,
Vrije Universiteit Brussel, Pleinlaan 2, Brussels, 1050, Belgium
e-mail: judujardin@gmail.com
Tel.: 003226293035

J. Dujardin
Cartography and GIS Research Unit Department of Geography,
Vrije Universiteit Brussel, Pleinlaan 2, Brussels, 1050, Belgium

J. Bronders · K. Hamonts · W. Dejonghe
VITO (Flemish Institute for Technological Research), Boeretang 200, Mol, 2400, Belgium

P. Jamin · S. Brouyère
Hydrology Unit Department ArGEnCo,
University of Liège, Building B52/3, Sart Tilman, 4000, Belgium

O. Batelaan
National Centre for Groundwater Research and Training School of the Environment,
Flinders University, GPO Box 2100, Adelaide, SA 5001, Australia

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Introduction

Since contaminated sites pose a significant risk to water resources (EC 2006; EEA 2007), national and international regulations like the European Water Framework Directive (EU 2000) mandate the protection of linked groundwater–surface-water systems and ask for a reliable assessment of fluxes across the groundwater–surface-water interface (Schmidt et al. 2008). Efficient remediation of polluted sites needs integrated management practices, especially for complex and large-scale pollution; risk management therefore is gaining importance in science and engineering (Bardos et al. 2002). Van Keer et al. (2009) describe a methodology for risk management for polluted sites as developed in the framework of the EU FP5 Welcome project. Another risk management plan for brownfield sites (or 'brownfields') has been introduced in Flanders, Belgium, by Bronders et al. (2007, 2008). The characterization phase of a risk management plan and the quality and reliability of its risk assessment relies strongly on the information on the water fluxes in the system. Cirone and Duncan (2000) and Smith (2005) for example outline the importance of proper estimates of the groundwater flux and groundwater–surface-water interaction of contaminated sites. In saturated conditions, the movement of water is the main vector for pollutant transport; its determination is needed to analyze the source–receptor pathway and to evaluate movement, behaviour and fate of pollutants. Hence, the measurement, calculation and modelling of groundwater fluxes are very important; great care has to be taken that proper methods are used and combined.

Many groundwater–surface-water related exchange processes take place in the hyporheic zone, i.e. the saturated sediments beneath and beside streams and rivers where groundwater and surface water is actively mixed (Hayashi and Rosenberry 2002). Characterized by relatively strong biogeochemical process rates (McClain et al. 2003; Triska et al. 1993), the hyporheic zone is seen as a potential sink or source of pollutants (Smith 2005). Due to hydrological connectivity, the hyporheic zone is connected with other landscape compartments including the aquifer and the riparian zone (Bracken and Croke 2007; Lexartza-Artza and Wainwright 2009); the exchange processes, therefore, are site dependent and may have a large variability in time and space (Sophocleous 2002).

Field methods providing spatial and/or temporal distributed estimates of groundwater–surface-water interaction have been comprehensibly described by Kalbus et al. (2006) and Rosenberry and LaBaugh (2008). They can be based on the following: Darcy's law, e.g. using piezometer nests and boreholes placed in the riverbed and/or in the adjacent riparian zone (Baxter et al. 2003; Cey et al. 1998); differential stream discharge gauging (Becker et al. 2004); numerical modeling (Cardenas and Zlotnik 2003; Fleckenstein et al. 2004); use of tracers like heat (Anderson 2005; Anibas et al. 2011), dye, salt, chloride or stable isotopes (Carey and Quinton 2005; Unland et al. 2013); and remote sensing (Dujardin et al. 2011; Loheide and Gorelick 2006).

Due to scale problems (Kikuchi et al. 2012), the heterogeneity of the underground (Schornberg et al. 2010) and limited possibilities for direct measurement, the quantification of the groundwater–surface-water flux is a challenging task. Moreover, all methodologies to quantify groundwater–surface-water interaction have distinct limitations and can only capture the exchange at a specific spatial or temporal scale (Kalbus et al. 2006). Methods based on Darcy's flux are hampered by difficulties in estimating the hydraulic conductivity (Chen 2000). This realization results in a focus on using different methods and combinations of them. Becker et al. (2004) combine stream and streambed temperature surveys with stream flow measurements to assess groundwater discharge to a stream. In another study, Unland et al. (2013) state that by combining several methods, including differential stream gauging and chemical mass balances, with temperature and electrical conductivity surveys, the applicability of each technique can be evaluated. Anibas et al. (2011), for example, state that using heat as a natural tracer as a stand-alone technique is possible, but it is preferably combined with other field methods. It can be concluded that because of the limitations and uncertainties associated with a single method, any attempt to reliably characterize groundwater–surface-water interactions will benefit from a multi-scale approach combining different techniques (Kalbus et al. 2006; Hyun et al. 2011; Kikuchi et al. 2012).

This case study, therefore, aims to determine the vertical and horizontal groundwater–surface-water interaction from a brownfield towards an urban river by independently applying and combining different field methods and a numerical groundwater flow model of the study area, Vilvoorde-Machelen and the Zenne River, Belgium. The field methods are (1) measurement of hydraulic gradients; and (2) thermal method (Anibas et al. 2009) to determine vertical groundwater fluxes; and (3) the finite volume point dilution method (FVPDM; Brouyère et al. 2008) which investigates the horizontal groundwater fluxes.

Study area

The different groundwater flux estimation techniques were applied on the industrial area of Vilvoorde-Machelen, located about 10 km north-east of Brussels, Belgium (Fig. 1). The study area of 10 km^2 is located in the Zenne catchment, which covers an area of about 600 km^2. Beside some minor rivers (including the Trawool River, the Woluwe River and the Vondelgracht), the Brussels-Scheldt Canal flows parallel to the Zenne River in a S–N direction through the study area (Boel 2008). Figure 2 illustrates the domain of the groundwater flow model with the positions of the boreholes SB1, SB2 and PB9, and point T, which are the locations where field measurements were performed. Figure 2 also shows the contours of the phreatic groundwater level indicating groundwater head gradients in the N and N–W direction and towards the Zenne River.

