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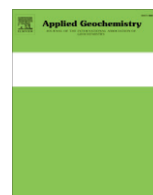
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The role of deuterium excess in determining the water salinisation mechanism: A case study of the arid Tarim River Basin, NW China

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ABSTRACT

Understanding the water salinisation mechanism is the basis for regional salt management. Mineral dissolution, evaporation and transpiration are the main factors controlling natural water salinity in arid inland basins; however, the two are difficult to differentiate. Because deuterium excess decreases during evaporation and is unrelated to the isotopic composition of the initial water, it is a potential tool for determining the contribution of the evapoconcentration of a given water body using the relationship between deuterium excess and salinity rather than between $\delta^{18}\text{O}$ (or $\delta^2\text{H}$) and salinity. In this paper, the relationship between the residual water fraction and deuterium excess was derived from the Rayleigh distillation equation. The contribution of evapoconcentration and mineral dissolution and/or transpiration for a given water body can be determined by comparing the residual water fraction and salinity between the initial water and the evapoconcentrated water. The extremely arid Tarim River Basin in NW China is taken as an example to demonstrate deuterium excess and salinity evolution from the source stream to river water, lake/reservoir water and groundwater. The results show that mineral dissolution contributes most of the salinity (67–77%) for Boston Lake and the Kongque and Tarim rivers relative to the source stream. Mineral dissolution and/or transpiration contribute greater salinity (73–99.6%) to the groundwater recharged by the river water in the middle and lower reaches of the Tarim River. The study provides a method for determining the salinisation mechanism and is important for salt movement and management.

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1. Introduction

As the total area of arid and semiarid regions accounts for 1/3 of the continental area globally, water salinisation is a major environmental issue in water resources, partly due to river regulation, land-use changes, irrigation and groundwater exploitation (Williams, 1999; Gaye, 2001). An understanding of the chemical evolution and salinisation mechanism of groundwater can provide insight into the interaction between water and the environment and can contribute to rational water resource management (Adams et al., 2001; Edmunds, 2009), especially in (semi)arid regions with a fragile ecosystem and intense anthropogenic interference (Bennetts et al., 2006). In the Nile delta, for example, an evaporation experiment (Simpson et al., 1987) showed that the increase in groundwater salinity is mainly caused by plant transpiration (unusable salts rejected by plants), and changes in the irrigation pattern to reduce water loss did not reduce the salinity. Land-use and hydrological regime changes have resulted in the mobilisation of large amounts of stored salts in the Murray River Basin, Australia (Allison et al., 1990; Simpson and Herczeg, 1991).

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Groundwater chemistry is also important for ecosystem restoration because water quality can control vegetation growth and soil characteristics if the salinity is high enough. The negative impacts of high-salinity water on vegetation growth include preventing vegetation from absorbing moisture and decreasing soil fertility (He et al., 2006; Manchanda and Garg, 2008). Plant features, including the richness, diversity and composition of species are significantly related to salinity, and deteriorate with increasing salinity (LyMBERG et al., 2003). Water quality issues can also emerge, threatening and further diminishing the amount of renewable water resources in (semi)arid areas (Edmunds, 2003).

In arid areas, mineral dissolution, evaporation and transpiration are the main salinisation mechanisms for water bodies in inland plain areas (Hao et al., 2000). However, quantifying the contribution of each of these factors to the total salinity is difficult, and no simple and reliable methods are currently available (Hao et al., 2000). Because evaporation enriches water bodies with heavy isotopes, some studies have used the relationship between $\delta^{18}\text{O}$ and $\delta^2\text{H}$, and $\delta^{18}\text{O}$ (or $\delta^2\text{H}$) and salinity to determine the sources of salinity (Gaye, 2001) and the evapoconcentration effect (Simpson et al., 1987; Hao et al., 2000; Phillips et al., 2003). For example, when a water body loses 1% of its mass relative to the initial water, the $\delta^2\text{H}$ in the remaining water can increase by 0.65‰, according to

an evaporation experiment in Egypt conducted by Simpson et al. (1987). In the study of two seasonal lakes in the northwestern Sahara, Fontes and Gonfiantini (1967) concluded that when the water bodies lost 1% of their mass relative to initial water, the $\delta^2\text{H}$ in the remaining water increased by 0.78‰. This value was 0.62‰ in the Murray River, Australia (Simpson and Herczeg, 1991). Thus, the contribution of evapoconcentration to salinity can be determined by the mass loss ratio.

However, the aforementioned method has some limitations, especially when the isotopic compositions of the initial water vary. A fixed isotopic composition at the basin scale is not common because precipitation with different isotopic compositions (monthly variations and spatial differences) may recharge rivers or groundwaters. For example, when water samples (river or groundwater) A2 and C2 with the same value of $\delta^{18}\text{O}$ are obtained, as shown in Fig. 1, it is difficult to identify the water loss and salinisation mechanism by the relationship between $\delta^{18}\text{O}$ and salinity unless their initial water (precipitation or source water, A1 and C1) is known. The water samples A2 and B2 show different values of $\delta^{18}\text{O}$, however, the mass loss ratios are same.

