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ORIGINAL ARTICLE

Characteristics of stable isotopes and hydrochemistry of river water in the Qinghai Lake Basin, northeast Qinghai-Tibet Plateau, China

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Abstract The integrated use of isotopic and hydrochemical tracers is an effective way to investigate hydrological processes on a range of spatial and temporal scales. This study investigated stable isotopes and hydrochemistry of the river water in the Qinghai Lake Basin, and discussed relationships between runoff and variations of air temperature and precipitation. Results indicated that all of the river water points lie close to the local meteoric water line (LMWL); and the slope of local evaporation line of river water samples (6.82) was smaller than that of the LMWL (7.98), indicating that the river water mainly originated from the precipitation in the catchments which underwent weak evaporation. The river water in the tributaries would undergo relatively stronger evaporation than that in the main stream. The hydrochemical type of river water was Ca–Mg–HCO_3, and the river water chemistry was mainly controlled by carbonate dissolution in the Qinghai Lake Basin. The river discharge was generated mainly from the middle and upper catchment. The runoff depths of Buha River Catchment and Shaliu River Catchment were 91.0 mm and 283.4 mm, respectively; and the runoff coefficients were 0.164 and 0.531, respectively. Because of a relatively longer channel, larger drainage area and smaller gradient, the surface water flowed more slowly and infiltrated more in the large river catchment; therefore, the runoff coefficient in the large river catchment was smaller than that in the relatively smaller catchment. The river runoff in the Qinghai Lake Basin was primarily influenced by precipitation. This study provides insights into the hydrological and geochemical processes of cold and alpine rivers, along with water resource management options in the Qinghai Lake Basin and northeast Qinghai-Tibet Plateau.

Keywords Stable isotope · Hydrochemistry · Runoff process · The Qinghai Lake Basin

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Introduction

Because of large areas and complex geological and geomorphic conditions, traditional methods are unsuitable for hydrological studies in high-altitude, cold and arid regions (Liu et al. 2008; Kong and Pang 2012); particularly in the Qinghai-Tibet Plateau, there are very few weather and hydrology observation stations. Recently, the integrated use of isotopic and hydrochemical tracers has emerged as an effective approach for investigating complex hydrological processes on a range of spatial and temporal scales (Pawellek et al. 2002; Phillips et al. 2003; Song et al. 2006; Ryu et al. 2007; Yuan and Miyamoto 2008; Liu et al. 2008; Meredith et al. 2009). For example, Pawellek et al. (2002) analyzed the isotopes and chemical composition of the Danube River from its source to 1,046 km downstream and its major tributaries to investigate the downstream evolution of the river hydrochemistry. Phillips et al. (2003) employed isotopic and hydrochemical tracers to investigate the causes of the salinization in the Rio Grande River. Ryu

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et al. (2007) analyzed the seasonal and spatial variations of the major chemical ions and isotopes of the Han River by monitoring 14–23 locations over a 1-year period, covering approximately 80 % of the entire drainage basin. Meredith et al. (2009) investigated the spatial and temporal hydrological processes in the Darling River between Bourke and Wilcannia using hydrochemical and stable isotope (^{18}O and ^{2}H) tracers. All these studies demonstrated the applicability of isotopes and ions as the conservative tracers for hydrologic studies. However, few studies have been carried out in the high alpine catchments versus the low-altitude regions (Liu et al. 2008; Zhang et al. 2009; Kong and Pang 2012).

