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# Determination of groundwater fluxes in the Belgian Aa River by sensing and simulation of streambed temperatures

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**Abstract** Measurements of streambed temperature profiles are used to determine the vertical advective flow of water in the hyporheic zone in a Belgian lowland river. Quasi-steady-state simulations were performed with a one-dimensional heat transport equation introduced in the ecosystem modelling platform FEMME and then spatially interpolated. Measurement campaigns were performed in the winter season when a quasi steady-state thermal condition can be assumed. This allows an estimation of the mass balance of groundwater–surface water exchange in the study area. The results were compared with a steady-state MODFLOW 2000 simulation for the winter of 2004. Results show higher discharge in the upper reaches of the section and lower values in the downstream section of the river. Dependent probably on the water level of the river, the downstream section changes in time from gaining to losing.

**Key words** hyporheic zone; groundwater–surface water interaction; heat transport; FEMME

## INTRODUCTION

The use of temperature measurements and heat flow in the subsurface to unveil the movement of groundwater dates back to the 1960s. Until the late 1980s, when Lapham (1989) published a methodology for using measured temperature profiles in the river bed for quantifying groundwater–surface water (GW–SW) interaction, the interest in the method was relatively low. Since the 1990s the literature within this field has been increasing substantially. Arriaga *et al.* (2004), Brewster Conant Jr (2004), Hatch *et al.* (2006) and Schmidt *et al.* (2006) describe different approaches for analysing thermal data sets in transient and quasi-steady-state. The growth might be related to the fact that data loggers and temperature sensors collecting reliable field data became cheaper and more user-friendly.

The frontier in GW–SW interactions research is the near-channel and in-channel exchange of water, solutes, particles and energy. An understanding of these processes is the key to evaluating the ecological structure of stream systems and their management. The focus of this paper is the quantification of the spatial and temporal dynamics of water through the hyporheic zone. Flanders is one of Europe's hotspots regarding severe nitrate pollution of its shallow groundwater bodies, hence the determination of the exchange processes between different parts of the ecosystem is an important motivation for this research.

## METHODOLOGY

To determine groundwater–surface water interaction on a local scale, a streambed temperature survey was chosen. By measuring vertical temperature profiles, quantitative estimates of groundwater discharge or recharge, hence fluxes, can be produced by applying a 1-D heat transport routine. Because the movement of water through the soil matrix distorts the natural temperature distribution in the subsurface, heat can be used as a tracer to follow the movement of water. Hence, temperature is used as surrogate for head and/or to supplement head measurements

(Anderson, 2005). The temperature gradient in the soil is assumed to be a result of the coupled heat transport processes of water advection or convection and heat conduction.

The simple concept of relating streambed temperatures to spatial differences of vertical water flux is subject to a number of assumptions concerning unknown boundary conditions and thermal characteristics of the streambed and aquifer material. More importantly, even when the discharge value is accurate it still represents a point estimate (Becker *et al*., 2004; Schmidt *et al*., 2006).

The temperature profiles beneath the river shift seasonally with changing surface temperatures forming an annual envelope. While groundwater temperature remains nearly constant near the mean annual air temperature at a sufficient depth, streamwater temperatures vary seasonally and diurnally. During the summer months, groundwater temperature is generally cooler than stream temperature in discharge areas, whereas in winter it is generally warmer.

This paper presents the application of a quasi-steady-state 1-D heat transport model where the vertical component of the advective flow is calibrated to temperature profiles recorded in winter. It is important to mention that transient influences reduce the practicability of the steady-state solution, especially under recharging conditions (water flow from the surface into the hyporheic zone). The annual, as well as the diurnal temperature changes on the land surface can be presented with a sinus function. In periods of the year or during the day when the sinus is reaching its turning point the transient influences are limited and a steady-state solution is acceptably accurate.