In addition to evaporation, mineral dissolution and/or transpiration are important factors controlling the salinity of water bodies in arid areas. Because deuterium excess (Dansgaard, 1964) decreases during evaporation and is unrelated to the isotopic composition of the initial water, it can be used to assess the effect of evaporation. The remaining factors, i.e. mineral dissolution and/or transpiration, cannot be distinguished by using stable isotopes. This is important for water resource and salt management, at least in arid areas where surface reservoirs are commonly used and a large amount of the water loss from a reservoir is caused by open-water evaporation. This scenario is especially true in Xinjiang and the Hexi Corridor, NW China. To the present authors' knowledge, this study is the first to use deuterium excess to quantify the salinisation mechanism. The extremely arid Tarim River Basin (TRB) is used as an example to demonstrate the main factors controlling groundwater salinity and how deuterium excess is useful in determining the salinisation mechanism. The data are taken from the authors' previous papers (Pang et al., 2010; Huang and Pang, 2010), which mainly focused on diminished groundwater recharge and circulation relative to degradation of riparian vegetation in the Middle Tarim River (MTR) as well as the response of groundwater to water diversion and the scope of modern recharge in the Lower Tarim River (LTR).

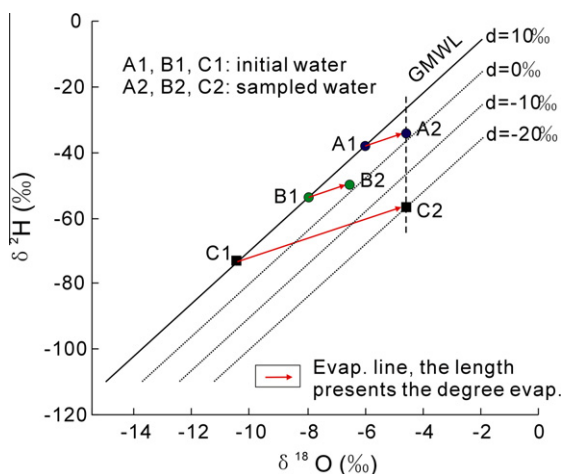


Fig. 1. Changes in stable isotopic composition and deuterium excess during evaporation.

2. Methodology

Globally, the stable isotopic composition of precipitation can be described by the following equation:

$$\delta^2\text{H} = 8\delta^{18}\text{O} + 10 \quad (1)$$

When precipitation undergoes evaporation on recharging rivers, lakes and groundwater through soil water infiltration, the slope of the evaporation line will be less than 8, commonly 4–6 for open water and even 2–3 for soil evaporation (Barnes and Allison, 1988). However, there is no isotopic fractionation during salt drainage, plant transpiration (Zimmermann et al., 1967; Foerstel, 1982) or mineral dissolution. Because vegetation usually excludes salt during water uptake, transpiration and mineral dissolution will increase the salinity of the residual water without isotopic fractionation. When the groundwater depth is deep and stable, the groundwater persistently evaporates from capillary moisture, causing no isotopic fractionation in the remaining groundwater. Therefore, obvious isotopic fractionation is only evident for open water evaporation compared with groundwater, especially for lake and water reservoirs.

In addition, one cannot use the relation between groundwater salinity and $\delta^{18}\text{O}$ (or $\delta^2\text{H}$) directly to examine the evaporation process because the isotopic compositions of initial water (precipitation or mountain rivers) exhibit seasonal variation. For example, the amount of precipitation was 38 and 43 mm for April and July, respectively, in Wulumuqi (GNIP station near Tianshan Mountain); however, the $\delta^{18}\text{O}$ was -11.6‰ and -5.9‰ , respectively (IAEA and WMO, 2010).

Dansgaard (1964) proposed the use of d to describe the deuterium excess for global precipitation. The value d is defined for a slope of 8 and is calculated for any water sample as

$$d = \delta^2\text{H} - 8\delta^{18}\text{O} \quad (2)$$

Globally, d averages 10‰ for precipitation. When a water body undergoes evaporation, the deuterium excess will decrease and the salinity will increase with a negative correlation.