The Qinghai Lake Basin, a closed basin with an area of 29,661 km^2 and an average altitude of 3,720 m above the sea level, lies in the cold and semiarid region of the northeast (NE) Qinghai-Tibet Plateau, China (Fig. 1). Qinghai Lake is the largest inland lake in China, with an area of 4,400 km^2 within the basin. The lake is one of China's national nature reserves and is important to the ecological security of the NE Qinghai-Tibet Plateau (Tang et al. 1992). In the recent decades, > 50 % of the rivers flowing into the Qinghai Lake have dried up due to climate change and human activity (LZBCAS 1994; Li et al. 2007); and the lake level has declined from 3,196.55 m in 1959 to 3,193.59 m in 2010, with an average decrease rate of 5.8 cm/year over 51 years (Li et al. 2007, 2012). These hydrological processes have contributed to environmental problems in the basin, such as desertification, loss of grazing grassland, water quality deterioration and a decrease in water supplies (Qin and Huang 1998; Zhang et al. 2003; Hao 2008). However, because of few meteorological and hydrological stations in the Qinghai Lake Basin (Fig. 1), previous studies on the river runoff only investigated the hydrochemistry composition of river water, changes in runoff, or impacts of landuse on the runoff (Qin and Huang 1998; Yan and Jia 2003; Li et al. 2005, 2007, 2009; Yi et al. 2010; Jin et al. 2010a, 2010b; Xu et al. 2010; Xiao et al. 2012). The runoff process and water cycle of the river are not yet completely understood. Therefore, to better understand the runoff process and water cycle of the Qinghai Lake Basin, it is essential to investigate the river water using stable isotopes and hydrochemistry.

The objectives of this study are as follows: (1) to investigate the characteristics of stable isotopes and hydrochemistry of the river water in the Qinghai Lake Basin; (2) to investigate the runoff characteristics based on isotope tracer techniques; and (3) to evaluate the sensitivity of river runoff to climate changes. The results may provide

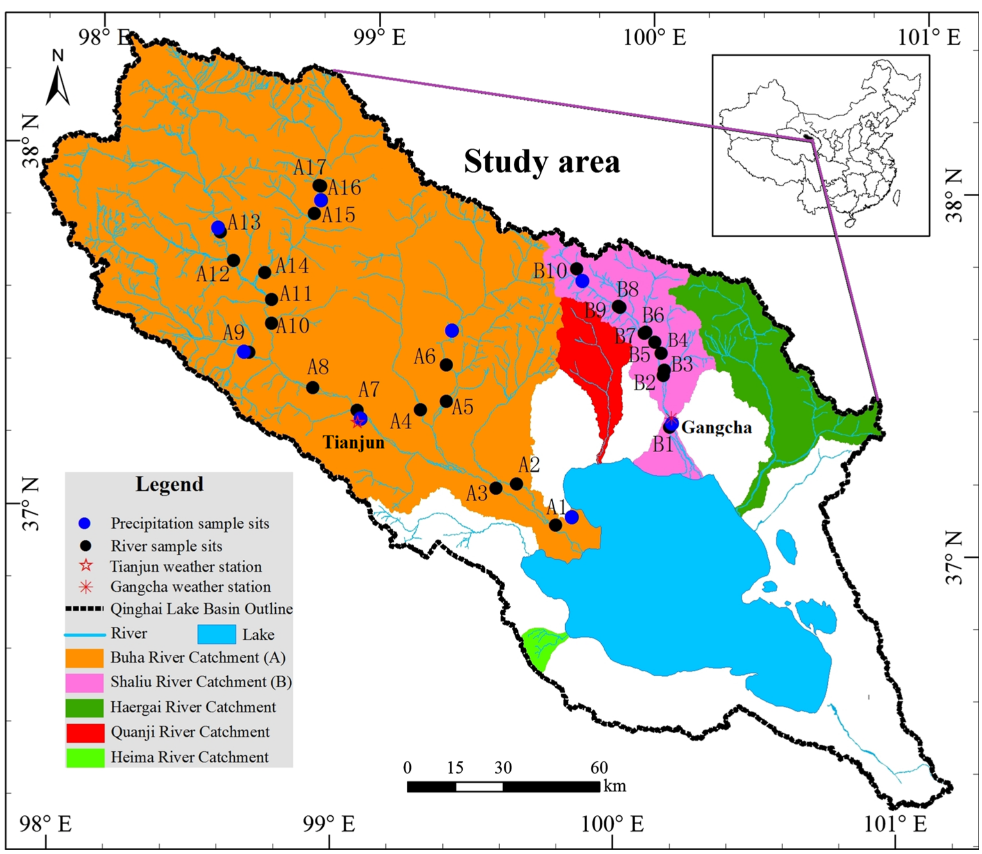

Fig. 1 Location of the Qinghai Lake Basin and sampling sites

insights into the hydrological and geochemical processes of cold and alpine rivers, along with water resource management options in the basin and NE Qinghai-Tibet Plateau.