The topography in the study area ranges from 10 to 50 m, with an average value of 16 m above sea level and with a mean slope of 1.3 %. The average precipitation is 852 mm/year (average values for 1981–2010; KMI 2013). The dominant soil type of the area is silty loam, while in the northeastern part some clay-loam occurs.

Since 1835, the Vilvoorde area has been a major industrial site, with considerable chemical industrial activity. The study area contains a number of well-known contaminated sources (OVAM 2003–2006; Boel 2008). Field investigations indicate the presence of an extensive regional contaminant plume, containing a mixture of BTEX (benzene, toluene, ethylbenzene and xylene), polycyclic aromatic hydrocarbons (PAH) and chlorinated aliphatic hydrocarbons (CAH). While Bronders et al. (2007) estimated the size of the contaminant plume as 1.2 by 0.6 km, Dujardin et al. (2011) identified the Zenne River as its final receptor.

Methods and measurements

Measured hydraulic gradients

Vertical hydraulic gradients across the streambed can be derived by comparing water level measurements above the streambed (level of the river) and the piezometric groundwater levels below the riverbed. This allows estimation of fluxes through the streambed using Darcy's Law (Kalbus et al. 2006). Vertical water fluxes (q) through

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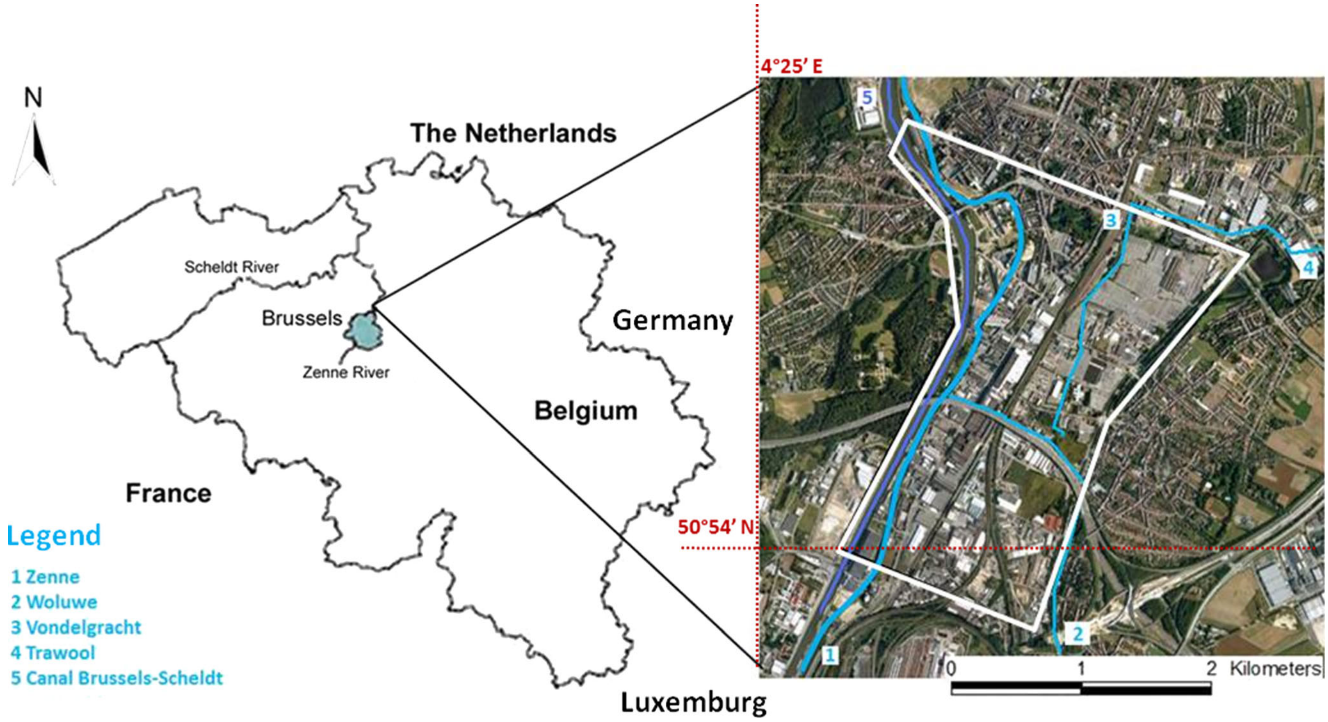

Fig. 1 Location of the study area within Belgium. The domain of the groundwater model is indicated by the *white line*; the major surface water bodies are shown as *blue lines*

the streambed can be estimated as

$$q = K_z.i \tag{1}$$

where K_z is the vertical hydraulic conductivity of the streambed and i is the vertical hydraulic gradient. Surface-water levels of the Zenne River were measured at location T (Fig. 2), near the right riverbank, while groundwater levels were measured at SB2 (8–9 m depth). The measurements were continuously measured, every 30 min, from November 2005 till April 2007.

Porous aquifers often show substantial differences between the horizontal and vertical hydraulic conductivity of one or two orders of magnitude (Freeze and Cherry 1979; Chen 2000). A horizontal hydraulic conductivity of 2.5 m/d was used in the groundwater model (Boel 2008);