When combining the δ notation (Eq. (3)) and Rayleigh distillation (Eq. (4)), the relation between deuterium excess (d) and the remaining fraction of that reservoir (f) can be obtained by the definition of deuterium excess (see Eq. (6)).

$$\delta = (R/R_{\text{VSMOW}} - 1) \times 1000 \quad (3)$$

$$R = R_0 f^{(\alpha_{v-l}-1)} \quad (4)$$

$$\delta = \left(\frac{R_0 f^{(\alpha-1)}}{R_{\text{VSMOW}}} - 1 \right) \times 1000 = \left[\left(\frac{\delta_0}{1000} + 1 \right) f^{(\alpha-1)} - 1 \right] \times 1000 = (\delta_0 + 1000) f^{(\alpha-1)} - 1000 \quad (5)$$

$$d = \delta^2\text{H} - 8\delta^{18}\text{O} = (\delta_0^2\text{H} + 1000) f^{(\alpha^2\text{H}-1)} - 8(\delta_0^{18}\text{O} + 1000) f^{(\alpha^{18}\text{O}-1)} + 7000 \quad (6)$$

where δ is the delta notation; R is the isotopic ratio; f is the remaining fraction of the reservoir; $\alpha_{(v-l)}$ is the fractionation factor between product (vapour) and reactant (liquid water), including equilibrium and kinetic fractionation; and $\delta_0^{18}\text{O}$ and $\delta_0^2\text{H}$ are the δ values for initial water. The total isotopic fractionation factor α_{l-v} ($1/\alpha_{v-l}$) for open water and air equals the sum of the fractionation factor for the equilibrium water-vapour exchange (ϵ_{l-v}) and the kinetic factor ($\Delta\epsilon_{bl-v}$) (Clark and Fritz, 1997):

$$\alpha_{l-v} = \epsilon_{l-v} + \Delta\epsilon_{bl-v} \quad (7)$$

The equilibrium water-vapour exchange (ϵ_{l-v}) is only dependent on temperature (T , Kelvin scale) (Majoube, 1971) and can be obtained by

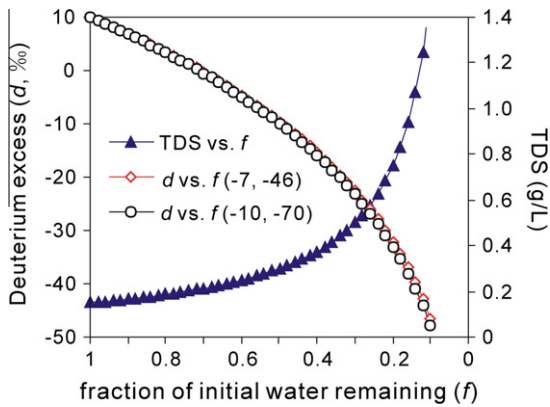


Fig. 2. Deuterium excess (d) and TDS versus remaining fraction of reservoir (f) during water evaporation with an initial water d of 10‰ and TDS of 0.15 g/L at 25 °C and 75% relative humidity.

$$10^3 \ln \epsilon_{l-v}^2\text{H} = 24.844(10^6/T^2) - 76.248(10^3/T) + 52.612 \quad (8)$$

$$10^3 \ln \epsilon_{l-v}^{18}\text{O} = 1.137(10^6/T^2) - 0.4156(10^3/T) - 2.0667 \quad (9)$$

Gonfiantini (1986) gave the following equations to determine the kinetic factor ($\Delta\epsilon_{bl-v}$):

$$\Delta\epsilon^2\text{H}_{bl-v} = 12.5(1-h)/1000 \quad (10)$$

$$\Delta\epsilon^{18}\text{O}_{bl-v} = 14.2(1-h)/1000 \quad (11)$$

At a given evaporation condition (temperature T and relative humidity h), the relationship between d and f can be obtained by Eq. (6). Here, the unknown $\delta_0^{18}\text{O}$ and $\delta_0^2\text{H}$ in Eq. (6), which do not affect the relationship, are not of concern as their deuterium excess values can be fixed for the initial water in a mountain area (average of ~10‰). If the relative humidity is unavailable, it can be obtained from the relationship between the slope of the evaporation line and the relative humidity (Gonfiantini, 1986). The relative humidities of 95%, 75%, 50%, 25% and 0% correspond approximately to slopes of 6.8, 5.2, 4.5, 4.2 and 3.9, respectively.

At a temperature of $T = 25\text{ °C} = 298.15\text{ K}$ and a relative humidity of $h = 75\%$, two water samples with different isotopic ($\delta^{18}\text{O}$, $\delta^2\text{H}$) compositions (-7‰ , -46‰) and (-10‰ , -70‰) with the same TDS of 0.15 g/L will undergo evaporation. Fig. 2 illustrates the d , f and TDS changes during evaporation. The results show that the

relationship between d and f is not related to the initial isotopic composition. When half the water is evaporated, the salinity will double due to evapoconcentration, and the deuterium excess will decrease from 10‰ to -9.7‰ .

When a given water body with a salinity of S_0 undergoes evaporation and the remaining fraction of the water body is f , the evapoconcentration makes the salinity increase to S_0/f and evaporation contributes a salinity of $(S_0/f - S_0)$. The contribution of mineral dissolution and/or transpiration (D) contributes the total salinity S (Fig. 3). The contribution ratio of evapoconcentration to the total salinity is $(S_0/f - S_0)/S = S_0(1-f)/(S \cdot f)$, and the contribution of the mineral dissolution and/or transpiration to the total salinity is $(S - S_0/f)/S = 1 - S_0/(S \cdot f)$, respectively.