Background of the rivers in the Qinghai Lake Basin

The Qinghai Lake Basin is a closed basin with an area of 29,661 km^2, and there are >50 rivers or streams flowing into the Qinghai Lake (LZBCAS 1994). These rivers are mainly located on the north and northwest of the lake, resulting in an asymmetric distribution in the Qinghai Lake Basin (Fig. 1; Yan and Jia 2003). Because > 77 % of the precipitation occurs during the monsoon season (June–October) (Cui 2011), most of the rivers are seasonal rivers; 85 % of the annual discharge occurs between June and September. The main rivers are Buha, Shaliu, Haergai, Quanji and Heima Rivers (LZBCAS 1994; Li et al. 2007), with a total discharge of 1.43×10^9 m^3/year (1960–2010) accounting for over 80 % of the total volume of the surface water flowing into the Qinghai Lake (Table 1).

The Buha and Shaliu Rivers are the two largest rivers in the Qinghai Lake Basin (LZBCAS 1994; Li et al. 2007; Jin et al. 2010b, 2013). Their total area is ~16,577 km^2 accounting for over 56 % of the Qinghai Lake Basin; and their total discharge is 1.21×10^9 m^3/year, accounting for over 63 % of the total volume of the surface water flowing into the Qinghai Lake (Table 1). Therefore, the hydrological characteristics of these two rivers would represent the hydrological characteristics of the rivers in the Qinghai Lake Basin and NE Qinghai-Tibet Plateau. Moreover, only these rivers have hydrological observation stations at their estuaries (Fig. 1, locations A1 and B1, respectively). Thus, there are only two weather stations (Tianjun and Gangcha weather stations) in the Qinghai Lake Basin, which lie in the Buha River Catchment and Shaliu River Catchment, respectively (Fig. 1). The monitoring data from these weather and hydrological observation stations could help to estimate runoff generation quantitatively. Thus, the Buha and Shaliu Rivers were selected for this study.

Sampling and methods

Sampling and data

The precipitation samples were collected monthly from July 2009 to June 2010 at eight locations, evenly distributed in the Buha and Shaliu River catchments (Fig. 1). A total of 45 samples were collected during the observation period. The precipitation samples were measured and stored in 100-mL high-density polyethylene square bottles for isotopic analyses. A total of 27 river water samples were collected along the main stream and tributary in July 2009 from 27 locations, among which 17 samples were collected along the Buha River and 10 samples were collected along the Shaliu River (Fig. 1). All the samples were collected by hand approximately 0.2 m below the water surface and 5–10 m from the river bank, where clean and flowing water was present. The samples were filtered through 0.45-µm nylon filters, and the samples for cation analysis were acidified with ultrapure HCl. The samples collected for anion and isotope analyses were transported in ice bags and refrigerated at approximately 4 °C until laboratory analysis. These water samples were stored in 100-mL high-density polyethylene square bottles for isotope analyses and 500-mL bottles for chemical analyses. The electrical conductivity (EC) of the river water was measured in situ using a handheld meter with a probe.

The monthly runoff records between 1961 and 2010 were collected from the Buha and Shaliu River hydrological stations (Fig. 1, locations A1 and B1, respectively), and the monthly precipitation and temperature records for the same period were collected from the Tianjun and Gangcha weather stations (Fig. 1). The annual data were derived from the monthly data.

