



A multi-method approach to quantify groundwater/surface water-interactions in the semi-arid Hailiutu River basin, northwest China

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Abstract Identification and quantification of groundwater and surface-water interactions provide important scientific insights for managing groundwater and surface-water conjunctively. This is especially relevant in semi-arid areas where groundwater is often the main source to feed river discharge and to maintain groundwater dependent ecosystems. Multiple field measurements were taken in the semi-arid Bulang sub-catchment, part of the Hailiutu River basin in northwest China, to identify and quantify groundwater and surface-water interactions. Measurements of groundwater levels and stream stages for a 1-year investigation period indicate continuous groundwater discharge to the river. Temperature measurements of stream water, streambed deposits at different depths, and groundwater confirm the upward flow of groundwater to the stream during all seasons. Results of a tracer-based hydrograph separation exercise reveal that, even during heavy rainfall events, groundwater contributes much more to the increased stream discharge than direct surface runoff. Spatially distributed groundwater seepage along the stream was estimated using mass balance equations with electrical conductivity measurements during a

constant salt injection experiment. Calculated groundwater seepage rates showed surprisingly large spatial variations for a relatively homogeneous sandy aquifer.

Keywords Groundwater/surface-water relations · Hydraulic head · Temperature method · Tracer tests · China

Introduction

Groundwater and surface water have been managed as isolated components for a long time, but they are hydrologically connected in terms of both quantity and quality (Winter 1999). Thus, a better understanding of the interactions between groundwater and surface water could provide crucial scientific insights for integrated management of water resources. Interactions between groundwater and surface water such as in headwaters, streams, lakes, wetlands, and estuaries have been studied since the 1960s (Winter 1995; Woessner 2000; Sophocleous 2002). Many methods of quantifying the interactions between groundwater and surface water have been applied by researchers all over the world. The measuring methods for groundwater and surface-water interactions were summarized by Kalbus et al. (2006), Brodie et al. (2007), Rosenberry and LaBaugh (2008). A number of studies have been conducted for identifying the interactions between groundwater and surface water using: multiple field investigations (Oxtobee and Novakowski 2002; Langhoff et al. 2006; Rautio and Korkka-Niemi 2011); differences in hydraulic heads between river and groundwater (Anderson et al. 2005) with supplemental chemical and stable isotope data (Marimuthu et al. 2005; Okkonen and Kløve 2012); temperature studies of streambeds (Conant 2004; Schmidt et al. 2007; Westhoff et al. 2007; Vogt et al. 2010; Lewandowski et al. 2011); and chemical and isotopic tracers (Wenninger et al. 2008; Ayenew et al. 2008; Didszun and Uhlenbrook 2008). Field surveys such as point measurements of hydraulic heads are normally used for interpreting the interaction between groundwater and surface water. Seepage meters can directly measure the exchange between the groundwater and surface water

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(Landon et al. 2001). The stable isotopes deuterium and oxygen-18 as well as hydrochemical tracers have been widely used for investigating runoff generation processes (Uhlenbrook and Hoeg 2003), determining the hydrological exchange (Ruehl et al. 2006), and inferring groundwater/surface-water exchanges (Rodgers et al. 2004). There are a number of publications dealing with the application of tracers such as hydrochemical components (Wels et al. 1991; Kirchner 2003; Uhlenbrook et al. 2002) or environmental isotopes (Sklash and Farvolden 1979; McDonnell et al. 1990) in hydrological investigations. Most of them have been employed for separating the stream flow into pre-event and event water using environmental isotopes or spatially distinct components of surface water and groundwater using hydrochemical compounds (e.g. Uhlenbrook and Hoeg 2003). As pointed out by Anderson (2005) and Constantz (2008), heat can be used as a tracer for investigating the groundwater and surface-water connectivity. The temperatures beneath the streambed have been interpreted for delineating and quantifying groundwater discharge zones, estimating the vertical velocity and fluxes in the streambed. Constantz (2008) distinguished the thermal pattern beneath the streambed for losing stream reaches and gaining reaches. The oscillating surface temperature signal results in both conductive and advective heat transfer, with higher infiltration rates associated with a losing stream reach resulting in greater advection, deeper penetration, and shorter lags in temperature extremes at a given depth. However, using streambed temperatures to quantify groundwater/surface-water interactions is limited to locations and time periods where groundwater and stream water have sufficient temperature differences (Schmidt et al. 2006). The interaction between groundwater and surface water can be quantified by directly measuring the differences of discharge along the reach with e.g. current meters (Becker et al. 2004). Discharge measurements using a continuous injection of a sodium chloride (NaCl) solution and integration of the electrical conductivity as a function of time is a traditional and well-documented method for turbulent streams (Hongve 1987). The NaCl dilution and mass recovery method were employed for estimating the channel water balance (Payn et al. 2009) as well as net change in discharge, gross hydrologic loss, and gross hydrologic gain in experimental reaches by assuming constant discharge and complete mass recovery.

In China, several studies have been conducted for quantifying interactions between groundwater and surface water in arid and semi-arid regions with tracer methods (Wang et al. 2001; Nie et al. 2005; Wang et al. 2009), water balance (Xiao et al. 2006), and model simulations (Wang et al. 2006; Hu et al. 2007, 2009). An integrated approach that consists of multiple measuring methods has not yet been applied to investigate groundwater and surface-water interactions.

Water level and temperature measurements indicate that groundwater discharges to the river or groundwater can be recharged by the river. Discharge measurements and hydrograph separation can estimate the total

groundwater discharge in the sub-catchment. However, determining spatial variations of groundwater discharges along rivers is still a challenge. The application of distributed temperature sensors along river reaches identified large spatial variations (Selker et al. 2006; Lowry et al. 2007); however, the method cannot provide estimations of groundwater discharges directly. Seepage meters were used to measure groundwater fluxes at multiple points (Paulsen et al. 2001; Rosenberry and Morin 2004; Rosenberry and Pitlick 2009; Hatch et al. 2010), but cannot estimate total groundwater discharge of the total stream reach. Temperature gradient measurements beneath the streambed combined with the analytical (Silliman et al. 1995; Becker et al. 2004) or numerical heat transport model (Constantz 1998; Constantz and Stonestorm 2003), can only estimate point groundwater discharges.

This study used multiple field measurements to identify and quantify groundwater and surface-water interactions in a small catchment located in the semi-arid Erdos Plateau in northwest China. Measurements of groundwater and surface-water levels could provide an indication of the directions of water exchange between groundwater and surface water. Measurements of temperature in groundwater, stream water, and streambed deposits at different depth give direct evidence of connectivity and flow directions between groundwater and stream water at specific locations. A hydrograph separation analysis of stream discharges using stable isotopes as a tracer estimates relative contributions of groundwater and direct surface runoff to the total stream discharges for a rainfall event. A new method to quantify spatial distribution of groundwater seepage in a gaining stream using electrical conductivity profile measurements along the stream course before and during a constant salt injection is developed and tested. The successive solutions of mass balance equations along the measured stream segments provide estimates of groundwater seepage for every 10-m reach. This method is cost-effective and can be widely applied to quantify the groundwater seepage to streams. The spatially distributed groundwater seepage provides also new sources of data to calibrate coupled groundwater/surface-water models.