3. Case study

3.1. Study area and data source

The TRB is located in the south of Xinjiang, NW China (Fig. 4) and is a typical inland basin for examining the evolution of salinity and deuterium excess from source streams, rivers, lake/water reservoirs and groundwater. The Middle and Lower Tarim River is dominated by a typical continental temperate arid climate. According to the Tikanlik meteorological station (TG in Fig. 4) data, the average precipitation is approximately 40 mm/a, the potential evaporation is 2590 mm/a, and the annual average temperature is 10.5 °C.

The groundwater samples are from five groundwater monitoring transects in the middle reaches of the Tarim River and nine transects in the lower reaches, namely, Shajilike (MA), Shazihe (MB), Wusiman (MC), Aqike (MD), Tiejizi (ME), Akdun (A), Yahopumarhan (B), Yengsu (C), Abudali (D), Karday (E), Tugmailai (F), Aragan (G), Yikanbujima (H) and Kargan (I). The surface water includes source streams (K1, K2, K3, TL in the Upper Aksu River and R11 in the Kaidu River), rivers (A1 in the Middle Aksu River, T1 and R6–R8 in the Tarim River and R10 in the Kongque River, which are defined as runoff below the mountain front in this study), lakes and water reservoirs (Boston lake (BL): L1, L4, L10; the Qiala (QL) and Daxihaizi (DX) water reservoirs) (Fig. 4). All samples were measured for stable isotopes and water chemistry. The results were published in the authors' previous papers (Pang et al., 2010; Huang and Pang, 2010) and can be found in Table S1. These stable isotopic data and the water chemistry are used to discuss the salinisation mechanism in the TRB.

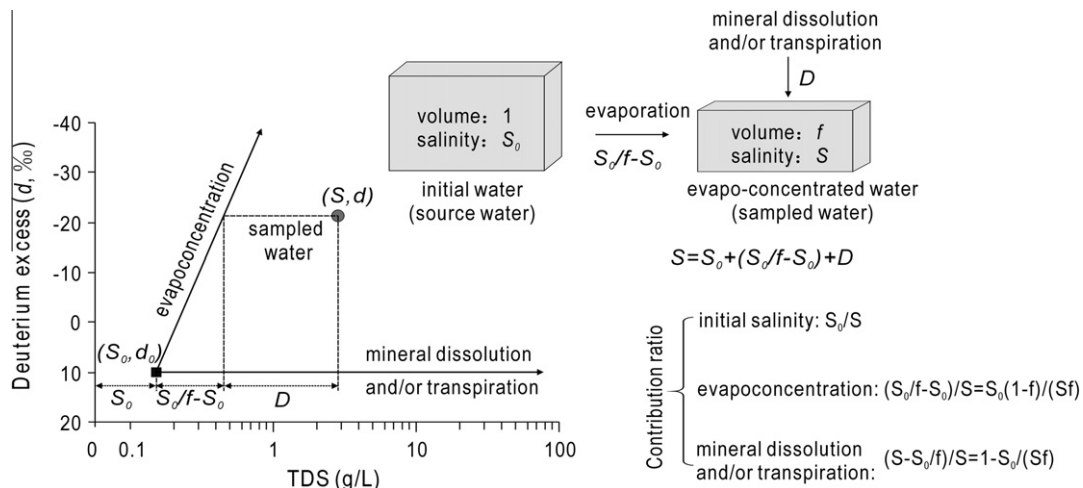


Fig. 3. The contribution of evapoconcentration and mineral dissolution (and/or transpiration) to total salinity using the deuterium excess method.

