

Groundwater storage changes and estimation of stream lateral seepage to groundwater in desert riparian forest region

Haiyang Xi, Qi Feng, Lu Zhang, Jianhua Si and Tengfei Yu

ABSTRACT

Field experiments were conducted in the lower reaches of the Heihe River basin in the northwest of China to determine relationships between stream and groundwater and to quantify stream lateral seepage. The water table fluctuation, water balance and statistical analysis methods were used to analyze long-term experimental data. Results showed that the groundwater depth along the experimental section responded strongly to the streamflow of the East River in the basin. The streamflow of the East River in all periods significantly influenced the amplitude of groundwater depth within the range of 300 m from the riverbank; the stream lateral seepage was mainly controlled by the streamflow and its durations. The evapotranspiration of riparian forest had used larger proportions of groundwater storage in summer, close to approximating 60%. This study quantified stream lateral seepage to groundwater for different water conveyance and provides support for regional water resources management in an arid inland river basin.

Key words | desert riparian forest, eddy covariance, groundwater, recharge, stream lateral seepage, water table fluctuation method

Haiyang Xi (corresponding author)

Qi Feng

Jianhua Si

Tengfei Yu

Key Laboratory of Ecohydrology of Inland River Basin,

Chinese Academy of Sciences, Lanzhou 730000, China

E-mail: xihy@lzb.ac.cn

Lu Zhang

CSIRO Land and Water Flagship,

Canberra,

ACT 2601, Australia

Haiyang Xi

Jianhua Si

Tengfei Yu

Alashan Desert Eco-hydrology Experimental Research Station,

Ejina 735400, China

Haiyang Xi

Qi Feng

Tengfei Yu

Northwest Institute of Eco-Environment and Resources,

Chinese Academy of Sciences, Lanzhou 730000, China

INTRODUCTION

Knowledge of aquifer and river water interaction is important for understanding the continuum of groundwater and surface water hydrology (Chou & Wyseure 2009). Quantification of interactions in surface-groundwater systems is an important research area because of its central role in conjunctive water management and efficient utilization of water resources (Sophocleous 2002). The interaction between surface water and groundwater affects the hydrological cycle in a river basin (Engeler *et al.* 2011a; Cai *et al.* 2016; Nakamura *et al.* 2017), and it is one of the critical components of hydrological processes (Wang *et al.* 2010). In arid and semi-arid regions of the world, water resources are limited, and under severe and increasing pressure due to population growth and increasing water use for irrigation (Ghazavi *et al.* 2010). Therefore, an understanding of

interactions between surface water and groundwater is needed for effective management of water resources (Sophocleous 2002). It is important to be able to quantify water exchanges between streams and aquifers (Osman & Bruen 2002). Recharge quantification is an important prerequisite for effectively managing groundwater resources as recharge estimates are needed to determine sustainable yields of groundwater aquifers for sustainable exploitation of the resource (Obuobie *et al.* 2012).

It is important to identify interactions between streamflow and groundwater for industry, agricultural water use, and environmental protection (Wan *et al.* 2012). Streams may gain water from aquifers or may lose water to aquifers depending on catchment characteristics (Doppler *et al.* 2007). Peterson & Wilson (1988) summarized four types of

relationships of stream-aquifer: connected gaining stream, connected losing stream, disconnected stream with a shallow water table, and disconnected stream with a deep water table. Identification of interactions between stream and aquifers is important for regional water budgets (Farber *et al.* 2005). Abdulrazzak & Morelseytoux (1983) studied the ephemeral stream recharge to groundwater, and found that the infiltration through ephemeral stream beds was the major source of aquifer recharge in arid regions. Voltz *et al.* (2013) determined the patterns of riparian hydraulic gradients and stream-groundwater exchange. Wan *et al.* (2012) revealed the hydraulic relationship between the Molin River and groundwater; the results showed that the Molin River was gaining from groundwater in the upper reaches and was recharged by groundwater in the lower reaches. In recent years, a number of methods and techniques have been used to explore the relationship between river and groundwater.

Many methods can be used to estimate the stream recharge to groundwater, including water balance method (Schulz *et al.* 2013), water chemistry method (Coleman *et al.* 2017), isotopes method (Becher Quinodoz *et al.* 2017), pumping experiment approach (Weatherington-Rice *et al.* 2000), water table fluctuation (WTF) method (Maasland 1959; Rai & Singh 1981), mathematical model (Vauclin & Vachaud 1977; Zhou 2011), and numerical simulation method (Rahayuntyas *et al.* 2014). Keating (1982) constructed a lumped parameter model of a chalk aquifer-stream system in Hampshire, UK, which revealed the presence of a thin, highly transmissive and storative zone in the chalk close to the water table. Dillon & Liggett (1983) built an ephemeral stream-aquifer interaction model; the results showed that the model predicted with sufficient accuracy the inflow and outflow from the stream to the aquifer during fluctuating water levels in the stream. Although analytical models can solve some of the problems concerning the interactions of stream-groundwater (Nawalany 1994; Workman *et al.* 1997; Serrano & Workman 1998), numerical models have become an important method to solve complicated questions about stream-groundwater interaction (Danierhan *et al.* 2013). Chemical methods also can be used to study stream-groundwater interactions (Majumder *et al.* 2013). Yang *et al.* (2012) used major ion chemistry to characterize the interactions between surface water and groundwater in the Jialu River

basin. Isotope tracing is also an important way to determine the interaction between river and groundwater (Yang *et al.* 2012). Ayenew *et al.* (2008) applied environmental isotopes to study the surface water and groundwater interaction in the Awash River basin, conceptualized the surface water and groundwater interaction, and explored the groundwater flow pattern in relation to the geological setting. In recent years, heat tracing method has been used to study surface-groundwater interactions (Duque *et al.* 2010). Lee *et al.* (2013) revealed the interaction of stream water and groundwater by temperature monitoring in the Haeon basin, Korea. Heat trace has been proven to be an effective method for determining the interaction between surface water and groundwater in many regions (Wu *et al.* 2013). The WTF method can effectively estimate the groundwater storage changes based on water table changes, which proved easy to obtain and observe (Delin *et al.* 2007). The above methods provide an important basis for the research of surface water-groundwater interaction. Using a combination of the above techniques can also effectively resolve the practical problems about the interaction between river and groundwater (Dahan *et al.* 2008; Engeler *et al.* 2011b). This study mainly used water balance and WTF methods to estimate the stream lateral seepage to groundwater based on the experimental data (ET eddy covariance) and observation data (groundwater level).

The environment has changed remarkably in recent decades with cease-to-flow in the Heihe River, drying-up of lakes, groundwater level declining, vegetation degeneration, atrophy of natural oasis, and desertification and frequent sand storms in the Ejina desert oasis of northwestern China under the conditions of global climate change and intensified human activities (Xi *et al.* 2010). Significant changes have taken place in desert riparian forest and availability of water resources in the Ejina desert oasis of the lower reaches of the Heihe River basin. The river and groundwater are the key factors that affect plant growth and the community succession of desert riparian forest in arid regions. The vulnerable ecosystem has been facing enormous threats in the Ejina desert oasis. A number of studies have reported the impacts of water environmental factors on the ecosystem in Ejina desert oasis. However, studies reporting on the river seepage and its effects on desert riparian forest are very limited for this region. The seepage from

the riverbed could provide water not only for riparian vegetation, but also for the aquifer. The stream lateral seepage could be vital for the Ejina desert riparian forest of this extremely arid inland region with an extremely fragile ecological environment.

