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Changes in groundwater induced by water diversion in the Lower Tarim River, Xinjiang Uygur, NW China: Evidence from environmental isotopes and water chemistry

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SUMMARY

The Lower Tarim River in NW China is under severe ecosystem degradation due to stopped stream flow and diminished groundwater recharge. Since year 2000, eight water diversions from the upper stream and from the neighboring Kaidu–Kongque River have been implemented to alleviate the ecosystem disaster. In order to assess the effectiveness of the water diversion project and to identify proper tracers of groundwater dynamics, we sampled the riparian groundwater system in 2007 and 2008 along the 350 km-long river channel through the 40 monitoring wells situated along nine transects perpendicular to the river and three soil profiles. Measurements on the samples have included environmental isotopes (^{18}O, ^{2}H, ^{3}H) and water chemistry. The results show that remarkable changes have been induced by the water diversions. The observed response of riparian groundwater system includes general decrease in total dissolved solid (TDS) and rise of water table. Scope with greater than 1 m rise in water table is within ~700 m from the riverbank in the upper segments and ~300 m in the lower ones. Greater rise of water table occurs near the river bank. Tritium data show that the extent of modern recharge (since 1960s), including that from the diverted water, is limited to 600 m from the riverbank at the upper segments and 200 m at the lower ones. Stable isotopes show that groundwaters, regardless of modern or pre-modern, are enriched in heavy isotopes and are plotted in parallel to the meteoric water line in the δ–δ plot, attributed to evaporation during recharge. Groundwater is generally of Na–Mg–Cl–SO_4 type and is formed by dissolution of minerals, such as halite, sulfate, and carbonates, based on component correlation matrices analysis. The salinity of groundwater is mainly affected by that of the diverted water and of the local antecedent groundwater, salts in the unsaturated zone, evapotranspiration during recharge. As the zone of smaller groundwater depth (less than 5 m) suitable for the most existing *Populus euphratica* and *Tamarix ramosissima,* the main species targeted by the rescue effort, restricts to 200 m from the riverbank, and narrows down towards downstream, long-term stability of the ecosystem cannot be achieved by the current water diversion scheme and regulating/saving water in source-streams and the Upper/Middle Tarim River is crucial for continuing water diversion.

1. Introduction

Under the dual impacts of anthropogenic activities and climate change, a common scenario in arid and semiarid catchments, particularly in the lower reaches of them, is severe ecological degradation, such as death of vegetation, intensified groundwater salinization, soil salinization and desertification, etc. (Feng et al., 2005; Gremmen et al., 1990; Ma et al., 2005; Richardson et al., 2007; Stromberg et al., 1996; Wang and Cheng, 2000). In groundwater-dependent ecosystems, dynamics of soil moisture relative to water table fluctuations control the overall ecosystem dynamics (Tamea et al., 2009), and when the riparian ecosystems are threatened by insufficient supply of the river, all the services provided by the ecosystems may be threatened (Stromberg et al., 2007). Therefore, ecological allocations of water should be considered for groundwater-dependent ecosystems (Eamus et al., 2006). Interest and investment in river restoration projects are growing worldwide (Klein et al., 2007 and references within). Better understanding the essential groundwater–surface water interaction and water chemistry evolution is of great significance to any restoration planning and implementation (Baillie et al., 2007; Loheide and Gorelick, 2007).

With a relatively stable inflow from headwater streams, Tarim River, Xinjiang Uygur, NW China as a typical catchment of its kind, water discharge to the lower reaches has decreased rapidly as a

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result of large scale agriculture development and irrational water resources utilization in the upper and middle reaches of the river (Feng et al., 2005; Song et al., 2000). The runoff has ceased to flow into the 350-km-long Lower Tarim River since the construction of the Daxihaizi Water Reservoir in 1972, causing severe damages to the riparian forest dominated by *Populus euphratica* (Song et al., 2000). The so-called "Green Corridor" and the high way along the Lower Tarim River (Kuala to Ruoqiang, part of National High Way G218) are endangered by degrading riparian vegetation and desertification.

Since 2000, the water diversion project (cost CNY 10.7 billion) has been implemented to alleviate the vegetation degradation. After the eight impulsive water diversions, the water tables have risen at differing degrees from the river bed and the composition, types, distribution and growth status of the riparian vegetation have changed correspondingly (Chen et al., 2008a,b; Deng, 2009; Hou et al., 2007a,b; Tao et al., 2008). Due to long-time runoff cutoff, the ecosystem is not sustainable as survival plants are dominated by the older ones. Although a total volume of 2.27 billion m^3 of water has been diverted to the lower reaches by eight water diversions, it is still at the stage of compensation for groundwater recession. The present impulsive linear water diversion through the river bed cannot radically preserve the local eco-environment, though ecological degradation is slowed down to some extent. Sustainable water utilization and ecosystem protection scheme are imperative for the degraded ecosystem restoration and prevention of the Taklimakan Desert and the Kuluk Desert from merging in the Lower Tarim River.

The study of water cycle evolution to solve water resources and relative environmental issues has become a significant task of the hydrological science (Zhang et al., 2000). Since climate, landforms, lithology and anthropogenic activities influence the crustal weathering and groundwater chemistry (Shen et al., 1993), an understanding of the chemical evolution of the groundwater can provide insight into the interaction between water and environment and can contribute to rational water resources management (Adams et al., 2001; Edmunds, 2009; Ma et al., 2009), especially in arid regions with fragile ecosystem and intense anthropogenic interferences (Bennetts et al., 2006). Groundwater chemistry is important for ecosystem restoration since water quality controls vegetation growth status and soil characteristics, though the most present studies focus on responses of water table and vegetation changes in the Lower Tarim River. One of the disadvantages of high salinity water to vegetations growth is to prevent vegetation from absorbing moisture and to plague soil fertility (He et al., 2006; Manchanda and Garg, 2008). Plant features, including species richness, species diversity and species composition are significantly related to salinity and deteriorate with increasing salinity (Lymbery et al., 2003). As groundwater depth, flood events and salt concentration are the key factors controlling the growth of the arbor and shrub (Thevs, 2007; Williams et al., 2006), the variations of water table and quality and their impacts on vegetation growth under water diversion circumstances need further study.

To increase the chance of success in riparian ecosystem restoration, it is important to gain knowledge of the hydrological and hydrogeochemical processes involved. The Xinjiang Water Conservancy Bureau and Ministry of Water Resources of China had investigated environmental background in the Lower Tarim River before water diversion (Deng, 2009). Based on monitoring data of water table and total dissolved solid (TDS) in groundwater since 2000 and two systematic sampling campaigns carried out in August 2007 and May to June, 2008, this paper attempts to assess the effectiveness of the water diversion project through understanding the mechanism of groundwater system changes, including groundwater residence times, the extent of groundwater recharge under water diversion, the changes of salinity and salinization mechanism so as to provide a basis for an optimal water diversion schemes using environmental isotopes (^{18}O, ^{2}H, ^{3}H), water chemistry and water table monitoring data. It is also a purpose of current study to identify suitable environmental tracers for the eco-hydrological processes.

2. Background

2.1. General setting

The Tarim River Basin is located in the south of Xinjiang, NW China. It has an area of 1.04×10^6 km^2 and is flanked by the Tianshan Mountains to the north and by the Kunlun Mountains to the south (Fig. 1). The Taklimakan Desert, the largest desert in China, is located in the center of the basin, occupying an area of 3.37×10^5 km^2 (Zhu et al., 1981). Aksu River, Yarkant River and Hotan River are three large rivers in the west of the basin, which feed the Tarim River at Aral. Due to river regulation, the later two presently recharge the Tarim River only during large floods. The Tarim River starts from Aral to Taitema Lake with a length of 1321 km. The upper stream of the Tarim River is from Aral to Yingbazha (495 km), while the middle reaches is from Yingbazha to the Qiala Water Reservoir (398 km), and the lower reaches is from the Qiala Water Reservoir to Taitema Lake (428 km) (Fig. 1). The lower reaches can be divided into three segments: the upper from the Qiala Water Reservoir to the Daxihaizi Water Reservoir, the middle from the Daxihaizi Water Reservoir to Aragan, and the lower from Aragan to Taitema Lake. The river is divided into two branches starting from the Daxihaizi Water Reservoir. The west is the old Tarim River (142 km) and the east is the Tarim River (205 km). The two branches combine at Argan. Further downstream from the Daxihaizi Water Reservoir, there is little irrigation and agricultural activities. The Kaidu River flows into Boston Lake, then is pumped into Kongque River at southwest of the lake, and finally reaches Lop Nur. There are two eolian deserts situated in both sides of the Lower Tarim River, the Taklimakan Desert on the west and the Kuluk Desert on the east, with moving and semi-moving sand dunes.

2.2. Hydrogeology

The Tarim River Basin is a Mesozoic–Cenozoic basin surrounded by folded mountains. Outcrop of the strata from late Paleozoic to Cenozoic is described in Fig. 1 and the Paleozoic is widely distributed in mountain areas. Archaeozoic and Proterozoic schist and gneiss, Paleozoic and Mesozoic sand stones, conglomerates and magmatic rocks occur in the sources area of the Tarim River (XETCAS, 1965; Zhu et al., 1981). The large fault between basins and mountains controls the tectonic evolution and forms a series of major depressions in the Tarim River Basin, such as Kuche Depression, North Depression, Southeast Depression and Southwest Depression (Cai et al., 1997). The major sediment is tertiary in those depressions, for instance, the Kuche Depression with depth of thousands of meters has a tertiary deposit with a maximum depth of 4500 m (Li et al., 2000). The quaternary has extensive distribution in those depressions (Fig. 1).

