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A simple thermal mapping method for seasonal spatial patterns of groundwater–surface water interaction

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SUMMARY

A simple thermal mapping method for simulating seasonal and spatial patterns of groundwater–surface water interaction is developed and tested for a segment of the Aa River, Belgium. Spatially distributed temperature profiles in the hyporheic zone of the river are measured in winter and summer seasons of three consecutive years. Inverse modeling of the one-dimensional heat transport equation is applied to estimate vertical advective fluxes using the numerical STRIVE model and an analytical model. Results of the study show that seasonal flux estimates for summer and winter can be derived with a minimum data input and simulation effort. The estimated fluxes are analyzed via non-parametric statistical tests, while spatial interpolation techniques are used to generate maps of distributed flux exchange. The estimated seepage is compared with volumetric flux obtained from piezometer measurements and output of a groundwater model. The thermal method shows higher discharge rates in winter and that the relative contribution of exfiltration to the river discharge is higher in summer. A higher flux and a more heterogeneous flow pattern are observed in the upper reach of the river compared to the lower reach. This spatial difference shows the importance of the local geomorphology and to a lesser extent the hydrogeologic setting on hyporheic flux exchange in the river. A significantly higher flux is noted on the banks than in the center of the river, which is driven by the relatively high hydraulic conductivity of the river banks. It is concluded that bank flow in groundwater–surface water interaction deserves more attention. The main channel of the Aa River alone accounts for about 15% of the total river discharge at its outlet. As the developed thermal method is cost-effective, simple and fast, it is recommended for use in identifying zones of interest in initial stages of field investigations of groundwater–surface water interaction.

1. Introduction

An important process in ecohydrology is groundwater–surface water interaction at the interface of lake or riverbeds (Stanford, 1998; Woessner, 2000). The accurate quantification of exchange of water and energy between groundwater and surface water systems remains a challenging task due to the problems of heterogeneity and scale (Becker et al., 2004; Kalbus et al., 2008).

The saturated zone beneath and beside streams, rivers, lakes and wetlands where groundwater and surface water actively mix and exchange is called hyporheic zone (Brunke and Gonser, 1997; Boulton et al., 1998; Hayashi and Rosenberry, 2002; Smith, 2005). The processes in the hyporheic zone are variable in space and time on relatively small scales (McClain et al., 2003). National and international regulations like the European Water Framework Directive (European Commission, 2000, 2006) mandate the protection of the links between groundwater and surface water systems. To do this, reliable and transferable conceptual models (NRC, 2004; Smith, 2005; Schmidt et al., 2008) are required to assess mass fluxes and related biochemical processes across groundwater–surface water interfaces.

Different methods have been discussed and used to quantify the exchange flux between groundwater and surface water systems (Kalbus et al., 2006). Seepage meters are the only means by which groundwater–surface water exchange is directly measured (Lee, 1977; Rosenberry, 2008). Although seepage meters are technically simple, there are uncertainties in the measured flux due to operational problems in the field (Shaw and Prepas, 1990; Murdoch and Kelly, 2003). Indirect methods estimate the groundwater–surface water interaction by deriving the fluxes across the interface of the hyporheic zone from other parameters. Groundwater head measurements from piezometers and boreholes in riverbeds or near streams (Cey et al., 1998; Baxter et al., 2003) can be used to

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derive vertical flow estimates. But the exact estimation of hyporheic-zone flux is hampered by difficulties in measuring in situ hydraulic conductivity (Chen, 2000). Piezometers installed in riverbeds are also prone to scouring and damage by floods or debris. Other indirect methods of estimating groundwater–surface water exchange include differential discharge gauging (Becker et al., 2004), numerical modeling (Cardenas and Zlotnik, 2003; Fleckenstein et al., 2004; Saenger et al., 2005) and the use of conservative tracers like dye, salt, chloride or stable isotopes (Carey and Quinton, 2005). Remote sensing, a relatively new technique in this field (Becker, 2006), offers a spatially continuous qualitative estimation of hyporheic-zone flux exchange (Loheide and Gorelick, 2006).

Heat can be used as a tracer to estimate groundwater–surface water interaction (Stonestrom and Constantz, 2003; Anderson, 2005; Constantz, 2008; Constantz et al., 2008). The exchange patterns can be inferred qualitatively from temperature variations at the groundwater–surface water interface (Lowry et al., 2007) or quantified by empirically relating mapped streambed temperatures to volumetric fluxes (Conant, 2004). Hatch et al. (2006) and Keery et al. (2007) presented methods to determine streambed seepage rates using phase shifts and amplitude damping of thermal time series data between pairs of sensors. Exchange rates have also been successfully quantified by inverse modeling of temperature profiles (Arriaga and Leap, 2006; Schmidt et al., 2006; Anibas et al., 2009).

Suzuki (1960) and Stallman (1965) presented analytical solutions for using temperature as a natural tracer. Bredehoeft and Papadopoulos (1965) introduced a graphical type curve solution method based on temperature–depth profiles for the estimation of vertical groundwater fluxes under steady-state thermal conditions. As temperature distributions in the subsurface are transient by nature, Lapham (1989) developed a numerical model that allows quantification of groundwater–surface water interaction using an annual temperature variation simplified by a sine function. Arriaga and Leap (2006) and Schmidt et al. (2006, 2007) presented applications of the estimation of groundwater–surface water interaction based on steady state heat transport, while Anibas et al. (2009) showed when the steady-state thermal assumption can be used.