## Study site

The field site is a 1425 m long stretch of the Aa River (Fig. 1) in Flanders, Belgium. The Aa River is a tributary of the Kleine Nete River, which is a tributary of the Scheldt River, and has a total length of 36.7 km and a drainage area of 237 km$^2$. At the field site the Aa River has an average width of 14 m and an average discharge of about 1.8 m$^3$ s$^{-1}$. The Aa is a typical Flemish lowland river with an average inclination of about 0.48 ‰ and a Manning coefficient of 0.060.

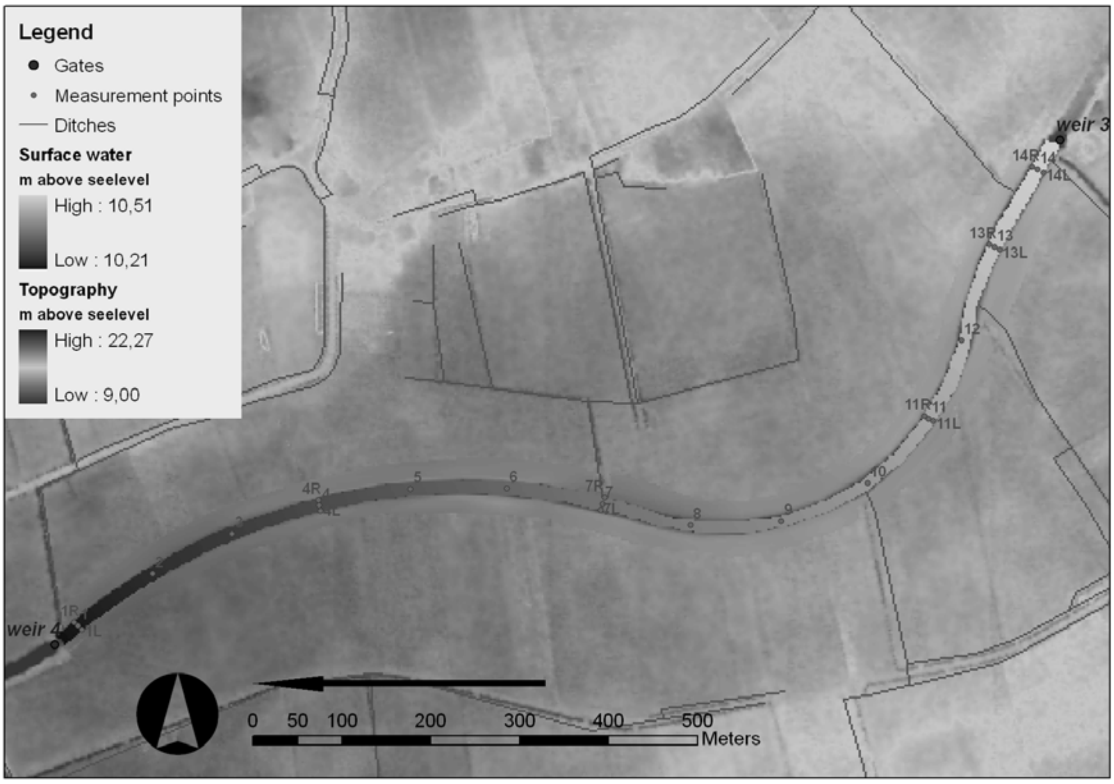


**Fig. 1** Digital elevation model showing the examined section of the River Aa close to the town of Herentals, Belgium and the location of the 26 measurement points.

The Aa River is a channelized river with an artificial sandy riverbed. The examined reach is located between two weirs; the upstream called number 3 and the downstream called number 4. The river section in between covers a surface area of about 20 400 m$^2$. The average water depth along the river stretch is about 1.0 m. Besides sand, the top layer of the riverbed has a variable fraction of organic matter. Different sandy formations dominate the geology of the river valley and constitute the top aquifer, which is underlain by the Boom aquitard.