Study site

The Bulang catchment is a sub-catchment of the Hailiutu River basin in the middle section of the Yellow River basin in northwest China (Fig. 1). The total drainage area of the Bulang sub-catchment is 91.7 km². The surface elevation of the Bulang sub-catchment ranges from 1,300 m at the northeastern boundary to 1,160 m above mean sea level at the catchment outlet in the southwest. The land surface is characterized by undulating sand dunes and a perennial river in the downstream area (Yang et al. 2012). The long-term annual average of daily mean temperature is 8.1 °C and the monthly mean daily air temperature is below zero in the wintertime from November until March. The mean annual precipitation

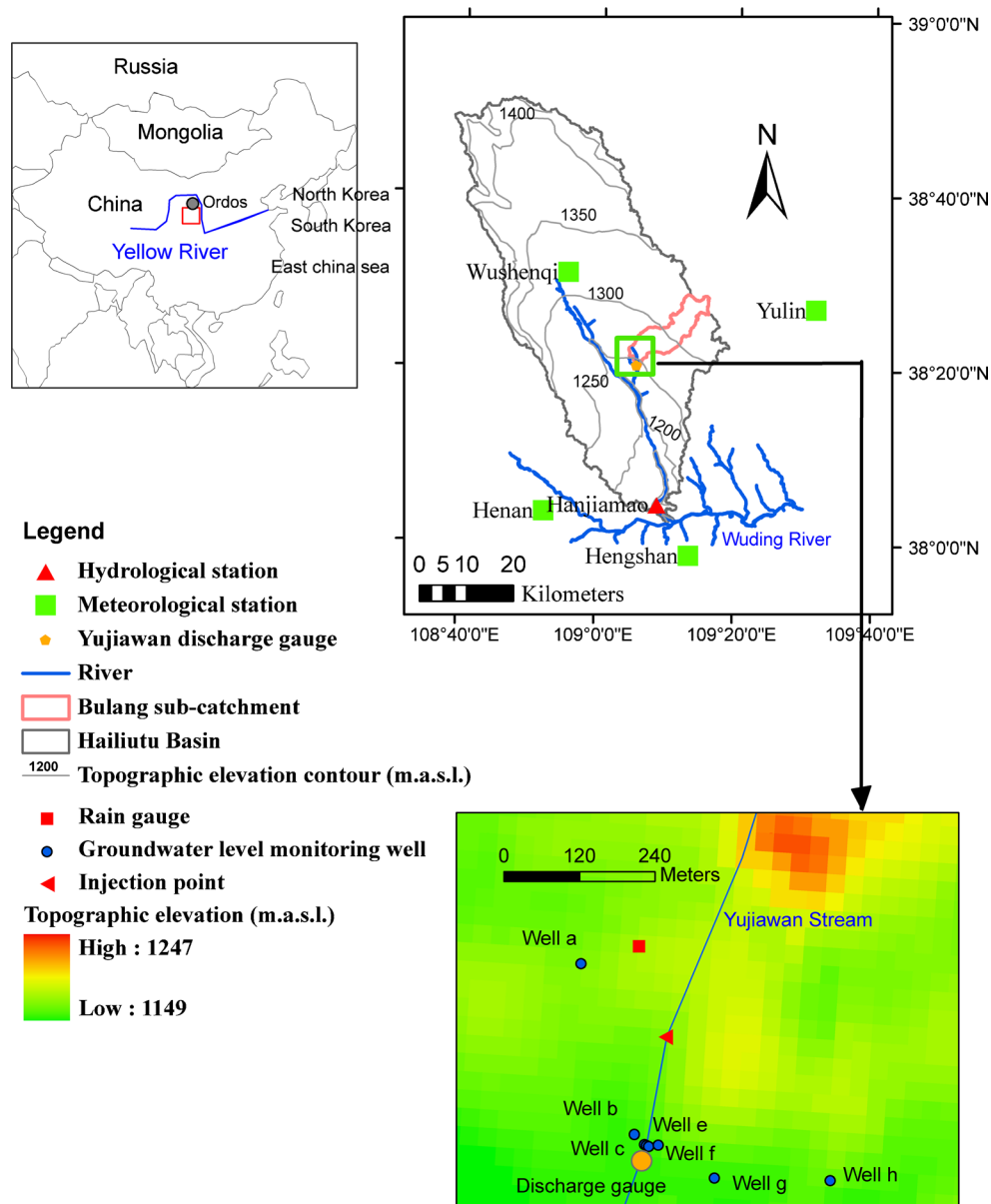


Fig. 1 Location of the Yujiawan discharge gauging station, groundwater monitoring wells, rain gauge, and the constant injection point at the Yujiawan stream in the Bulang sub-catchment inside the Hailiutu River basin

for the period 1985 to 2008 was 340 mm/year, measured at the nearby meteorological station in Wushenqi. The majority of precipitation falls in June, July, August and September. The mean annual pan evaporation (recorded with an evaporation pan with a diameter of 20 cm) is 2184 mm/year (Wushenqi metrological station, 1985–2004). The monthly pan evaporation significantly increases from April onwards, reaches highest values from May to July, and decreases from August onwards. The geological formations in the Bulang sub-catchment mainly consist of four strata (Hou et al. 2008): (1) the Holocene Maowusu sand dunes with a thickness from 0 to 30 m; (2) the upper Pleistocene Shalawusu sandstone with a thickness of 5 to ~90 m; (3) the Cretaceous Luohe

sandstone with a thickness of 180 to ~330 m; and (4) the bedrock, which consists of Jurassic impermeable sedimentary formations. No continuous semi-permeable formations exist, so that the Quaternary and Cretaceous formations form a continuous regional aquifer system in the Bulang sub-catchment. The natural vegetation in the sub-catchment is dominated by salix bushes (*Salix Psammophila*), while the livelihood of the local people depends on growing maize on croplands. The low density shrub, high density shrub, and cropland in the Bulang sub-catchment occupy 69, 23, and 8 % of the total area, respectively. Groundwater is abstracted for irrigation in the growing season from April to October in the Bulang sub-catchment. The total length of the investigated

segment of the Yujiawan stream is 900 m, while the width of the stream varies from 0.5 m at the upstream section to 2.5 m at the downstream section. The depth of the water in the stream is smaller than 5 cm except during flood events. The stream bank and the streambed are formed by sand.

elevation of wells, height of water column above the MiniDiver in the wells, and the depth of the MiniDiver in the boreholes. The rainfall was recorded by the a rain gauge (HOBO RG3, Onset Corporation, Bourne, USA) near the discharge gauging station (Fig. 1).

Methods

Groundwater and stream stage monitoring

A set of groundwater-level monitoring wells was installed at the streambed, stream banks, flood plain, and the terrace at the upstream of the discharge gauge in the Bulang sub-catchment (Figs. 1 and 2). Groundwater levels in the shallow aquifer were measured in 10-min intervals using a MiniDiver submersible pressure transducer (Eijkelkamp Agrisearch Equipment, Giesbeek, The Netherlands) installed in the monitoring wells. Barometric compensation was carried out using air pressure measurements from a BaroDiver (Eijkelkamp Agrisearch Equipment, Giesbeek, The Netherlands) installed at the site. Groundwater levels were reported as the height of the water table above mean sea level with the calibration of the land

Temperature

Streambed temperatures at depths of 10, 30, 50 and 80 cm were measured by HOBO Pro v2 water temperature sensors (U22-001, Onset Corporation, Bourne, USA) installed in a 5-cm diameter polyvinyl chloride (PVC) pipe (Fig. 2) upstream of the discharge gauge. In order to avoid vertical fluxes in the pipe, the two ends of the PVC pipe were sealed and several PVC blockers were put in the pipe between the sensors. The temperature sensors can register the temperature of the surrounding streambed through horizontal holes at different depths. Water temperature at the surface of the stream bottom is recorded by a temperature sensor on the surface of the streambed. One piezometer, equipped with a MiniDiver, was installed in the streambed for measuring the groundwater level below the stream. The stream stage was registered by a