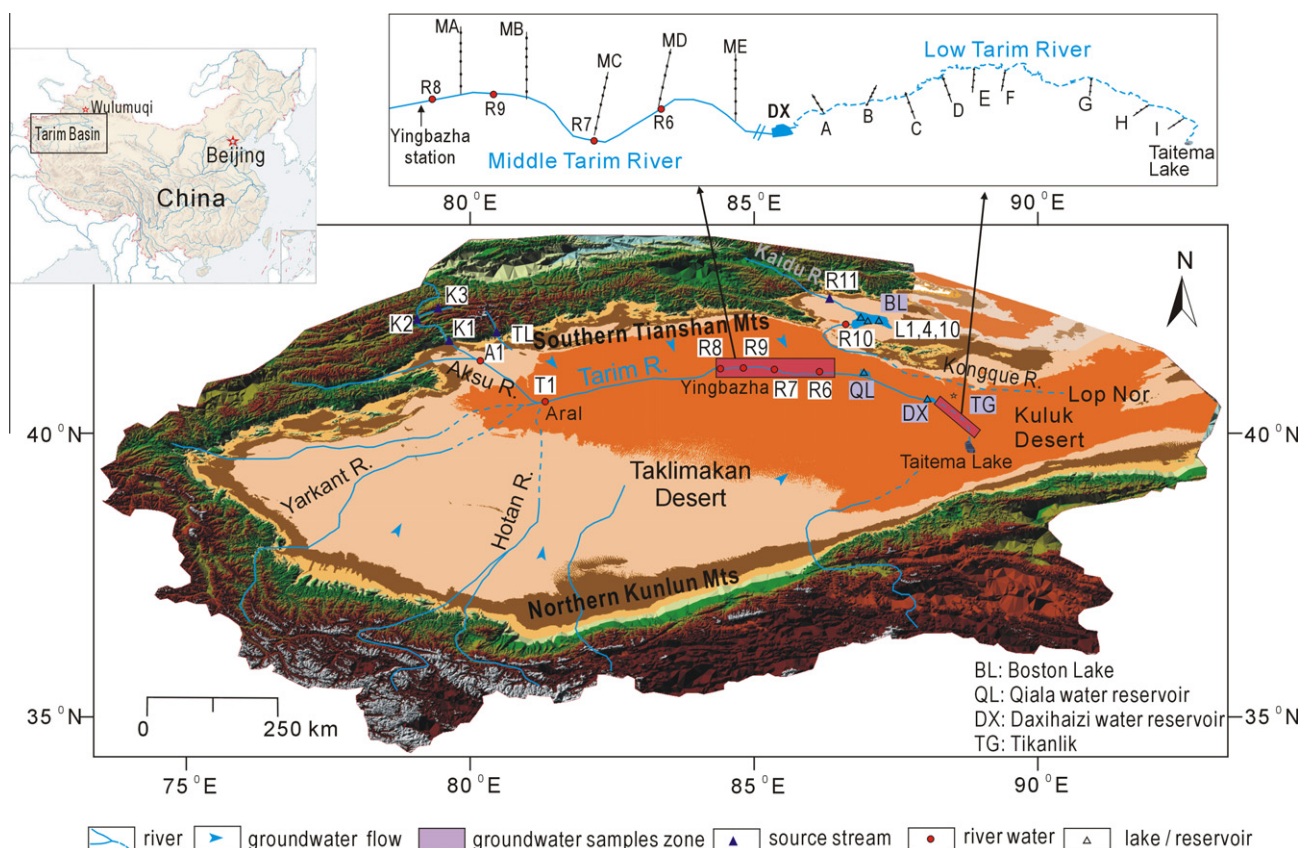


Fig. 4. The Tarim River Basin and sampling locations. The detailed distribution of groundwater samples can be found in Fig. 6.

Table 1
The statistics of salinity and deuterium excess of the four types of water in the Tarim River Basin.

	TDS (g/L)			Deuterium excess (‰)		
	Min.	Max.	Average	Min.	Max.	Average
Source stream	0.129	0.181	0.144	8.4	10.8	9.9
River	0.217	0.666	0.577	5.1	8.6	7.2
Lake and reservoir	0.754	1.053	0.877	-13.6	-4.2	-8.7
Groundwater	0.500	55	2.88 ^a	-0.1	5.2	2.3

^a Not including the extremely high value (55 g/L) at section I.

3.2. Salinity and d value

The source streams have very low TDS salinity, ranging from 0.129 to 0.181 g/L, with an average of 0.144 g/L (Table 1). Although the stable isotopes vary widely from -11.2‰ to -7.8‰ for δ¹⁸O and -78.9‰ to -54.0‰ for δ²H, respectively, the deuterium excess of the source streams shows inconspicuous variation, with an average of 9.9‰, which is very similar to the global average of 10‰ (Dansgaard, 1964). The rivers show a slight decrease in deuterium excess, with an average of 7.2‰, compared with the source streams, and the TDS of the rivers increases to an average of 0.577 g/L. However, the stable isotopes in Boston Lake and the Qiala and Daxihaizi reservoirs are significantly enriched (Fig. 5), with an average deuterium excess of -8.7‰.

The groundwaters show a wide range of stable isotopic composition from -9.7‰ to -5.2‰ for δ¹⁸O and from -75.9‰ to -41.1‰ for δ²H, respectively. The regression line with a slope of 7 is sub-parallel to the Global Meteoric Water Line (GMWL) (Craig, 1961). The phenomenon is commonly observed in dry climates (e.g., Chapman et al., 2003; Huang et al., 2012) and is attributed to the

same evaporation capacity from river water with different isotopic compositions. The deuterium excesses of groundwaters range from -0.1‰ to 5.2‰, with an average of 2.3‰, and are relatively clustered (Fig. 5). However, the TDS of all groundwater varies widely, ranging from 0.5 to 55 g/L (Table 1, Fig. 6).

