Damming Weakens the Groundwater and Surface Water Exchange and Its Nitrate Removal in Semi-arid Prairie Rivers

Jing Huang

China Agricultural University

Xixi Lu

National University of Singapore

Xin Luo

Hong Kong University: University of Hong Kong

Haizhu Hu (✉ huhaizhu@imu.edu.cn)

Inner Mongolia University <https://orcid.org/0000-0002-6029-8465>

Jiu Jimmy Jiao

Hong Kong University: University of Hong Kong

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Abstract

Surface water (SW)-Groundwater (GW) exchange plays a vital role in a prairie aquatic system and the biogeochemical cycling in such a system. Considering the inadequate understanding of damming on SW-GW exchange, a damming prairie river in Southeast Eurasian steppe was chosen to investigate variations of the SW-GW exchange and its influences on the fate of nitrate (NO_3^-). Both hydraulic and hydrochemical methods were applied to precisely depict the daily and seasonal exchange processes. The upstream and downstream reaches of the dam were observed to be upwelling and downwelling conditions respectively within a hydrologic year. Results obtained from multiple tracer methods and hydraulic method indicate that damming contributed to transfer the stream from the upwelling to the downwelling condition and weaken the SW-GW exchange in the downstream. The patterns of SW-GW exchange modulated the NO_3^- uptake or production between the SW and the GW. NO_3^- was mainly removed in the SW-GW exchange zone (SW-GW EZ) of the upwelling segment, while produced in the downwelling segment. Both the removal and production of NO_3^- were enhanced during snowmelt period, which might be an active period for the SW-GW exchange and NO_3^- fate. This study underscores the negative effect of damming on the SW-GW exchange and accompanied NO_3^- removal in prairie river systems.

1. Introduction

It is increasingly recognized that surface water (SW) and groundwater (GW) are not isolated components in the hydrologic system, and there have been developments in the studies of SW-GW exchange in the past decades (Winter 1995, Wang et al. 2018, Toran 2019). The SW-GW exchange zone (SW-GW EZ), a critical ecological interaction area, has gained considerable research interest from both hydrogeologist and hydrologists (McClain et al. 2003, Lewandowski et al. 2011, Marmonier et al. 2012). The SW-GW EZ is recognized as a key unit controlling the fate of nutrients within the fluvial networks and a critical regulator of biogeochemical cycling, due to the complex reactions in the zone (Bardini et al. 2012, Harvey et al. 2013). Nitrogen pollution is globally prevalent, and the natural nitrogen cycling has been seriously disturbed by human activities (Galloway et al. 2002, Fields 2004). Increases of nitrogen inputs have modulated the processes of nitrogen transport in the SW-GW EZ (Zarnetske et al. 2011). Denitrification in SW-GW EZ is widely regarded as a major mechanism of nitrate removal in river systems but highly variable (Grimaldi and Chaplot 2000). Biogeochemical reactions associated with nitrate production and removal can be strongly controlled by SW-GW exchange (Naranjo, Niswonger and Davis 2015, Lewandowski et al. 2011, Wallace and Soltanian 2021).

SW-GW exchange usually involves two processes: SW infiltration into the GW and GW exfiltration back to the SW. When the SW infiltration overwhelms the GW exfiltration, the stream is defined as a downwelling condition, and vice versa as an upwelling condition (Chen et al. 2013). In downwelling streams, abundant organic matters and dissolved oxygen (DO) are taken into river sediments which may favor the aerobic reactions. In upwelling streams, available nitrogen and reductive materials are brought into the sediments by upwelling GW and thus, the redox environment is modified (Gomez-Velez, Krause and Wilson 2014,

Hogsett and Goel 2013). Heterogeneity in alluvial hydrogeologic properties can lead to the variation of SW-GW exchange and nutrient retention (Ward, Schmadel and Wondzell 2018, Gomez, Wilson and Cardenas 2012). Although the linkages between SW-GW EZ hydrology and nitrogen biogeochemistry have been recognized (Harvey and Gooseff 2015), there have been limited studies investigating how the SW-GW EZ influences nitrogen dynamics at reach scales with both upwelling and downwelling segments.