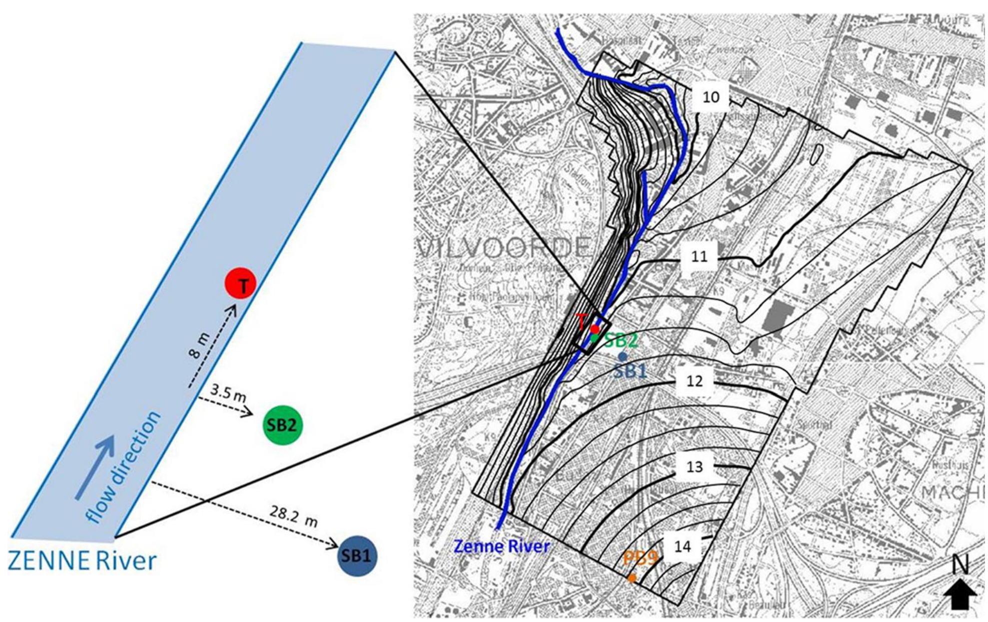

Fig. 2 The model area with phreatic groundwater level contours (*black lines*; m above sea level) showing a gradient in the northern direction and towards the Zenne River. Position of the boreholes *SB1*, *SB2* and *PB9*, as well as the locations where temperatures and hydraulic heads were measured in the river (point *T* indicated by a *red dot*; *highlighted on the left*); slug tests and FVPDM tests were performed in *SB2* and *PB9*

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following Anibas et al. (2011), an anisotropy factor of 9 was used to obtain a vertical hydraulic conductivity K_z of 0.28 m/d. In case no other estimates are available to characterize riverbed conductivities, the sandy aquifers occurring in Flanders are assumed to have anisotropies of about 10 (e.g. Woldeamlak 2007). Here, an anisotropy value was used which was derived for the Flemish lowland Aa River (Anibas et al. 2011). This approach is justified by the fact that the Aa River and the Zenne River have a comparable hydrology and hydrogeology.

Thermal method

At position T (Fig. 2), temperature measurements in the riverbed were performed every hour between November 2005 and April 2007 using probes at about 0.2, 0.6 and 1.20 m depth. For the heat transport model, however, temperatures averaged over 6-h intervals were used. Between May 2006 and September 2006, no temperature measurements were available.

Temperature gradients measured in the riverbed can be used to estimate interaction between groundwater and surface water (Anderson 2005; Constantz 2008). Given the fact that groundwater temperatures are relatively stable throughout the year and stream temperatures vary on a seasonal and daily basis, the flow of groundwater causes disturbances of the natural temperature-depth distributions. By applying combined one-dimensional (1D) heat and fluid transport modeling, these variations can be used to compute point estimates of vertical groundwater–surface-water exchange fluxes. This thermal method has proved to be reliable (Anibas et al. 2009; Lautz 2010), not least because gathering of thermal data, the parameter estimation, the establishment of model boundary conditions and the model calibration are relatively simple.

Different implementations of the thermal method exist (Anderson 2005); most commonly exchange rates are quantified by inverse modeling of measured temperature profiles (Schmidt et al. 2007; Anibas et al. 2011) or their time series (Keery et al. 2007; Anibas et al. 2009) in riverbeds composed of unconsolidated sediments.

In this research STRIVE (Stream RIVer Ecosystem), a package of subroutines introduced in the FEMME ecosystem modeling platform (Soetaert et al. 2002; Anibas et al. 2009) was applied to determine groundwater–surface-water interaction. Based on Lapham (1989) STRIVE contains a numerical 1D heat transport module.

$$\frac{\kappa_e}{\rho c}\nabla^2 T - \frac{\rho_w c_w}{\rho c}\nabla\cdot(Tq) = \frac{\partial T}{\partial t} \tag{2}$$

Equation (2) describes the combined heat and fluid transport of an incompressible fluid through a homogeneous porous media, where T is the temperature at depth z [m] and time t [s] in the soil in K [°C], c_w the specific heat capacity of the fluid [J/kgK], ρ_w the density of the fluid [kg/m^3], c and ρ are the specific heat capacity and density of the sediment-fluid matrix [J/kgK] and [kg/m^3] respectively. q is the seepage velocity or specific discharge vector [m/s]. For presentation purposes, however, the units mm/d are used to indicate fluxes. Fluxes with a negative sign stand for movement of groundwater in direction to the surface (i.e. a gaining reach), whereas fluxes of water from the river into the soil (i.e. a losing reach) have a positive sign. κ_e is the effective thermal conductivity of the soil-water matrix [J/smK].

The main advantage of the thermal method is its simple parameterization, since the physical parameters (e.g. κ_e or c) have a limited range (Stonestrom and Constantz 2003; Anderson 2005). The thermal parameters, based on Anibas et al. (2011), who successfully simulated groundwater–surface-water interaction at the Aa River, Belgium, were applied in this study. The Aa River is in its thermal and hydrogeologic characteristics comparable to the Zenne River. Table 1 shows thermal parameters used for the STRIVE model for the Zenne River.

A time series of surface-water temperatures and a constant groundwater temperature at 5 m depth constitute the boundary conditions of the heat transport model. The groundwater temperature is based on the mean annual surface temperature of the study area (i.e. average 1981–2010 was 10.5 °C; KMI 2013). At a depth of 5 m the groundwater temperature shows a relative constant behavior in time, which is about 1–2 °C above the mean annual surface temperature (Anderson 2005). In this case, a groundwater temperature of 12.0 °C was used.

STRIVE contains routines to calibrate the vertical flux until a best fit is obtained between the simulated temperature profiles and the measured temperature distributions (Anibas et al. 2009). Transient thermal modeling is applied in which the first temperature profile of a measured time series at three depths (0.2, 0.6 and 1.2 m) is used to initialize the model, while the others are used for calibration. The computation of the temporal resolution of groundwater–surface-water interaction is possible with STRIVE by splitting the time series in equal parts of the length of 1 week (Dujardin et al. 2011).