Table 1 Characteristics of the five main rivers in the Qinghai Lake basin

River name	Catchment area (km^2)	Main stream length (km)	Mean annual runoff (10^8 m^3/year)	Percent of all basin discharge (%)
Buha River	14,932	286.0	8.09	45.5
Shaliu River	1,645	105.8	3.12	17.5
Haergai River	1,572	109.5	2.42	13.6
Quanji River	599	63.4	0.54	3.0
Heima River	123	17.2	0.11	0.6
Total	18,872	581.9	14.28	80.3

Methods

The stable isotopic composition of the samples was analyzed using a Picarro L1102-i water isotope analyzer in the Stable Isotope Laboratory, Institute of Geology and Geophysics, Chinese Academy of Sciences. The isotopic values were reported using the standard δ notion relative to the standard V-SMOW (Vienna Standard Mean Ocean Water); the precisions were ± 0.5 and ± 0.1 ‰ for δ^2H and $\delta^{18}O$, respectively. The major cations in the samples were determined using a Dionex-600 ion chromatograph; and the anions were determined using a Dionex-500 ion chromatograph in the Analytical Laboratory, Beijing Research Institute of Uranium Geology.

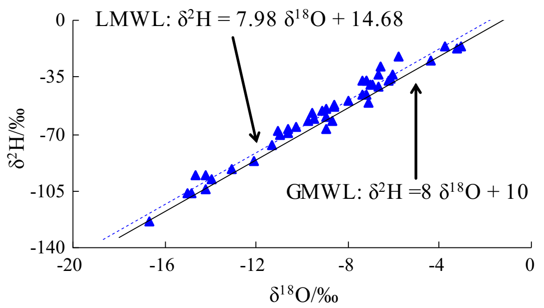

Fig. 2 Relationship between the δ^2H and δ^{18}O of the precipitation in the study area

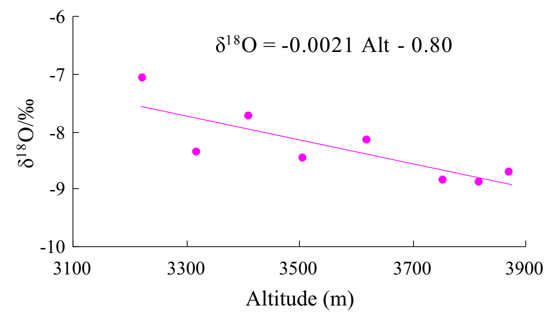

Fig. 3 Altitude effect of precipitation in the study area

Two and three-component isotope hydrograph separation (Pearce et al. 1986; DeWalle et al. 1988; Ryu et al. 2007) was used to distinguish the origin of the water coming from different Buha and Shaliu River tributaries. Both the two and three-component methods can be described as a uniform equation (Kong and Pang 2012):

$$Q_t = \sum_{m=1}^{n} Q_m, \quad Q_t \times C_t^j = \sum_{m=1}^{n} Q_m \times C_m^j, \qquad j = l, \ldots, k \tag{1}$$

where Q_t is the total discharge, Q_m is the discharge of component m, and C_m^j is tracer j incorporated in component m.

In previous studies, various types of tracers were used, including geochemical tracers such as TDS, EC, Cl^-, and SiO_2, and stable isotopic tracers such as δ^{18}O, δ^2H, and deuterium excess (d-excess, defined as d = δ^2H − 8 δ^{18}O) (Laudon and Slaymaker 1997; Gibson et al. 2005; Kong and Pang 2012). In this study, δ^{18}O and TDS were used as the tracers based on their significant differences in recharging sources.

Moreover, the boundary, altitude, and area of the catchments or sub-watersheds were processed and overlaid on the spatial distribution using the ArcGIS 9.0 software from the Environmental Systems Research Institute (ESRI).

Results and discussion

Stable isotope in precipitation

According to the precipitation samples collected in the study area monthly from July 2009 to June 2010, the δ^{18}O and δ^2H values of the precipitation ranged from −16.69 to −3.07 ‰ and from −123.28 to −15.65 ‰, respectively. The values were within previously reported ranges for China: −35.5 to +2.5 ‰ for δ^{18}O and −280.0 to +24.0 ‰ for δ^2H (Tian et al. 2001). The local meteoric water line (LMWL) was defined by the δ^{18}O and δ^2H content of precipitation (Fig. 2):

$$\delta^2\mathrm{H} = 7.98\delta^{18}\mathrm{O} + 14.68\,\mathrm{VSMOW}\,(n = 45, P < 0.001, R = 0.99) \tag{2}$$

The LMWL deviated slightly from the Global Meteoric Water Line (GMWL) because of the local variation in the moisture sources and evaporation modes (Craig 1961; Clark and Fritz 1997).