Accurate quantification of SW-GW flow exchange has always been a challenge, especially in arid/semi-arid regions showing large temporal and spatial variations in SW-GW exchange. The hydraulic method based on field survey has been normally used for investigating the SW-GW exchange (Banks et al. 2011, Banzhaf, Krein and Scheytt 2011, Zhu et al. 2020). The physical properties and chemical compositions of natural water record the history of water flow or migration to some extent (Winter 1995, Chitsazan et al. 2019). Thus, tracer-based approaches such as environmental solute (electrical conductivity (EC) and stable isotopes) and temperature tracing are widely employed in SW-GW exchange (King, Raiber and Cox 2014, Martinez, Raiber and Cox 2015, Irawan et al. 2015). Simulation of SW-GW interaction has also been advanced significantly in recent years. However, due to the demand of high accuracy and stability, numerical models still face the challenges (Fleckenstein et al. 2010). Thus, a combination of both hydraulic method and natural tracer method is more reliable in exploring the SW-GW exchanges in upwelling/downwelling streams under various climates (Kong et al. 2019).

In this study, an inland river in the east of Eurasian steppes, the Xilin River was chosen as an example to investigate the variation of SW-GW exchange and its impact on the NO_3^- fate in prairie rivers. The study went as follows: (1) quantification of daily and seasonal SW-GW exchange in the prairie river, (2) comparison of the patterns between SW-GW exchange in both upstream and downstream reaches of the Xilin River dam, and (3) revealing the NO_3^- production and removal mediated by the SW-GW exchange in the Xilin River.

2. Materials And Methods

2.1. Regional settings

The Xilin River, located in the semiarid regions of north China, is a typical inland meandering river in the grassland of Inner Mongolia (Figure 1(a)). The main stream is over 174 km long and the basin covers a total area of $1.1 \times 10^4 \text{ km}^2$. This region is characterized to be semiarid continental temperature climate, with a mean annual temperature of 2.8 °C. Winter is cold and frozen. Based on the records from 1968 to 2015 at the Xilinhote meteorological station, the mean annual precipitation is 286 mm, most of which occur during the wet season from July to September. The mean annual evaporation of 1862.9 mm with slowly declining rate of 0.0191 mm per year (Yang, Qian and Hao 2015). The spring snowmelt period starts at the end of March with duration of about 20 days. About 34.2 % of the annual runoff is derived from the spring snowmelt (Figure 2(a)).

The region is dominantly covered by grassland, with the main soil type to be castanozems. Quaternary loose sediments with a thickness up to 100 m are widely distributed in alluvial plain areas of the Xilin River valley and gully depression between hills. Cenozonic basalts are distributed in the southern area of the Xilinhot city. The phreatic aquifer has been developed within Quaternary loose sediments with thickness of 30-60 m. The flowing yield in the upper phreatic aquifer is $500-1000 \text{ m}^3 \text{ d}^{-1}$, while the flowing yield of lower semi-confined aquifer is $1000-3000 \text{ m}^3 \text{ d}^{-1}$ (Cui 2017). Along the streamflow, the aquifer increases in thickness, the groundwater flow rates are increasing, and the water quality is deteriorating. The water depth of the shallow unconfined aquifer on both sides of the river channel is generally about 5 m and declines to about 1 m on the flood plains (Xu 2017). The Xilin River reservoir is built at the middle stream of the Xilin River. The contributing area of the reservoir is 3852 km^2 and the dam affects natural surface runoff. The river discharge in the downstream of the dam is basically equally to the reservoir seepage rate (Figure 1(c)), only about $1500 \text{ m}^3 \text{ d}^{-1}$ (3.16% of the mean annual streamflow $46397 \text{ m}^3 \text{ d}^{-1}$).