Groundwater flow model

The groundwater flow in the study area is simulated with a steady-state groundwater flow model (Touchant et al. 2007; Dujardin et al. 2011) built in MODFLOW-2000 (Harbaugh et al. 2000). Figure 3 shows a hydrogeological

Table 1 Applied input parameters for the STRIVE model

Parameter	Symbol	Value	Unit
Density of the saturated sediment	ρ	1,965	kg/m^3
Specific heat capacity of the saturated sediment	c	1.365	J/kgK
Thermal conductivity of the saturated sediment	κ_e	1.833	J/smK
Density of water	ρ_w	1,000	kg/m^3
Specific heat capacity of water	c_w	4,180	J/kgK

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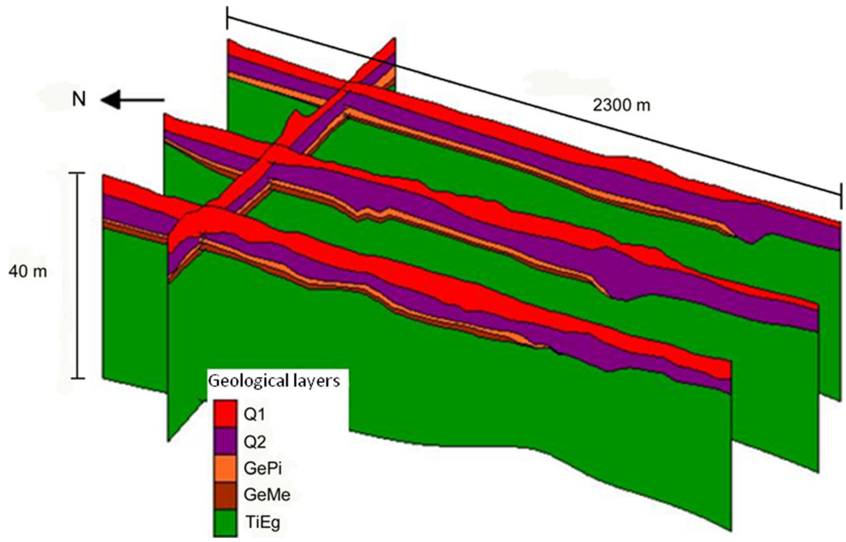

Fig. 3 N–S and E–W hydrogeological cross-sections of the subsurface of the study area (*Q1* silty eolian sands; *Q2* gravel; *GePi–GeMe* Ghent Formation; *TiEg* Tielt Formation; Dujardin et al. 2011, with permission from Elsevier)

fence diagram of the five model layers (from top to bottom):

1. Silty eolian sands (Q1; 2–12 m thick)
2. Gravel in a silty sand matrix (Q2; 3–10 m thick)
3. Wedge-shaped Ghent Formation of which the upper part consists of densely packed sand layer (GePi; 4 m thick)
4. The lower part of the Ghent Formation consists of clay-containing silty sand (GeMe; 2 m thick)
5. Tielt Formation (TiEg; 20–30 m thick) consisting of silty, glauconite fine sands

The hydrostratigraphy was interpreted by 68 borehole loggings and 179 cone penetration tests, available from the Geological Database for the Subsoil of Flanders (DOV 2008). Parameters for hydraulic conductivity of the model layers are described in Table 2; the storage coefficient S_s= 1×10^{-5} 1/m was obtained from literature (i.e. Morris and Johnson 1967).

The model boundary conditions are a combination of the employed MODFLOW packages: River, General Head, Specified Head and Recharge. The lower boundary is defined at the top of the Tertiary clay-rich Kortrijk Formation; considered as almost impervious, it is thus implemented as a no-flow boundary. Beginning from the north, the other model boundaries are defined in clockwise direction as follows (Fig. 1): no flow and river boundary (at the Trawool River), constant head at the eastern boundary (based on head measurements inside and outside the model area), no flow boundary in the south and constant head due to the Brussels-Scheldt Canal in the west. The no flow boundaries are justified by the regional groundwater flow which is parallel to the northern and southern model boundaries.

The upper boundary is defined by groundwater recharge. Because the spatial distribution of the groundwater recharge has an impact on the modeled groundwater–surface-water interaction, distributed recharge estimations were acquired from Dujardin et al. (2011) using the WetSpass methodology (Batelaan and De Smedt 2007). The average recharge used in this model is 159 mm/year with a standard deviation of 91 mm/year.

To study groundwater–surface-water interaction in the Zenne River the River package (RIV) is employed. Field investigations delivered average values for river water level h=9.6 m and river depth d=1.2 m. The conductance of the river sediments C= 2.66 m^2/d is calculated as

$$C = KA/b \tag{3}$$

where, K the hydraulic conductivity of the bottom sediments [m/d], A the surface area of the bottom sediments [m^2] and b is the thickness of the sediment layer [m]. A more extensive description of the applied groundwater model and its internal and external boundary conditions can be found in Touchant et al. (2007), Boel (2008) and Dujardin et al. (2011).

Table 2 Applied hydraulic conductivity (K) values for the groundwater flow model

Layer	Designation	K_{xy} [m/d]	K_z[m/d]
1	Q1	2.5	2.5
2	Q2	12.5	6
3	GeMe	0.5	0.5
4	GePi	4	3
5	TiEg	6	4

FVPDM tests

Slug tests, hydraulic gradients and pumping tests in piezometers have traditionally been used to estimate hydraulic conductivities and to constrain groundwater

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fluxes (Butler 1998). Another technique applicable on piezometers is the finite volume point dilution method (FVPDM; Brouyère et al. 2008).

This novel method was developed to overcome the difficulties of implementing classical point dilution methods (PDM), like instantaneous and uniform mixing of tracer into the well without disturbance of groundwater fluxes (Drost et al. 1968; Haveley et al. 1967). The FVPDM is a tracer technique that has been introduced in 2003 (Brouyère 2003; Brouyère et al. 2008), generalizing all the single-well point dilution tests to almost any tracer injection scenario. This method is based on a mathematical and numerical model for the tracer injection into a well and considers the mass balance of the injection of tracer fluid and transiting groundwater flow passing through the well screen (Brouyère et al. 2005). The analytical solution of this model applied to a single well tracer technique enables the accurate measurement of transit flow rate and, thus, of the Darcy fluxes (Eq. 4).