The occurrence of groundwater is similar between the Southern Tianshan watershed and Northern Kunlun watershed. The sink of the two groundwaters systems is centered in the south of the Tarim River. The diluvial aquifer from the northern mountains is composed of sand deposits some 100–300 m thick forming an unconfined aquifer in which the present day water table ranges between 20 m and 200 m below surface (Fig. 2). This allows a certain amount of surface runoff in the piedmont fan to infiltrate and recharge the aquifer. At the southern edge of this diluvial fan, the

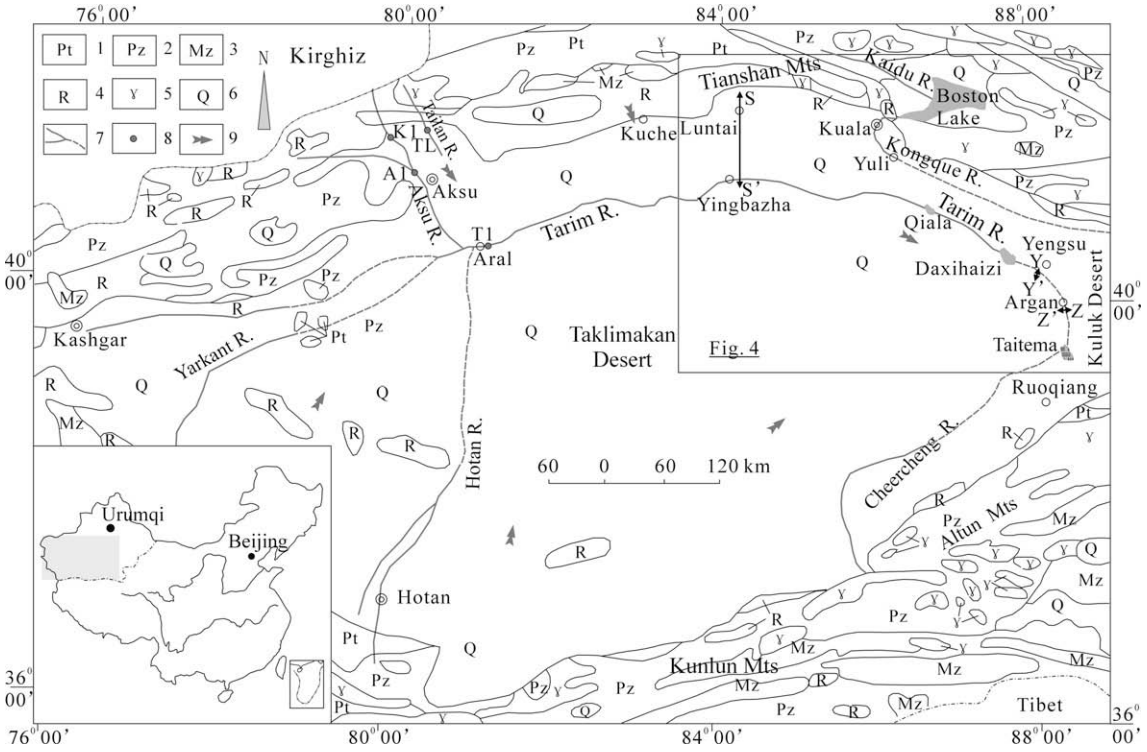

Fig. 1. Sketch map of the Tarim River Basin (modified from Li et al. (2000)) 1 – Proterozoic; 2 – Paleozoic; 3 – Mesozoic; 4 – tertiary; 5 – granite; 6 – quaternary; 7 – river; 8 – surface water in the Upper Tarim River and the Aksu River; 9 – regional groundwater flow.

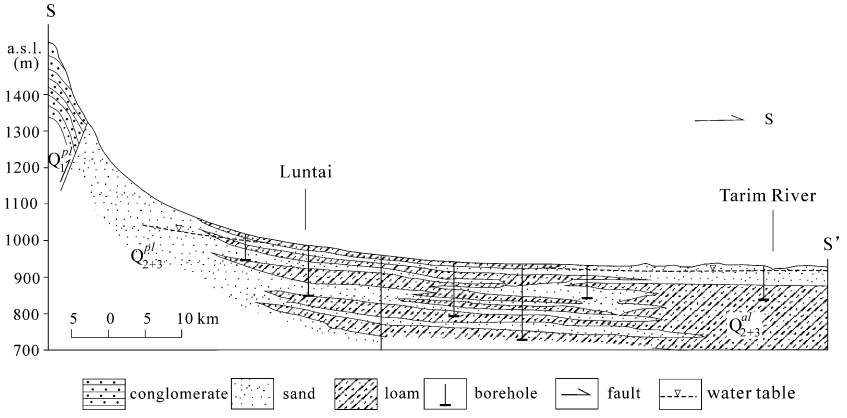

Fig. 2. Hydrogeological cross section in the Middle Tarim River, modified from Li et al. (2000), see Fig. 1 for the location.

aquifer comprised of alluvial loam becomes confined or semi-confined and the thickness is about 200 m. The regional flow is from north to south (Li et al., 2000). In the middle and lower reaches of the Tarim River, local groundwater system gets recharge from surface water through bank infiltration.

The Lower Tarim River is located at a flat alluvial plain and the hydrogeological condition is simple. The multi-layered aquifer with homogenous lithology is dominated by fluvial and lacustrine facies of fine sand and silty fine sand, and locally by eolian sand (Fig. 3). The typical porous aquifer can be divided into phreatic and confined ones. The shallow groundwater is closely related to surface water, serving as the main water sources for the riparian vegetation. The phreatic aquifer mainly consists of fine sand and silty fine sand and has a relatively low permeability and poor water yield with the thickness, hydraulic conductivity and specific yield of 30–40 m, 1.2–4.8 m d^{-1} and less than 150 m^3 d^{-1} m^{-1}, respectively. The aquifer is underlain by a clay bed, which extends continuously and horizontally and almost has little hydraulic connection to the confined aquifer underneath (Alim and Xu, 2003; Dong and Deng, 2005).

The minerals contained in the granites and metamorphic rocks are difficult to be dissolved. In the alpine areas, the soluble rocks are mainly carbonate (Zhu and Yang, 2007). The dissolution begins with eroding host rocks with CO_2 from precipitation and soil zone and generally forms low mineralization type water of HCO_3–Ca in mountainous area. The originating water for the Tarim River has low TDS (less than 0.5 g/L) at river closure mouths. In the tertiary sediments at the low elevations, more soluble deposits are found (Zhu et al., 1981). Under inland dry climate conditions, those deposits make considerable contribution to the high TDS in river and groundwater in middle and lower reaches of the river. At the Tarim River sections, Fan et al. (2002) showed that the average

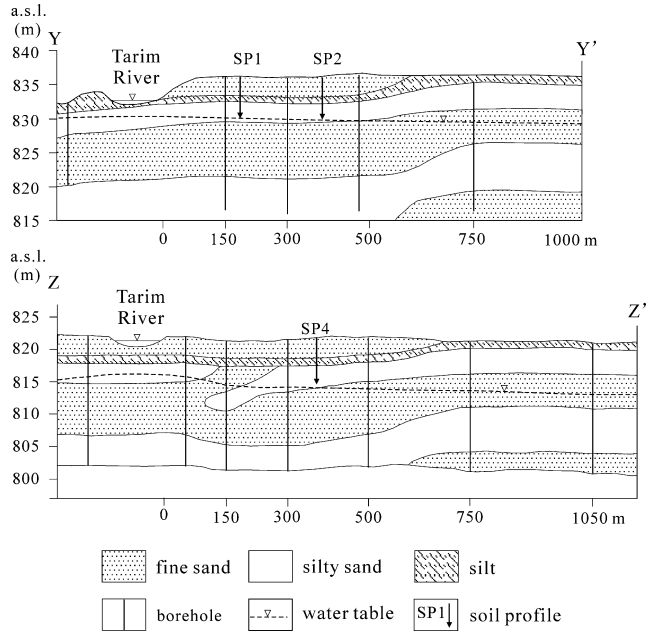

Fig. 3. Hydrogeological cross sections in the Lower Tarim River, modified from Deng (2009), see Fig. 1 for the location; also showing the location of the three soil profiles.

monthly mineralization since 1958 was greater than 1 g/L over the entire year except for the wet season (August), due to runoff decreasing and salt drainage of the main canals for irrigation. Zhang et al. (1995) studied water chemistry of source-rivers of the Tarim River using conventional hydrochemistry and strontium isotopes ($^{87}Sr/^{86}Sr$) and pointed out that weathering of silicates, carbonates and evaporite contributed most of the TDS in those rivers. The Lower Tarim River, as the terminal of water and salt of the Tarim Basin, has high salt concentration due to long-term evaporation (Cheng, 1993). The background of the regional TDS distribution in groundwater in the Lower Tarim River before the water diversions is that, the scale of low-mineral groundwater region with TDS of 1–3 g/L distributes in riparian area with the width from 1000 to 2000 m to the riverbank due to river dilution, however, beyond this zone, TDS in groundwater is greater than 5 g/L and more than 10 g/L in the lower segments of the lower reaches (Cheng, 1993; Deng, 2009).

Table 1
Runoff changes at each station on the Tarim River (10^8 m^3).

	1950s	1960s	1970s	1980s	1990s
Aral	49.4	51.7	44.4	44.8	42.0
Qiala	13.5	11.4	6.7	3.9	2.8
Tikanlik	8–9	2.9	0.5	0.4	0.1
Argan	Persist	Discontinue	Nil	Nil	Nil
Lop Village	5–4	0.2	Nil	Nil	Nil

2.3. Climate

The Lower Tarim River is dominated by typical continental temperate arid climate. According to Tikanlik meteorological station in the Lower Tarim River, the average precipitation is about 40 mm/yr; the potential evaporation is 2590 mm/yr; the annual average temperature is 10.5 °C and average sandstorm days are 8.2 days per year, causing severe wind-sand hazards.