In successive precipitation samples, the altitude effect was temperature-dependent and can be explained by the rainout from adiabatic cooling (Coplen 1993). The altitude effect of δ^{18}O was significant in the annual precipitation, and the best-fit equation can be expressed as follows (Fig. 3):

$$\delta^{18}\mathrm{O} = -0.0021\,Alt - 0.80\,(n = 8,\ P < 0.01,\ R = 0.82) \tag{3}$$

The equation indicated that the δ^{18}O was depleted at a rate of −0.21 ‰ per 100 m elevation (Fig. 3), within the range of the δ^{18}O-altitude gradients, from −0.10 to −0.50 ‰/100 m, obtained from different areas of the world (Clark and Fritz 1997). The calculated value of the altitude effect for δ^{18}O was similar to that of the high-altitude Qilian Mountains (−0.18 ‰/100 m; Wang et al. 2009) and Tianshan Mountains (from −0.16 to −0.23 ‰/100 m; Pang et al. 2011), and relatively smaller than the effect calculated in other regions (Niewodniczanski et al. 1981; Stowhas and Moyano 1993; Saravana et al. 2010).

Stable isotope in river waters

The δ^{18}O and δ^2H values of the river water samples collected in July 2009 ranged from −9.78 to −6.38 ‰ and from −63.62 to −38.65 ‰, respectively (Table 2; Fig. 4). For each river, the δ^{18}O and δ^2H values in the Buha River ranged from −9.78 to −6.49 ‰ and from −63.62 to −41.29 ‰, respectively; and the δ^{18}O and δ^2H values in Shaliu River ranged from −8.11 to −6.38 ‰

Table 2 Stable isotope and hydrochemistry compositions of river water in the Qinghai Lake Basin

Sites	Channel type	$\delta^{18}O$	$\delta^{2}H$	TDS	EC	Ca^{2+}	Mg^{2+}	Na^{+}	K^{+}	HCO_3^{-}	Cl^{-}	SO_4^{2-}	Controlling area
A1	Main stream	−7.52	−46.69	375.65	0.198	69.9	14.6	18.5	1.55	219	18.1	34	14,611.15
A2	Tributary	−6.49	−41.29	346.55	0.168	59.7	12.7	16.9	1.35	218	17.6	20.3	994.54
A3	Main stream	−7.6	−47.27	355.53	0.200	61.4	13.6	17.4	1.63	209	17.7	34.8	13,303.95
A4	Tributary	−7.08	−43.56	327.99	0.161	64.1	11.4	10.8	1.16	209	9.63	21.9	1,116.31
A5	Tributary	−6.74	−42.39	336.48	0.157	62.8	11.9	10.1	1.08	215	12.1	23.5	1,832.59
A6	Tributary	−6.82	−42.21	355.05	0.175	68	12.2	10.2	0.95	225	11.9	26.8	1,439.49
A7	Main stream	−7.85	−49.38	367.2	0.227	63	14.8	19.6	1.8	204	20.3	43.7	8,658.61
A8	Main stream	−8.01	−50.08	380.45	0.217	70.7	16	21.3	1.75	205	19.2	46.5	8,318.65
A9	Tributary	−7.5	−47.14	332.36	0.192	57.8	12.5	24.6	1.66	178	28.1	29.7	1,309.28
A10	Tributary	−7.56	−47.07	488.13	0.344	92.2	22.1	26.3	2.03	229	40	76.5	276.30
A11	Main stream	−8.91	−56.05	389.91	0.225	73.9	15.5	21.3	2.21	215	17.7	44.3	5,891.38
A12	Main stream	−9.78	−63.62	401.94	0.235	78.2	13	21.4	3.04	230	20.5	35.8	3,309.21
A13	Main stream	−9.73	−62.21	401.82	0.235	74.7	13.5	24	3.22	224	22.6	39.8	2,733.23
A14	Main stream	−8.43	−51.84	327.68	0.148	62	13.7	9.95	2.16	196	7.67	36.2	2,179.43
A15	Main stream	−8.58	−51.83	308.61	0.124	60.4	12.7	9.99	1.65	187	5.17	31.7	1,454.69
A16	Tributary	−7.32	−44.29	257.95	0.072	55.3	7.4	6.06	0.8	168	8.89	11.5	338.76
A17	Main stream	−8.76	−52.41	302.77	0.124	56.9	12.9	10.4	1.46	181	4.81	35.3	1,023.44
B1	Main stream	−6.96	−41.97	344.73	0.160	57.4	12.5	14.8	1.4	232	4.93	21.7	1,446.17
B2	Main stream	−7.04	−41.53	325.89	0.142	54.1	12.1	12	1.32	220	4.67	21.7	1,252.66
B3	Tributary	−6.38	−38.65	356.08	0.161	63.7	10.5	9.63	2.08	250	7.27	12.9	138.25
B4	Tributary	−7.35	−44.41	226.27	0.050	40.9	6.5	5.46	1.13	157	3.78	11.5	136.17
B5	Main stream	−7.14	−42.35	321.43	0.142	52.1	12.5	13.4	1.28	218	4.65	19.5	917.93
B6	Tributary	−7.02	−42.36	329.11	0.132	51.9	15.2	7.35	1.29	229	5.87	18.5	351.24
B7	Main stream	−7.38	−43.08	314.75	0.122	49.1	10.5	16.5	1.27	211	3.98	22.4	517.84
B8	Tributary	−7.06	−42.13	359.29	0.163	45.8	11.1	30.5	1.82	236	4.87	29.2	108.14
B9	Main stream	−7.61	−45.05	291.7	0.104	48.3	10.1	12.1	1.09	197	3.61	19.5	297.92
B10	Main stream	−8.11	−47.96	303.05	0.127	39.3	11.4	22.5	1.64	196	3.41	28.8	104.91