In general the cross-section of the river resembles an almost rectangular shape with a flat river bottom and steep, almost vertical banks. Along a significant part of the section between weirs 3 and 4 the river bottom sands are less compacted along the centreline than along the banks. This characteristic might have an influence on local exchange processes (local flow in riffles). The changes in geometry may also have an influence on the groundwater flow and relevant exchange processes (Thibodeaux *et al*., 1987). The surveyed section of the river shows an extreme growth of macrophytes in the growing season, strongly reducing the discharge capacity of the channel.

## Field measurements

We performed temperature measurements with a self-made "temperature stick" ("T-stick"). The overall length of the stick is 2 m. It consists of a metal rod, a T-shaped handle at one end to push the instrument into the soil, and a pointed tip on the other end of the instrument (Fig. 2). Close to the tip there is a slotted hole holding the temperature sensor, a thermistor (Davis Instruments Model 7817, Hayward, California, USA). The sensor is connected to a common multimeter.

In this research, 26 spatially distributed measurement points (Fig. 1) were used to measure temperature profiles in the hyporheic zone along the section between weirs 3 and 4. Because of the time needed for the measurements the campaigns were split up into two consecutive days, presuming that the conditions along the river will remain almost constant. In the first day of a campaign 14 points along the centreline of the river were measured, beginning around 30 m upstream from the lower weir 4, and propagating upstream in steps of 100 m between the measurement points. At most of the points five temperature measurements in depth have been

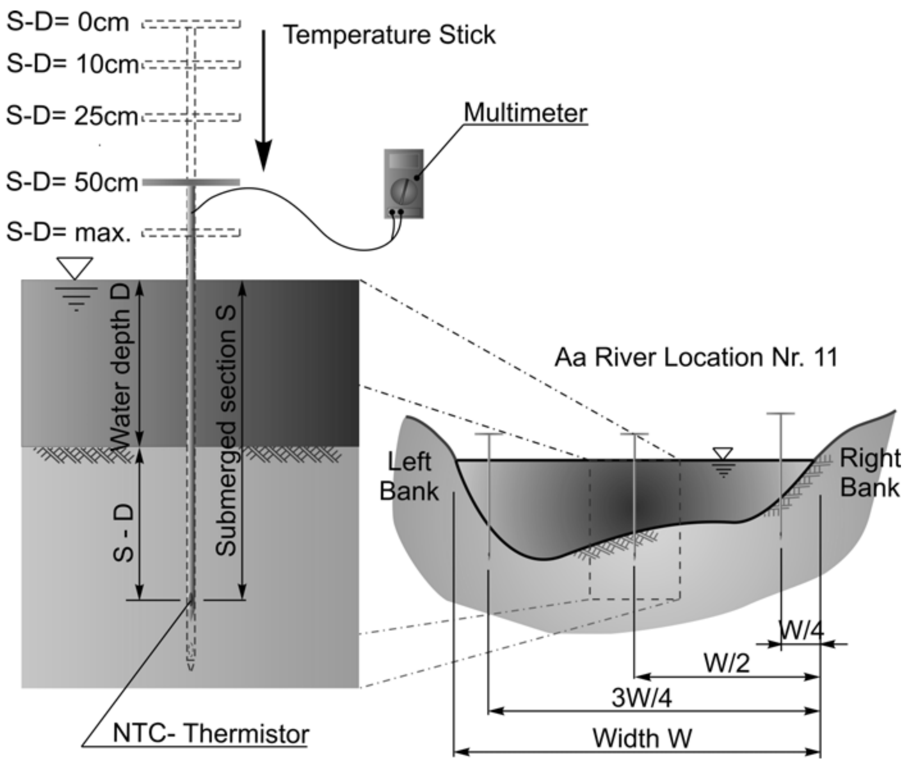


**Fig. 2** Measurement scheme for temperature profiles in the Aa River with the temperature stick.

conducted, at 0 (representing the interface between water and river bottom), 10, 25, 50 cm, and finally at the maximum depth of penetration, which is generally about 50–120 cm, mostly the depth where the probe faced an impenetrable compacted sandy soil layer (Fig. 2).