As a field method, the FVPDM is based on a controlled continuous injection of a tracer into a well and on the monitoring of its concentration over time. During the whole experiment, the water column within the well is mixed to insure a homogeneous repartition of the tracer mass. The tracer concentration is proportional to the groundwater fluxes that flush the tracer out of the well. The FVPDM experiment can be maintained as long as tracer fluid and power supply for injection and mixing are available, in which case a continuous temporal monitoring of the variations of groundwater fluxes is possible.

$$C_{\mathrm{w}}(t) = \frac{Q_{\mathrm{in}}C_{\mathrm{in}} - \left(Q_{\mathrm{in}}C_{\mathrm{in}} - Q_{\mathrm{out}}C_{\mathrm{w},0}\right)\exp\left(-\frac{Q_{\mathrm{out}}}{V_{\mathrm{w}}}(t-t_0)\right)}{Q_{\mathrm{out}}} \quad (4)$$

with

$$Q_{\mathrm{out}} = Q_{\mathrm{in}} + Q_{\mathrm{t}}^{\mathrm{in}} \quad (5)$$

C_{w}, C_{in} and $C_{\mathrm{w},0}$ are tracer concentrations [M/L^3] in the well, in the injection water and in the well at time t_0 respectively. Q_{in}, $Q^{\mathrm{in}}{}_{\mathrm{t}}$ and Q_{out} [L^3/T] are the injection rate, the transit flow rate corresponding to the groundwater flow intercepted by the well screen that is directly related to the apparent Darcy's flux (vD) and the flow rate leaving the well through the screen carrying tracer at concentration C_{w}. V_{w} is the volume of water in the injection well, assumed to be constant.

The FVPDM tests have been executed in SB2 and PB9, beginning with piezometer SB2_F2 in September 2008 (Fig. 2). This piezometer with a diameter of 2 inches (5.1 cm) consists of two well screens, F1 (7–8 m depth) and F2 (9–10 m depth). The tracer uranine was injected with two flow steps: T_1 at 0.01 L/min for 181 min and T_2 at 0.02 L/min for 128 min. Piezometric head, temperature, turbidity and uranine concentrations were monitored for 350 min from the beginning of the test using a LevelTroll probe (In-Situ Inc., Ft. Collins, CO, USA) and with a field fluorimeter (GGUN-FL30#1370; Geomagnetism Group, University of Neuchâtel, Neuchâtel, Switzerland) respectively. A second FVPDM test was performed on PB9_F1 (screen 4.3–6.3 m depth) in November 2008. Uranine tracer was injected at 0.006 L/min for 120 min, with piezometric head, temperature, turbidity and uranine concentration monitored for 180 min from the beginning of the test respectively using the LevelTroll probe and the field fluorimeter.

Results

Measured hydraulic gradients

Based on hydraulic gradients between the piezometric groundwater level and the river stage, Fig. 4 shows the flux estimates as a blue line. The fluxes are negative (i.e. a gaining river), varying around –60 mm/d. The lowest vertical fluxes occur in February 2006, with an average of –36 mm/d, the highest values are observed in December 2006 and March 2007, with averages of –84 mm/d.

Thermal gradients

Figure 4 also indicates average weekly vertical water fluxes (red line) simulated with STRIVE. It can be observed that only groundwater discharge to the Zenne River occurred, on average about –45 mm/d; missing parts in the line graph indicate periods were no reliable output could be generated or where no temperature data were available. This is especially the case for periods where no strong temperature gradients are present between the groundwater and the surface water; hence, in spring and autumn (Anibas et al. 2009).

Groundwater flow model

The groundwater model was calibrated using observed heads from 27 observation wells with measured heads between 1999 and 2006. The calibration resulted in a root mean squared error (RMSE) of 0.32 m and an absolute error (AE) of 0.02 m. The general orientation of the groundwater flow is north-west (Fig. 2). In the western part of the area, the groundwater flow is oriented towards the east, because of the draining effect of the Zenne River. A water budget for the Zenne River was calculated; Fig. 5 and Table 3 show the global inflow and outflow of the Zenne River in the study area.

It is clear that the Zenne River is receiving groundwater from:

1. The Brussels-Scheldt Canal (B-S Canal), producing a flux from the west
2. The regional groundwater flow, coming from the south-east

The B-S Canal discharges 7,054 m^3/d (over a length of 3,830 m), representing 72 % of the total inflow into the

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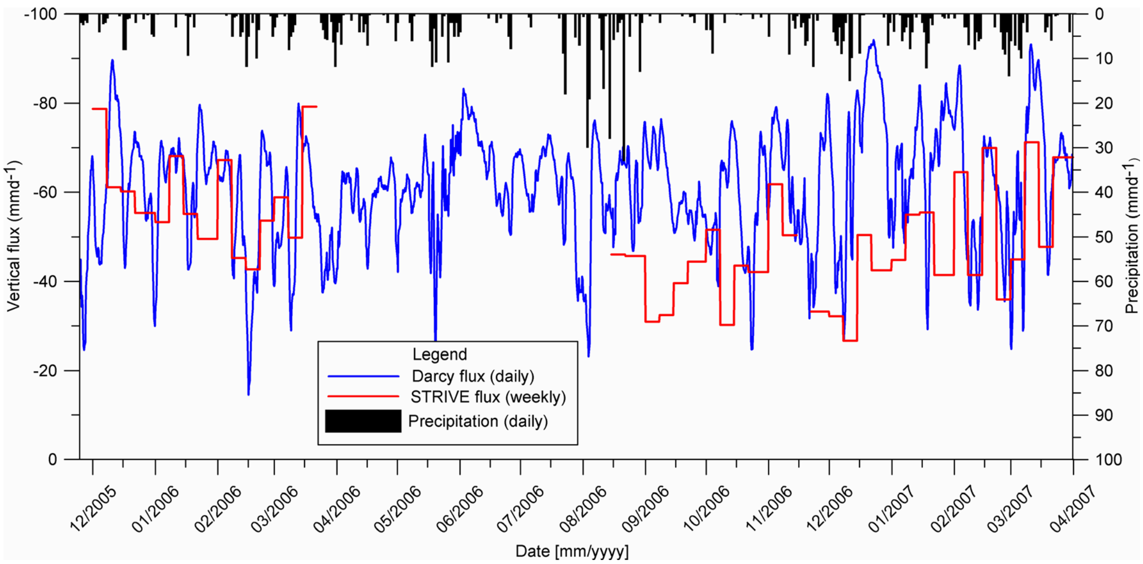