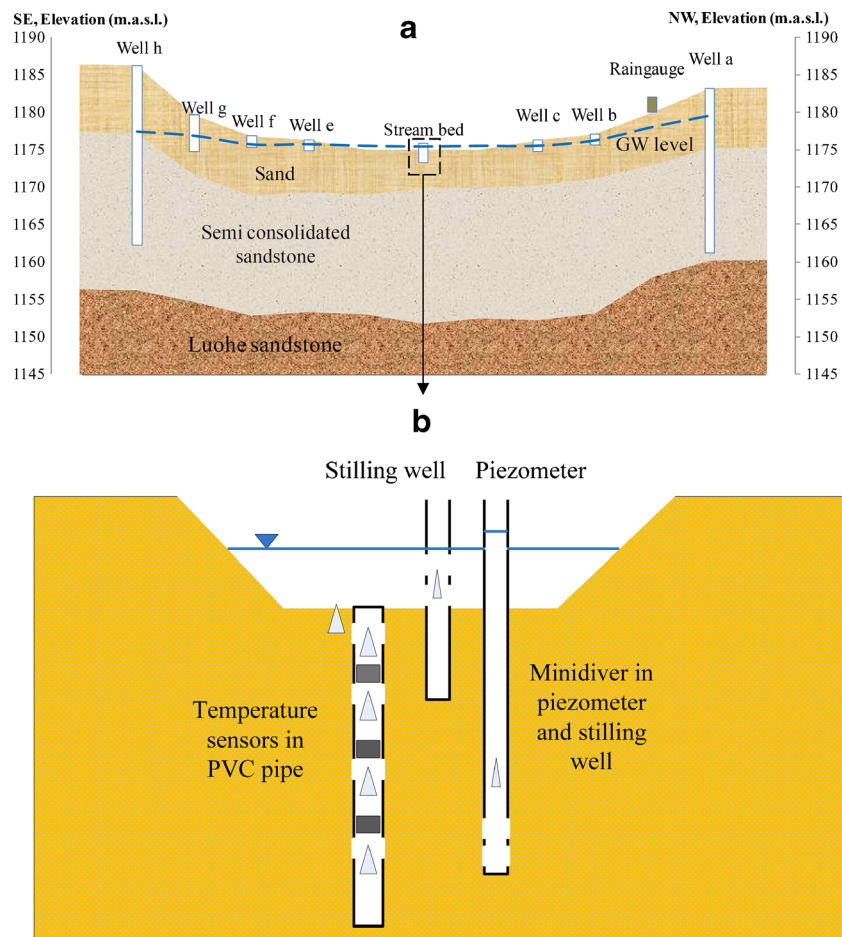


Fig. 2 a Schematic plot of the groundwater monitoring wells installed in the Bulang sub-catchment; the dotted blue line indicates groundwater heads in the monitoring wells and the piezometer in the streambed at 20:00 14 June, 2011; b the installation of temperature sensors, piezometer in the streambed, and the stilling well for the stream stage registration

MiniDiver in steel stilling pipe installed in the stream, where the water level inside the steel pipe is equal to the stream stage. The sampling frequency for groundwater level, stream stage, and temperature at different depth beneath the streambed was set to 10 min.

Discharge measurements

In order to measure the discharge of the Bulang sub-catchment, one discharge gauging station was constructed at the outlet of the sub-catchment (Fig. 1). The Yujiawan gauging station consists of one permanent rectangular weir equipped with an e+WATER L water level logger (type 11.41.54, Eijkelkamp Agrisearch Equipment, Giesbeek, The Netherlands) where water levels are recorded with a frequency of 30 min. Water depths are converted to discharges using a rating curve based on regular manual discharge measurements carried out with a current meter and the velocity-area method.

Event sampling for hydrograph separation

A hydrograph is the time-series record of water level, water flow or other hydraulic properties, and can be analyzed to gain insights into the relationships between rivers and aquifers (Brodie et al. 2007). The runoff event of 2 July 2011 in the Bulang sub-catchment was intensively sampled to carry out a hydrograph component separation. Water samples of precipitation, groundwater, and the stream water were collected from 1 July to 5 July 2011 in hourly intervals and analyzed for their hydrochemical and isotopic compositions. The analyzed isotopes are the stable water isotopes oxygen-18 (^{18}O) and deuterium (^2H). The chemical analysis includes the major anions and cations. The hydrograph separation using tracers is based on the steady-state mass balance equations of water and tracer fluxes.

Seepage calculation with electrical conductivity (EC) profile

Electrical conductivity (EC) profile measurements

A constant salt tracer injection experiment was conducted on 21 June 2011 along a 180-m river profile upstream of the discharge gauge in the Yujiawan stream (Fig. 3). A Mariotte bottle was used for injecting the salt tracer at a constant rate (Moore 2004). The constant injection rate from the Mariotte bottle was manually calibrated with a

stopwatch and beaker in the field. A sodium chloride (NaCl) solution was used for the constant injection experiment. The electrical conductivity was measured along the 180-m profile with intervals of 10 m from the injection point to the downstream discharge gauge before and during the constant injection experiment. The electrical conductivity values of stream water were measured with a portable water-quality multi-meter (18.28 and 18.21.SA temperature/conductivity meter, Eijkelkamp Agrisearch Equipment, Giesbeek, The Netherlands). The background EC profile was measured from 10:00 to 11:00 on 21 June 2011. The constant injection experiment started at 11:10, the stable plateau value of the EC at the discharge gauging station was observed around 11:51. From 11:40 to 14:00, the EC profile of the stream under the constant injection experiment was measured every 10 m along the 180 m stream reach. The width of the stream varies from about 0.5 m at the first 70 m downstream of the injection point to 2 m at the discharge weir. The discharge remained approximately constant during the experiment. The mixing length is less than the 10 m from the theoretical estimation, and was confirmed by measuring the EC value at different locations along the downstream river cross section.

Estimation of seepage with the natural EC profile

Two mass balance equations were formulated with EC measurements under natural conditions. The mass balance of the total 180 m stream reach (Fig. 4a) was used to calculate the upstream inflow and total seepage. The mass balance per stream segment was used to calculate the seepage every 10 m (Fig. 4b).

The mass balance equations for the total 180-m reach (Fig. 4a) are:

$$Q_{\text{in}}C_{\text{in}} + Q_{\text{g}}C_{\text{g}} = Q_{\text{y}}C_{\text{y}} \quad (1)$$

$$Q_{\text{in}} + Q_{\text{g}} = Q_{\text{y}} \quad (2)$$

Where the Q_{in} is the inflow from the upstream in m^3/s ; Q_{y} is the outflow at the Yujiawan gauging station in m^3/s ; Q_{g} is the total groundwater seepage along the 180-m reach in m^3/s ; C_{in} is the EC value of the upstream inflow water in $\mu\text{S}/\text{cm}$; C_{y} is the EC value at the Yujiawan gauging station in $\mu\text{S}/\text{cm}$, and C_{g} is the average EC value of groundwater in $\mu\text{S}/\text{cm}$. Eqs. (1) and (2) can be solved for

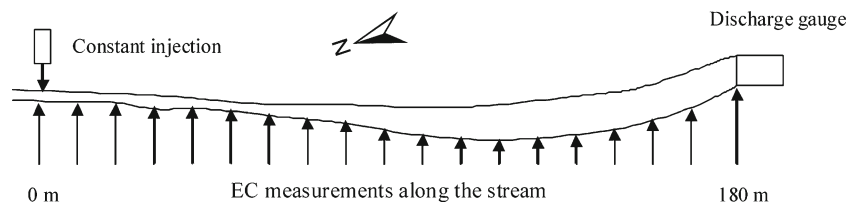


Fig. 3 Plan view of locations of electrical conductivity (EC) measurements in the Yujiawan stream

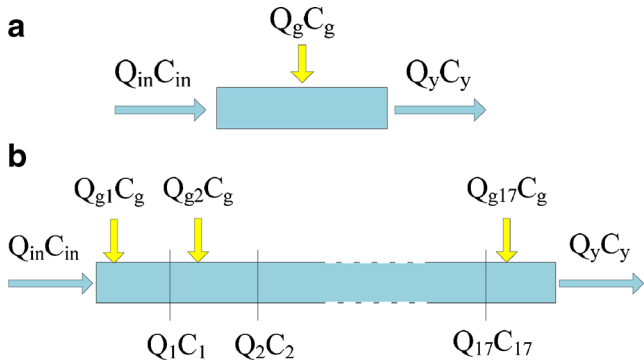


Fig. 4 Schematic plot of mass balance calculations under the natural situation: **a** for the total 180 m reach and **b** for 10-m segments

Q_{in} and Q_g when the groundwater EC value is known. The average EC value of the groundwater was calculated with seven groundwater samples collected from the observation wells near the stream. The EC values vary from 313 to 823 $\mu\text{S}/\text{cm}$ with average value of 545 $\mu\text{S}/\text{cm}$. Therefore, the Q_{in} , Q_g , and the average groundwater seepage rate along the 180-m reach can be calculated.