2.4. Ecosystem degradation

Under the impact of anthropogenic activities, runoff of three source streams (Aksu River, Hotan River and Yarkant River) to the Tarim River has decreased gradually in the last 50 years due

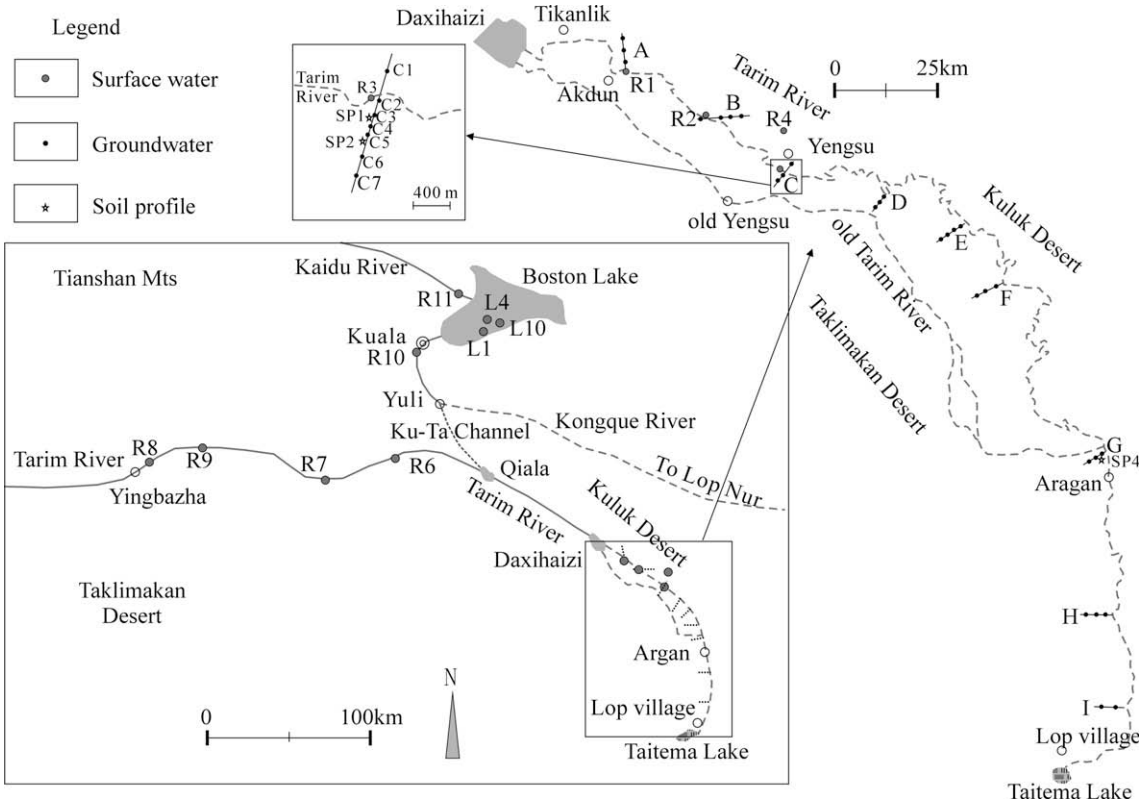

Fig. 4. Sampling locations.

Table 2
Statistics of eight water diversions to the Lower Tarim River.

No.	Delivering duration	Boston Lake (10^8 m^3)	The Tarim River (10^8 m^3)	Total (10^8 m^3)	Transection for water reach
First	2000.5.14-7.12	0.98	0	0.98	Karday (E)
Second	2000.11.3-2001.2.5	2.25	0	2.25	Kargen (I)
Third	2001.4.1-7.6	1.84	0	1.84	Taitema Lake
	2001.9.12-11.18	1.63	0.35	1.98	Taitema Lake
Fourth	2002.7.20-11.10	2.45	0.86	3.31	Taitema Lake
Fifth	2003.3.3-7.11	2.4	0.97	3.37	Taitema Lake
	2003.8.4-11.3	0	2.85	2.85	Taitema Lake
Sixth	2004.4.23-6.22	0.74	0.29	1.03	Taitema Lake
Seventh	2005.4.18-6.7	0.52	0	0.52	Yikanbujima (H)
	2005.8.30-11.2	0	2.28	2.28	Taitema Lake
Eighth	2006.9.25-11.21	0.26	2.07	2.33	Taitema Lake
Total	13 (57%)	9.7 (43%)	22.7 (100%)		

to extensive oasis agriculture with increasing water utilization (Feng et al., 2005). With gradually decreased inflow to the Tarim River, proportion of water consumption in the upper and middle reaches increased gradually from 1970s to 1990s, while flow to the lower reaches reduced significantly (Table 1). The groundwater depth has increased to 8–12 m. As a result, serious ecosystem degradation in the Lower Tarim River has been observed and the major manifestation is described as follow (Deng, 2009):

(1) The area of the groundwater-dependent *P. euphratica* forests has decreased from 54,000 hm^2 in 1950s to 7.3000 hm^2 at present. The Green Corridor of great strategic significance for the Lower Tarim River between Taklimakan and Kuluk Deserts has been brought to the verge of destruction, as the two deserts are taking on a trend of merging and some flowing sand has already intruded into the river bed and the national highway G218.
(2) The desertification has intensified. According to the aerial photographs in 1959–1983, desertification has developed most seriously in the Lower Tarim River. The area of the desertified land has increased by 22% during 24 years. Particularly, desertification has developed dramatically since 1972 as the Lower Tarim River dried out. As a result, the vegetation coverage has been declining, and many roads, farmlands and villages have been buried by sand, causing a serious threat to the survival and development of the oasis.

2.5. Water diversion

To protect the Green Corridor, the riparian vegetation restoration is imperative. Taking advantage of the wet period of the Kaidu River, the Ku–Ta Channel (Yuli to Qiala, Fig. 4) was constructed for diverting water from Kongque River to the Lower Tarim River. The source water for the first five water diversions is mainly Boston Lake (Table 2) while that for the later water diversions is mainly the Tarim River, which was all stored in the Daxihaizi Water Reservoir before being released to the lower reaches. Altogether 2.27 billion m^3 of water (1.30 billion m^3 from the Boston Lake and 0.97 billion m^3 from the Tarim River) has been diverted to the Lower Tarim River from the Daxihaizi Water Reservoir from year 2000 to the end of 2006. Six impulsive water diversions out of the eight reached the Taitema Lake, the terminal lake of the Tarim River. Groundwater system has been changed by the water diversions, resulting in redistribution of water salinity, water table and soil moisture content, which affect the evolution of the ecosystem.

3. Sampling and analyses

The first sampling campaign was carried out during the wet season on 26–31, August, 2007, after the eighth water diversion. Surface waters from Kaidu River, Boston Lake, Kongque River, the middle reaches of the Tarim River and residual waters of water diversion in dry riverbed in the Lower Tarim River were collected. Two rainfall samples were also collected during a storm event. Groundwater samples were collected from boreholes at varying distances from the riverbed along nine groundwater monitoring transects, namely: Akdun (A), Yahopumarhan (B), Yengsu (C), Abudali (D), Karday (E), Tugmailai (F), Aragan (G), Yikanbujima (H), and Kargan (I) (Fig. 4), respectively. Along each transect, monitoring wells with depth of 8–17 m were dug at intervals of 100–200 m. In total, 40 monitoring wells were measured and sampled. Water table and TDS of groundwater have been monitored three times a month during the water diversions and once a month in between the water diversion events.

A further sampling campaign was implemented in May 28 to June 3 in 2008 in Aksu River, the Upper Tarim River, the Qiala

Table 3
Site measurements and isotopic composition of surface waters.

Sample	TDS (g/L)	EC (ms/cm)	Temp (°C)	pH	δ^{18}O (‰)	δ^2H (‰)	^{3}H (TU)
Aksu River							
K1	0.132	0.28	12.7	8.0	−11.1	−78.0	
TL	0.149	0.31	10.5	8.6	−10.8	−77.6	20.0
A1	0.217	0.45	30.7	7.8	−10.0	−74.9	
Tarim River							
T1	0.608	1.24	31.1	7.6	−9.9	−73.7	
R8	0.666	1.35	23.0	8.8	−8.4	−58.7	18.4
R9	0.642	1.30	24.8	8.0	−8.3	−57.9	
R7	0.631	1.28	27.7	8.5	−7.8	−54.9	20.3
R6	0.617	1.24	24.8	8.1	−8.5	−59.4	18.8
Kaidu–Kongque Basin							
R11	0.181	0.38	19.6	8.4	−7.8	−54.0	27.7
L1	1.053	2.09	27.4	8.6	−1.0	−17.1	28.3
L2	0.981	1.49	24.5	8.6			
L3	0.949	1.89	24.6	8.6			
L4	0.840	1.68	24.3	8.6	−2.7	−26.9	
L5	0.844	1.69	23.6	8.7			
L6	0.838	1.68	23.8	8.6			
L7	0.838	1.67	23.8	8.6			
L8	0.769	1.55	24.3	8.6			
L9	0.792	1.59	23.9	8.6			
L10	0.844	1.70	23.5	8.5	−2.4	−23.4	26.9
R10	0.661	1.34	23.2	8.1	−6.7	−46.7	28.1
Water reservoir							
Qiala	0.754	1.49			−4.6	−48.3	
Daxihaizi	0.894	1.79	24.0	9.2	−3.6	−42.4	
Residual water in the Lower Tarim River							
R1	3.20	6.06	23.3	9.0	3.5	−5.8	27.0
R2	2.64	5.00	25.5	8.8	2.0	−17.1	
R4	3.82	7.19	26.0	8.2	2.4	−13.1	32.6
R3	2.93	5.56	27.3	9.8	2.5	−13.8	22.7
Rainfall							
Rn1					−2.2	−12.7	
Rn2					−1.1	−6.6	29.8

Water Reservoir and the Daxihaizi Water Reservoir. The data of these surface water bodies will serve as source water composition and major modifications through evaporation. Soil samples were collected from three sampling profiles in the Lower Tarim River. The soil samples were obtained using a hollow-stem hand auger with interchangeable 1.5 m aluminum rod from three profiles (5.8–7.7 m depth, SP1 and SP2 in section C and SP4 in section G, Figs. 3 and 4). Bulk soil samples of ~400 g were collected at intervals of 0.25 m. Samples were homogenized over the sampled interval and immediately sealed in polyethylene bags. Gravimetric moisture content was determined by drying a minimum of 80 g of soil at 110 °C for 12 h. To determine chloride content, double-deionised water (40 mL) was added to the oven-dried soil sample (40 g) (Scanlon, 1991). Samples were agitated on a reciprocal shaker table for 8 h. The supernatant was filtered through 0.45 μm filters. Chloride was then analyzed by ion chromatography. The chloride concentration of the soil solution is then calculated by dividing the measured concentration by gravimetric moisture content and by multiplying the mass ratio of solution to oven-dry soil.