Tritium is used to distinguish modern groundwater (recharge occurring since 1960) from pre-modern groundwater (recharge occurring prior to 1960). The ³H input sequence of precipitation in the study area was reconstructed by Huang and Pang (2010). The ³H input in the study area is relatively high. The ³H content for current precipitation was as high as 29.8 TU in 2007, and for surface water (rivers, lakes, water reservoirs), it ranges from 18.4 to 28.3 TU (Table S1). The decayed results show that groundwater a ³H with content below 10 TU should have been recharged before 1960, or at least most of it is pre-modern water when mixing with modern water is considered. However, the ³H content for pre-modern groundwaters in the study area mostly ranges from <0.3 to 3 TU, while it ranges from 18.2 TU to 110.4 TU for modern groundwaters (Table S1). The extremely high ³H contents of 94.4 and 110.4 TU in the MTR and 85.3 TU in the LTR must represent bomb-pulse water from aboveground thermonuclear tests in the 1960s. The modern groundwaters are relatively low in salinity of TDS, ranging from 0.8 to 3.1 g/L (Fig. 6). These modern groundwaters are located near rivers and can receive river recharge, in which the salinity is relatively low, with a TDS less than 1 g/L (Table 1). However, the TDS of pre-modern groundwaters has a wide range, from 0.5 to 15.3 g/L, except for section I. Section I is located near the Taimema Lake, which is the terminal of water and salt for the whole inland basin. Long-term mineral dissolution and evapotranspiration cause high salinity in the local groundwater. Beyond the riverbank (1000–2000 m), the TDS in the groundwater is greater than 5 g/L and more than 10 g/L in the lower segments of the lower reaches (Cheng, 1993).

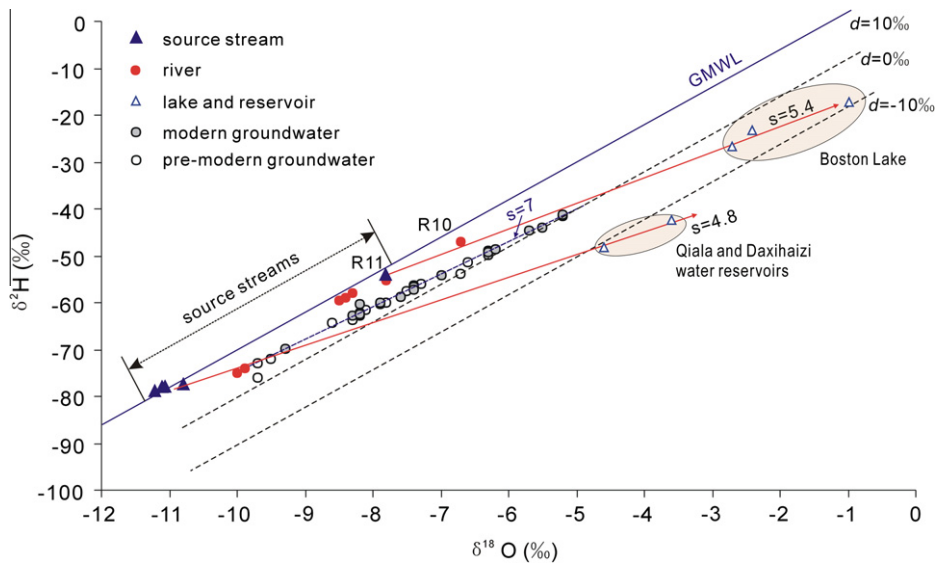


Fig. 5. Stable isotopic composition for four types of water (source stream, river, lake/water reservoir, and groundwater) from the Tarim River Basin.