The mass balance equations for the first 10-m segment (Fig. 4b) can be written as follows:

$$Q_{in}C_{in} + Q_{g1}C_g = Q_1C_1 \tag{3}$$

$$Q_{in} + Q_{g1} = Q_1 \tag{4}$$

where the Q_{g1} and Q_1 are the groundwater seepage in the first 10-m segment and the discharge at the end of the first segment in m^3/s ; C_g and C_1 are the EC values of the groundwater seepage and total discharge at the end of the first segment in $\mu\text{S}/\text{cm}$, respectively. Combining Eqs. (3) and (4) can solve for Q_1 and Q_{g1} with the formula:

$$Q_1 = Q_{in} \frac{C_{in} - C_g}{C_1 - C_g} \tag{5}$$

$$Q_{g1} = Q_1 - Q_{in} \tag{6}$$

The stream discharge and groundwater seepage at the remaining segments can be calculated successively with Eqs. (5) and (6).

Estimation of seepage using the EC profile under a constant injection

The EC profile measurements were taken when a constant plateau value was reached at the Yujiawan gauging station. Because of the measurement error of the extremely high EC value of the injection solution and the mixing effect, the

upstream boundary is taken from 10 m downstream from the injection point. The mass balance for the downstream 170 m reach (Fig. 5a) can be calculated as follows:

$$Q'_1C'_1 + Q'_gC_g = Q'_yC'_y \tag{7}$$

$$Q'_1 + Q'_g = Q'_y \tag{8}$$

Where, Q'_1 is the discharge at 10 m downstream from the constant injection point in m^3/s ; Q'_y is the discharge at the Yujiawan gauging station in m^3/s ; Q'_g is the total groundwater seepage along the 170-m reach in m^3/s ; C'_1 is the EC value 10 m downstream in $\mu\text{S}/\text{cm}$ and C'_y is the EC value at Yujiawan gauging station in $\mu\text{S}/\text{cm}$.

The mass balance equations for the second segment from 10 to 20 m (Fig. 5b) can be written as follows:

$$Q'_1C'_1 + Q'_{g2}C_g = Q'_2C'_2 \tag{9}$$

$$Q'_1 + Q'_{g2} = Q'_2 \tag{10}$$

Where the Q'_{g2} and Q'_2 are the groundwater seepage rates in the segment and the discharge at the end of the segment in m^3/s ; C_g and C'_2 are the EC values of the groundwater seepage and stream water at the end of the segment in $\mu\text{S}/\text{cm}$, respectively. With the estimated Q'_1 in Eqs. (7) and (8), Q'_{g2} and Q'_2 can be computed with the formula:

$$Q'_2 = Q'_1 \frac{C'_2 - C_g}{C'_1 - C_g} \tag{11}$$

$$Q'_{g2} = Q'_2 - Q'_1 \tag{12}$$

The stream discharge and groundwater seepage at the remaining segments can be calculated consecutively with Eqs. (11) and (12).

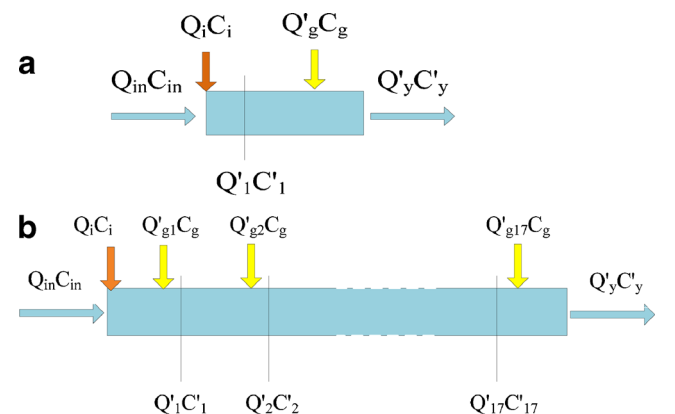


Fig. 5 Schematic plot of mass balance calculations under constant injection: **a** for the total 180-m reach; and **b** for 10-m segments

Estimation of seepage with the combined EC profiles

Stream discharge and groundwater seepage can be calculated by using both the natural EC profile and the EC values under the constant injection. The mass balance equations for the total 170-m reach are:

$$Q_1 C_1 + Q_g C_g = Q_y C_y \quad (13)$$

$$Q'_1 C'_1 + Q'_g C'_g = Q'_y C'_y \quad (14)$$

$$Q_1 + Q_g = Q_y \quad (15)$$

$$Q'_1 + Q'_g = Q'_y \quad (16)$$

Equations (13)–(16) can be solved for Q_1 and Q_g when assuming that the stream discharge and groundwater seepage remain the same during the natural and constant injection EC measurements. Indeed, the difference of stream discharge at the Yujiawan gauging station is very small between the natural and constant injection EC measurements. The average discharge at the Yujiawan from 10:00 to 13:00 can be used and is calculated to be $0.0314 \text{ m}^3/\text{s}$. Therefore, Q_1 and Q_g along the 170-m river reach can be calculated. The mass balance equations for the second 10-m segment from 20 to 30 m downstream under the natural and constant injection conditions are:

$$Q_1 C_1 + Q_{g2} C_g = Q_2 C_2 \quad (17)$$

$$Q'_1 C'_1 + Q'_{g2} C'_g = Q'_2 C'_2 \quad (18)$$

Equations (17) and (18) can be solved to calculate Q_2 and Q_{g2} . Stream discharge and groundwater seepage at the remaining segments can be calculated successively.

Sensitivity analysis

The reliability of tracer methods for estimating groundwater and surface-water exchange was evaluated by Ge and Boufadel (2006) and Wagner and Harvey (2001). They concluded that the stream tracer approach had minimal sensitivity to the surface-subsurface exchange at high baseflow conditions. As indicated by the hydraulic gradients and thermal methods, the groundwater discharge to the surface water dominates the main interaction between groundwater and surface water in this 180 m reach, which reduces the probability of overestimation of groundwater discharge caused by losing solute in the process. For the seepage calculation methods in this case study, uncertainties may result from the estimation error of the groundwater EC value along the stream bank, EC measurement errors along

the reach, and discharge measurement error at the gauging station. Given unknown distribution of the error distribution of the calculation, a simple sensitivity analysis was carried out to investigate relative errors in seepage calculation caused by the likely EC and discharge measurement errors. The range of the EC values from the groundwater monitoring wells in the stream valley is $510 \mu\text{S}/\text{cm}$, which is very large and most likely caused by irrigation. Groundwater EC values along the stream bank are expected to be lower than the measured EC values of natural stream water. Therefore, a 5% variation with respect to the average value is assumed for the sensitivity analysis. The sensitivity of EC measurement errors on the calculation of groundwater seepage and discharge was investigated by systematically increasing and decreasing of EC values along the river by 5%.

Results of the measurements

River discharge measurements

Several peaks corresponding with rainfall events can be observed in Fig. 6. The maximum daily discharge is $0.231 \text{ m}^3/\text{s}$ on 2 July 2011 when a heavy rainfall event occurred. The minimum daily discharge is $0.0105 \text{ m}^3/\text{s}$ on 17 May 2011, which was caused by water diversions and groundwater abstraction for irrigation from the end of April till the beginning of October. The stream flow keeps relatively stable during the winter, since there is hardly any precipitation and no anthropogenic water use during this period. Furthermore, the constant discharge indicates groundwater sustaining the base flow. The average discharge on 21 June 2011 was $0.0316 \text{ m}^3/\text{s}$ from 10:00 to 11:00, and $0.0321 \text{ m}^3/\text{s}$ from 13:00 to 14:00 when the constant injection experiment was carried out with an injection rate of $89.7 \text{ ml}/\text{min}$. The discharge increase after 18:00 was caused by the stopping of water diversions and groundwater abstraction for irrigation in the adjacent flood plain.