Fig. 4 Calculated daily vertical groundwater flux (*blue line*) based on measured hydraulic gradients and weekly measured thermal gradients using the STRIVE model (*red line*) and daily precipitation (*black lines*). Short-term fluctuations, sensitive to precipitation, dominate. The river is gaining throughout the 16-month period; no clear seasonal pattern is visible

Zenne River; the remaining 28 % of the total inflow is coming from the regional groundwater flow (2,690 m^3/d). The upper geological layer (Q1) is characterized by horizontal fluxes between 0 and 80 mm/d. Figure 6 shows vertical groundwater fluxes in the study area as simulated with the numerical groundwater flow model. The colored cells show upward fluxes, while the white cells have downward fluxes. From Fig. 6 it is clear that the Zenne is gaining, the fluxes vary from –12 to –40 mm/d.

FVPDM tests

The results of the first FVPDM test (SB2_F2) showed that the tracer injection rate was too high for the very low groundwater flow prevailing during the test. A better dimensioning was achieved in the second FVPDM test PB9_F1. Figure 7 shows the comparison of measured and simulated concentration of tracer using FVPDM method. The groundwater flow velocity was determined by simulating the tracer elution curve and comparing this curve with experimental data. Simulated concentrations were adjusted by modifying only the apparent Darcy's flux (vD). The other terms of Eq. (3) are based on experimental conditions.

In Fig. 7, the solid line corresponds to the best adjustment of Darcy's flux for the experimental conditions. The dashed and dotted lines indicate the sensitivity of the method to the magnitude of Darcy's flux. The ascending part of the simulated curves matches almost perfectly with the experimental measurements. A small gap can be observed at the beginning of the test on PB9_F1. This is probably due to a longer homogenization time of the tracer over the whole height of the well. The FVPDM tests allowed calculating a horizontal Darcy's flux of 2600 mm/d for SB2_F2 and a Darcy's flux of 500 mm/d for PB9_F1. The differences observed between the horizontal Darcy's flux of SB2_F2 and PB9_F1 can be explained by local variation of aquifer hydraulic conductivity as is frequently observed in such alluvial aquifers.

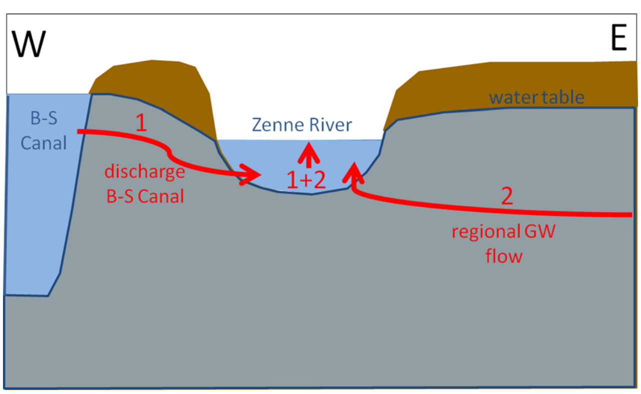

Fig. 5 Schematic view of the different components of the water budget of the Zenne River (*B-S Canal* Brussels-Scheldt Canal, *GW* groundwater). See Table 3

Discussion

Table 4 summarizes the different estimations of vertical fluxes using three methodologies: hydraulic gradients, the thermal method and the groundwater flow model. To compare the resulting vertical fluxes from the former with the vertical fluxes obtained from the groundwater flow model, Table 4 contains the vertical flux simulated in the model cell located around SB2.

Table 4 indicates that all three methods give fairly similar vertical flux estimates, observing an upward flux from the groundwater towards the Zenne River. Method 1 gives an average of –61 mm/d of water discharge; method

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Table 3 Absolute (m^3/d) and relative (%) inflows and outflows for the Zenne River (length 3,830 m). The values in the table are related to the values in Fig. 5 (*B-S Canal* Brussels-Scheldt Canal, *GW* groundwater)

	Absolute inflow (m^3/d)	Relative inflow (m^3/d)	Absolute outflow (m^3/d)	Relative outflow (m^3/d)
West B-S Canal discharge (‘1’ in Fig. 5)	7,054	72	0	0
South-eastern regional GW flow (‘2’ in Fig. 5)	2,690	28	0	0
River drainage (‘1+2’ in Fig. 5)	0	0	9,744	100

2, the thermal method, gives a bit less, around –45 mm/d; and method 3, the groundwater flow model, gives around –40 mm/d. Since the results of the three independent methodologies show similar estimates, it can be expected that they give a realistic and reliable value of the vertical groundwater–surface-water interaction at the Zenne River.

Notice that a complete similitude is unlikely; the two methods, the hydraulic gradient method and the thermal method, are methodologically distinct and are based on different assumptions. The disagreement in exchange fluxes between the hydraulic gradient method and the thermal method can be attributed to the uncertainty regarding the hydraulic conductivity of the riverbed and the alluvial sediments.

Regarding the thermal method, the magnitude of the estimated exchange fluxes is comparable with other studies of Flemish rivers as described by Anibas et al. (2009, 2011), but are fairly low in comparison with some other works like Keery et al. (2007). The fact that STRIVE integrates the exchange fluxes over the vertical model domain of 5 m depth may partly explain the big differences in flux estimates.

It is possible to use the flux estimates of the thermal method together with the measured hydraulic gradient for the estimation of the vertical hydraulic conductivity. By doing so, the obtained flux estimates of the hydraulic gradient method would be similar to those of the thermal method; both methods are not applied independently anymore. For the Zenne River, since the hydraulic gradient method yields higher results then the others the hydraulic conductivity is reduced to 0.21 m/d. It can be stated that the used vertical hydraulic conductivity was overestimated. However, keeping in mind the large uncertainties regarding the estimation of (vertical) hydraulic conductivities, the initial value of 0.28 m/d seems to be well chosen. In fact, Anibas et al. (2011) derived their estimate of the vertical riverbed hydraulic conductivity with a similar approach.