In this context, the main objectives of this study were: (1) to quantify interactions between stream and aquifer; (2) to determine the effect of ephemeral streams on groundwater in the riparian zone; (3) to assess the extent of river–groundwater interaction in a riparian forest region, especially to find the relationship between stream lateral seepage and riverbank storage; and (4) to determine the difference between groundwater storage changes in different locations away from the riverbank and to estimate the lateral seepage from stream to groundwater.

STUDY AREA

The field experiments were carried out in Alashan Desert Eco-hydrology Experimental Research Station of Northwest Institute of Eco-Environment and Resources, Chinese Academy of Sciences (N 42°02′02.79″, E 101°03′07.26″), located near Ejina County, 390 km northeast of Jiuquan City (Figure 1(a)). The experimental site is situated in the lower reaches of the Heihe River, northwest China. The Heihe

River originates in the Qilian Mountains and flows through Zhengyixia, Shaomaying, and Langxin Mountain hydrological sections and finally flows into the terminal of East Juyan Lake and West Juyan Lake (Figure 1(b)). The Heihe River has a total length of 330 km and is divided into the East River and West River downstream of Langxin Mountain. The downstream of Langxin Mountain becomes ephemeral and only flows for 154–269 days per year (average 214 days for the period 1999–2013), with most flow events occurring during August to October and December to the following April. The average annual flow volume is approximately $531 \times 10^6 \text{ m}^3$ at the Langxin Mountain station during the period of 1999–2013. The stream channels are wide and shallow with mostly Quaternary unconsolidated sediments, coupled with an intensive network of tributaries; leakage occurs easily, potentially providing direct recharge to the underlying aquifers. Annual average precipitation is approximately 34 mm (1960–2012) in this study area, with maximum and minimum of 101 mm and 7 mm, respectively. The precipitation mostly occurs in June to September. The average annual pan evaporation is about 3,218 mm (1960–2012).

Hydrogeologically, the study area belongs to the western Alashan highland of Inner Mongolia, and is made up of a series of mid or low denudation mountains, proluvial fan dry delta and basin. Its altitude is 890–1,200 m above sea level. The lowest points are located in East Juyan Lake

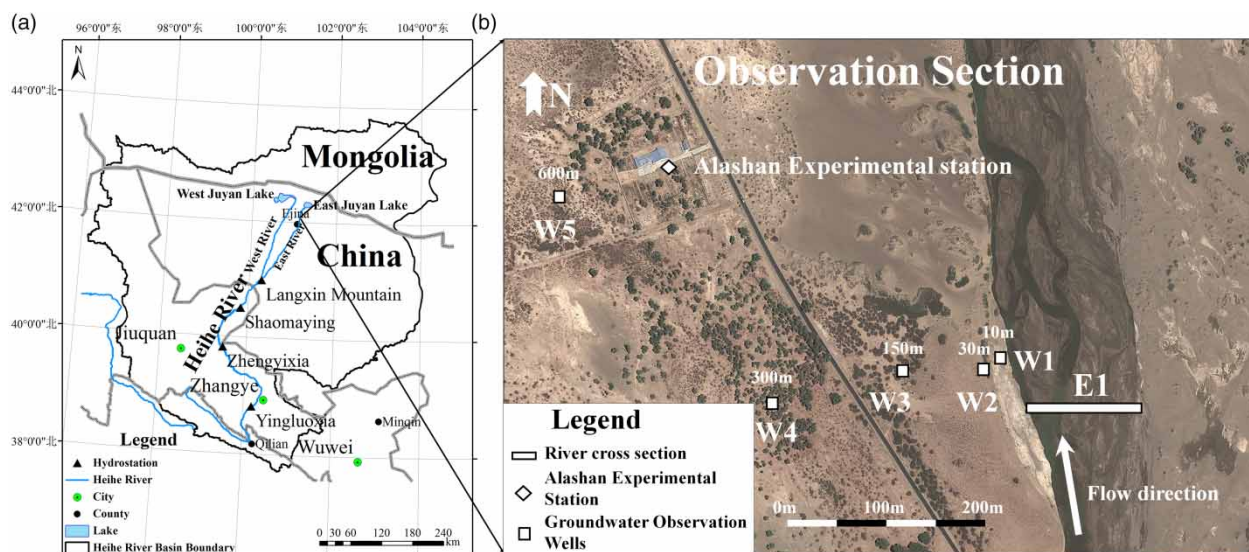


Figure 1 | Sketch map of the lower reaches of Heihe River basin (right of plot from Google Earth).

and West Juyan Lake; the highest point is in the southern Langxin Mountain. This study area has a multilayered aquifer system with Quaternary unconfined aquifer overlying confined sand and limestone aquifers. The lithologic characteristics of the aquifer system vary gradually, from gravel to fine sand, from the south to the north of the basin, and the water table gradually becomes shallow in the north. The aquifer system transforms from a one-layer phreatic aquifer to a several-layered unconfined–confined aquifer system. The one-layer aquifer system is in the south of the basin, and the several-layered unconfined–confined aquifer system is in the northern part. The depth of these layers is 50–500 m, the depth of aquifers is 150–200 m. The groundwater depth is less than 7 m for most of the area. The salinity of shallow groundwater is greater than 1 g/L. Groundwater is primarily used for drinking water and riparian forest growth (Xi *et al.* 2010).

The stream channel is 150 m wide and approximately 1 m deep at the experimental observation section. The water table in the Quaternary aquifer is less than 2 m below the streambed, so the stream has direct hydraulical connection with the aquifer. The main types of soils at the oasis and around East and West Juyan Lakes are alluvial soil, saline–alkali soil, and zonal forest meadow soil. The main soil types on both sides of the stream are fixed and semi-fixed aeolian sandy soil and alkali soil. Vegetation along the stream is predominantly desert riparian forest and riparian shrubbery, such as *P. euphratica* forest, *P. euphratica*–*T. chinensis* mixed forest, and *T. chinensis* forest.

METHODS

Field measurements

Precipitation and streamflow data

The precipitation data of Ejina weather station over the period of 2012 to 2014 was obtained from the China Meteorological Data Sharing Service System. The streamflow data of the East River (Langxin Hydrological station) during 2012 to 2014 was obtained from the Management Bureau of the Heihe River basin. The flow velocity and streamflow

was measured by portable flow meter (model: Flowwatch) from May 2013 to December 2014 at the East River observation section (E1) (Figure 1(b)).

Groundwater depth and temperature data

Five observation wells, Well 1 (W1), Well 2 (W2), Well 3 (W3), Well 4 (W4), and Well 5 (W5) have been installed since 2012 in the direction perpendicular to the streamflow of the East River at 10 m, 30 m, 150 m, 300 m and 600 m away from the river bank (Figure 1(b)). Automatic pressure and temperature transducers (HOBO U20-001-04, Onset Computer Corporation, USA) were installed into each of the observation wells to monitor water head and temperature from June 2012 to December 2014. The recording interval was set as 10 minutes for each HOBO transducer. A HOBO pressure transducer was also used to monitor changes in atmospheric pressure outside the wells. All water pressure measurements by HOBO transducers subtracted the recorded atmospheric pressure to obtain the pressure that was exerted only by the water column above the sensors.