$\delta^{18}O$ and $\delta^{2}H$ are in ‰ V-SMOW; all cations, anions and TDS concentrations are in mg/L; controlling area is in km^2

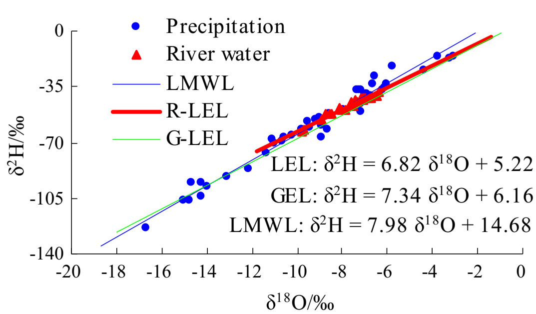

Fig. 4 Relationship between the $\delta^{2}H$ and $\delta^{18}O$ in river water, LMWL, and G-LEL (*LMWL* local meteoric water line, *R-LEL* local evaporation line of river water, *G-LEL* local evaporation line of Ground water)

and from −47.96 to −38.65 ‰, respectively. The ranges of $\delta^{18}O$ and $\delta^{2}H$ in the Buha River were larger than that in the Shaliu River. Comparing river water samples with the LMWL was useful in determining the water source in regional hydrology investigations (Clark and Fritz 1997). Figure 4 shows the relationship between the river water samples and LMWL using a bivariate plot of $\delta^{18}O$ versus $\delta^{2}H$. All the isotopic data points lay close to the LMWL; and the slope (6.82) of the local evaporation line [LEL: $\delta^{2}H = 6.82\ \delta^{18}O + 5.22$ ($n = 27$; r = 0.98)] of the river water samples was smaller than that of the LMWL (7.98), indicating that the river water mainly originated from the precipitation in the catchments that underwent weak evaporation (Friedman et al. 1962; Paul and Wanielista 2000).