The thermal method and the hydraulic gradient method also resolve the temporal behavior of the groundwater–surface-water interaction. In Fig. 4, it can be seen that the results of both methods correspond and show comparable values and trends. It is clear that hydraulic gradients deliver flux estimates with the highest temporal resolution.

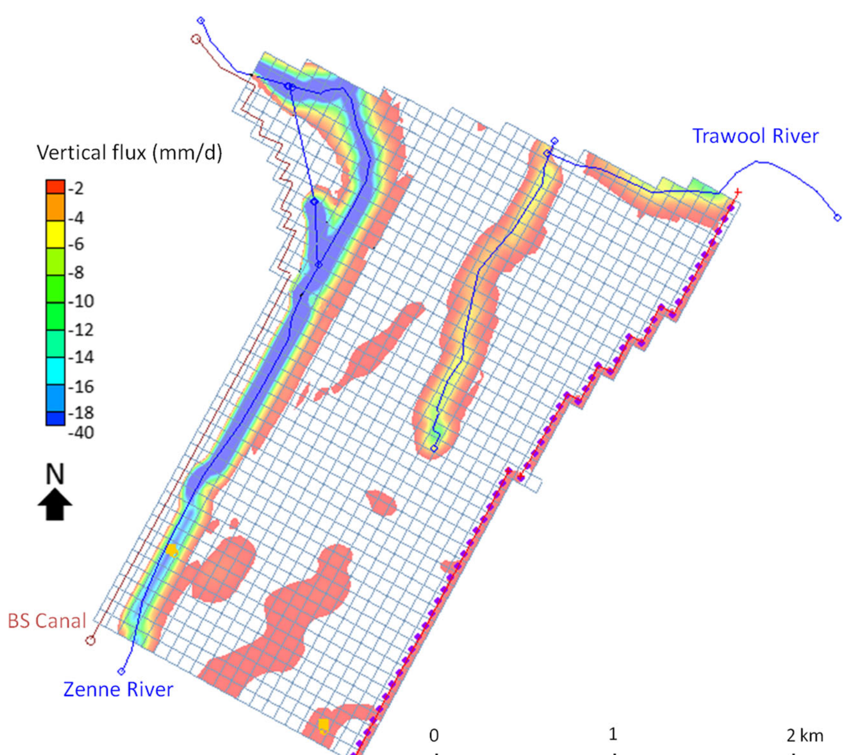

Fig. 6 Vertical groundwater fluxes in the study area as simulated with the numerical groundwater flow model. The *colored cells* show upward fluxes, while the *white cells* have downward fluxes

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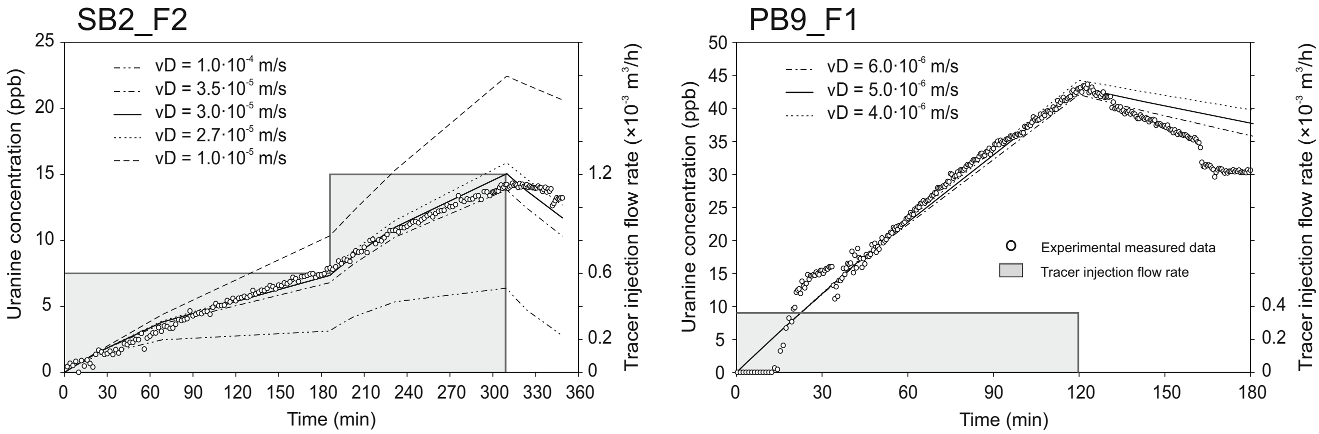

Fig. 7 Monitored and modelled concentration evolution and injection flow rates. The *solid black curve* provides the Darcy's flux with the fit. The first FVPDM test was at piezometer *SB2_F2*, the second text at *PB9_F1*

Figure 4 shows a fairly constant flux for the whole simulation period, and strong seasonal variations are not indicated. The short-term fluctuations in groundwater–surface-water interaction can be explained by the large variations in stream discharge of the Zenne River. As an urban river, it is quite sensitive to precipitation; hence, the stream discharge shows a quick response in function of rainfall events, which has an impact on the vertical groundwater–surface-water interaction of the Zenne River. Because of changing hydraulic gradients, the vertical discharge will decrease when river water levels rise; when the stream discharge decreases again, the vertical discharge into the Zenne River will increase again; however, the river remained gaining for the whole investigated period.

The left side of Table 5 compares horizontal Darcy's fluxes simulated by the groundwater flow model for two locations, while the right side presents horizontal Darcy's fluxes estimated by the FVPDM for the same locations. The groundwater flow model estimated a Darcy's flux of 150 mm/d in SB2_F2, and of 100 mm/d in PB9_F2. The FVPDM estimated a Darcy's flux of 2,600 mm/d in SB2_F2 and of 500 mm/d in PB9_F2, which is much higher, 5–17 times, than obtained from the groundwater flow model. This can be explained by the fact that FVPDM, as a single-well dilution technique, only provides a groundwater flow representative of the moment when the test is performed and at close vicinity of the tested well. FVPDM fluxes are instantaneous estimations, which depend strongly on experimental conditions, like e.g. river stage variations. These variations generate local pressure changes in the aquifer and thus changes in groundwater fluxes around the injection wells (Brouyère et al. 2008). This is especially the case for SB2_F2 where the injection well is situated at 3.5 m of the Zenne River, making it very sensitive to river stage variations and the differences in pressure due to these variations. The groundwater flow model on the other hand is a steady-state model, which smooths out the temporal variation of groundwater flow.