This paper presents a simple and fast method for quantifying seasonally distributed groundwater–surface water exchange. The proposed approach is a robust, adjustable and first-hand field investigation tool for detailed monitoring of zones of interest. The underlying hypothesis of the methodology is that steady-state solutions for simultaneous vertical transport of heat and water in riverbeds are able to detect groundwater–surface water interaction with sufficient spatial and temporal resolution. Measured riverbed temperature profiles from the Aa River in Belgium are used to test the proposed method. Furthermore, the analytical solution method of Arriaga and Leap (2006) is used in conjunction with the numerical heat transport model STRIVE to statistically verify the proposed method. The STRIVE model is based on the approach of Lapham (1989) and the FEMME ecosystem modeling platform (Soetaert et al., 2002).

2. Methodology

The temperature in the hyporheic zone varies seasonally and diurnally as a consequence of heating and cooling of the land surface. During summer months the groundwater temperature is generally cooler than the river temperature, whereas in winter it is generally the opposite. The groundwater flow in the riverbed is assumed to obey Darcy's law consequently the natural temperature distribution in the riverbed is influenced by the movement of water. The geothermal gradient in the shallow subsurface is neglected for our analysis hence the one dimensional, vertical, anisothermal transport of heat through homogeneous, porous media is described as (Suzuki, 1960; Stallman, 1965; Lapham, 1989).

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

where k is the thermal conductivity of the soil–water matrix in $J\,s^{-1}\,m^{-1}\,K^{-1}$, T the temperature at point z at time t in the sediment in K (°C), c_w the specific heat capacity of the fluid in $J\,kg^{-1}\,K^{-1}$, ρ_w the density of the fluid in $kg\,m^{-3}$, v_z the vertical component of the groundwater velocity in the sediment in $m\,s^{-1}$, c the specific heat capacity of the rock-fluid matrix in $J\,kg^{-1}\,K^{-1}$, and ρ the wet-bulk density (density of the rock-fluid matrix) in $kg\,m^{-3}$. The terms $c_w\rho_w$ and $c\rho$ represent the volumetric heat capacity of the fluid and the rock-fluid matrix in $J\,m^{-3}\,K^{-1}$ respectively. The first term of the left hand side of Eq. (1) represents the conductive and the second term the advective part of the heat transport. A positive sign of the vertical groundwater velocity stands for water moving from the surface into the hyporheic zone (e.g. groundwater recharge or losing stream reach) and a negative sign represents water moving from the hyporheic zone into the river (e.g. groundwater discharge or gaining stream reach). The groundwater velocity is expressed in $mm\,d^{-1}$, equivalent with a unit flux in $L\,m^{-2}\,d^{-1}$.

In the case of a thermal steady state, the temperature distribution in the riverbed is supposed to be constant over time, which reduces the right hand side of Eq. (1) to 0:

$$k\frac{\partial^2 T}{\partial z^2} - v_z c_w \rho_w \frac{\partial T}{\partial z} = 0 \tag{2}$$

Notice that in Eq. (2), the thermal properties of the fluid–sediment matrix are only described by the thermal conductivity k. For sediments as different as peat or gravel the thermal conductivity k varies less than an order of magnitude (Freeze and Cherry, 1979; Domenico and Schwartz, 1998); e.g. saturated sands usually have k values between 1.4 and 2.2 $J\,s^{-1}\,m^{-1}\,K^{-1}$ (Lapham, 1989; Schön, 1998; Stonestrom and Blasch, 2003). This small range of k is a major advantage of the thermal method compared to the hydraulic head method in which the hydraulic conductivity can vary over orders of magnitude (Chen, 2000). Hence, in many cases, the thermal conductivity can be parameterized by taking its value from literature.

The thermal method is an indirect method, field data must be processed with a model in order to derive quantitative estimates of the flow velocity. We apply inverse modeling of the thermal steady state by solving Eq. (1) numerically and Eq. (2) analytically. Fig. 1 shows the concept of the steady state 1D vertical fluid and heat transport for a saturated sediment column of the riverbed, in which measured temperatures define the upper and lower boundary conditions. The concept of the model does not allow to determine lateral or longitudinal flow vectors (Fairley and Nicholson, 2005; Hoehn and Cirpka, 2006). One or more measured temperatures within the domain are needed in order to fit the simulated temperature distribution to the measured profile (Fig. 1a). When groundwater flow occurs, the thermal profile shows a stronger curvature with increasing flow velocity (Fig. 1b). For practical reasons the presented methodology is limited to discharging flow conditions if applied to shallow groundwater flow systems (Schmidt et al., 2007). Mathematically Eqs. (1) and (2) allow estimation of infiltration, a reversal of the flux under steady state conditions however is producing a uniform temperature distribution with no significant thermal gradient. Hence it is not possible to obtain a realistic fit of the model with the applied boundary conditions (Fig. 1c).

To solve the thermal steady state problem a hydraulic steady state has to be assumed, diurnal, seasonal and irregular