According to the distributions of the sampling sites and the isotopic values (Figs. 1; 5; Table 2), the $\delta^{18}O$ of the river water increased consistently from the upper to the lower reaches (Fig. 5), i.e., the lowest (or highest) $\delta^{18}O$ values were generally observed in the upper (or lower) part of the rivers. This indicated that the altitude effect (Fig. 3) caused the river water to originate from depleted precipitation or snow melt in the upper parts and received

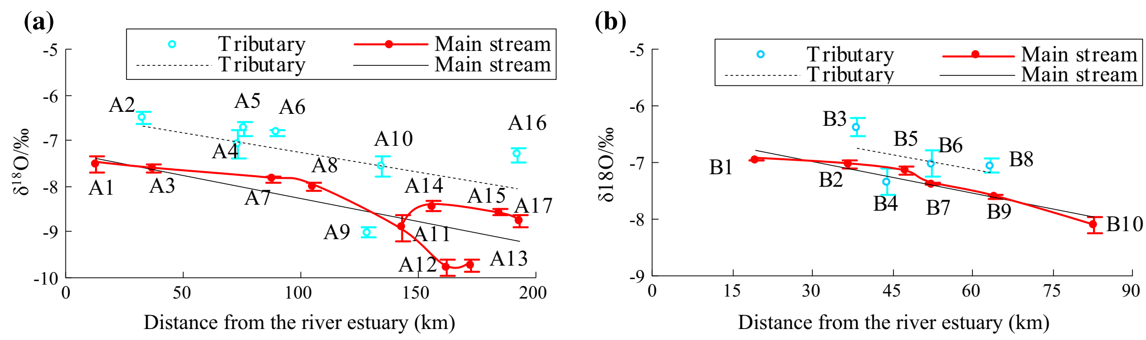

Fig. 5 Changes of $\delta^{18}O$ in the main stream water and tributary water along the Buha River (**a**) and Shaliu River (**b**)

relatively enriched precipitation in the lower parts (Rank et al. 1998; Miljević et al. 2008; Ogrinc et al. 2008). Moreover, the $\delta^{18}O$ values in the tributaries were relatively higher than those in the main stream at the same altitude (Fig. 5). The slopes of the river water evaporation line were smaller than those of the groundwater (Table 3; Fig. 4; Xiao et al. 2013), and the slopes of the tributary water evaporation line were smaller than those of the main stream water (Table 3). These results indicated that unevaporated (or depleted) groundwater contributed to the main stream flow in the watershed's lower reaches.

Table 3 Evaporation lines of the river water in the main stream and tributary

Type of river water	Evaporation line
Main stream water in Buha River	$\delta^2H = 7.05\ \delta^{18}O + 6.93$ ($n = 10$, $P < 0.001$, $R = 0.976$)
Tributary water in Buha River	$\delta^2H = 5.50\ \delta^{18}O - 5.10$ ($n = 7$, $P < 0.001$, $R = 0.959$)
Main stream water in Shaliu River	$\delta^2H = 5.60\ \delta^{18}O + 0.72$ ($n = 6$, $P < 0.001$, $R = 0.996$)
Tributary water in Shaliu River	$\delta^2H = 5.46\ \delta^{18}O - 3.97$ ($n = 4$, $P < 0.001$, $R = 0.991$)

River water hydrochemistry

Table 2 shows major ion concentrations in the river water samples. The total dissolved solids (TDS) and EC values of the river waters ranged from 226.27 to 488.13 mg/L and from 0.050 to 0.344 ms/cm (Table 2), respectively, with average of 341.79 mg/L and 0.167 ms/cm, respectively. The ternary diagrams were useful in exploring the relative contribution of different weathering regimes (Hu et al. 1982; Edmond et al. 1996; Zhu et al. 2013). The hydrochemical type of the river water was Ca–Mg–HCO_3 (Fig. 6), with the Ca^{2+} and Mg^{2+} concentrations accounting for > 81 % of the cations (Table 2). This indicated that the river water chemistry was mainly controlled by the carbonate weathering in the Qinghai Lake Basin (Xiao et al. 2013). Comparing the precipitation and groundwater around the Qinghai Lake (the average TDS of precipitation and groundwater were 68.10 and 539.05 mg/L, respectively; the EC values of groundwater were 0.356 ms/cm; Hou et al. 2009; Cui and Li 2014), the TDS and EC of the river waters were relatively higher than those of the precipitation and lower than those of the groundwater, indicating that the interactions between water and rocks were less in the river water than in the groundwater. Furthermore, the TDS and Ca^{2+} and Mg^{2+} concentrations increased consistently along the river due to carbonate dissolution (Table 2), indicating that the carbonate weathering by groundwater contributed to the main stream flow in the watershed's lower reaches.