The hydraulic conductivity of the aquifer in the vicinity of the tested well can be deduced from the groundwater flux measured by the FVPDM and the local hydraulic gradient. A gradient of 0.0024 around well PB9 and 0.0033 around SB2 has been measured at the time of the experiment based on head measurements in neighboring piezometers, which gives a hydraulic conductivity of 208 and 788 m/d respectively for PB9 and SB2. The hydraulic properties of the groundwater flow model are average aquifer values, representing a much larger area than the ones investigated by a single well dilution test, leading to the much lower mean hydraulic conductivity of layer (Q2) of 12.5 m/d. Hence, the differences between the groundwater fluxes measured by the FVPDM in the field and the ones used in the groundwater flow model are caused by spatial heterogeneities of the porous media of the aquifer and by the temporal dynamics of groundwater flow close to the Zenne River.

The results show that horizontal fluxes might be larger than the vertical ones, and they might be much more variable as well. Not considering the extreme value of 2,600 mm/d, it is expected that the contribution of a

Table 4 Comparison of estimated vertical fluxes obtained from various methods [unit: mm/d]. Negative values indicate upward fluxes

	Method 1: hydraulic gradients Time-dependent	Method 2: temperature gradients Time-dependent	Method 3: GW flow model Steady-state
Period	November 2005–April 2007	November 2005–April 2007	Calibration period 2005
Results	Vertical flux at the river	Vertical flux at the river	Vertical flux at the river
Location	SB2	8 m upstream from SB2	Around SB2
Flux (mm/d)	–61	–45	–40

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Table 5 Comparison of estimated horizontal Darcy's fluxes obtained from the groundwater flow model and the FVPDM method [unit: mm/d]

	Method 3: GW flow model		Method 4: FVPDM	
	Steady-state		Steady-state	
Period	Calibration period 2005		Sept 2008	Nov 2008
Results	Darcy's flux (horizontal)		Darcy's flux (horizontal)	
Location (depth)	SB2_F2 (9.5 m)	PB9_F1 (5.3 m)	SB2_F2 (9.5 m)	PB9_F1 (5.3 m)
Flux (mm/d)	150	100	2,600	500

horizontal flux component to groundwater–surface-water exchange will still be smaller than the vertical component. As is described in literature (e.g. Rosenberry and LaBaugh 2008), the results from the groundwater flow model (Boel 2008; Dujardin et al. 2011) show that the flow lines bend strongly below the river cells, leading to predominantly vertical groundwater–surface-water exchange at the river–riverbed interface. Hence, with respect to an assessment of contaminant transport towards the Zenne River, the vertical groundwater–surface-water interaction is the most important pathway. Boel (2008) and Dujardin et al. (2011) show that the pollution sources are at such a distance from the river that their potential horizontal pollution pathways are predominantly located in model layers Q2 and TiEg (Fig. 3).

The differences between the horizontal and vertical estimates and the differences between the estimates of the different methods show that there is a possibility to over or underestimate the fluxes if one makes use of only a single method. While the groundwater model in principal simulates the total groundwater flux towards the river, these estimates are based on assumptions made for the whole model domain, including the more or less realistic definition of the river–riverbed interface. Especially for small-scale groundwater models, the simplification of this interface may lead to sub-optimal estimates of groundwater–surface-water interaction.

Field methods have the advantage to better describe the interaction, since they are based on actual field measurements at discrete locations. As point measurements, on the other hand, they lack the possibility to extrapolate these values on spatial scales; thus, only the combination of different methods will deliver a reliable perception of the variation of groundwater–surface-water interaction in space and time and enables understanding of the uncertainty and heterogeneity of the investigated physical processes.

Conclusion

Knowledge about water fluxes is important for the assessment of the contaminant transport from the surrounding brownfields towards the Zenne River. This study, therefore, compared four independent methodologies in order to characterize groundwater–surface-water interactions between a river, the adjacent hyporheic zone and aquifer.

As the final receptor, the flux estimates show that the Zenne River receives a little less than one third of the discharging groundwater from the eastern part of the study area. Using the different methodologies, vertical and horizontal exchange in the study area were investigated. The hydraulic gradient and the thermal method yield temporal variations and the steady-state groundwater flow model indicates the spatial variation in the whole study area, including horizontal fluxes. Since the hydraulic gradient and the thermal method are limited to vertical exchange rates, a fourth methodology, the FVPDM was chosen to give an independent estimation of horizontal Darcy's fluxes.

The vertical flux estimates resulted in comparable groundwater discharge fluxes of –61 to –40 mm/d to the receiving Zenne River. Being in line with studies of other Flemish lowland rivers, these values have therefore a strong reliability. The hydraulic gradient method and the thermal method also show comparable results and trends in the temporal distribution of groundwater–surface-water exchange. Regardless of different rainfall events, the river remained gaining throughout the entire investigated period of 16 months (Fig. 4). With a weak seasonal pattern, short-term variations dominate the vertical exchange, which is explained by the sensitivity of the groundwater discharge on the water level of the mostly urban Zenne River in connection with heavy rainfall events.

The obtained results for horizontal flow showed differences, explained by the strong sensitivity to field conditions of the transient FVPDM method. These differences in estimations indicate how strong the uncertainties are when relying solely on estimations of a single field method or on a groundwater flow model and also emphasize the importance of investigating temporal variations of groundwater fluxes and spatial heterogeneities of the hydraulic properties of an aquifer. A risk management plan, hence, should not rely on site characterizations from a single field campaign, since these measurements and results might not be representative for the fluctuations of the groundwater flow in particular and aquifer dynamics in general.

The combination of field techniques, therefore, improves the capacity of a risk management plan for brownfields and adjacent surface water and groundwater bodies. By application of different techniques, uncertainties of the estimates are reduced, while the confidence as well as the credibility of the applied methods and the risk management is improved.

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