Depth of boreholes, water table, location (GPS), water temperature, pH, TDS and electrical conductivity (EC) were measured on the site. Fifty milliliters of water was collected for stable isotopes and was measured in the Stable Isotopes Laboratory, Institute of Geology and Geophysics, Chinese Academy of Sciences. $^{2}H/H$ and $^{18}O/^{16}O$ were measured on isotope ratio mass spectrometry (MAT253™) by chrome reduction and equilibrium with CO_2, respectively. Results are reported as $\delta^{2}H$ and $\delta^{18}O$ ($\delta = (R_{sample}/R_{standard} - 1) \times 1000$) using the Vienna Standard Mean Ocean Water (VSMOW) as standard. The analytical precision is ±1‰ for $\delta^{2}H$ and ±0.1‰ for $\delta^{18}O$. Five hundred milliliters of water samples were collected for tritium measurement in the Groundwater Tracing Laboratory, Institute of Geology and Geophysics, Chinese Academy of Sciences through electrolytic enrichment with a tritium enrichment factor of about 20 and the liquid scintillation counting (Quantulus 1220™) method with a detection limit of 0.3 TU (Tritium Unit). Water chemistry was measured using ion chromatography (Dionex-500™) at the Beijing Research Institute of Uranium Geology. The methods for cation measurements are taken from the National Analysis Standard DZ/T0064.28-93 while for anion are from DZ/T0064.51-93. Alkalinity was measured on automatic titrator (785 DMP™). Analytical precision was 3% of concentration based on reproducibility of samples and standards and detection limit was 0.1 mg/L. The charge balance error for 24 samples with low TDS ranges from −4% to 4% and that for 11 samples with high TDS ranges from −13% to 8%, which may be caused by large dilution ratio and disturbance of organic matters. The results are shown in Tables 3–5.

Table 4
Site measurements and isotopic composition for groundwaters at each section.

Sample	Local well	Distance to the river (m)	Welldepth (m)	TDS (g/L)	SEC (ms/cm)	Temp (°C)	pH	$\delta^{18}O$ (‰)	$\delta^{2}H$ (‰)	^{3}H (TU)
Akdun (A)										
w1	A2	50 (east)	2.97	2.33	4.47	15.3	7.2			
w2	A3	150 (east)	4.40	2.43	4.62	16.9	7.3			
w3	A4	250 (east)	5.09	2.07	3.99	16.7	7.3			
Yahopumarhan (B)										
w4	B1	50	3.92	1.284	2.53	17.9	8.1	−5.7	−44.6	22.2
w5	B2	50 (east)	4.42	1.170	2.31	19.0	7.8			
w6	B3	150 (east)	5.93	1.383	2.71	17.5	7.8	−6.3	−49.2	21.4
w7	B4	250 (east)	5.96	2.04	3.94	19.0	7.9	−6.3	−48.9	21.8
Yengsu (C)										
w14	C1	250 (east)	2.17	3.72	7.02	18.2	6.4	−6.3	−49.9	37.0
w13	C2	50	4.62	0.884	1.77	17.6	9.6	−7.4	−56.3	18.9
w12	C3	150	5.66	1.588	3.11	15.8	8.3	−5.2	−41.6	25.6
w11	C4	250	5.46	1.737	3.38	16.7	7.9	−6.3	−48.8	43.8
w10	C5	350	5.50	1.631	3.18	17.9	7.7	−5.5	−44.1	46.3
w9	C6	450	6.03	1.986	3.84	17.9	7.5	−7.4	−56.7	22.5
w8	C7	750	6.17	15.32	26.20	18.0	7.2	−6.7	−53.7	1.1
Abudali (D)										
w15	D1	50	4.87	1.116	2.21	20.8	8.1			
w16	D2	150	4.86	0.973	1.94	18.6	7.7			
w17	D3	250	4.50	2.25	4.32	19.6	7.7			
Karday (E)										
w21	E2	150	5.92	1.160	2.29	18.0	7.8	−5.2	−41.1	22.7
w20	E3	250	6.01	0.750	1.51	18.3	7.9	−7.4	−57.1	19.1
w19	E4	550	5.57	1.192	2.36	19.7	7.4	−8.2	−62.9	<0.3
w18	E5	850	9.54	5.90	10.77	19.2	7.4	−8.1	−61.4	0.8
Tugmailai (F)										
w22	F1	50	4.52	1.605	3.13	21.4	8.0			
w23	F2	150	4.60	1.363	2.68	18.6	7.9			
w24	F3	250	4.98	3.11	5.89	19.5	7.9			
Aragan (G)										
w27	G2	50	4.80	1.693	3.32	18.0	9.2	−7.0	−54.1	19.6
w26	G4	500	9.22	0.878	1.75	21.1	7.9	−7.9	−60.1	29.2
w25	G5	800	6.57	1.039	2.05	19.2	7.9	−7.5	−57.4	2.4
Yikanbujima (H)										
w32	H1	100	5.60	1.355	2.65	20.3	7.6	−6.3	−49.2	29.0
w31	H2	300	5.88	1.279	2.51	22.2	7.5	−6.2	−48.7	30.0
w30	H3	500	6.56	0.742	1.49	18.9	7.6	−7.6	−58.9	4.1
Kargan (I)										
w29	I2	300	10.22	>55						
w28	I3	500	12.26	55	107.60	22.2	6.8	−6.6	−51.3	<0.3

Table 5
Chemical composition of the surface water and groundwater sections (mg/L).

Sample	Ca^{2+}	Mg^{2+}	Na^{+}	K^{+}	HCO_3^-	Cl^-	SO_4^{2-}	F^-	NO_3^-
Aksu River									
K1	39	11	3.7	2.3	92	4.0	66	0.4	2.0
TL	39	7.6	13	2.0	98	16	52	1.0	2.3
A1	44	20	17	2.8	120	25	94	0.2	3.1
Tarim River									
T1	82	45	127	7.6	153	193	226	0.4	3.4
R8	68	33	153	9.2	132	287	247	1.0	2.6
R9	67	33	160	10	128	272	233	1.1	2.5
R7	73	36	150	14	115	238	265	0.9	0.8
R6	65	31	150	10	130	264	234	0.7	2.7
Kaidu–Kongque Basin									
R11	64	16	12	3.9	194	7.7	47	0.1	2.0
L1	60	91	250	20	255	339	547	0.5	1.1
L10	57	76	195	16	246	274	438	0.4	1.6
R10	73	50	138	8.2	261	181	285	0.3	1.1
Water Reservoir									
Qiala	77	71	204	12	251	235	331	0.5	1.7
Daxihaizi	67	67	239	14	94	353	404	0.9	2.2
Residual water in the Lower Tarim River									
R1	142	214	887	40	101	1705	1609	1.9	3.1
R4	150	249	1125	54	660	1721	1477	6.5	<0.1
R3	115	177	799	37	222	1511	1191	1.6	4.1
Yahopumarhan (B)									
w4	126	84	329	23	324	525	503	1.3	6.1
w6	126	105	350	27	398	573	553	1.0	11.5
w7	151	159	485	31	436	906	875	1.0	15.1
Yengsu (C)									
w13	47	64	270	21	192	451	222	1.0	1.6
w12	92	103	454	23	445	665	430	0.6	3.8
w11	145	142	422	46	321	770	662	1.0	2.8
w10	136	135	415	29	362	728	687	1.4	10.7
w8	571	1095	4550	83	754	9291	5912	1.9	11.2
Karday (E)									
w21	107	62	338	17	323	457	433	0.7	6.0
w20	94	59	174	26	304	277	235	0.3	5.8
w19	96	80	321	18	302	548	396	0.7	4.2
w18	183	204	1949	53	353	3748	980	3.7	3.7
Aragan (G)									
w27	32	110	547	23	487	682	563	1.0	<0.1
w26	80	105	167	18	314	304	320	1.8	4.8
w25	103	70	268	21	336	415	340	1.4	4.1
Yikanbujima (H)									
w32	85	98	373	24	253	734	420	0.8	4.9
w31	160	109	249	20	260	581	530	0.6	5.5
w30	94	66	151	16	282	247	291	0.7	6.2

4. Results and discussion

4.1. Tritium

Tritium with half-life of 12.32 years carries the information of water itself and has severed as an important tracer in determining modern groundwater recharge, movement and ages (Michel, 2005). Tritium is used to distinguish pre-modern recharge from modern recharge in this study. The approximate tritium input sequence of precipitation in the study area is based on the tritium sequence for the neighboring Lop Nur from 1952 to 1996 reconstructed by Jiao et al. (2004), Urumqi meteorologic observation station from 1986 to 2001 (IAEA and WMO, 2006), river samples collected in 2001 by Li et al. (2006), and river samples plus one rainfall sample collected by this study. The data for the overlapping period (1986–1996) is similar (Fig. 5). In 2001, the content of tritium in surface water from the Aksu River and the Tarim River ranges from 20 to 56 TU, while in 2007, from 18 to 21 TU, and that in Boston Lake and residual water is more concentrated ranging from 22 to 33 TU. Although glacial melt water contributes about 40% to river runoff in the Tarim River Basin (Shen and Wang, 2002), tritium in precipitation in Tarim Basin is still higher than other areas. The content of tritium for the rainfall sample is 29.8 TU. This high tritium content may relate to nuclear tests, atmospheric circulation in high latitude (Liu, 2001). The results show a tritium content decrease from 2586 TU (1963) to 20~30 TU (2007). Using an exponential decay equation, the decayed tritium contents for 2007 in precipitation, which would represent tritium concentrations in groundwater that had infiltrated between 1952 and 2007, ranges from 50 TU to 225 TU from 1962 to 1966, and ranges from 10 to 35 TU from 1967 to 2007 (Fig. 5). Therefore, groundwater with tritium content less than 10 TU is regarded as pre-modern water, or at least most part is pre-modern water when mixing with modern water is considered.