Hydrochemistry analyses of water samples

The water samples were collected from the East River, unconfined and confined groundwater near the East River from upstream to downstream, springs as well as Juyan Lake during 10–15 October, 2012. In total, 26 water samples were taken. Laboratory analyses were carried out to determine the salinity of water samples. Milli-Q water was used for cleaning containers and sample-processing equipment. Samples were diluted in closed polyethylene containers and analyzed for cations (Na^+ , K^+ , Ca^{2+} , and Mg^{2+}) and anions (Cl^- , NO_3^- , and SO_4^{2-}) with a Dionex ICS-5000 ion chromatograph. The analyzing columns used for the cations and anions were AS11-HC and CS12A, and protecting columns were AG11-HC and CG12A, respectively. The injected volume was 25 μL according to the concentration of sample. Eluents used for cations and anions were methanesulfonic acid and sodium hydroxide with concentrations of 20 mM and 30 mM, respectively. Blanks were regularly monitored during the sample analysis, and all blanks were found to be lower than detection limits. Analytical

imprecision was calculated (by replicate measurements of standards) to be within 10% of the average concentration levels found in the samples.

Eddy covariance flux data

An eddy covariance system was installed on a 10 m tower above the canopy and has operated continuously since the autumn of 2012 at Alashan Desert Eco-hydrology Experimental Research Station. The wind speed and concentration measurements were measured with CSAT3 Three Dimensional Sonic Anemometer by CR3000 and CR1000 dataloggers. Air temperature and humidity were measured on the towers with Vaisala HMP45C probes. The CO₂ and H₂O were measured with a LI-7500 analyzer. Net radiation was measured with CNR4. The detailed method and process of *ET* calculation is referenced by the researches of Scott (2010), Williams *et al.* (2004), and Wilson *et al.* (2001).

Analysis methods

Groundwater storage changes

The groundwater storage was estimated by WTF method. The WTF method is based on relating changes in measured water-table elevation in groundwater with changes in the amount of water stored in the aquifer (Delin *et al.* 2007):

$$\Delta S = S_y \times \Delta h \quad (1)$$

where ΔS is change in groundwater storage in a defined time interval (e.g., t_0 to t) (mm), S_y is specific yield of the aquifer (dimensionless), and Δh is water level rise in observation wells at a defined time interval (e.g., t_0 to t) (mm). This method assumes that: (1) the observed well hydrograph depicts only natural water-table fluctuations caused by groundwater recharge and discharge; (2) S_y is known and constant over the interval of the water-table fluctuations; and (3) the pre-recharge water-level recession can be extrapolated to determine the Δh .

In this study, the Δh is estimated by a master recession curve approach in observation wells (Heppner & Nimmo 2005). The S_y of soil was determined on the

base of soil mechanical composition analysis at experiment sites by the Mastersizer-2000 laser particle size analyzer. The average percentages of clay, silt, and sand material of soil are 5.37%, 24.17%, and 70.84% in the study area, respectively. The S_y of soil was calculated using the method of Johnson (1967). The estimated S_y is 0.169 and is consistent with the value recommended by the Gansu Bureau of Geology and Mineral Investigation (1990).

Estimation of lateral seepage from stream to groundwater

Stream lateral seepage was estimated at the W5 site with a point scale based on the observational data on the experimental section of riverbank in this study, so the two-dimensional water balance method was used to calculate the stream lateral seepage to groundwater control volume. The schematic diagram of the conceptual model is shown in Figure 2. The water balance formulation of groundwater control volume can be expressed as:

$$P + RF + Q_{in} = ET + PG + Q_{out} + \Delta S \quad (2)$$

where P is precipitation; and RF is irrigation return flow (mm); Q_{in} is stream lateral seepage recharge to the

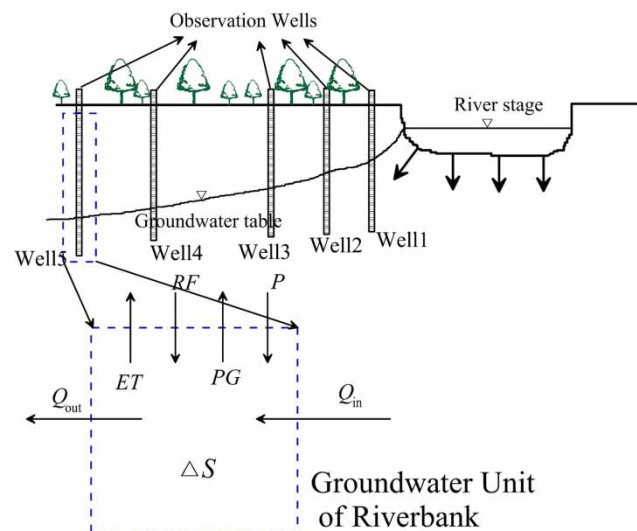


Figure 2 | The water balance framework of groundwater unit at the W5 site on the riverbank.

groundwater unit (mm); ET is evapotranspiration from the groundwater unit (mm); PG is groundwater extraction by pumping (mm); Q_{out} is groundwater lateral discharge from groundwater unit of riverbank to stream (mm); and ΔS is change in groundwater storage (mm).

For this study, there was no irrigation return flow and groundwater extraction, so that RF and PG equal zero; therefore, Equation (2) can be simplified as:

$$Q_{in} - Q_{out} = ET + \Delta S - P \quad (3)$$

If the $Q_{in} - Q_{out}$ equals a positive value it indicates stream lateral seepage to groundwater, while a negative value of $Q_{in} - Q_{out}$ represents groundwater lateral discharges to the stream. Thus, the difference $Q_{in} - Q_{out}$ is defined as Q_{net} , that is:

$$Q_{net} = Q_{in} - Q_{out} \quad (4)$$

where Q_{net} is net lateral seepage (mm).

The formulations (3) and (4) are combined as:

$$Q_{net} = ET + \Delta S - P \quad (5)$$

Other analyzing methods

The Pearson's correlation approach was used to analyze the relationships between groundwater level and stream streamflow. The Kriging interpolation method was used to calculate the spatial and temporal changes in groundwater depth in this study area.

RESULTS

Streamflow variability of East River

The streamflow of the East River is mainly controlled by the water diversion plan of the Heihe River Basin implemented since 2000. The aim of the plan is to ensure a certain flow is allocated into the lower reaches from the upper and middle reaches of the Heihe River under different reliabilities on the basis of annual streamflow. The East River has become ephemeral in past years. There was considerable variability in the streamflow in the period of 2012–2013. A hydrograph of the East River is shown in Figure 3. It can be seen that the streamflow became discontinuous during 2012–2013 with

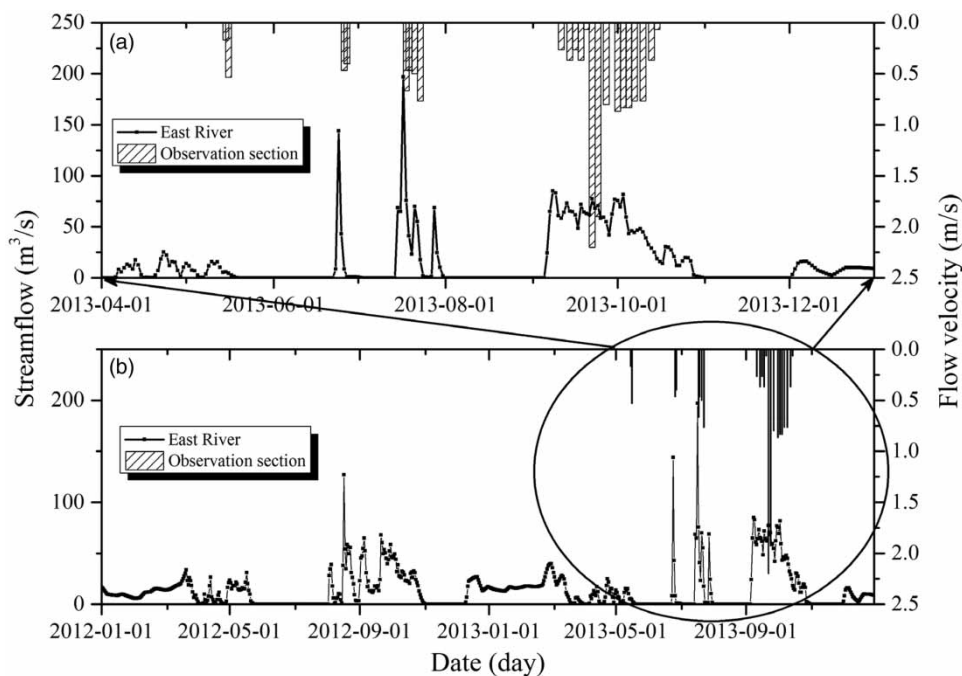


Figure 3 | Average daily streamflow and average flow velocity of East River in the lower reaches of Heihe River basin from January 2012 to December 2013.