2.2 Monitoring well settings

The monitoring wells were installed in June 2016 at the upstream (site A) and downstream (site B) of the reservoir, respectively (Figures 1(a) and 1(b)). Site A is located at 9 km upstream ($43^\circ 49' 13.5'' \text{N}$, $116^\circ 9' 23.9'' \text{E}$), and site B is located at 5 km downstream ($43^\circ 51' 46.7'' \text{N}$, $116^\circ 4' 51.3'' \text{E}$) of the reservoir. For each site, three monitoring wells (inner diameter or ID: 50 mm) were installed to a depth of 1 m beneath the riverbed surface along the streamflow. The in-stream monitoring wells were named as HS-RC1, HS-RC2 and HS-RC3 for site A, and RB-RC1, RB-RC2 and RB-RC3 for site B. Two upland monitoring wells, namely HS-RB1 and HS-RB2 for site A, and RB-RB1 and RB-RB2 for site B, were also installed perpendicular to the stream (Figures 3(a) and 3(b)). Upland monitoring wells were created down to about 2 m beneath the ground surface. HS-RB1 and HS-RB2 were 2.9 m and 11 m away from the channel along the topographic gradient, while RB-RB1 and RB-RB2 were 9 m and 10.9 m away from the channel.

2.3 Field sampling and laboratory analysis

Monthly samplings and measurements were conducted from July 2016 to June 2017 in the non-frozen seasons (frozen seasons: about six months from October 2016 to March 2017 when the stream was not flowing). Daily samplings were conducted during the spring snowmelt period with the largest streamflow over a hydrologic year from April 6th to April 25th in 2017. GW depth and SW level were manually measured and calibrated to the local water table based on the elevation obtained from electronic distance measuring device. The hydraulic conductivity ($K=0.0003 \text{ m s}^{-1}$) of the sediments was determined by slug test. For each sampling activity, water samples were collected from each monitoring wells and stream water at each site. Each sample was filtered through a $0.45\text{-}\mu\text{m}$ filter and stored in pre-cleaned 100-mL bottle. Besides, in situ measurements of temperature, pH, electrical conductivity (EC), and DO were conducted with portable multi-parameter water quality probes (Multi3420, Mettler Toledo SevenGo).

The major ions of K^+ , Ca^{2+} , Na^+ , Mg^{2+} , SO_4^{2-} , Cl^- , and NO_3^- were analyzed with ion chromatography (Metrohm 881Compact IC Pro-03030615). As the Xilin River water is subject to sub-alkaline water, the

concentrations of CO_3^{2-} were rather negligible compared to other major ions (Chen 1987). Thus, HCO_3^- concentrations were estimated from the charge in-balance between the major ions and anions (Murakami et al. 2007). During the installation of the monitoring wells, soil samples in boreholes were collected in the order of depth and at an interval of 10 cm. Each soil sample was stored in a plastic bag and brought back to the laboratory for soil grain analysis by the laser grain-size analyzer (MicrotracS3500). However, there was a lack of measurement at site B as the sandy soils were quite difficult to collect.

2.4 Hydraulic approaches, water and solute mass balance models

2.4.1. SW-GW exchange flux

SW-GW exchange ($Q_{\text{SW-GW}}$: $\text{m}^3 \text{ s}^{-1}$) were estimated based on Darcy's law using the following equation:

$$Q_{\text{SW-GW}} = K \left(\frac{\Delta h}{\Delta l} \right) A$$

where K is the hydraulic conductivity of the riverbed sediments, and obtained to be 0.0003 m s^{-1} by slug test both at site A and site B. This value is further cross checked according to tabulation based on the grain analysis (Table 1) (Sun 2006), Δh is the head difference between SW and GW level (m), Δl is the length of the seepage path (m); A is the cross-sectional area (m^2).