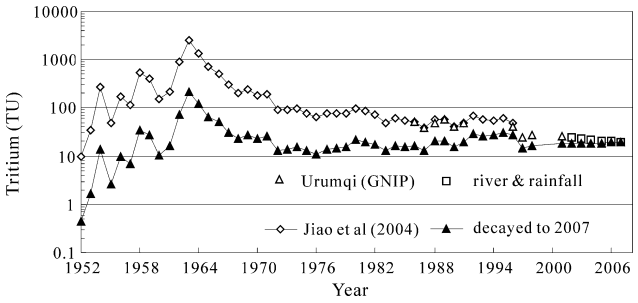

Fig. 5. The precipitation tritium input from 1952 to 2007 and the decayed value for 2007.

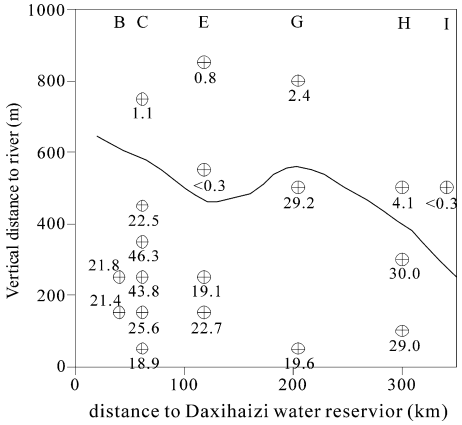

Fig. 6. Post map of Tritium content (TU) for groundwaters from the Lower Tarim River with a solid line showing the scope of modern recharge.

Fig. 6 shows the tritium contents in the riparian groundwater from the Lower Tarim River. The solid line in the figure is the boundary between groundwater that is recharged by modern water (19.1–46.3 TU) and that by pre-modern water (less than 4.1 TU). The extent of modern recharge (including the diverted water recharge) is limited to 600 m in the upper segments and to 200 m in the lower segments with a decreasing trend. There is a rising extent at section G (the confluence area of two tributaries) due to flush recharge.

According to the groundwater samples collected in the vicinity of the river bed of the Lower Tarim River in 1989 by Liu et al. (1997), 30% of them have the tritium contents more than 67 TU after the 18 years' decay by 2007, thus the groundwater in those regions is a mixing of the diverted water from year 2000 and the antecedent groundwater.

4.2. Stable isotopes

4.2.1. Stable isotopes in surface waters

Only two stations (Urumqi and Hotan, Fig. 1) are involved in the Global Network of Isotopes in Precipitation (GNIP) and the precipitation data are close to the Global Meteoric Water Line (GMWL, Craig, 1961). Therefore, for simplicity, the GMWL has been used as the reference meteoric water line for this study. In the field sampling in the Lower Tarim River, two rainfall samples were also collected during a rainfall event. The isotopic composition is shown with a slope of 5.5 (Fig. 7), indicating evaporation during precipitation.

The Kaidu River at the entrance to the Boston Lake has isotopic composition of −7.8‰ for $\delta^{18}O$ and −54.0‰ for $\delta^{2}H$. Isotopic enrichment occurs in the Boston Lake and the Kongque River as a result of evaporation. However, the Boston Lake water is not mixed completely with the deviation of 0.7‰ in $\delta^{18}O$ of the three water samples collected in the south part of the lake (Fig. 7). Those five samples form an evaporation line:

$$\sigma^2H = 5.4 \times \sigma^{18}O - 11.3, \quad R^2 = 0.998, n = 5 \tag{1}$$

with the slop of 5.4, indicating humidity of ~75% during evaporation (Gonfiantini, 1986). Four water samples were collected in 2007 from the middle reaches of the Tarim River. The isotopic composition ranges from −7.8‰ to −8.5‰ for $\delta^{18}O$ and from −59.4‰ to −54.9‰ for $\delta^{2}H$, respectively. These data points are closely plotted with an average deuterium excess ($\delta^{2}H$-8 × $\delta^{18}O$, Dansgaard, 1964) of 8.2. However, the isotopic composition of the Tarim River has considerable seasonal and annual variations. According to Liu et al. (1997) and Li et al. (2006), oxygen isotope of the river water from the Upper Tarim River is as depleted as −10.5‰. In the 2008 sampling campaign, the oxygen isotope for the Upper Tarim River (Aral, T1) is −9.9‰. Aksu River, located at a higher altitude, heavy stable isotopes are more depleted with respect to the Kaidu River. The samples from Aksu River (K1, TL, and A1, Fig. 1) show $\delta^{18}O$ between −10.0‰ and −11.1‰ with the average of −10.6‰. In all, the river water recharging the Lower Tarim River has the isotopic composition ranging from −11.1‰ to −7.8‰ with respect to $\delta^{18}O$. The stable isotopes in the water reservoirs in the lower reaches are enriched in heavy isotopes due to evaporation: the Qiala Water Reservoir has the $\delta^{18}O$ of −4.6‰ and the Daxihaizi Water Reservoir has the $\delta^{18}O$ of −3.6‰ due to extended evaporation. The residual waters collected from the river bed at the Lower Tarim River show $\delta^{18}O$ from 2.0‰ to 3.5‰ and $\delta^{2}H$ from −17.1‰ to −5.8‰, as a consequence of intensive evaporation.

4.2.2. Stable isotopes in groundwaters

Riparian groundwater along the Lower Tarim River used to be fed by the river before the cutoff. After the construction of the Daxihaizi Water Reservoir in 1972, the river ceased to flow. The pre-modern waters with tritium less than 4.1 TU recharged by the Tarim River show $\delta^{18}O$ value of −8.2‰ to −6.6‰ with an average of −7.5‰ and the $\delta^{2}H$ values of −62.9‰ to −51.3‰ with an average of −57.6‰ (Fig. 7).These groundwater samples are located relatively far away from the river bed at each section as compared to the modern water samples (Fig. 6).

The modern groundwaters with tritium contents ranging from 18.9 TU to 46.3 TU have been recharged by the Tarim River since the 1960s and some of them have mixed with the diverted water. The modern groundwaters show $\delta^{18}O$ values of −7.9‰ to −5.2‰ with an average of −6.4‰ and $\delta^{2}H$ values of −41.1‰ to −60.1‰ with an average of −50.0‰, heavier than pre-modern waters.

The modern and pre-modern groundwaters exhibit a similar behavior to fall in slight parallel to the meteoric water line, but enriched relative to the recharging river water. The phenomenon is commonly observed in dry climate (Chapman et al., 2008; Prasanna et al., 2009) and attributed to evaporation during river recharge to the riparian groundwater system in a rather uniform manner.

4.3. Hydrochemical characteristics

4.3.1. Surface water and groundwater chemistry

Hydrochemical data for surface water samples can be divided into four groups: source streams (Aksu and Kaidu River: K1, TL, A1 and R11), rivers (Tarim and Kongque River: T1, R6-9, R10), lakes and water reservoirs (Boston Lake, Qiala and Daxihaizi Water Reservoir: L1, L10, Qiala and Daxihaizi) and residual waters in the Lower Tarim River (R1, R3 and R4) and is all plotted in piper trilinear diagram (Fig. 8). The source streams have very low TDS ranging from 0.132 to 0.217 g/L and are mainly characterized by Ca–Mg–HCO_3–SO_4 type of chemistry. Hydrochemical type of the Tarim River and Kongque River has changed to Na–Ca–(Mg)–Cl–SO_4 facies, and the TDS changes slightly from 0.609 to 0.666 g/L. Boston Lake, Qiala and Daxihaizi Water Reservoir have Na–Mg–Cl–SO_4 facies with TDS ranging from 0.754 to 1.053 g/L. The residual waters in the river bed of the Lower Tarim River has a high TDS between 2.64 and 3.82 g/L, with hydrochemical type of Na–Mg–Cl–SO_4 and high pH of 9.0 ± 0.2, due to long-time intensive evaporation and dissolution of the salt from the near surface layer of the river bed.