On the second day, several transects of the river were measured, including several points in the middle of the river, which were also measured the day before. In total, 5 or 6 cross-sections were measured with the T-stick located at the points 1, 4, 7, 11, 13 and 14. The transect measurements are called 1L(eft), 1C(entre), and 1R(ight), respectively. The procedure shown in Fig. 2 is then repeated at all 5 or 6 depicted points.

## Model

The temperature information collected in the field does not directly lead to a result of groundwater exchange flux. Therefore a heat transport model based on the analytical solutions of Suzuki (1960), Stallmann (1965) and Lapham (1989) was introduced in the model platform FEMME (Flexible Environment for Mathematically Modeling the Environment, Soetaert *et al.*, 2002). The ecosystem model consists of several different modules, which can be linked together if sufficient data sets are available.

One part of FEMME, the so-called *hyporheic zone module*, is able to simulate heat transport. It will be used to determine the vertical advective component of the Darcian flow velocity through the sediment matrix. The water flux between the different zones defined in the ecosystem model then determines the transport, retention and reaction processes of nutrients, other solutes and particulate matter in the soil. The governing equation of the model (equation (1)) describes the 1-D, vertical, an isothermal flow of an incompressible fluid through homogeneous, porous media.

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

where $k$ is the thermal conductivity of the soil–water matrix in J s$^{-1}$ m$^{-1}$ K$^{-1}$, T the temperature at depth $z$ at time $t$ in the soil, $c_w$ the volumetric heat capacity of the fluid in J m$^{-3}$ K$^{-1}$, $\rho_w$ the density of the fluid in kg m$^{-3}$, $v_z$ the vertical component of the Darcian fluid velocity in the soil expressed in mm day$^{-1}$ (or as flux equal to L m$^{-2}$ day$^{-1}$), $c$ the volumetric heat capacity of the rock–fluid matrix in J m$^{-3}$ K$^{-1}$, and $\rho$ the wet-bulk density (density of the rock-fluid matrix) in kg m$^{-3}$.

The quasi steady-state solution presented in this paper needs a single temperature profile as input. By setting the right hand side of equation (1) d$T$/d$t$ to zero, the module will deliver a single scalar value for every measured profile representing the water flux at a certain point and time. The algebraic sign indicates the direction of the groundwater flow, where + indicates that water is moving from the surface into the hyporheic zone (e.g. groundwater recharge), and – represents water moving upwards from the hyporheic zone into the river (e.g. groundwater discharge). The hyporheic zone module can also be used to solve transient simulations with different set ups of boundary conditions and other exercises, which are not discussed in this paper.

# RESULTS AND DISCUSSION

The data of the measurement campaigns of January 2006 and February 2007 was used to create spatially distributed maps of the groundwater–surface water interaction of the river section. The conditions for the cases of 2006 and 2007 especially differ in the average river water depth, which was 103 and 92 cm, respectively. The winter of 2005/2006 was cold compared to 2006/2007 (Table 1).

The last row in Table 1 shows the difference in $T$ between the uppermost measurement and the deepest measurement. The average depth, which could be reached by the T-stick instrument is 84 cm.

A required parameter is the thermal conductivity $k$ of the fluid soil matrix. The range of the thermal conductivity is in general small; $k$ can be estimated for a sandy riverbed as 1.8 J s$^{-1}$m$^{-1}$K$^{-1}$

**Table 1** Average temperatures measured at the 14 points composing the longitudinal transect along the Aa River.

| Depth in cm | 13 January 2006 T in °C | 8 February 2007 T in °C |
|---|---|---|
| 0 | 6.1 | 5.6 |
| 10 | 6.7 | 6.7 |
| 25 | 7.2 | 7.9 |
| 50 | 8.1 | 9.2 |
| 84 | 8.9 | 10.5 |
| Δ min–max | 2.8 | 5.0 |