2.4.2. NO_3^- flux

The net removal or loading rate of NO_3^- for river water in each segment was calculated with:

$$F_{\text{RW}}(\text{NO}_3^-) = Q_{\text{dnRW}} C_{\text{dnRW}}(\text{NO}_3^-) - Q_{\text{upRW}} C_{\text{upRW}}(\text{NO}_3^-)$$

where Q_{dnRW} ($\text{m}^3 \text{ s}^{-1}$) and Q_{upRW} ($\text{m}^3 \text{ s}^{-1}$) are the downstream and upstream streamflow of each river segment, calculated based on the flow velocity, the depth and the width of the river. C_{dnRW} (mg L^{-1}) and C_{upRW} (mg L^{-1}) are the NO_3^- concentrations in the upstream and downstream river water of each segment.

The NO_3^- load derived from SW flow and GW flow were calculated by the following equations:

$$F_{\text{SW}}(\text{NO}_3^-) = Q_{\text{SW}} C_{\text{SW}}(\text{NO}_3^-)$$

$$F_{\text{GW}}(\text{NO}_3^-) = Q_{\text{GW}} C_{\text{GW}}(\text{NO}_3^-)$$

where $F_{\text{SW}}(\text{NO}_3^-)$, $F_{\text{GW}}(\text{NO}_3^-)$ are the NO_3^- fluxes derived from SW, GW flow. Q_{SW} ($\text{m}^3 \text{ s}^{-1}$) and Q_{GW} ($\text{m}^3 \text{ s}^{-1}$) are the SW and GW flow. $C_{\text{SW}}(\text{NO}_3^-)$ and $C_{\text{GW}}(\text{NO}_3^-)$ (mg L^{-1})

3. Results

3.1 Hydraulic dynamics of SW-GW exchange

Figure 4 shows the water level differences between stream water and groundwater in monitoring wells. The water level differences (WLDs) were recorded in both seasonal and daily time scales during the non-frozen seasons over a hydrologic year. Over the seasonal scale, the negative values of WLDs (e.g., RC1: -21.2 to -25.4 cm; RC2: -11.9 to -26.4 cm; RC3: -11 to -25.8 cm) between SW and GW indicated the upwelling condition at site A (Figure 4(a)). The WLDs in the dry season were lower than the wet season. At site B, the WLDs were predominantly positive (e.g., RC1: approaching zero; RC2: 38 to 41 cm in the wet season, close to zero in the dry season; RC3: 0 to 85 cm), suggesting the downwelling condition of site B (Figure 4(b)). Over the daily scale during the spring snowmelt period, the WLDs at site A were also negative (e.g., RC1: -5.4 to -24.9 cm; RC2: -7 to -18 cm; RC3: -1.7 to 13.4 cm), suggesting the upwelling conditions (Figure 4(a)). The upwelling condition dominated in the whole hydrological year. For site B, the WLDs were mainly positive showing a downwelling condition, which is consistent with other seasons. However, upwelling conditions were also observed at the beginning stage of the spring snowmelt period (Figure 4b, RC1: -19.5 to 21.5 cm; RC2: -23.9 to 3 cm; RC3: -40 to 85.6 cm).

3.2 Hydrochemical characteristics SW and GW

3.2.1. Electrical Conductivity (EC)

Figures 5(a) and 5(b) showed variations of EC in the SW and GW during the non-frozen seasons. Generally, the EC at site A was lower than site B, and the EC in the SW was significantly lower than the GW for both sites. The seasonal variation trend of EC was similar between sites A and B. The lowest EC of both sites was shown in August 2016, because the intensive rainfall may have a dilution effect (Figure 3(b)). Except for August, the EC values were higher in wet seasons than dry seasons at site A. However, the dry seasons had higher EC values at site B.