The TDS in groundwater samples in the eastern river bed (sections A and B) and in the western (sections C, D, E, F, G, H and I) are all plotted in Fig. 9. The TDS in groundwaters has a large variation ranging from 0.7 g/L to more than 55 g/L (exceeding the detect limit of the field device). However, TDS in the groundwater within 600 m to the river bed ranges from 0.8 g/L (E3) to 3.7 g/L

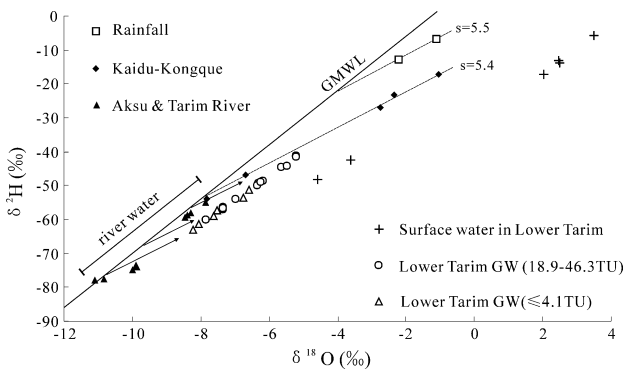

Fig. 7. Stable isotopic composition for surface waters and groundwaters.

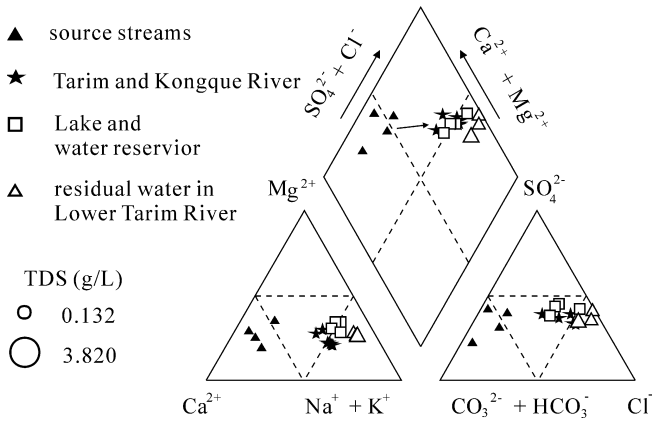

Fig. 8. Piper trigraph for surface waters.

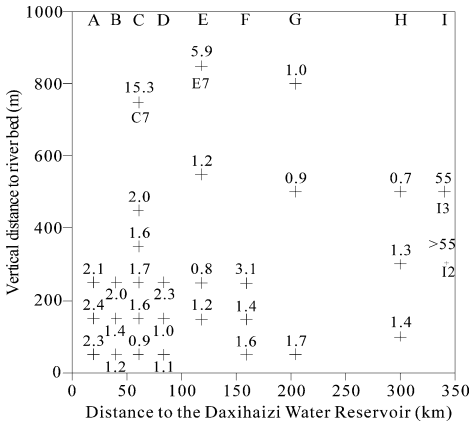

Fig. 9. TDS distribution (g/L) in the Lower Tarim River.

(C1) with an average of 1.6 g/L, except for section I, which is located in the terminal of the water flow and salt with very high TDS. Beyond the distance, TDS in groundwater is also high except for sample G5. The piper plot for the groundwaters (Fig. 10) shows that the dominant anions are Cl and SO_4 while the main cations are Na and Mg, forming geochemical facies of Na–Mg–Cl–SO_4 type.

The fluoride concentrations are high in the Tarim River Basin. In Aksu River, the fluoride concentrations range from 0.2 to 1.0 mg/L; in the Tarim River those become higher, ranging from 0.4 to 1.1 mg/L and groundwaters in the Lower Tarim River have the higher concentrations ranging from 0.3 (w20) to 3.7 mg/L (w18) with an average of 1.2 mg/L. The high fluoride concentrations in metasomatic deposit with intrusive contact in carbonate and acid igneous, Jurassic coal layers in some cases, and the high frequency of surface water–groundwater interactions in mountainous areas respond for high fluoride concentration in the Aksu River (Chen, 2008). The ratios of F to Cl for source stream (the Aksu River) range from 0.008 to 0.100 and that for the groundwaters in the Lower Tarim River range from 0.001 to 0.003 except for w26 of 0.006. Therefore, the high fluoride concentration in groundwater is mainly caused by evapoconcentration of river with high fluoride concentration rather than local water–rock interaction. The water type conversion from Ca–Mg–HCO_3–SO_4 with low TDS in source stream to Na–Mg–Cl–SO_4 with high TDS in the lower reaches of the basin and fluoride concentration evolution show the one of the most typical water evolution of arid water system, going along with evaporation and dissolution.

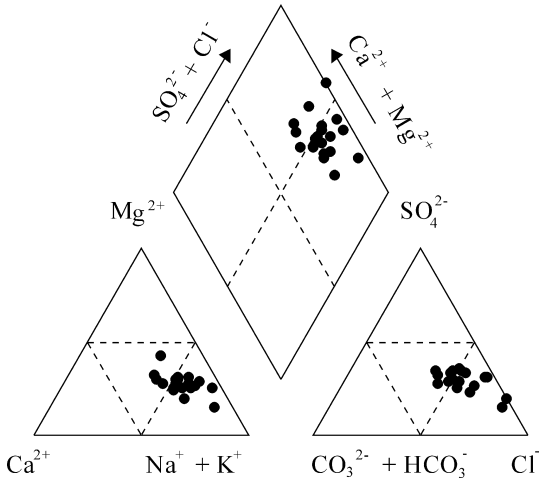

Fig. 10. Piper trigraph for groundwaters.

4.3.2. Groundwater salinization mechanism

The groundwater salinity in mainly controlled by dissolution, evapoconcentration and evapotranspiration in arid inland plain regions. Large quantities of salt in the unsaturated zone and aquifers in some cases can cause high salinity, even more than 55 g/L. The soluble salt in the Lower Tarim River is mainly evaporite (NaCl, $Ca_XMg_{(1-X)}SO_4$) and carbonate minerals ($Ca_XMg_{(1-X)}CO_3$) due to long-time concentration and accumulation (Zhu et al., 1981; Zhang et al., 1995). Correlation matrices for TDS and the main ions (Table 6) are used to find relationship between every two of the variables. There are strongly correlations between the TDS and ions ($r > 0.73$) except for NO_3 and F. Strong correlations exist between TDS and Cl, SO_4 and Na ($r > 0.95$), which only precipitate at very high salinity. The strong correlations between Na and Cl ($r = 1.00$) and an approximately 1:1 trend (Fig. 11a) suggest Na and Cl mainly comes from halite. Ca and Mg have strong correlations with SO_4 (0.97, 1.00) and relatively weak with HCO_3 (0.77, 0.84). The equivalent ratios for (Ca + Mg) against HCO_3 are more than 1 (Fig. 11b), suggesting that, besides dissolution of carbonates, dissolution of

Table 6
Correlation matrices for showing marked correlation at a signification level of 0.05 ($n = 18$) for groundwaters.

	TDS	Ca^{2+}	Mg^{2+}	Na^+	K^+	HCO_3^-	Cl^-	SO_4^{2-}	F^-	NO_3^-
TDS	1									
Ca^{2+}	0.95	1								
Mg^{2+}	0.98	0.97	1							
Na^+	1.00	0.93	0.96	1						
K^+	0.92	0.89	0.88	0.92	1					
HCO_3^-	0.81	0.77	0.84	0.80	0.73	1				
Cl^-	1.00	0.95	0.97	1.00	0.92	0.79	1			
SO_4^{2-}	0.97	0.97	1.00	0.96	0.87	0.84	0.96	1		
F^-	0.50	0.37	0.35	0.54	0.57	0.26	0.53	0.33	1	
NO_3^-	0.31	0.48	0.37	0.28	0.29	0.41	0.29	0.39	0.02	1

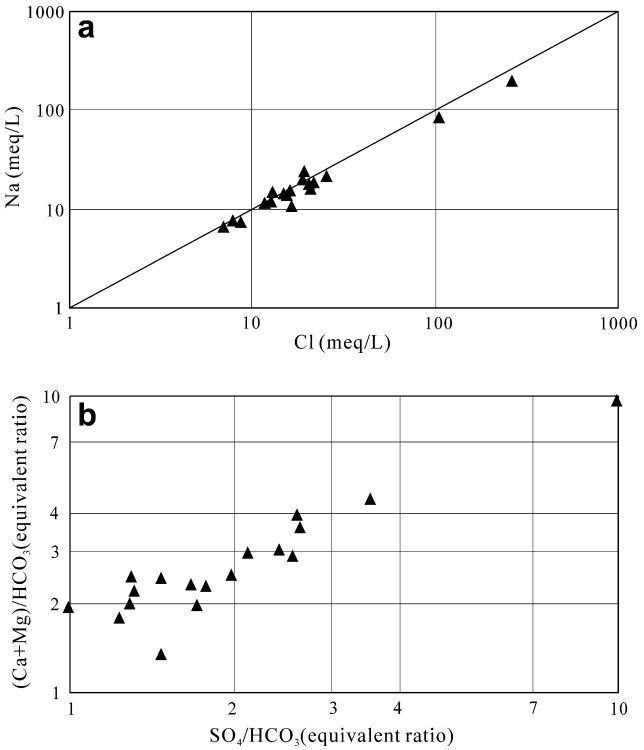

Fig. 11. Ionic ratio for the groundwaters.

sulfate ($Ca_XMg_{(1-X)}SO_4$) contributes the additional (Ca + Mg) when concentration of (Ca + Mg) increases with that of SO_4. The equivalent ratios for SO_4 against HCO_3 are all higher than 1, suggesting that quantity of (Ca + Mg) from dissolution of sulfate is more than that from carbonates. There is no further systematic increase in HCO_3 and Ca concentration in groundwaters with higher TDS, although the TDS increases significantly. The saturation indices for the groundwaters are calculated with PHREEQC software (Parkhurst and Appelo, 1999). All groundwaters are undersaturated with respect to gypsum and supersaturated with respect to calcite and dolomite in the Lower Tarim River. Such a pattern indicates that saturation of a relevant salt controls the concentration of the respective ions.