**Table 2** Results of average, maximum recharge and maximum discharge values of vertical flux $v_z$ obtained with FEMME.

| $v_z$ in mm day$^{-1}$ | 13 and 14 January 2006[(1)] | 8 and 9 February 2007[(2)] |
|---|---|---|
| Max. | 26 | 13 |
| Min. | –334 | –407 |
| Average | –86 | –116 |

[(1)]13 longitudinal measurements; 5 cross-sections with 15 measurements.
[(2)]13 longitudinal measurements; 6 cross-sections with 17 measurements.

(Stonestrom & Blasch, 2004; Schmidt *et al*., 2006). The value is a little lower than published in other sources, trying to take the changing content of organic material in the layer adjacent to the surface water–groundwater interface into account. The boundary conditions of the homogeneous model domain are determined by the obtained depth and the values of the respective T-profiles. $c_w$ was estimated with 4.18 $10^6$ J m$^{-3}$ K$^{-1}$ for the model, $\rho_w$ as 1000 kg m$^{-3}$.

The measurements from the two campaigns include values from all 26 measurement points along and across the river (Table 2). However, a few measured profiles are significantly distorted by transient processes and/or measurement errors and were discarded during the modelling process.

Lapham (1989) stated that the method is of limited value if upward fluxes are exceeding 305 mm day$^{-1}$, but Schmidt *et al*. (2006) measured fluxes up to 455 mm day$^{-1}$ and concluded that with decreasing measurement depth even higher fluxes can be accurately quantified. The lower detection limit of the method is ±10 mm day$^{-1}$. However, FEMME converges well at the whole range of magnitude of fluxes we encountered.

The estimation of correct recharge values is in general problematic. A profile indicating a downward flux might just show transient influences coming from recent meteorological events, instead of actual downward movement of water. A quasi-steady-state assumption with constant recharge would furthermore show a homogeneous *T*-distribution in the soil, which was not measured in our discharge dominated system. Recharge is occurring temporarily under certain hydrological conditions, like high discharge and water levels, and cannot be captured properly with our approach. Thus the error in the calculated recharge values might be high. Nevertheless the results show at least the spatial extend of recharging zones along the river stretch.

With the combined information of both the longitudinal survey and the cross-sectional data, we calculated spatially distributed maps where the point values of the simulation were extrapolated to the river's surface area (Figs 3 and 4). The interpolation was performed with the Surfer software using anisotropic kriging.

In general the fluxes calculated from the measurements in the middle of the river are significantly lower than those which are observed close to the banks of the river. The lateral inflow of shallow groundwater could be dominant in this system, which is determined by shallow groundwater tables. Lateral movement of water from the groundwater into the river cannot be covered with this method and thus the results from the T-stick likely underestimate the real GW–SW interaction.