six zero-flow periods. These zero-flow events happened in June, July, or November for the East River. The streamflow mainly occurred during the periods of January–May and September–October. During 2012–2013, the average streamflow was $14.2 \text{ m}^3/\text{s}$ and the maximum streamflow appeared on 17 July 2013 with $197 \text{ m}^3/\text{s}$. The coefficient of variation in streamflow is 1.385, indicating a high variability. There was little streamflow during April–August and this coincides with the period of active plant growth.

Groundwater depth variations

The groundwater depths of observation wells presented dramatic fluctuations during June 2012 and October 2014. The amplitude of the change in groundwater depth at W1 was larger than at other observation wells. The highest groundwater level occurred in March, while the lowest occurred in September. As can be seen from Figure 4, large increases in groundwater level occurred at W1 in response to surface water diversions during the period of June 2012 to October 2014, and the maximum increase was more than 1.5 m in September 2014. The amplitudes of increases in groundwater depth at other observation wells were smaller and there were delays compared with W1. The groundwater level at W1 showed greater temporal variation compared with those of the other observation wells. The patterns of groundwater level at different observation wells were similar except for

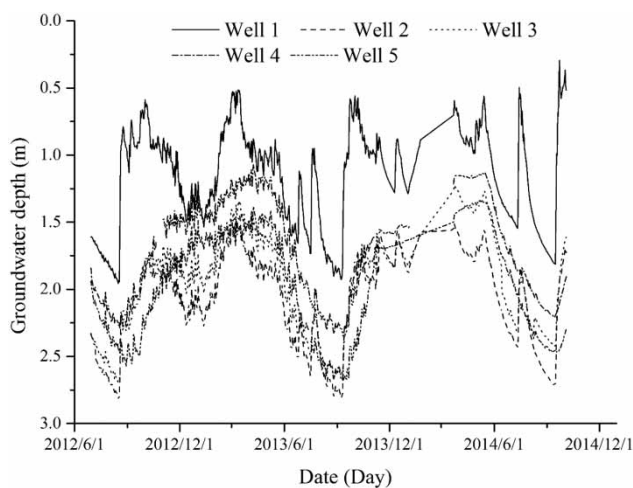


Figure 4 | Daily change of groundwater depth of different observation wells in the observation section.

W1. The mean of the groundwater depth was 1.16 m, 2.06 m, 1.93 m, 1.69 m, and 1.95 m at W1, W2, W3, W4, and W5, respectively. The range (maximum–minimum) of groundwater depth was 1.66 m, 1.37 m, 1.47 m, 1.29 m, and 1.35 m, respectively. The coefficient of variation of groundwater depth was the largest at W1 with a value of 0.32.

There were large differences between the changes of groundwater depth at W1 and those of the other observation wells. Figure 5 shows the scatter plots of groundwater depth at W1 against those of the other observation wells. The results indicate that the correlation coefficient is decreasing with the distance of the observation wells away from the riverbank.

Response of groundwater change to streamflow

The correlation between streamflow and groundwater depth

The observed groundwater depth responded strongly to the streamflow events and the responses were lagged by several hours to days depending on the distance away from the riverbank. As shown in Figure 6, the groundwater level started to rise after a period of time when the streamflow passed the section. The streamflow is larger, the rise of groundwater level is higher. For example, the streamflow of the East River lasted about 92 days from 3 August 2012 to 2 October 2012, the groundwater levels started rising sharply from

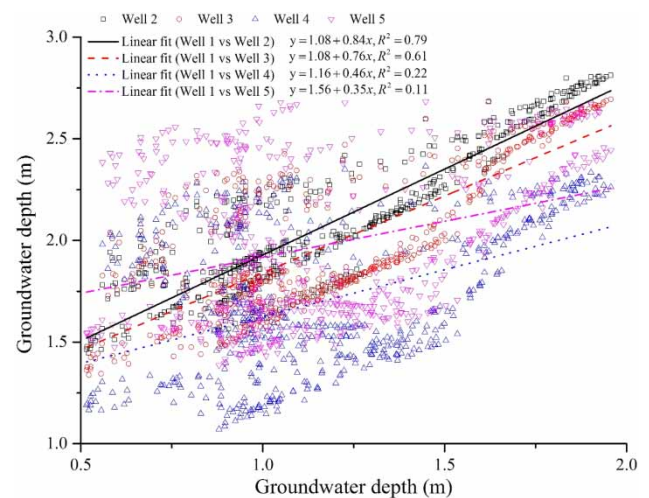


Figure 5 | Scatter plot between groundwater depths at W1 with other observation wells (W2, W3, W4, and W5) in the observation section.

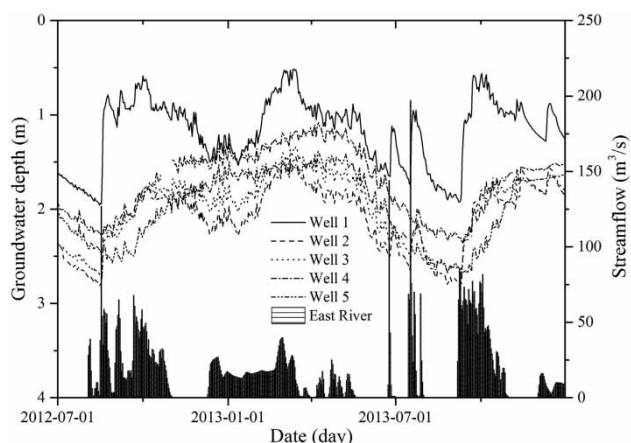


Figure 6 | Comparison between the streamflow of East River and the dynamic change of groundwater depth at observation wells from July 2012 to December 2013.

17 August 2012 at W1 and W2, similarly, the groundwater levels at other wells also began to rise gradually from 18 August 2012. The groundwater levels continued to rise after the streamflow events and then started to fall in all the observation wells.