Over daily scale in the spring snowmelt period, the EC at site A was significantly lower than site B for both SW and GW, as a low flow rate downstream the reservoir (site B) prolonged the retention time and thus accumulated salts. EC in surface water was maximum at the very beginning of snowmelt and then decreased to a stable level, because accumulated salt and sand from the winter was flushed into the river at first (Figures 5(a) & 5(b)). The EC was stable in both SW and GW at site A, suggesting the chemostatic characteristic of EC (Godsey, Kirchner and Clow 2009, de Barros et al. 2020). At site B, the EC in the SW also performed a stable value over time. However, the EC in the GW increased at first, reached the peak of $3000 \mu\text{S cm}^{-1}$ and then decreased to around $1700 \mu\text{S cm}^{-1}$. This pattern might be mainly ascribed to the change of the SW-GW exchange. At the beginning of the snowmelt season, the EC value in groundwater was relatively high caused by increased solids and salts, and the upwelling groundwater with high EC was brought into river water from April 6th to April 10th (Figure 4(b)). After April 10th, the downwelling river water with relatively low EC was gradually recharged into groundwater and thus leading to a decrease

3.2.2. pH

Figures 5(c) and 5(d) showed seasonal variations of pH from July 2016 to June 2017. At site A, SW pH ranged from 7.45 to 8.09, and GW pH ranged from 6.22 to 8.35. At site B, SW pH ranged from 7.12 to 8.49, and GW pH ranged from 6.76 to 8.33. Generally, higher pH was observed during the wet seasons than the dry seasons, and the lowest pH was recorded in April at two sites. Figures 5(c) and 5(d) showed daily variations of pH during the spring snowmelt month of April 2017. GW pH was higher at site B than at site A. SW pH was steady over time at site A, which confirmed that there was no significant process of acid and alkali production. There was a slight decreasing trend for GW pH at both sites A and B. The GW of site A changed from alkaline to acidic condition since April 2017. The GW of site B showed acidic condition.

3.2.3. Dissolved Oxygen (DO)

Seasonal dynamics of DO in 2016 and 2017 were shown in Figures 5(e) and 5(f). At site A, the DO concentration ranged from 6.27 to 10.98 mg L⁻¹ in SW and from 3.91 to 7.30 mg L⁻¹ in GW. At site B, the DO concentration ranged from 7.13 to 11.26 mg L⁻¹ in the SW and from 3.57 to 8.00 mg L⁻¹ in the GW. The DO concentration showed a high level in the snowmelt season among the hydrological year. During the spring snowmelt period, a similar daily fluctuation of DO concentration was observed for SW and GW in both two sites, with a rising trend from April 6th to April 10th, and decreasing to a minimum around April 15th and then increasing to a relatively stable level.

3.2.4. Major ions

The major ion concentrations of SW and GW in non-frozen season were shown in Figure 6. Cl⁻ was the most enriched ion and Ca²⁺ was the most abundant cation in both site A and site B. The groundwater can be classified into Ca-Cl type (Figure 7) and this might be ascribed to the dissolution of gypsum mineral in the sediments or aquifer grains. Generally, the major ion concentrations in GW were higher than that in the SW, and the concentrations in GW increased with distance away from the river. The concentration of groundwater major ions in river bank is larger in site B than in site A, because there is less exchange between river water and groundwater far away from the river.

The concentrations of each ion in the SW and the GW showed similar trends and values upstream of the dam (Figure 6). Bush et al. (2006) demonstrated in their study that in the upwelling, the concentration of major cation and anion in the intercalation water of sediments was close to that of GW. The results of our study also indicated that the concentration of major ions at site A was closer to that of ambient GW.

3.3 Seasonal dynamics of the NO₃⁻ concentrations in SW and GW

The distributions of NO₃⁻ concentration in different water endmembers of the two sites were shown in Figure 8. The concentration of NO₃⁻ in the SW and GW at site A were higher than those at site B, while the

concentration of NO_3^- in the downstream ambient GW was higher than the upstream GW. For both sites, the NO_3^- concentrations in SW and GW were relatively high in April because of the accumulated materials released from snowmelt, while the relatively low concentrations were found in September mainly due to the less input NO_3^- (non-tourist season) and rainfall dilution.