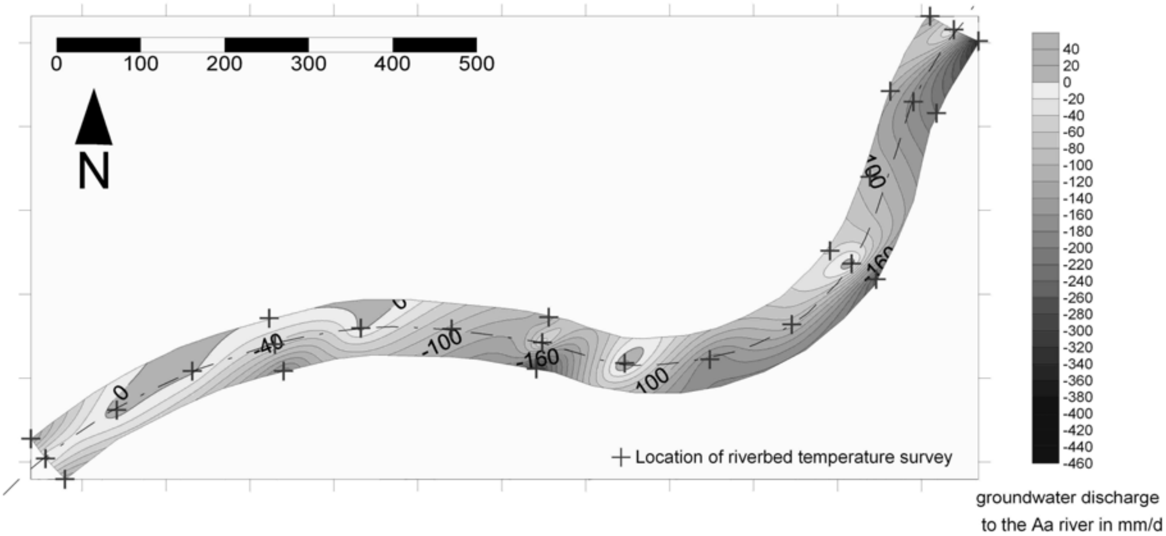


**Fig. 3** Spatial interpolation of the measurements of 13 and 14 January 2006, the width of the river is exaggerated for better visibility.

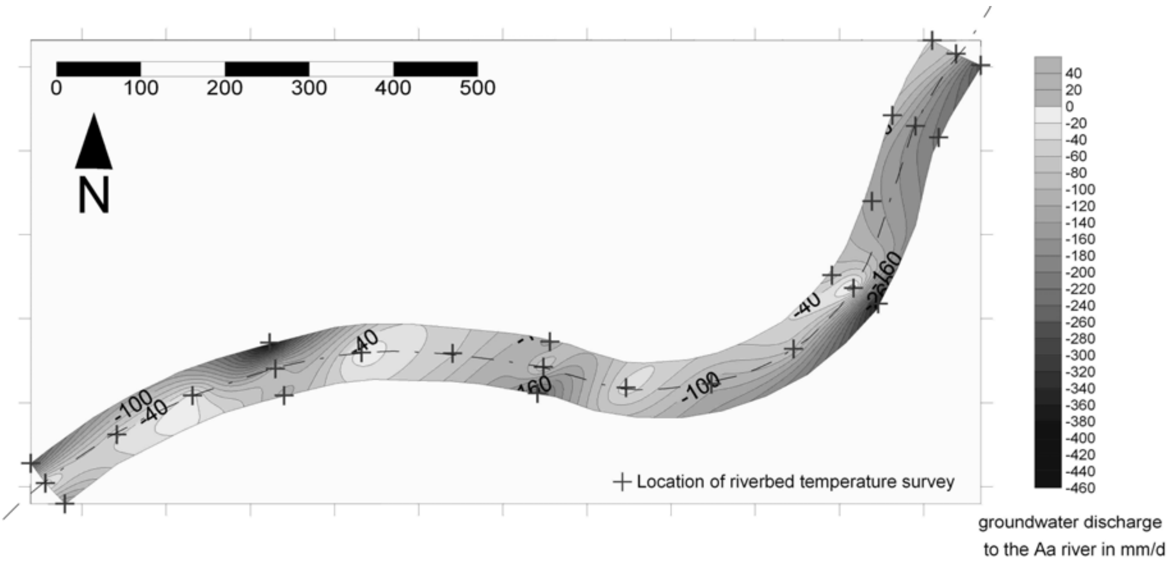


**Fig. 4** Spatial interpolation of the measurements of 8 and 9 February 2007, the width of the river is exaggerated for better visibility.

The sections indicating recharge could also be small scale local phenomena. The recharge zones are predominantly located in the shallow, soft sediments in the middle of the river. Downward flow may thus occur in local pool or riffle structures (Thibodeaux & Boyle, 1987) or other obstructions (Hutchinson & Webster, 1998).