There was significant negative correlation between streamflow of the East River and groundwater depth at W1 with correlation coefficient -0.26 , p -value <0.001 at 95% confidence level, as shown in Table 1. The groundwater depth at W2 had no significant relationship with the streamflow of East River, with the correlation coefficient -0.03 . Similarly, there was no significant relationship between the groundwater

depth at W3 and streamflow of East River; the correlation coefficient was 0.08. The groundwater depths at W4 and W5 showed significant positive correlations with the streamflow of the East River; the correlation coefficient values were 0.26 and 0.32 with p -value <0.001 , respectively. The correlation coefficient values of groundwater depth and streamflow gradually increased as the distance of the observation wells away from the riverbank increases. However, there were significant positive correlations among the changes of groundwater depth at all observation wells, which indicated that there were consistent or similar changes of groundwater depth in different observation wells. If the lagging effect of groundwater depth changes behind the streamflow of the East River can cause the changes of correlation coefficient between the streamflow of the East River and groundwater depth at different observation wells, then the groundwater depth would have stronger negative correlation with the time-series of streamflow moved backwards. When the streamflow of the East River was moved backwards 6 days by the time-series, there was significant negative correlation between the streamflow of the East River and the groundwater depth at W1 with correlation coefficient -0.473 (p -value <0.001) at 95% confidence level. When the streamflow of the East River was moved backwards 15 days, 33 days, 70 days, and 77 days by the time-series, the adjusted streamflow had maximum significant negative correlations with the groundwater depth at W2, W3, W4, and W5 for the first

Table 1 | Correlations between streamflow of East River and changes of groundwater depth at all observation wells in the observation section

Streamflow		Groundwater depth				
		Well 1	Well 2	Well 3	Well 4	Well 5
East River streamflow	Pearson corr.	-0.26^*	-0.03	0.08	0.26^*	0.32^*
	Sig.	1.32×10^{-9}	0.47	0.07	9.12×10^{-10}	1.62×10^{-14}
East River streamflow (time series was moved backwards 6 days)	Pearson corr.	-0.473^*	-0.216^*	-0.082	0.188^*	0.28^*
	Sig.	0	3.37×10^{-7}	0.055	1.43×10^{-5}	2.83×10^{-11}
East River streamflow (time series was moved backwards 15 days)	Pearson corr.	-0.399^*	-0.229^*	-0.147^*	0.083	0.189^*
	Sig.	0	8.74×10^{-8}	6.21×10^{-4}	0.059	1.06×10^{-5}
East River streamflow (time series was moved backwards 33 days)	Pearson corr.	-0.233^*	-0.206^*	-0.208^*	-0.065	-0.002
	Sig.	8.60×10^{-8}	2.28×10^{-6}	1.76×10^{-6}	0.146	0.971
East River streamflow (time series was moved backwards 70 days)	Pearson corr.	-0.115^*	-0.288^*	-0.241^*	-0.22^*	-0.254^*
	Sig.	0.012	1.31×10^{-10}	8.82×10^{-8}	1.81×10^{-6}	1.64×10^{-8}
East River streamflow (time series was moved backwards 77 days)	Pearson corr.	-0.069	-0.247^*	-0.23^*	-0.22^*	-0.258^*
	Sig.	0.131	4.86×10^{-8}	4.11×10^{-7}	2.10×10^{-6}	1.25×10^{-8}

Two-tailed test of significance used.

*Correlation is significant at 0.05.

time; the correlation coefficient extreme values were -0.23 , -0.21 , -0.22 , and -0.26 , respectively. These show that the lagging effect caused the changes of correlation relationships between groundwater depth and streamflow of the East river.

Groundwater response time to streamflow events

The ranges in groundwater table depths gradually decreased with the distance away from the riverbank. The response time of groundwater table to streamflow events increased with the distance away from the riverbank. As shown in Table 2, the maximum coefficient of variation was 2.16 for the streamflow of the East River. The maximum streamflow occurred in July 2013 with $197 \text{ m}^3/\text{s}$, the maximum and minimum of the average streamflow were $45.18 \text{ m}^3/\text{s}$ and $9.00 \text{ m}^3/\text{s}$ in the periods of 6 September 2013 to 31 October 2013 and 2–31 December 2013, respectively. The longest duration of streamflow event was 161 days from 10 December 2012 to 19 May 2013 and the shortest one was 17 days from 15 July 2013 to 31 July 2013 in the six water conveyances of the East River. For example, during the period from 3 August 2012 to 2 November 2012, the ranges of groundwater depth at W1, W2, W3, W4, and W5 were 1.37 m, 1.05 m, 1.02 m, 0.99 m, and 0.99 m, respectively. The response times in which groundwater depth changed from maximum to minimum were 47 days, 49 days, 80 days, 137 days, and 137 days, respectively. Similarly, during the period 10 December 2012 to 19 May 2013, the ranges of groundwater depth were 1.02 m, 0.84 m, 0.69 m, 0.47 m, and 0.32 m, respectively, for all observation wells according to ascending order in distance to the riverbank of the East River. The response times in which groundwater depth changed from maximum to minimum were 60 days, 62 days, 62 days, 86 days, and 86 days, respectively.

Magnitude of groundwater responses

The amplitudes of the groundwater depth showed stronger correlation with the total streamflow than the maximum and mean streamflow and persistent period of the East River during each water conveyance. The amplitudes of groundwater depth at all observation wells except W5 had significant positive correlations with total streamflow of the East River. The correlation coefficient values between

streamflow of the East River and amplitudes of groundwater depth at W1, W2, W3, and W4 were 0.96, 0.99, 0.97, and 0.93, respectively, as shown in Table 3. These results indicated that streamflow of the East River in all periods significantly influenced the amplitude of groundwater depth within a range of 300 m away from the riverbank. Total streamflow of the East River in all periods also impacted on the amplitude of groundwater depth at W5, but not significantly. There were positive correlation relationships between the mean streamflow of the East River and all observation wells; the correlation coefficient values were 0.56, 0.46, 0.48, 0.33, and 0.25, but not significant. The streamflow persistent period had positive correlations with the amplitudes of groundwater depth at all wells except W5, and similarly, not significant. There were no obvious correlations between the maximum streamflow of the East River and the amplitudes of groundwater depth at all wells. These results indicate that the rising amplitude of groundwater depth has a close relationship with total streamflow of each water conveyance.

Groundwater storage change and stream lateral seepage

The stream lateral seepage fluctuated strongly and the amplitudes of groundwater storage change decreased as the distance away from the riverbank increases. According to the results that were calculated by formulation (1), the trends of groundwater storage change were consistent with the streamflow of the East River, as shown in Figure 7. When the streamflow suddenly increased, the groundwater storage evidently increased at all wells, but magnitude of groundwater storage increase declined as the distance away from the riverbank increases. The changes of groundwater storage at all wells also lagged behind the changes of streamflow. There was a total opposite trend between the groundwater storage and *ET* in the study area (Figure 7), that is, when the *ET* was relatively larger, the groundwater storage would be relatively smaller for all observation wells. It also confirms the observed results that the East River dried up when the desert riparian vegetation grew from April to August, so that the vegetation growth only relied on the groundwater, leading to a sinking groundwater table. There was no significant relationship between the precipitation and the groundwater storage.

Table 2 | Characteristics of streamflow and the response of groundwater change to the streamflow of East River during each period of water transfer from August 2012 to December 2013