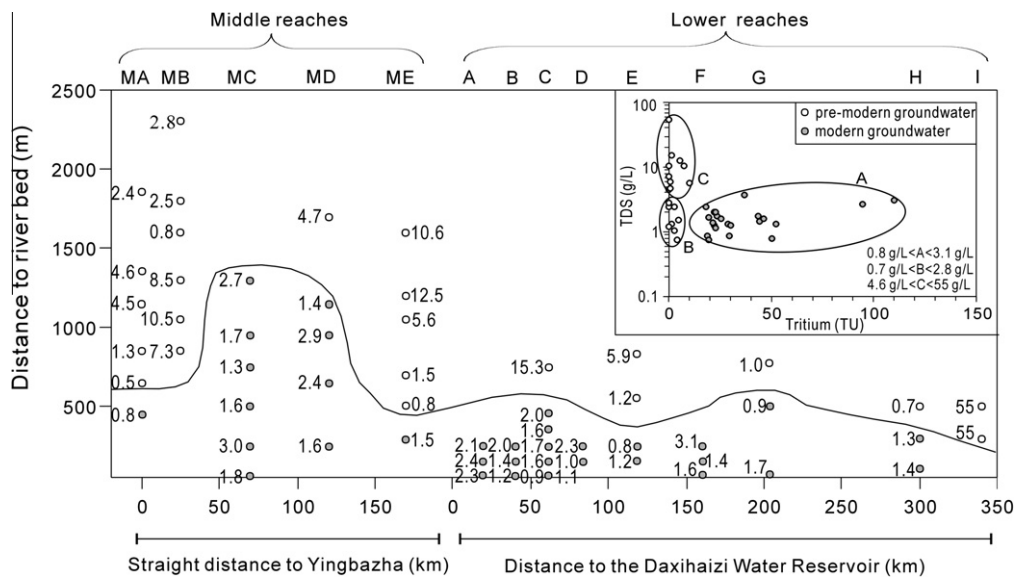


Fig. 6. TDS distribution (g/L) and the relationship between ³H content and TDS of groundwater along the Middle and Lower Tarim River. The black solid line represents the boundary of modern and pre-modern groundwater based on ³H.

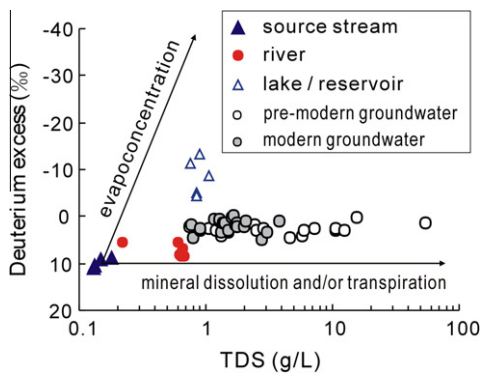


Fig. 7. Relationship between deuterium excess and TDS for surface waters and groundwaters in the Tarim River Basin.

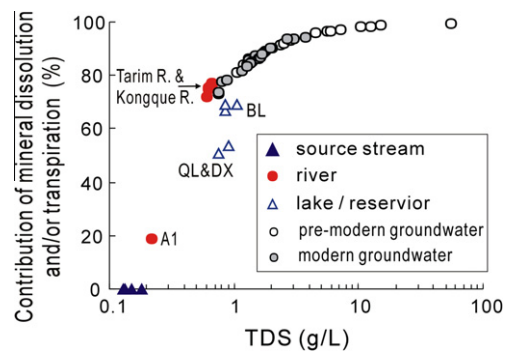


Fig. 8. Contribution of mineral dissolution and/or transpiration to total salinity for surface waters and groundwaters in the Tarim River Basin.

Table 2

Deuterium excesses (d), salinity (TDS), the remaining fraction (f), evapoconcentration ($S_0/f - S_0$) contribution and ratios, mineral dissolution (D) contribution and the ratio for river, lake, water reservoir and groundwater in the Tarim River Basin.