4.3.3. Soil moisture and salt content

Near surface soil moisture is low (0.05–2.16%), but generally increases with depth until coming into contact with silt layer (Fig. 3 and Fig. 12), with the depth of 0.55 m for SP1, 1.80 m for SP2 and 3.95 m for SP4. Below the silt layer, the moisture content decreases rapidly before reaching the water table (i.e. SP1 and SP2) or meeting the other silt layer (6.7 m, SP4). It can be concluded that the soil type controls its moisture content.

Since chloride is a conservative tracer in water cycle process that can only precipitate in very high concentration and is the major ion in the groundwaters in the Lower Tarim River and has strong correlation with the TDS (Table 6), it is a good tracer to check the salt content in the unsaturated zone. The chloride concentrations in capillary samples for SP1 and SP2 are 487 mg/L and 679 mg/L, less than that in groundwater samples, which are 665–770 mg/L and ~728 mg/L, respectively. There is an increasing trend in chloride concentration from the bottom to the top of the profiles. The presence of chloride peak in the three profiles, reaching 6.1 g/L for SP1, 87.5 g/L for SP2 and 9938 g/L for SP4 (actually most chloride in solid phase), respectively (Fig. 12), is widespread in (semi) arid condition and shown in numerous locations worldwidely (Gates et al., 2008; Phillips, 1994; Scanlon, 1991). Persistently removing the moisture near surface by evapo-transpiration contributes the accumulation of solutes. By comparison with similar environments, such as the Badain Jaran Desert, the diffuse recharge is expected to be very little and less than 1 mm/yr (Gates et al., 2008; Ma and Edmunds, 2006) under such arid condition where annual precipitation is only 40 mm/yr. If the volume weighted average chloride concentration in rainfall of 1.7 mg/L (from 2005 to 2007) for rural monitoring station in the vicinity of Xi'an, China (EANET, 2008) and the value of 1.5 mg/L measured by Ma and Edmunds (2006) in Badain Jaran Desert and 1.7 mg/L measured by Xu et al. (2009) are adopted, and the aerosol flux deposition is negligible, which is acceptable provided the long-term aerosol flux is near steady state (Goni et al., 2001), the annual chloride input from atmosphere is expected to be 70–100 mg m^{-2} yr^{-1}, contributing little to the salinity of the region of high salt content.

When considering the potential impact on water diversion and ecological restoration, soil salt and solution give better idea of salt evolution. Fig. 12 shows the chloride concentration in dry soil, appearing to be highly variable. The SP1 has less chloride in the whole profile, ranging from 24 to 862 mg/kg, compared to the SP2 and SP4, which show maximum chloride concentration of 4416 mg/kg (2.05–2.30 m depth) and 11,925 mg/kg (0.35–1.00 m depth), respectively. However, at the bottom of the profiles, chloride concentration is rather low. When groundwater table rises within the depth, the chloride concentration in groundwater table would be acceptable for the Tugai vegetation (Song et al., 2000; Thevs, 2007). Taking profile SP4 for an example and assuming groundwater chloride concentration of 400 mg/L, soil bulk density of 1.5 g/cm^3, sand porosity of 0.35 and silt porosity of 0.41, the chloride concentration would be 780 mg/L when groundwater table rises to the depth of 5.0 m. However, if flooding the SP4, it would be another story. The flood water would flush the salt into groundwater and the chloride concentration would exceed 6000 mg/L.

4.4. Water table and salinity changes

The first (from 14/05/2000 to 12/07/2000) and the second water diversion (from 03/11/2000 to 05/02/2001) got to Karday (E) and Kargan (I), respectively, and that of the later six water diversions all got to Taitema Lake except for the seventh, which only made it to Yikanbujima (H) (Table 2). The four sections have been used to discuss the changes of the water table and salinity, which are sections B and C, representing the upper segments and G and H, representing the lower segments. Fig. 13 illustrates the changes of the water table and salinity (TDS) at sections B, C, G, and H after the eight water diversions and the data from May, 2000 represents the inventories of pre-diversion.

The variations of water tables have the following characteristics: (1) before the water diversion, the water tables were similar at each section and the flow field was stable; (2) after the water

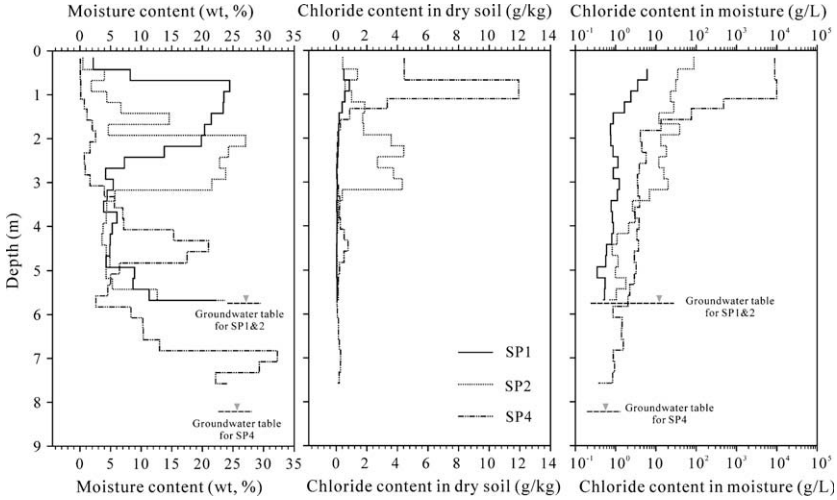

Fig. 12. Moisture and chloride content for the three soil profiles.

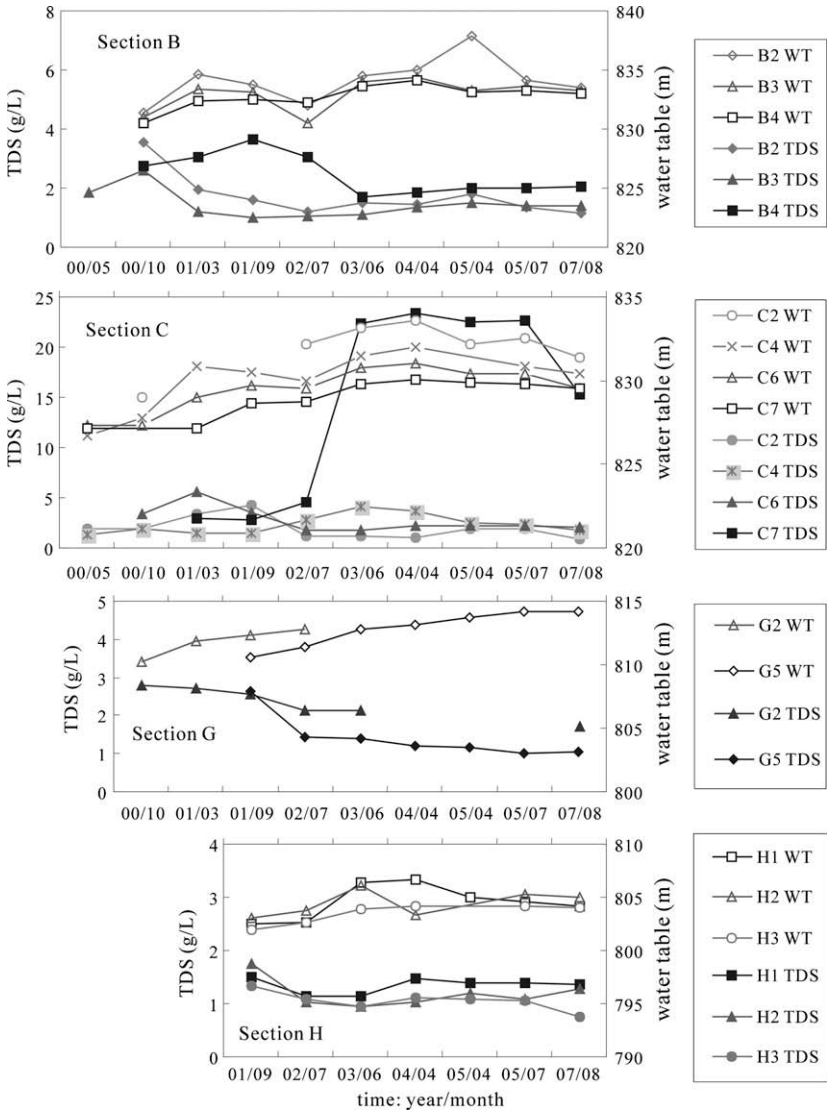

Fig. 13. Water tables and TDS changes at certain groundwater monitoring section in the Lower Tarim River under water diversions.

diversion, the water tables rose to different extent. The closer to the river bed, the faster the groundwater table rose. The unstable infiltration takes place at river bed (Deng, 2009), as a result, the high groundwater hydraulic head moves towards both river-sides. At the beginning of diversion, the water table close to the river bed rose quickly and after the diversion it decreased due to recharge the adjacent aquifer. Compared with the groundwaters far away from the river, the groundwaters close to river have larger changes in water table. Therefore, from the end of one diversion to the beginning of the next one, redistribution of flow field and evapotranspiration result in water table drops. Taking July 2002 (before the fourth diversion) and August 2007 (after the eighth diversion) as an example, water tables at sections B and C all decreased.

Changes of TDS in groundwaters show the following characteristics: (1) TDS in groundwater from sections B and C has increased after the first diversion; (2) after the second and third water diversions, TDS in groundwaters from all sections has decreased. (3) The TDS for samples C7 (w8, 850 m away from the river) tempestuously increased from 4.57 g/L to 22.37 g/L after the fourth water diversion, to 15.32 g/L in August, 2007.