For site A, the NO_3^- concentrations peaked in August 2016, which is probably caused by the human activities in the local tourist seasons. The seasonal variation trend of NO_3^- concentration was similar in the SW and GW. The highest concentration was observed in the GW, which could be due to the upwarding discharge of high concentration of NO_3^- . For site B, both SW and GW NO_3^- concentrations decreased during the summer flood period possibly because of the combination effect of the dilution effect and the infiltration of SW.

4. Discussion

4.1. The effects of damming on SW-GW exchange

Upwelling and downwelling conditions were precisely observed upstream and downstream of the dam (Figure 4). The riverbed and riparian sediment are subject to relatively coarse particles in the study area (Table 1), favoring the SW-GW exchange. As shown in Table 2, the vertical SW-GW exchange rates were about ten times larger than the transverse exchange rate. The exchange rate in the upstream (site A) was distinctly higher than the downstream (site B) for both the vertical and the transverse directions, except in the snowmelt season (Table 2).

Dam-induced changes to river stage and local GW flow can significantly alter the hydrologic regime and surface-subsurface hydrologic exchange (McCallum et al. 2014, Salem et al. 2020). After the Xilin River reservoir was built, the river discharge downstream the dam was largely reduced due to the interception of the natural water flow by the dam. There was nearly no water released from the reservoir during the experiment period, because the inflow water could not even reach the reservoir normal storage (Figure 1(c)). Even in the wet year, the reservoir would no longer release water to the downstream channel, as the impounded water is delivered into municipal water supply pipes.

The limited river discharge after the dam changed the pattern of SW-GW exchange and weakens the exchange in the downstream site of the dam. The Xilin River transitioned from the upwelling condition at the upstream site to the downwelling condition at the downstream site of the dam. The results indicate that damming could weaken the SW-GW exchange by altering the flow regimes of prairie rivers. The continuous, significant, and variable SW-GW exchange, such as reversal of hydraulic head gradients and weakening of exchange rate were also found in other dam-regulated rivers, especially in the semi-arid small rivers (Ferencz, Cardenas and Neilson 2019, Francis, Francis and Cardenas 2010, Gomez-Velez et al. 2017). However, the SW-GW exchange could be enhanced for those regulated large rivers in which river stage fluctuation could reach a few meters downstream the dam (Shuai et al. 2019, Song et al. 2020).

4.2. Nature tracer of EC and Cl⁻ indicating the SW-GW exchange

Water samples of different sources showed a more clustered distribution on the piper diagrams in the upstream than in the downstream. This may suggest that the hydraulic exchange between SW and GW in site A is stronger than site B. EC values and Cl⁻ concentrations in the SW and the GW can be used as natural tracer to indicate the SW-GW exchange. As shown in the Table 3, significant correlations of EC between SW and GW ($r > 0.66$, $p < 0.01$) was presented in the upstream during the study period, implying relatively strong SW-GW exchange. In contrast, the results in the downstream ($p > 0.05$) probably indicated weak SW-GW exchange. The correlation of EC in the SW and the GW also indicated that the different patterns of SW-GW exchange in the upstream and downstream. The Cl⁻ concentrations in the SW were significant correlated with the concentrations in the GW in site A (Table 4). In the upstream, the negative correlation implies that the discharge of groundwater with relatively high Cl⁻ concentration into the river can cause the increase of Cl⁻ concentration in river water which initially has low Cl⁻ concentration. However, there were no significant correlation in the downstream of the dam (Table 4). From this perspective, the SW-GW exchange in the upstream is more active than the downstream.