Item	Period of water transfer					
	2012.8.3– 2012.11.2	2012.12.10– 2013.5.19	2013.6.23– 2013.7.14	2013.7.15– 2013.7.31	2013.9.6– 2013.10.31	2013.12.2– 2013.12.31
East River						
Maximum streamflow (m ³ /s)	127	39.8	144	197	85.1	16.1
Mean streamflow (m ³ /s)	28.64	14.29	20.89	42.61	45.18	9
Streamflow lasting time (day)	92	161	10	17	56	30
Total streamflow (10 ⁸ m ³)	2.28	1.99	0.18	0.63	2.19	0.23
Coefficient of variation of streamflow	0.73	0.64	2.16	1.15	0.54	0.43
Well 1						
Maximum groundwater depth (m)	1.96	1.54	1.66	1.73	1.93	1.28
Date of maximum	2012.8.15	2013.1.12	2013.6.25	2013.7.16	2013.9.8	2013.12.10
Minimum groundwater depth (m)	0.59	0.52	1.12	0.95	0.56	0.88
Date of minimum	2012.10.1	2013.3.13	2013.6.27	2013.7.23	2013.10.2	2013.12.15
Range of groundwater depth (m)	1.37	1.02	0.54	0.78	1.37	0.4
Response time (min–max) (day)	47	60	2	7	24	5
Well 2						
Maximum groundwater depth (m)	2.81	2.28	2.52	2.64	2.81	1.84
Date of maximum	2012.8.15	2013.1.12	2013.6.25	2013.7.16	2013.9.8	2013.12.10
Minimum groundwater depth (m)	1.76	1.45	2.24	2.14	1.71	1.56
Date of minimum	2012.10.3	2013.3.15	2013.6.30	2013.7.24	2013.10.7	2013.12.15
Range of groundwater depth (m)	1.05	0.84	0.28	0.5	1.1	0.28
Response time (min–max) (day)	49	62	5	8	29	5
Well 3						
Maximum groundwater depth (m)	2.69	2.03	2.34	2.51	2.68	1.85
Date of maximum	2012.8.15	2013.1.12	2013.6.25	2013.7.16	2013.9.10	2013.12.10
Minimum groundwater depth (m)	1.67	1.34	2.16	2.11	1.66	1.63
Date of minimum	2012.11.3	2013.3.15	2013.6.30	2013.7.24	2013.10.14	2013.12.16
Range of groundwater depth (m)	1.02	0.69	0.18	0.41	1.02	0.21
Response time (min–max) (day)	80	62	5	8	34	6
Well 4						
Maximum groundwater depth (m)	2.29	1.54		2.14	2.36	1.58
Date of maximum	2012.8.18	2013.1.12		2013.7.22	2013.9.11	2013.12.10
Minimum groundwater depth (m)	1.3	1.07		1.99	1.61	1.52
Date of minimum	2013.1.2	2013.4.8		2013.7.24	2013.11.2	2013.12.21
Range of groundwater depth (m)	0.99	0.47		0.16	0.76	0.07
Response time (min–max) (day)	137	86		2 ^a	52	11
Well 5						
Maximum groundwater depth (m)	2.47	1.71		2.47	2.68	
Date of maximum	2012.8.18	2013.1.12		2013.7.22	2013.9.11	
Minimum groundwater depth (m)	1.49	1.39		2.35	1.57	
Date of minimum	2013.1.2	2013.4.8		2013.7.24	2013.11.9	
Range of groundwater depth (m)	0.99	0.32		0.12	1.11	
Response time (min–max) (day)	137	86		2 ^a	59	

^aRepresents the special value in the results calculated.

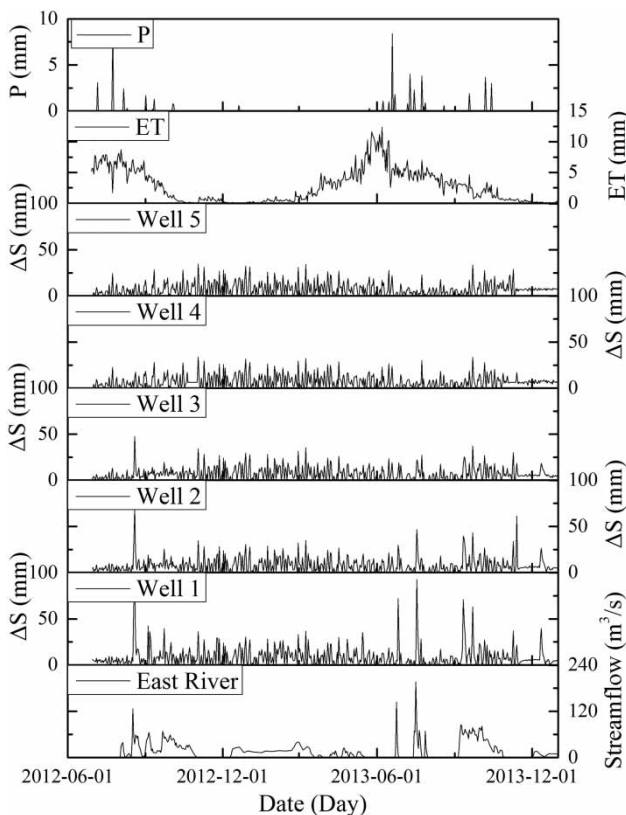
Table 3 | Correlation relationships between the amplitude of groundwater depth and the characteristics of streamflow of East River during each period of water transfer from August 2012 to December 2013

Amplitude of groundwater depth		Maximum streamflow	Mean streamflow	Streamflow persistent period	Total streamflow
Well 1	Pearson Corr.	0.1	0.56	0.54	0.96*
	Sig.	0.85	0.25	0.27	0.003
Well 2	Pearson Corr.	-0.06	0.46	0.62	0.99*
	Sig.	0.92	0.35	0.19	3.35×10^{-4}
Well 3	Pearson Corr.	-0.04	0.48	0.57	0.97*
	Sig.	0.95	0.34	0.24	0.001
Well 4	Pearson Corr.	0.13	0.33	0.44	0.93*
	Sig.	0.84	0.59	0.46	0.02
Well 5	Pearson Corr.	-0.29	0.25	-0.003	0.79
	Sig.	0.72	0.75	0.997	0.21

Two-tailed test of significance used.

*Correlation is significant at 0.05.

The results of groundwater storage estimated during different periods at all observation wells are shown in Table 4. The results calculated by Equation (5) indicated

**Figure 7** | Groundwater storage changes estimated at all observation wells based on a WTF method and comparison with streamflow of East River, *ET* and precipitation in the observation section from June 2012 to December 2013.

that the mean and total groundwater storage became smaller with the increased distance of observation wells away from the riverbank. For example, during the period 3 August 2012 to 2 November 2012, the mean groundwater storage was 9.99 mm at W1, and it became smaller, from 8.26 mm to 6.55 mm, at W2 and W5. The mean groundwater storage was maximum with 12.11 mm at W1 during the period of 15–31 July 2013. The maximum and minimum of daily *ET* occurred in the periods 23 June–14 July 2013 and 2–31 December 2013, respectively, with 5.07 mm and 0.12 mm. There was little precipitation during the periods of 10 December 2012 to 19 May 2013 and 2–31 December 2013.

Stream lateral seepage (Q_{net}) can lead to increases in groundwater level and *ET* of riparian forest. The stream lateral seepage was mainly controlled by the streamflow events and their durations. The results suggest that the stream lateral seepage contributed mostly to increased groundwater storage (i.e., increased ΔS) in winter and spring. However, during summer and autumn, the stream lateral seepage is mainly consumed by riparian forest as *ET*. For example, at the experimental site of observation W5, during the period of water conveyance from 10 December 2012 to 19 May 2013, ΔS accounted for 82.6% of the stream lateral seepage, the *ET* of riparian zone accounted for 17.4% of the stream lateral seepage. During the period of water conveyance from 23 June to 14 July 2013, the ΔS and *ET* were 43.3% and 59.2% of the stream lateral seepage, respectively.