In both 2006 and 2007 the upstream part shows higher discharge values than the lower reach of the river stretch. Beside different hydrological reasons, this phenomenon could simply be linked to the presence of the two weirs, which create a relatively high hydraulic gradient over a short distance. As a result the groundwater tries to bypass the weirs and a higher discharge could be expected downstream of the weir. At the upstream section of the river the bulk of the discharge is on the left bank of the river, in contrast the lower reach seems to have the higher influx on the right bank, which can also be explained with lateral movement of water under changing exposure of the river banks to the regional groundwater flow. The water depth of the river could have an influence on the exchange pattern; high water level means lower GW–SW discharge, and low water level higher GW–SW discharge. This phenomenon might also be linked to the discharge history of the river. Despite the fact that 26 point measurements cannot cover all the heterogeneities in the river bed, the spatial interpolation might be used to estimate a simple mass balance of the groundwater–surface water exchange.

Table 3 shows the results gained from the spatial interpolation. It can be seen that the surface water–groundwater exchange characteristics of the river can change from an entirely gaining reach to a partly gaining, partly losing one. Thus, when the water level as in January 2006 is higher, the discharge is not only generally lower, but the direction of flow is also changing.

For comparison, Table 3 also shows the results of a local groundwater model based on data sets from 2004. Although from different years, both MODFLOW and FEMME simulations show the same characteristic features. The flux rates calculated with MODFLOW are higher, and the area occupied by a recharging environment is almost as big as the discharging one. However, the discharge occurs in the upstream part, whereas the recharge zone is located in the downstream part which corresponds with the results from the temperature survey.

**Table 3** Groundwater- Surface water interaction along the Aa River.

| parameter | Unit | 13 and 14 January 2006[(1)] | 8 and 9 February 2007[(1)] | 10 and 20 January 2004[(2)] |
|---|---|---|---|---|
| Mean net flux | mm day$^{-1}$ | –89 | –109 | –164 |
| Mean net flux | L s$^{-1}$ | –21.1 | –25.8 | –28.5 |
| Discharge flux | mm day$^{-1}$ | –99 | –109 | –175 |
| Recharge flux | mm day$^{-1}$ | 7 | 0 | 11 |
| Discharge area | m$^2$ | 18533 | 20403 | 7950 |
| Recharge area | m$^2$ | 1870 | 0 | 7050 |
| Water level | m | 1.03 | 0.92 | – |

[(1)] Spatial interpolation of results derived by FEMME and interpolated with anisotropic kriging.
[(2)] Results from MODFLOW 2000 steady state simulation; setup: 300 rows x 360 columns, cell size 5 m, River surface: 15 000 m$^2$, 3 layers, 6 hydrogeologic units, usage of DRAIN, RECHARGE, WELL, RIVER packages.

## CONCLUSIONS

Measurements of river bed temperature profiles on a 1425 m long stretch of the canalized Aa River have been conducted. Vertical advective fluxes were estimated by applying a 1-D analytical solution of the heat transport equation introduced in a FEMME model platform. The temperature method provides the possibility to map the spatial heterogeneities of the groundwater–surface water interaction on a local scale at resolutions where other methods are well below their detection limit.

The survey showed that considerable differences in discharge values can be expected along and across a river stretch, even in an environment which seems to be relatively uniform. The upstream part of the examined river section shows in general higher discharge values than the downstream one.

Significant recharge in the river could not be proven, although the results show that in the downstream part of the section the direction of flow might change from gaining to losing. Local flow phenomenon cannot be excluded as a reason for this result. The interpolation of point results to a spatially distributed map gives an idea of the mass balance of exchange between the groundwater and the river. In general, the agreement between the MODFLOW and the FEMME simulation is good.

The mapping of temperature profiles can be adjusted according to the demands of the research, and on a local scale it can detect heterogeneities more economically than other methods. The overall reliability of temperature mapping can be improved by combining it with other measurements, which provide additional information or an independent validation of the simulated results.

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