4.3. NO₃⁻ uptake and production

Figure 9 shows the variations of the sources and sinks of NO₃⁻ in different water endmembers in the upstream and downstream of the dam. As aforementioned, the upstream of the dam was characterized by upwelling conditions. NO₃⁻ in the upstream SW showed a net uptake (-0.288, -0.139, -0.208 and -0.699 g s⁻¹ for four sampling times respectively), probably ascribed to the dilution effect caused by the upwelling groundwater, combined with denitrification and uptake by the aquatic organism. In contrast, the NO₃⁻ exhibited a net production (0.121, 0.077, 0.083 and 0.623 g s⁻¹ for four sampling times respectively) in the downstream of the dam where downwelling conditions dominated, probably due to the nitrification and organic matter mineralization occurred in the SW-GW exchange zone. Additionally, as site B is closer to the urban region than site A, the SW in site B likely receives more nitrogen input affected by anthropogenic activities.

Globally, about 50% of the total nitrogen input to rivers is removed by denitrification (Mulholland et al. 2008, Yan et al. 2010). Studies have shown that denitrification is the best way to remove nitrogen from rivers and address riverine nitrogen pollution, and is a key process in regulating the nitrogen cycle on a regional or even global scale (Moomaw and Birch 2005). The temporal and spatial variability of denitrification rates in rivers is relatively high (Heiss, Fields and Fulweiler 2012), and this variability may also be manifested in different reaches of the same river (Wu et al. 2013). The results for nitrate in the river water upstream of the dam indicated that the upstream may be effective in removing nitrogen, and the NO₃⁻ removal was the most effective in spring snowmelt season.

It was suggested that denitrification was readily to occur in the anoxic zones of downwelling stream and was primarily limited by nitrogen and carbon (Holmes et al. 1996, Zenagui et al. 2020). In this study, the

relatively reducing substances in the GW enter the EZ through upwelling in the upstream, possibly leading to the enhancement of denitrification in the EZ. The downstream EZ was dominated by downwelling. In the downwelling zone, the injection of DO enriched SW may favor the ammonification and nitrification and thus results in the increase of NO_3^- concentrations in the GW (Xia et al. 2013). It is therefore concluded that upwelling or downwelling conditions will regulate the redox status and nitrogen supply of EZ and therefore affect nitrogen biogeochemical reactions (Heathwaite et al. 2021, Naranjo et al. 2015).

4.4. Limitations and future works

The nitrogen cycling depends on a series of reactions, such as denitrification, nitrification, anaerobic ammonium oxidation, ammonification, aquatic plant uptake, etc. (Haddou et al. 2018, Quick et al. 2019, Kollah et al. 2021). Denitrification is one of the most important reactions among them and a vital process for NO_3^- removal in the SW and GW (Li et al. 2021, Seitzinger et al. 2006). In the future, precise description of the related reactions involved in nitrogen cycling needs to be fully investigated. Lack of relevant species, such as dissolved organic carbon (DOC) and other forms of nitrogen limit the accurate characterization of hydro-geochemical cycling in the study. In addition, the geo-microbiological settings in the sediment, such as biofilm could potentially affect the biogeochemical reactions (Writer, Ryan and Barber 2011, Xia et al. 2018). How the micro-biotic processes mediate nitrogen cycling will also be considered in the future works.

5. Conclusion

Both hydraulic and hydrochemical approaches were applied to investigate the SW-GW exchange in the Xilin River. The pattern of SW-GW exchange and distribution of NO_3^- concentration in SW and GW show distinctly spatiotemporal variation. Upwelling conditions and NO_3^- net uptake in SW were observed in the upstream while downwelling conditions and NO_3^- net production in SW was shown in the downstream of the dam. The largest net removal rate of NO_3^- for river water was found in spring snowmelt period suggesting that snowmelt season can be a hot moment for SW-GW exchange among the whole hydrologic year in such cold region in North China. Results derived from multiple tracer methods and hydraulic method confirm that damming contributes to the transition from upwelling to downwelling conditions and weakens the SW-GW exchange. This study highlights the negative effect of damming on the SW-GW exchange and nitrate removal in prairie river systems.