Table 4 | Calculation results of groundwater storage change ΔS at all observation wells and stream lateral seepage Q_{net} and ET at W5 during different periods of water transfer from June 2012 to December 2013

Item	Period of water transfer					
	2012.8.3–2012.11.2	2012.12.10–2013.5.19	2013.6.23–2013.7.14	2013.7.15–2013.7.31	2013.9.6–2013.10.31	2013.12.2–2013.12.31
Well 1						
Mean ΔS (mm/day)	9.99	9.3	7.15	12.11	11.72	7.15
Total ΔS (mm)	919.11	1,498	157.3	205.94	656.06	214.39
Well 2						
Mean ΔS (mm/day)	8.26	8.55	5.65	8.74	9.9	7.03
Total ΔS (mm)	760.28	1,376.66	124.32	148.5	554.39	211.04
Well 3						
Mean ΔS (mm/day)	7.03	8.07	4.2	6.89	8.73	6.39
Total ΔS (mm)	646.38	1,299.28	92.41	117.07	488.77	191.66
Well 4						
Mean ΔS (mm/day)	6.97	8.53	4.13	4.21	7.72	6.87
Total ΔS (mm)	641.03	1,373.44	90.92	71.64	432.34	206.03
Well 5						
Mean ΔS (mm/day)	6.55	8.46	3.71	3.58	7.54	7.16
Total ΔS (mm)	602.26	1,362.64	81.56	60.86	422.12	214.86
Mean ET (mm/day)	2.87	1.78	5.07	4.88	1.82	0.12
Total ET (mm)	264.34	286.95	111.55	82.88	101.87	3.61
Mean P (mm/day)	0.08	0	0.22	0.45	0.15	0
Total P (mm)	7.4	0.6	4.8	7.6	8.6	0
Mean Q_{net} (mm/day)	9.34	10.24	8.56	8.01	9.2	7.28
Total Q_{net} (mm)	859.21	1,649	188.31	136.14	515.39	218.47

Hydrochemical characteristics of different water bodies

As shown in Figure 8, hydrochemical characteristics of different water bodies vary considerably. The differences reflected that the different water bodies were influenced under different hydrological processes or by different water–rock interactions. The sampling points that included unconfined water and river water are concentrated in the solid circle. The main hydrochemical types of unconfined water were Na-SO_4 , Mg-SO_4 , and Mg-HCO_3 , and the hydrochemical types of river water Mg-SO_4 , Mg-HCO_3 , and Na-SO_4 . The unconfined water samples were more similar to the river water samples than the other types of samples, especially the types of cations in this study area. The HCO_3^- percentages of unconfined water samples were lower than that of river water samples. Conversely, the SO_4^{2-} percentages of unconfined water samples were

higher than that of river water samples. The total dissolved solids (TDS) of unconfined water samples was higher than that of river water samples. This could be due to the concentration caused by transpiration of groundwater in an unconfined aquifer after river water recharging unconfined water. That is to say, the unconfined groundwater has a close connection with river water in this study.

DISCUSSION

Method of recharge between groundwater and stream as well as the response of groundwater changes to different water conveyance

There are mainly four types of relationships between groundwater and stream, including connected gaining

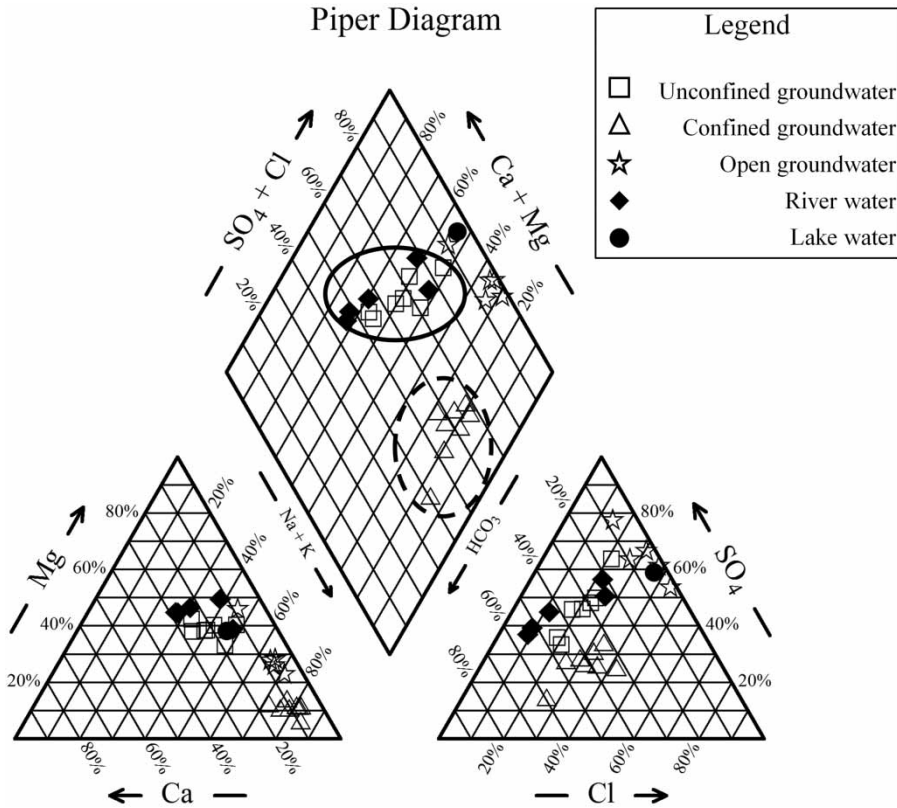


Figure 8 | Piper diagram of hydrochemistry composition in different water bodies, including river (RIV), unconfined water (UNC), confined water (CON), exposed groundwater (L), and lake (LAK) samples.

stream, connected losing stream, disconnected stream with shallow water table, disconnected stream with deep water table (Osman & Bruen 2002). In this study, the water table was below the stream water level and water flows from the stream into the aquifer, and the unconfined groundwater table was shallow, so it was connected to the losing stream, i.e., the groundwater was recharged by the stream. The groundwater change was significantly influenced by the streamflow of the East River, which was reflected in the amplitude of groundwater change corresponding to the streamflow changes and the number of days that groundwater responds, as shown in Figures 9 and 10. The amplitude of groundwater depth decreased with the distance away from the riverbank, and the response time of groundwater depth change to streamflow change increased with the distance away from the riverbank. These results indicate that the rate of stream lateral seepage to groundwater gradually became weaker as the distance from the riverbank increases.

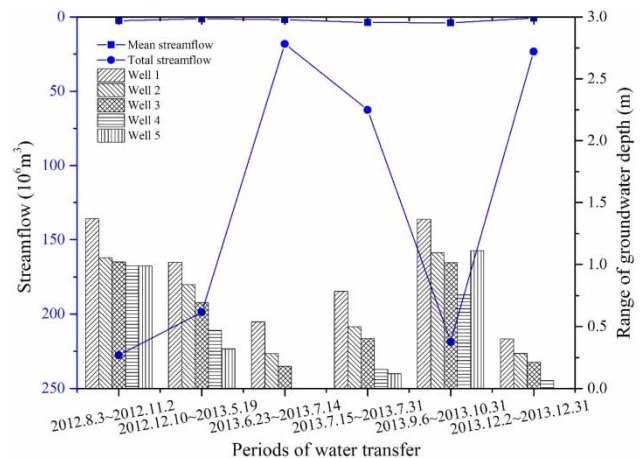


Figure 9 | Response of amplitudes of groundwater depth at different observation wells to streamflow of East River during different periods of water conveyance.