Sample	d (‰)	TDS (g/L)	f	Evapoconcentration		Mineral dissolution	
				$S_0/f - S_0$ (g/L)	Ratio (%)	D (g/L)	Ratio (%)
<i>River</i>							
A1	5.1	0.217	0.82	0.032	14.6	0.041	19.1
T1	5.5	0.608	0.84	0.027	4.5	0.437	71.8
R8	8.5	0.666	0.94	0.009	1.4	0.513	77.0
R9	8.5	0.642	0.94	0.009	1.4	0.489	76.1
R7	7.5	0.631	0.91	0.014	2.3	0.473	74.9
R6	8.6	0.617	0.95	0.008	1.2	0.465	75.4
R10	6.9	0.661	0.89	0.018	2.7	0.499	75.5
<i>Lake and water reservoir</i>							
L1	−9.1	1.053	0.44	0.183	17.4	0.726	68.9
L4	−5.3	0.840	0.52	0.133	15.8	0.563	67.0
L10	−4.2	0.844	0.55	0.118	14.0	0.582	69.0
Qjala	−11.5	0.754	0.39	0.225	29.9	0.385	51.0
Daxihaizi	−13.6	0.894	0.35	0.267	29.9	0.483	54.0
<i>MA: Shajilike groundwater wells</i>							
W56	4.6	0.794	0.81	0.034	4.3	0.616	77.6
W58	4.0	1.318	0.79	0.038	2.9	1.136	86.2
W60	4.8	4.56	0.81	0.034	0.7	4.382	96.1
W61	1.7	2.39	0.72	0.056	2.3	2.190	91.6
<i>MB: Shazihe groundwater wells</i>							
W50	3.0	7.26	0.056	0.045	0.6	7.071	97.4
W51	3.2	10.47	0.76	0.043	0.4	10.283	98.2
W54	2.8	2.47	0.75	0.048	1.9	2.278	92.2
W55	2.3	2.82	0.73	0.053	1.9	2.623	93.0
<i>MC: Wusiman groundwater wells</i>							
W44	2.1	1.773	0.73	0.053	3.0	1.576	88.9
W45	3.5	3.02	0.77	0.043	1.4	2.833	93.8
W47	3.3	1.322	0.77	0.043	3.3	1.135	85.9
W49	5.2	2.70	0.83	0.029	1.1	2.527	93.6
<i>ME: Teyizi groundwater wells</i>							
W33	3.1	1.485	0.76	0.045	3.1	1.296	87.2
W35	3.1	1.503	0.76	0.045	3.0	1.314	87.4
W36	4.6	5.64	0.81	0.034	0.6	5.462	96.8
W37	2.8	12.46	0.75	0.048	0.4	12.268	98.5
W38	2.3	10.55	0.73	0.053	0.5	10.353	98.1
<i>B: Yahopumarhan groundwater wells</i>							
w4	1.0	1.284	0.69	0.065	5.0	1.075	83.7
w6	1.2	1.383	0.70	0.062	4.5	1.177	85.1
w7	1.5	2.04	0.71	0.059	2.9	1.837	90.1
<i>C: Yengsu groundwater wells</i>							
w14	0.5	3.72	0.68	0.068	1.8	3.508	94.3
w13	2.9	0.884	0.75	0.048	5.4	0.692	78.3
w12	0.0	1.588	0.67	0.071	4.5	1.373	86.5
w11	1.6	1.737	0.71	0.059	3.4	1.534	88.3
w10	−0.1	1.631	0.66	0.074	4.5	1.413	86.6
w9	2.5	1.986	0.74	0.051	2.5	1.791	90.2
w8	−0.1	15.32	0.66	0.074	0.5	15.102	98.6
<i>E: Karday groundwater wells</i>							
w21	0.5	1.16	0.68	0.068	5.8	0.948	81.7
w20	2.1	0.75	0.73	0.053	7.1	0.553	73.7
w19	2.7	1.192	0.75	0.048	4.0	1.000	83.9
w18	3.4	5.9	0.77	0.043	0.7	5.713	96.8
<i>G: Aragan groundwater wells</i>							
w27	1.9	1.693	0.72	0.056	3.3	1.493	88.2
w26	3.1	0.878	0.76	0.045	5.2	0.689	78.4
w25	2.6	1.039	0.74	0.051	4.9	0.844	81.3
<i>H: Yikanbujima groundwater wells</i>							
w32	1.2	1.355	0.70	0.062	4.6	1.149	84.8
w31	0.9	1.279	0.69	0.065	5.1	1.070	83.7
w30	1.9	0.742	0.72	0.056	7.5	0.542	73.0
<i>I: Kargan groundwater wells</i>							
w28	1.5	55	0.056	0.059	0.1	54.8	99.6

3.3. Salinisation mechanism

The slope of the evaporation line for Boston Lake and the Qjala and Daxihaizi reservoirs is 4.8 and 5.4, respectively (Fig. 5), with an

average of 5.1, corresponding to a relative humidity of 75%. Using the temperature of 10.5 °C, the relationship between d and f can be determined by Eq. (6). When source stream water with an average TDS of 0.144 g/L is considered as the initial water, the salinity

caused by evapoconcentration can be estimated. The remaining salinity is contributed by mineral dissolution and/or transpiration. Fig. 7 shows the relationship between the TDS and d for the four types of water in the basin. Fig. 8 and Table 2 show the contribution ratio of mineral dissolution and/or transpiration to the total salinity. For rivers, lakes and water reservoirs, the transpiration effect can be ignored. Mineral dissolution contributes most of the salinity (67–77%) to Boston Lake, the Kongque River and the Tarim River. The exception is sample A1 (Xidaqiao) in the Middle Aksu River, with a TDS of 0.217 g/L and a deuterium excess of 5.1‰, which is only 19%. The mineral dissolution contribution ratio is 51% and 54% for the Qiala and Daxihaizi water reservoirs, respectively. The mineral dissolution and/or transpiration contribute more to the salinity (73–99.6%) in the groundwaters, which are recharged by river water in the middle and lower reaches of the Tarim River.

4. Conclusions

The use of deuterium excess does not require the knowledge of initial stable isotopic composition of water, and is a physically based, effective and reliable approach to determining the contribution of evapoconcentration to salinity. In the TRB, mineral dissolution contributes most of the salinity (67–77%) in Boston Lake, the Kongque River and the Tarim River. For groundwater, the increase in salinity is not accompanied by a significant evaporative isotopic signal. Mineral dissolution and/or transpiration contribute more to the salinity (73–99.6%) in the groundwaters, which are recharged by river water. Thus, the increase of salinity is explained by mineral dissolution, where soluble salt is abundant in the middle and lower reaches of the arid inland basin.

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Appendix A. Supplementary material

Supplementary data associated with this article can be found, in the online version, at <http://dx.doi.org/10.1016/j.apgeochem.2012.08.015>.

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