Declarations

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Data Availability

The raw/processed data required to reproduce these findings cannot be shared at this time as the data also forms part of an ongoing study.

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Tables

Table 1. Vertical distribution of soil particle size at the experiment sites A and B

Sampling Sites		Grain size			Soil type
		> 0.05 mm	0.002-0.05 mm	<0.002 mm	
Upstream	HS-RC1	76.79	22.89	0.32	Sandy loam
	HS-RC2	79.02	20.68	0.30	Sandy loam
	HS-RC3	81.09	18.49	0.42	Sand
	HS-RB1	76.61	23.08	0.31	Sandy loam
	HS-RB2	72.01	27.62	0.31	Sandy loam
Downstream	RB-RC1	94.20	5.8	0	Sand
	RB-RC2	86.18	13.76	0.06	Sand
	RB-RB1	66.58	32.72	0.70	Sandy loam
	RB-RB2	87.72	12.25	0.03	Sand

Table 2. The dynamics of SW-GW exchange rates ($10^{-3} \text{ m}^3 \text{ s}^{-1}$) in the two study sites. The symbol of “_” denotes no observation.

	Vertical exchange rates		Transverse exchange rates	
	site A	site B	site A	site B
Jul-2016	-9.46	7.21	1.23	-0.99
Aug-2016	-9.43	6.71	1.15	-0.71
Sep-2016	-9.97	1.28	1.38	-1.28
Apr-2017	-6.23	0.30	0.57	-1.47
May-2017	–	–	1.04	-1.54
Jun-2017	–	–	–	-0.98

Table 3. The correlation of EC in the SW and the GW at site A and site B.

Site A			Site B		
EC	r	p	EC	r	p
HS-RC1	0.824	<0.01	RB-RC1	0.156	> 0.05
HS-RC2	0.777	<0.01	RB-RC2	0.503	<0.01
HS-RC3	0.835	<0.01	RB-RC3	0.260	<0.01
HS-RB1	0.691	<0.01	RB-RB1	0.386	<0.01
HS-RB2	0.994	> 0.05	RB-RB2	0.568	<0.05

Table 4. The correlation of Cl⁻ concentrations in the SW and the GW at site A and site B.

Site A			Site B		
Cl ⁻	r	p	Cl ⁻	r	p
HS-RC1	-0.583	<0.05	RB-RC1	0.822	> 0.05
HS-RC2	-0.552	<0.05	RB-RC2	0.837	>0.05
HS-RC3	-0.612	>0.05	RB-RC3	0.976	>0.05
HS-RB1	-0.790	<0.05	RB-RB1	0.420	>0.05
HS-RB2	-0.840	<0.01	RB-RB2	0.323	>0.05

Figures

Figure 1

The Xilin River, located in the semiarid regions of north China, is a typical inland meandering river in the grassland of Inner Mongolia

Figure 2

The spring snowmelt period starts at the end of March with duration of about 20 days. About 34.2 % of the annual runoff is derived from the spring snowmelt

Figure 3

Two upland monitoring wells, namely HS-RB1 and HS-RB2 for site A, and RB-RB1 and RB-RB2 for site B, were also installed perpendicular to the stream

Figure 4

water level differences between stream water and groundwater in monitoring wells.

Figure 5

variations of EC in the SW and GW during the non-frozen seasons. Generally, the EC at site A was lower than site B, and the EC in the SW was significantly lower than the GW for both sites.

Figure 6

The major ion concentrations of SW and GW in non-frozen season

Figure 7

Cl^- was the most enriched ion and Ca^{2+} was the most abundant cation in both site A and site B. The groundwater can be classified into Ca-Cl type

Figure 8

The distributions of NO_3^- concentration in different water endmembers of the two sites

Figure 9

variations of the sources and sinks of NO_3^- in different water endmembers in the upstream and downstream of the dam.