Groundwater level measurements

The groundwater levels in boreholes (Fig. 1) at the terrace (well a), the flood plain (well b), stream bank (well c), and below the streambed of the Yujiawan stream are shown in Fig. 7. Groundwater levels are relatively stable through the year. Heavy rainfall events in July cause an abrupt rise of the groundwater levels in all wells.

Temperature measurements

Figure 8 shows the temperature measurements at the streambed surface and at different depths beneath the streambed from September 2010 to October 2011. The gap in April was caused by the limitation of data storage capacity of the data loggers. The temperature at a depth of 80 cm beneath the streambed was constant throughout the year. The temperature at depths of 50 and 30 cm show some small fluctuation, while the temperature at the surface of the streambed shows large daily and seasonal variations.

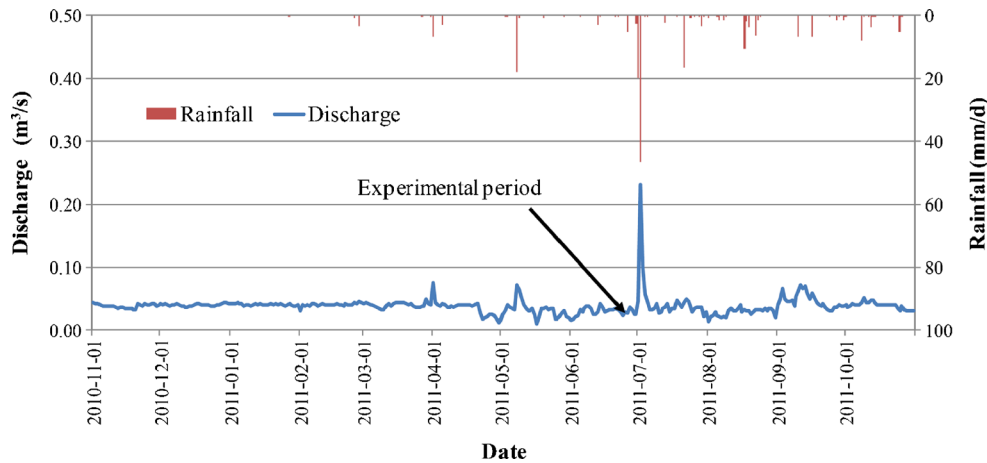


Fig. 6 Stream discharge at Yujiawan gauging station and rainfall at the rain gauge from 1 November 2010 to 31 October 2011

Selected hydrochemical parameters

Figure 9 illustrates changes of stable isotope values for deuterium and oxygen-18, and the concentration changes of the cations potassium (K^+) and sodium (Na^+) as well as the anions chloride (Cl^-) and nitrate (NO_3^-) in river water samples from 1–5 July 2011, which included the heavy rainfall on 2 July 2011. After the heavy rainfall event on 2 July, stable isotope values decreased, while concentrations of both cations and anions increased. The concentrations of the cations (K^+ and Na^+) and anions (Cl^- and NO_3^-) in stream flow samples increased during the discharge event caused by the heavy rainfall event at the beginning of July (Fig. 9).

measured for the natural status (before conducting the constant injection experiment) and under the constant injection condition. Figure 10 illustrates the difference in EC values along the Yujiawan stream for natural and injection conditions. The EC values for natural stream water vary from 640 $\mu S/cm$ at the constant injection point to 581 $\mu S/cm$ at 180 m downstream at the gauging station. The highest EC value of 882 $\mu S/cm$ was measured 10 m downstream of the injection point during the constant injection condition and the EC values decreased gradually to 660 $\mu S/cm$ at 180 m downstream.

Constant injection tracer experiment

The electrical conductivity (EC) values along the Yujiawan stream from the constant injection point to the discharge gauging station located 180 m downstream were

Discussion

Hydraulic method

The typical distribution of groundwater levels in the sub-catchment at 20:00 14 June 2011 is plotted in Fig. 7. It

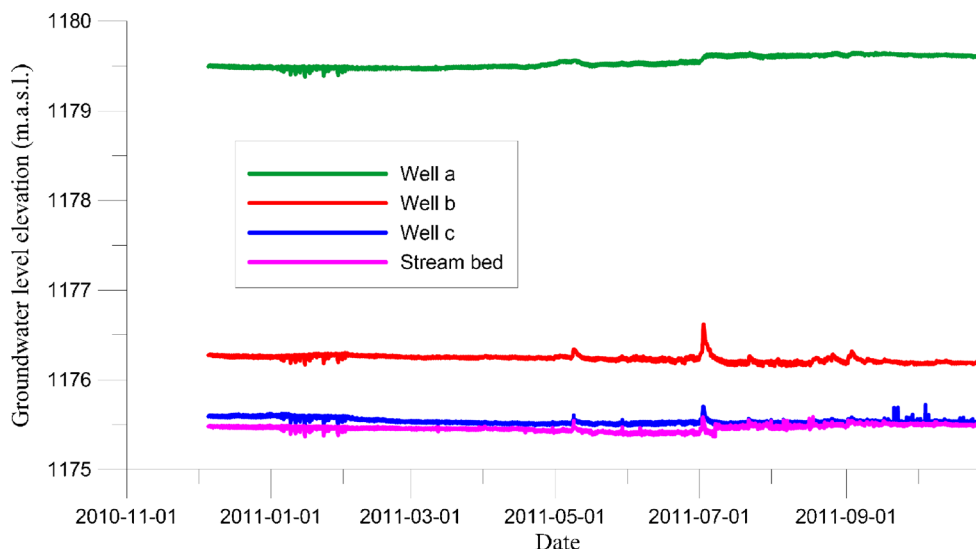


Fig. 7 Groundwater levels below the terrace (*well a*), flood plain (*well b*), stream bank (*well c*), and streambed from 1 November 2010 to 31 October 2011