The TDS in groundwater at certain location is affected by TDS in the diverted water and of the antecedent groundwater, salt content of aquifer and the unsaturated zone as well as groundwater movement in the Lower Tarim River under impulsive water diversions. The TDS in water from the Daxihaizi Water Reservoir was as high as 4.2 g/L and 3.7 g/L before water diversion and during the first diversion, respectively (Fig. 14), due to dry and salty river bed for long time, as a consequence of the increased TDS close to the river at sections B and C after the first water diversion. As diverted water continually reached to the Daxihaizi Water Reservoir, the TDS decreased to 1–2 g/L (Fig. 14). The recharging diverted water with relative low TDS diluted the groundwater during the later water

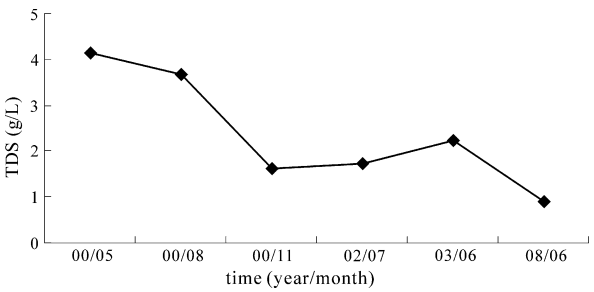

Fig. 14. TDS changes for the Daxihaizi Water Reservoir during water diversions.

diversions. The varying TDS in groundwater at sections B and C had a general decreasing trend except for sample C7 (w8). The tritium contents are between 18.9 and 46.3 TU for B2, B3, B4 at sections B and C2, C4 and C6 at section C, which belong to modern water and suggest some of mixing between the diverted water and the antecedent groundwater. The water table for sample C7 has increased caused by water head pressure transfer rather than diverted water molecules recharge, since groundwater at C7 is pre-modern water with tritium content of 1.1 TU. From July 2002 to June 2003, the water table had increased about 1 m while the TDS increased from 4.57 to 22.37 g/L for samples C7 dramatically (Fig. 13). The increasing TDS was very likely caused by dissolving salt in unsaturated zone as water table rose, something discussed in detail in Section 4.3.3.

For section G, tritium content for G2 (50 m away from the river bed) is 19.6 TU and G5 (800 m away from the river bed) is 2.4 TU as pre-modern water. The water tables of both boreholes have risen. The TDS for G2 decreased rapidly at the earlier stages, and stabilizing at 1 g/L. The rising water table of G5 is also caused by water head pressure transfer. The decreasing TDS for G5 suggests that the closer to the river bed, the lower TDS in groundwater is. The groundwater was recharged by the Tarim River with low TDS in the past, when groundwater depth was deeper (more than 8 m) and evapotranspiration was less intensive before the water diversions. Section G is the confluence reaches of the old Tarim River and the Tarim River (Fig. 4), therefore relative fluent runoff condition and deep groundwater depth of about 11 m before diversion with little evaporation mostly respond for low TDS in groundwater. As a result, the decreasing TDS at section G occurs under water diversion.

The same process occurs at section H. Boreholes H1 and H2, 100 m and 300 m away from the river, with tritium content of 29.0 TU and 30.0 TU, respectively, have a variation from 0.94 to 1.75 g/L in TDS. The water table for sample H3 with tritium content of 4.1 TU has risen about 2 m. Meanwhile, TDS has a little decreasing trend, as well as for sample G5.

4.5. Groundwater table changes relative to vegetation

The variation of groundwater table between that before water diversion (05/2000) and after the eighth the water diversion (08/2007) and the groundwater depth (08/2007) are plotted in Fig. 15 using the Inverse Distance Weighting Method. Groundwater depth in three boreholes (E5, I2, I3), located far way from the river and in the lower segments, has unusually increased, most probably being high water table observation (the perched water) before the diversion since the water table were persistently decreasing during water diversion. The left groundwater depths are all decreasing via the implementation of eight intermittent water diversions, water tables rising by more than 1 m within 700 m from the river in the upper segments to 300 m in the lower (Fig. 15a). Greater water tables rising occurs nearby the river channel, which can reach 6.7 m (G2). The scopes of water tables rise are wider than modern recharge limit, due to water head pressure transfer. Since the vegetation coverage exhibits a continued declining trend with dropping water table (Hao et al., 2010) and modern water recharge scope (Pang et al., 2010) in the area, Niu and Li (2008) concluded there was 7345 hm^2 increased vegetation area from 2000 to 2006 after water diversion based on satellite images of the main area where vegetation distributed in the Lower Tarim River. However, the groundwater depth suitable for the growth of *P. euphratica* is less than 5 m in the Lower Tarim River, according to the relationship among groundwater quality (Chen et al., 2008b; Xu et al., 2007) and proline accumulation (Chen et al., 2003). If the target of the ecological restoration is to maintain the vegetation dominated by *P. euphratica*, the optimal groundwater depth would be less than 5 m. However, the territory with the suitable groundwater depth is narrow, within 200 m from the river bed and becomes even narrower towards downstream (Fig. 15b), far way from most of the existing ecosystem at 1000–1500 m. The desert riparian ecological forest formed in the historical period under larger river recharge is still diminishing.

According to the Program of 'Recent Tarim River Basin Comprehensive Management' announced by the Xinjiang Government and Ministry of Water Resources of China in 2001 (XGMWRC, 2002), the aims of ecosystem recovery are to convert 22,000 ha agriculture filed to natural ecosystem in the Upper/Middle Tarim River, to reduce the area of high water consuming crop (e.g. paddy), to ensure that the runoff to the Daxihaizi Water Reservoir increases to 3.5×10^8 m^3/yr and the diverted water flows into the Taitema Lake, so as to improve the eco-environment in the Lower Tarim River. Although groundwater depth, water salinity, status of the riparian vegetation has measurable positive changes, it is still far away from meeting the requirements of the suitable groundwater depth (less than 5 m within riparian zone of 1000–1500 m from the river bed) for the survival of the existing *P. euphratica* and *Tamarix ramosissima,* the main species targeted by the rescue effort in the Lower Tarim River. As the groundwater depth less than 5 m is spatially limited to within 200 m from the river bank, and it narrows down towards downstream, long-term stability of the ecosystem cannot be achieved by the current water diversion scheme. Due to climate variation (e.g. the runoff of Aral in the Tarim River deceased by

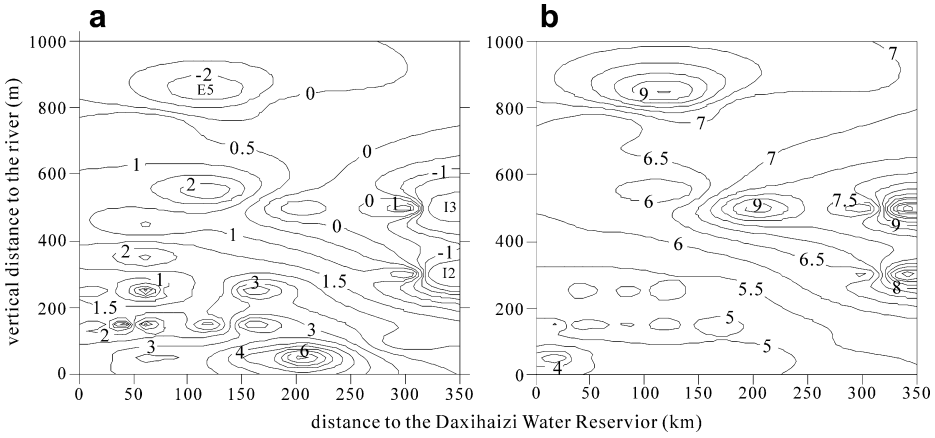

Fig. 15. The variation of groundwater table (m) before water diversion and after the eighth water diversion (a) and groundwater depth (m) in the Lower Tarim River (August, 2007).

~50% from October 2007 to September 2008), not decreased agricultural area and not saving water in source-stream and the Upper/Middle Tarim River, the water diversion project to the Lowe Tarim River has been interrupted, especially since 2008. The embarrassment for the water diversion is mainly caused by imbalance water utilization between upper streams and lower streams, unfair water allocation between eco-environmental water requirement and agriculture water use. This is not only a scientific, but also a socioeconomic issue in the Tarim River, since the Program has given more specific investment and relevant measurements.

5. Conclusions

Stable isotopes, tritium and water chemistry in the Lower Tarim River provide insight into basic hydrological and geochemical processes. The isotopic composition of shallow groundwater forms a trend line that is almost in parallel to the GMWL but is enriched in heavy isotopes compared with the recharging river water. This can be attributed to evaporation during the river recharge to the riparian groundwater system in a rather uniform manner. Tritium data show that the extent of modern recharge (since 1960) is limited to 600–200 m from the river bank with a descending trend towards downstream. The Lower Tarim River is the terminal for both water and salt of the basin, and the dissolution of salts, such as evaporites and carbonate minerals, is the main geochemical process controlling groundwater salinity.

Water table has risen after the eight water diversions except for three unusual boreholes where perched water probably exists. The scope with more than 1 m of water table rise is from 700 m from the riverbank in the upper segments to 300 m in the lower segments. Groundwater salinity at certain locations is mainly affected by the salinity of the diverted river water and of the local antecedent groundwater, salts in the unsaturated zone, evapotranspiration during the diverted water relocation. The TDS of groundwater has generally decreased after the water diversions.

The thirsty Lower Tarim River needs more water. Regulating and saving water in the Upper/Middle Tarim River is crucial for continuing water diversion. Furthermore, monitoring of groundwater should be continued to assess how fast the 2.27 billion m^3 of water diverted will evaporate and how the hydrological regime will change. Extend investigations, especially the soil moisture, soil salt content, stable isotopes and tritium should be deployed to study the hydrological, geochemical and biological processes to evaluate water and soil conditions simultaneously and to test various plant communities for anti-salt and anti-aridity species.

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