Influencing range of stream on groundwater

The stream plays an important role in recharging groundwater near the riverbank. The closer to the riverbank, the

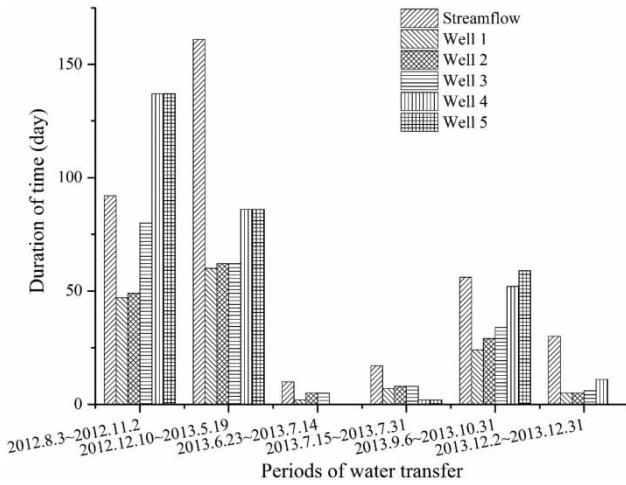


Figure 10 | Duration of streamflow and response duration of groundwater depth at different observation wells to streamflow change of East River during different periods of water conveyance.

more frequently the groundwater depth changes the in study area. As shown in Figure 11, the density of groundwater depth contours reflects the degree of groundwater change. This means that the groundwater changes sharply closer to the riverbank, changes steadily between 50 m and 300 m away from the riverbank, and changes slowly in the area where distance to the riverbank is larger than 300 m. The slope of groundwater contours also illustrated that change rate of groundwater depth. The groundwater

contours were skewed to the right, indicating that there was the lag effect of stream lateral seepage to groundwater. Similarly, the change of groundwater temperature also explained the problem with stream lateral seepage to groundwater, in Figure 12. The stream temperature changes more intensively than the groundwater temperature, because the stream temperature is changing with the air temperature. The change of groundwater temperature is similar to the stream temperature near the riverbank; it is quite intense in the position where the distance is less than 50 m to the riverbank. When the distance is larger than 300 m away from the riverbank, the change of groundwater temperature is fairly stable. This illustrates that the influence range of stream temperature change on groundwater temperature was less than the influence range of stream on groundwater fluctuation in this study section. Constantz et al. (1994) presented the influence of diurnal variations in temperature on streamflow loss and groundwater recharge. Anzai et al. (2014) have researched the influence of seepage from canals and paddy fields on the groundwater level in the Ili River basin, Kazakhstan, and reported that the groundwater depth was influenced for up to 300 and 400 m from the canals and paddy fields, respectively. Luan & Deng (2013) discussed the influences of tidal processes and river runoff on the shallow groundwater

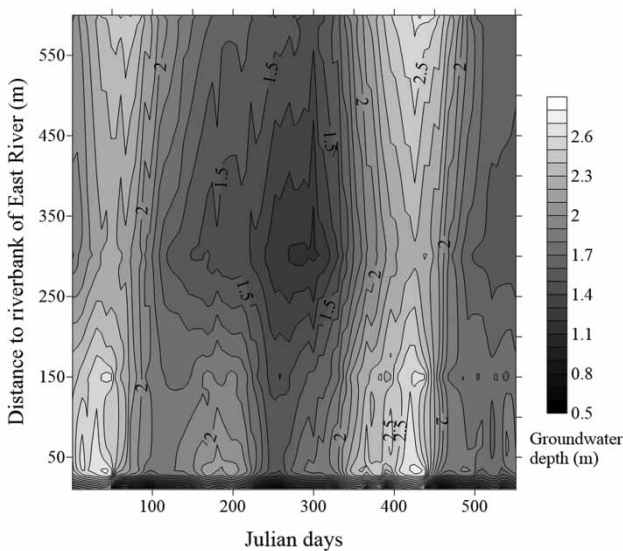


Figure 11 | Change of groundwater depth with the streamflow of East River based on Kriging spatial interpolation method in the observation section from 29 June 2012 to 31 December 2013.

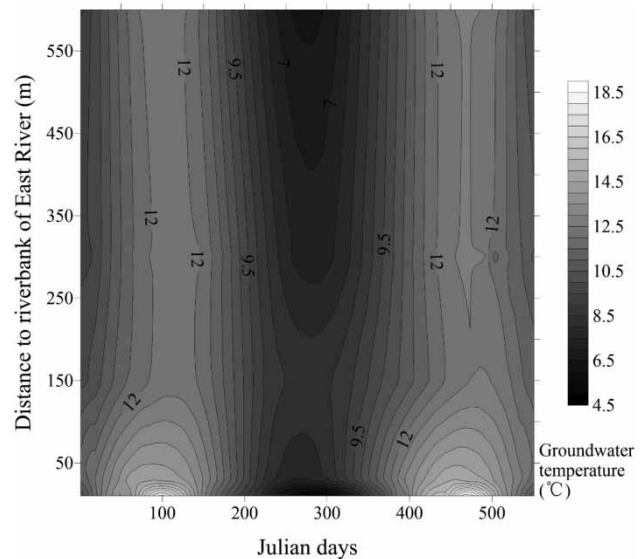


Figure 12 | Change of groundwater temperature with the streamflow of East River based on Kriging spatial interpolation method in the observation section from 29 June 2012 to 31 December 2013.

dynamic in coastal wetlands of the Yellow River Delta, arriving at the influence distances of 14,700 m and 11,600 m, respectively. Mallard *et al.* (2014) quantitatively studied the interaction between stream and groundwater, and produced the hydrologic gain and loss as well as influence range using tracer experiments. One can infer that the influence range for the losing stream on groundwater or aquifer will be different depending on streamflow and its duration for a specific region. Meanwhile, the distribution of vegetation is able to illustrate the long-term relationship between stream and groundwater in the riparian area. In an arid region, the riparian vegetation strongly depends on stream to grow. For instance, in this study section, *Tamarix chinensis* was scattered around the riparian area, but the forest of *Populus euphratica* mainly was distributed between 150 m and 600 m away from the river, as shown in Figure 1(b). When the distance was large than 600 m from the riverbank, there were small amounts of low shrubs dotted around the area, mainly including *Tamarix chinensis*, *Nitraria tangutorum*, and *Lycium ruthenicum*. This proved that the distribution of vegetation was able to reflect the range of more active interaction between stream and groundwater indirectly. Baattrup-Pedersen *et al.* (2013) discussed the effects of stream flooding on the distribution and diversity of groundwater-dependent vegetation in riparian areas. Stromberg *et al.* (2005) studied the effects of streamflow intermittency on riparian vegetation of a semiarid region river, thus explaining how the riparian vegetation reflected streamflow.

CONCLUSIONS

This study analyzed the relationship between stream and groundwater. We explored the dynamic changes of groundwater and temperature under water conveyance, obtained the correlations between streamflow and groundwater depth, and revealed the response of groundwater change to streamflow, and estimated the stream lateral seepage and change of groundwater storage. The following conclusions were obtained:

1. The groundwater depth of observation wells presented dramatic fluctuations during June 2012 to October 2014 by moving up sharply and falling slowly.
2. The correlation relationships between streamflow of the East River and groundwater depth at different observation wells are different.
3. The amplitudes of groundwater depth gradually decreased as the distance away from the riverbank increases.
4. The amplitudes of groundwater depth at observation wells showed stronger correlations with total streamflow volume.
5. The stream lateral seepage was mainly controlled by the streamflow and its durations. The evapotranspiration of riparian forest used larger proportions in total stream lateral seepage for the water conveyance, especially in summer.
6. The groundwater changes sharply closer to the riverbank, changes steadily between 50 m and 300 m away from the riverbank, and changes slowly in the region further away from the riverbank.

Although the stream lateral seepage has been estimated quantitatively and the response of groundwater to the streamflow has been discussed in this study, we are not able to quantify the uncertainty in seepage estimation. In the future, we hope to use a simulation model method to analyze the interactive process between stream and groundwater in this study area, so that the quantitative relation between stream and groundwater can be uncovered.

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