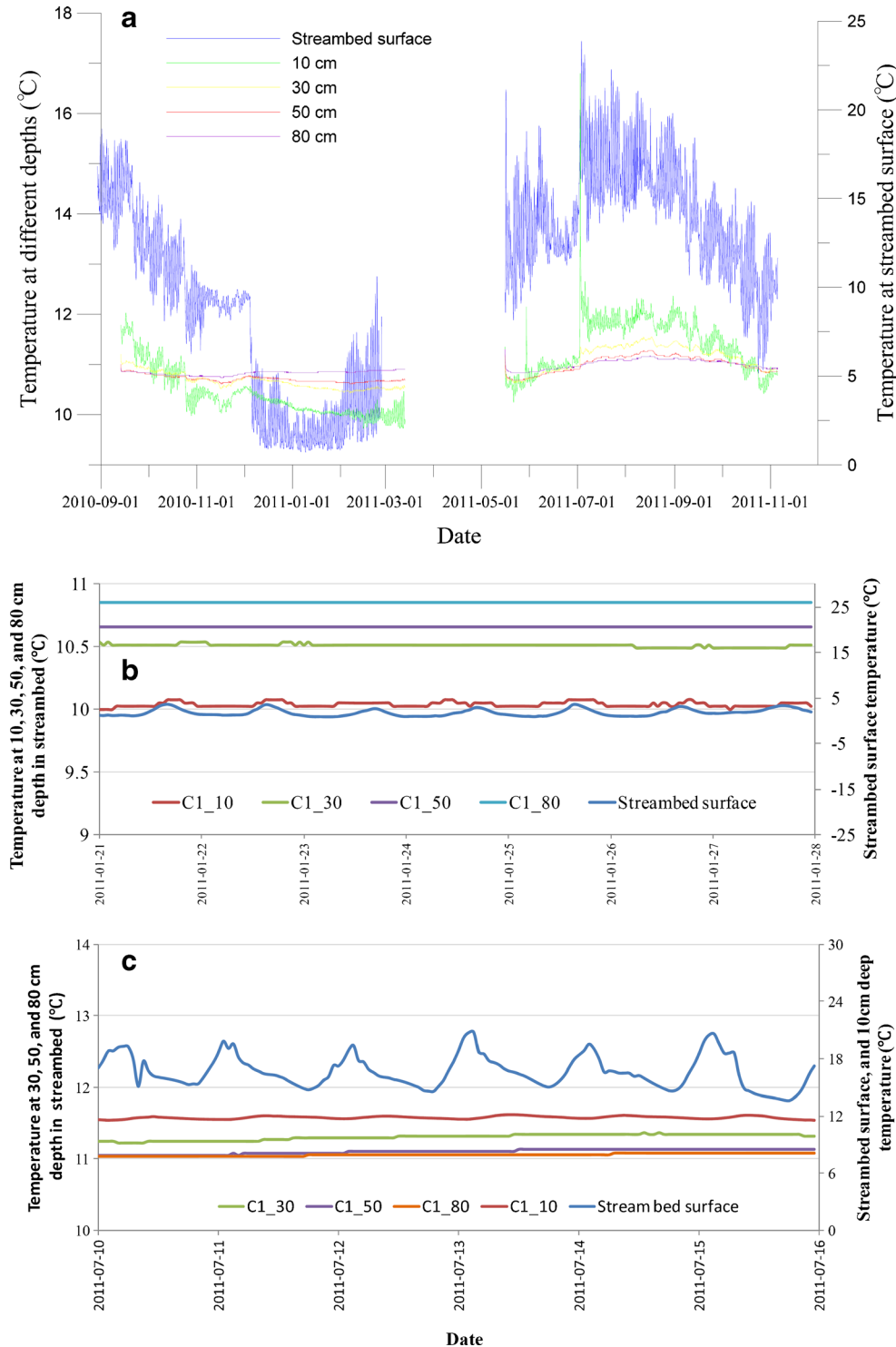


Fig. 8 a Temperature of stream water and stream sediment at 10, 30, 50, and 80-cm depth beneath the stream bed from 1 September 2010 to 31 October 2011. Temperature at the surface of the streambed (*right y-axis*) and at different depths beneath the streambed in **b** winter and **c** summer in 2011

indicates that groundwater flows from both hill slopes towards the stream. Groundwater discharge to the stream occurs during the whole year since groundwater levels in the valley are always higher than the stream stage during the period from December 2010 to October 2011 (Fig. 7).

Temperature method

For a gaining stream reach with upward advection, the oscillating surface temperature signal is attenuated at shallow depths, such that the greater the discharge, the greater the attenuation of temperature extremes, and the

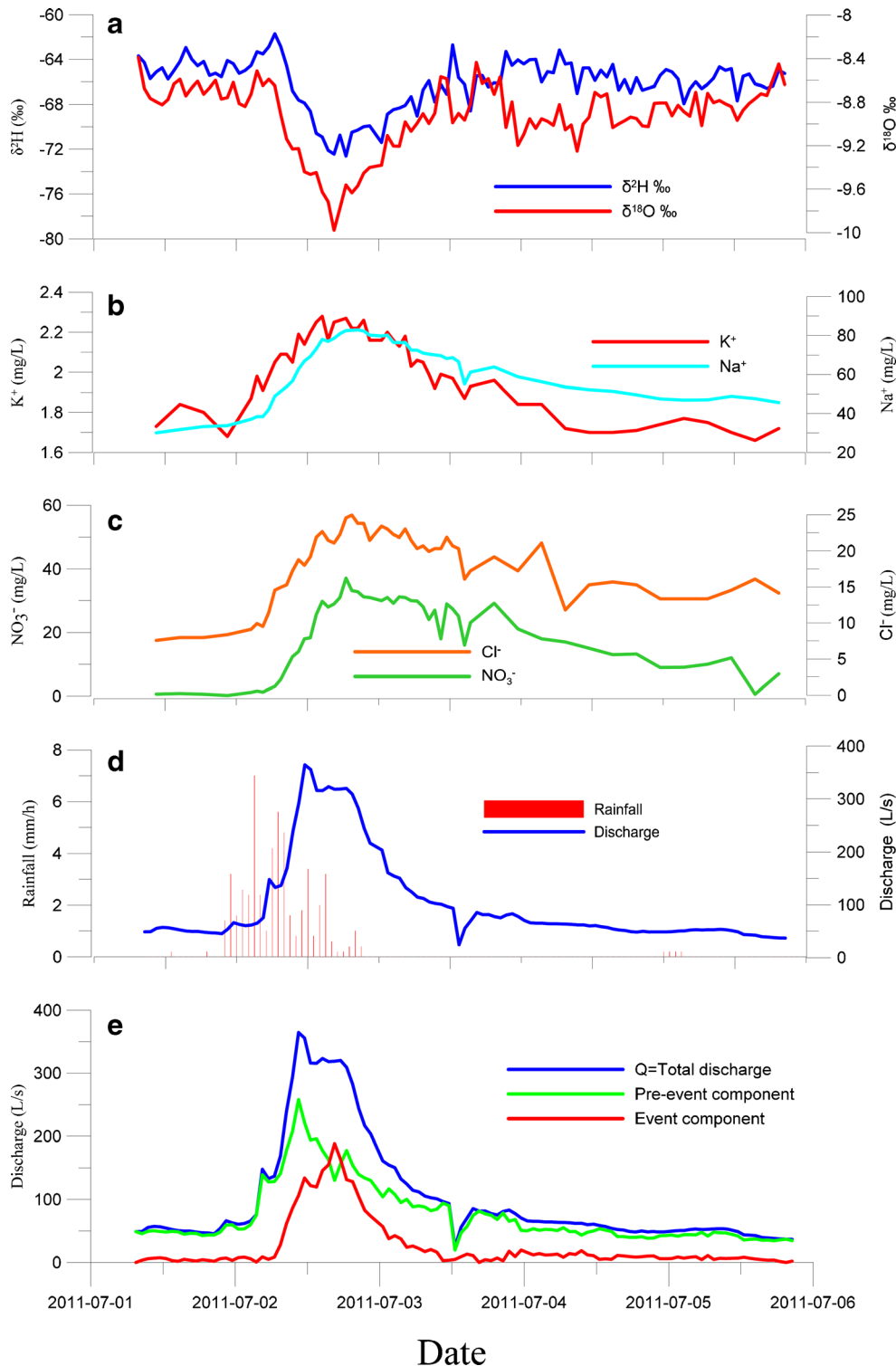


Fig. 9 a–c Stable isotope values, hydrochemical behavior of streamflow samples, **d** rainfall and discharge from 1–5 July 2011, and **e** a two-component hydrograph separation using oxygen-18 for the rainfall event that occurred on 2 July 2011

greater the lag in temperature extremes in the sediments. Temperature measurements (Fig. 8) also show groundwater discharge to the stream for the whole investigated period. Variations of temperatures during a week in winter (21–28 January 2011) and in summer (10–16 July 2011)

are plotted to view temperature differences at different depths. During winter (January), the air temperature is below zero, and the stream-water temperature at the stream bottom is still above the freezing point with small diurnal fluctuations (Fig. 8b). Temperature increases with

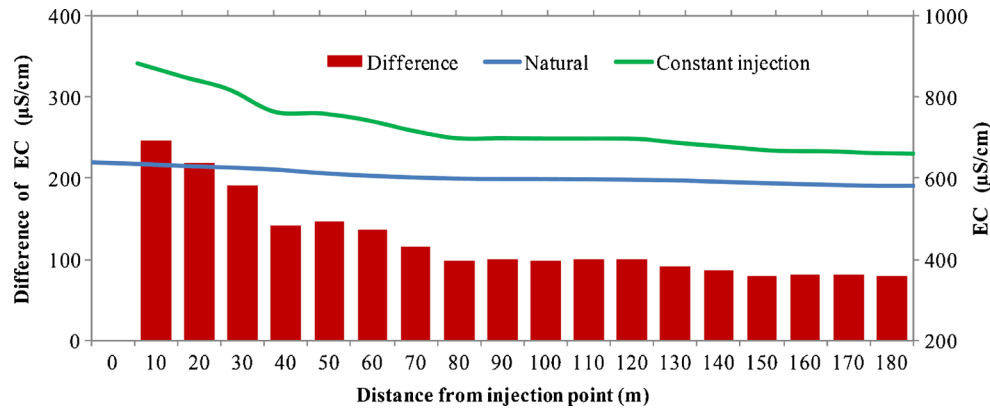


Fig. 10 Measured natural electrical conductivity (*EC*) profile and *EC* values during the constant injection experiment on 2 June 2011

the increase in depth, and the temperature at 80 cm remains constant at around 11 °C, which represents groundwater temperature. These conditions show that both convective and dispersive heat transport are from groundwater upward to the stream. The higher temperature of the groundwater prevents stream water from freezing. In the summer period (July), the stream-water temperature at the stream bottom is much higher than the groundwater temperature (Fig. 8c). The large diurnal fluctuations of stream-water temperature do not penetrate into the streambed deposits. However, temperature decreases with the increase of depth. The lack of diurnal fluctuations in sediment temperature indicates that convective heat transport is from the groundwater upward to the stream, while dispersive heat transport is from the stream downward to the groundwater.

Hydrograph separation

The increased concentrations of chemical components after the start of the rainfall event could be caused by the dissolution of salts which have accumulated in the top soil of the agricultural fields or from the fertilizers used for agriculture in the sub-catchment. Discharge suddenly decreased from 3 July, as did the concentrations of several chemical constituents, which could be due to the decrease of event-water with high chemical constituents discharged into the stream after the stop of rainfall on 2 July. It was not possible to carry out hydrograph separations using these tracers, because it was not possible to measure the representative end-member concentration of the runoff components or to assume that they are constant in time. Thus, the stable isotope oxygen-18 was used for a two-component hydrograph separation to separate pre-event and event water (Fig. 9). The groundwater discharge to the stream could be estimated through the proportion of pre-event water to the total stream flow while the event water represents the surface runoff to the stream. Light rainfall occurred at the end of June, followed by the heavy event starting from 1 July. The event samples were collected from 1 July till 5 July. There was intensive rainfall during

the 12 h on 2 July within the Bulang sub-catchment. The isotopic components in the rainfall and groundwater were assumed to be constant in space and time of the duration of the investigated event. The results of hydrograph separation illustrate that the pre-event component accounts for 74.8 % of the total discharge during this heavy rainfall event. The response of the pre-event component to the rainfall is faster than the event water component that might be caused by the so-called kinematic wave effect (Buttle 1994; Cey et al. 1998; Krein and De Sutter 2001), which is caused by a faster flood wave propagation velocity compared to the flow velocity of water. This results in a high groundwater discharge component in the stream also during discharge events.

Seepage calculation with *EC* profile

Calculated results

Figure 11 and Table 1 present the results of seepage estimation by the three methods. For a relatively homogeneous sandy aquifer like in this test case, groundwater seepage along this river reach shows large spatial variations that might be caused by the variations of the streambed topography. Large spatial variations of groundwater discharge to rivers were also found in other studies using temperature surveys (Becker et al. 2004; Lowry et al. 2007). It can be seen that the groundwater seepages calculated by the constant injection and combined method are very close (Fig. 11). The natural *EC* profile method calculates slightly lower seepage rates. All three methods calculate a comparable discharge value at the gauging station (Table 1). The calculated seepage under a bridge (at 100 m distance) is zero since the concrete wall of the bridge is not permeable. Groundwater seepage is not uniform along the reach. In segments with very low seepage rates, the combined method calculated negative seepage rates which are possibly caused by local-scale streambed topography variations or by measurement errors of the *EC* values. The application of natural and constant injection methods requires the estimation of *EC*

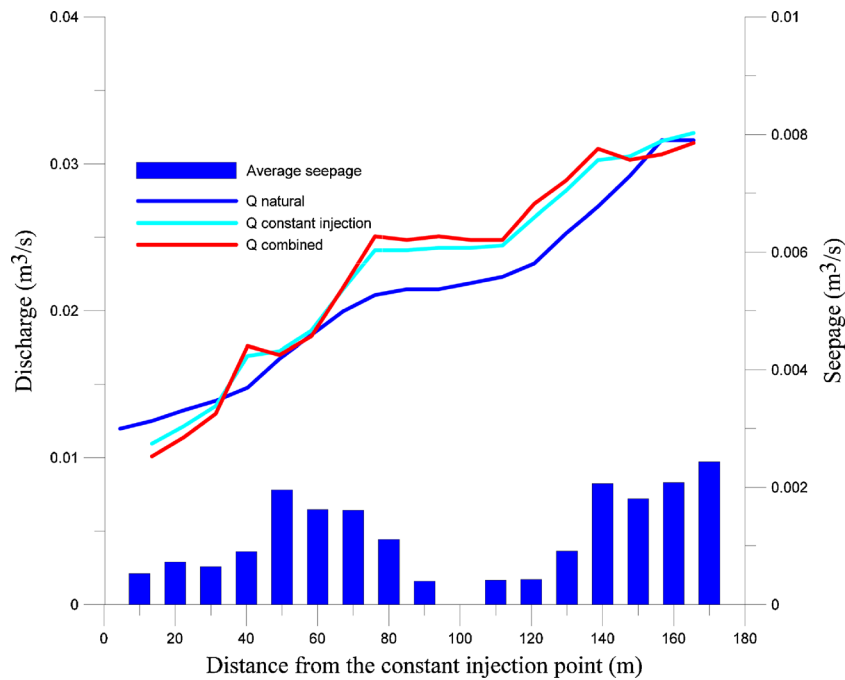


Fig. 11 Estimation of seepage and discharge along the stream reach

values of groundwater along the stream bank. The combined method eliminates the unknown groundwater EC values, therefore, it is more convenient to use.

Sensitivity analysis

The results of sensitivity analysis are shown in Fig. 12. It is clear that the natural EC profile method is also more sensitive to EC measurement errors. The measured mean discharge at the gauging station from 10:00 to 14:00 was used for the estimation of the groundwater seepage with the combined method. The standard deviation of the discharge during the experiment is $0.00143 \text{ m}^3/\text{s}$. Thus, the sensitivity analysis to the discharge measurement errors was analyzed by increasing and decreasing the discharge with two standard deviations. Figure 13 shows that the sensitivity of the estimated river discharge along the reach to the discharge measurements at the downstream gauge station is smaller than the likely measurement error at the gauging station.

Table 1 Seepage calculation for the 180-m reach by the three EC-profile methods

	Natural	Constant injection	Combined
Inflow from upstream (m^3/s)	0.01197	0.01095	0.01009
Average seepage (m^3/s)	0.0196	0.02115	0.02133
Discharge at 180-m downstream (m^3/s)	0.0316	0.0321	0.0314
Groundwater seepage rate ($\text{m}^3/\text{m}/\text{d}$)	9.41	10.75	10.84

The sensitivity analysis indicates that the combined method for determining groundwater seepage provides the most reliable estimations. Another advantage of the combined method is that it does not need to measure EC values of groundwater along the stream bank, therefore, reducing measurement costs.

The hydraulic method indicates the interaction between the groundwater and surface water by providing the difference between groundwater heads and river stage. The temperature method could be applied in losing or gaining reaches at carefully selected locations, which identifies the direction of water flow in the streambed. However, either the hydraulic method or temperature method should only be employed for quantitative estimation with external hydraulic information such as the hydraulic conductivity, and head difference between groundwater and river stage. Hydrograph separation could only be applied during rainfall events with extensively field work, isotopic and chemical analysis in the laboratory. The EC profile method can be finished within hours, which is more efficient compared with other methods.

Conclusions

Multiple field measurements for an investigation period of 1 year indicate that groundwater discharges to the river during the entire period in the Bulang sub-catchment. Even during heavy rainfall events, river discharge is composed of more groundwater discharge than direct surface runoff, which is in line with other field observations and the high infiltration rates in the catchment

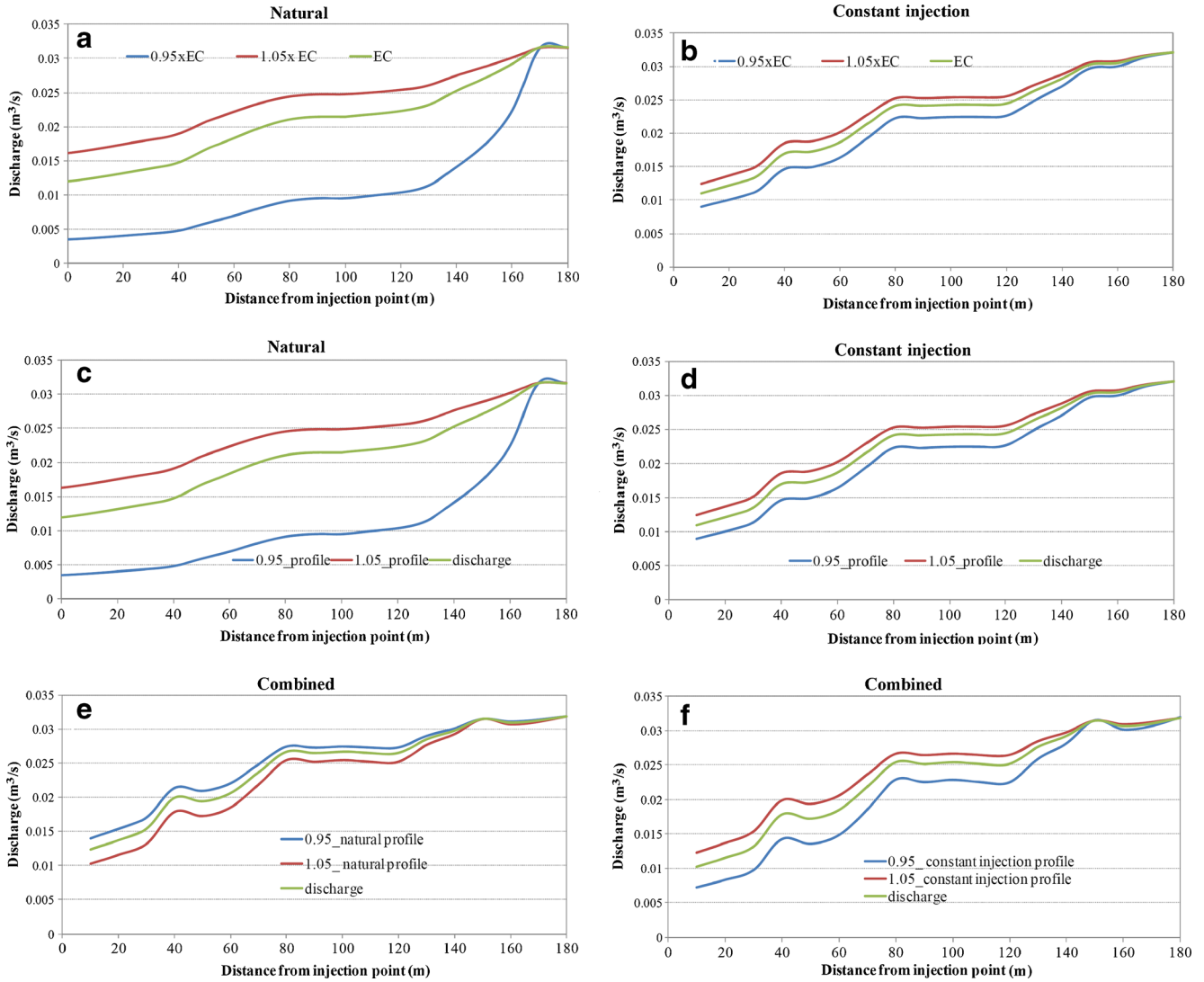


Fig. 12 Sensitivity of estimated stream discharge to groundwater EC value along the stream bank **a** with the natural EC profile method, **b** with the constant injection method. Sensitivity of estimated stream discharge to EC measurement errors **c** with the natural EC profile method, **d** with the constant injection method, and **e-f** with the combined method

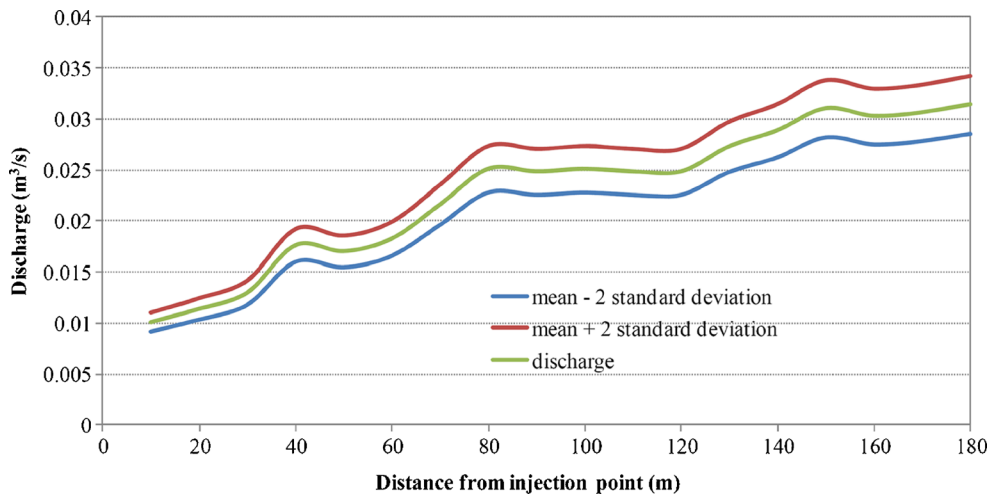


Fig. 13 Sensitivity of estimated stream discharge along the reach to the discharge measurement error at the gauging station using the combined method

(dominated by sand dunes). Consequently, groundwater and stream water are essentially one resource and need to be managed conjunctively.

Groundwater level measurements in a cross-section in relation to stream stage measurements provide indication of flow directions of groundwater and stream interactions. The measurements in the Bulang sub-catchment show that groundwater levels are higher than the stream stage for the whole measurement period, a clear evidence of groundwater discharge to the stream all the year round. The temperature measurements provide additional information on groundwater and stream interactions. Groundwater has a relatively constant temperature, while stream-water temperature has not only seasonal changes, but also diurnal fluctuations. Temperature measurements in the streambed deposits at different depths can identify the direction of water exchange between groundwater and surface water. The Bulang River never freezes despite the cold temperature in the winter, due to groundwater seepage with a higher temperature. Both convective and dispersive heat transports occur in the same upward direction from groundwater to the stream in the winter period. In summer, stream-water temperature is much higher than groundwater temperature. The dispersive heat transport is downward, but the convective heat transport is upward. The large diurnal fluctuation of stream-water temperature is dammed by the upward cold groundwater flow. Therefore, temperature of streambed deposits below 30 cm depth remains stable, while the temperature at shallow depth shows small fluctuations.

River gauging is indispensable for quantifying groundwater and stream interactions. First, hydrograph separation of measured stream discharge provided an estimate of total groundwater discharge to the stream. Second, mass balance equations with EC measurements were solved more accurately with the measured discharge at the gauging station. The combined use of EC profile measurements under the natural situation with the constant injection can result in spatially distributed groundwater seepage estimates along the stream. Sensitivity analysis demonstrates that the combined method is neither very sensitive to the EC measurement errors, nor to the measurement errors of the discharge measurements, and provides the most reliable estimation of groundwater seepage.

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