Meanwhile, the TDS, Na^+, Mg^{2+} and SO_4^{2-} concentrations in the main stream were higher than those in the tributaries, indicating that human activity might be an important factor for alkaline earth enrichment in the main stream water (Pawellek et al. 2002). Most residential areas and roads were concentrated along the Qinghai Lake Basin's main stream. For example, the Tian-Mu road followed the Buha River's main stream from Tianjun city to Muli town; and thousands of trucks transported coal daily from the Muli Coal Mine to Tianjun. Thus, the garbage and coal dust increased the ion contents, particularly Na^+, Mg^{2+}, and SO_4^{2-}, influencing the main stream's hydrochemical composition.

Gibbs (1970) developed a boomerang envelope model describing three mechanisms that control the dissolved salt composition of the world's waters: atmospheric precipitation, rock dominance, and the evaporation-crystallization (Machender et al. 2014). As shown in Fig. 7, all the samples were within the mid-upper branch of the Gibbs boomerang envelope, suggesting that the river water's chemical compositions were dominated by rock-weathering rather than evaporation. This indicated that rock-weathering, ion exchange, and precipitation are the major geochemical processes responsible for the solutes in the river water within the Qinghai Lake Basin.

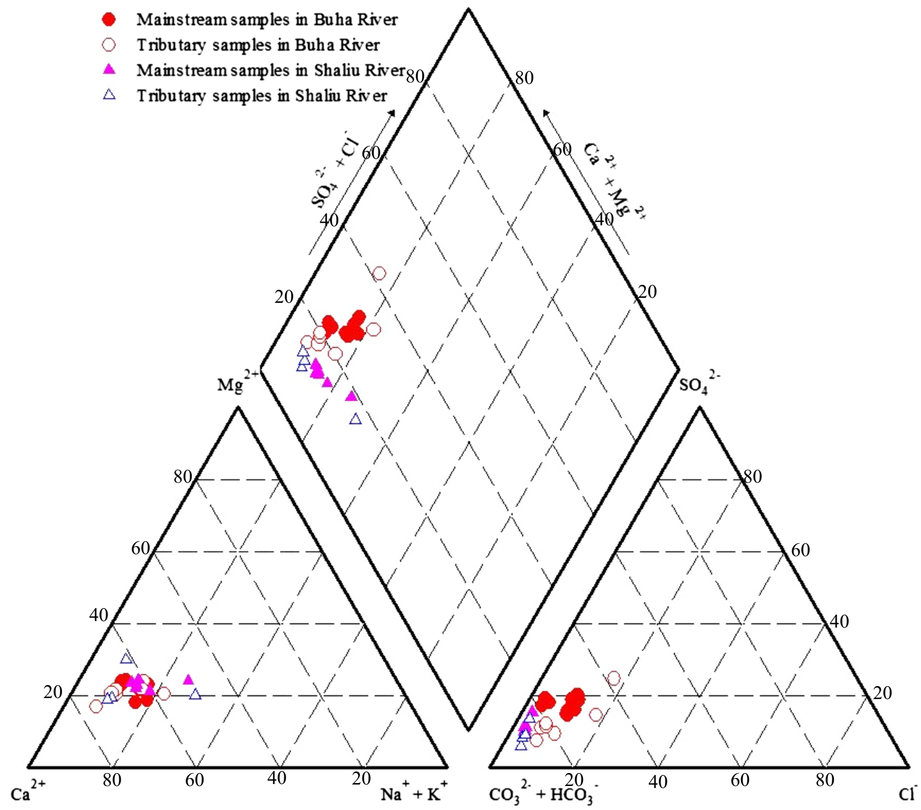

Fig. 6 Piper plots of cations and anions in the river water in the Buha River and Shaliu River

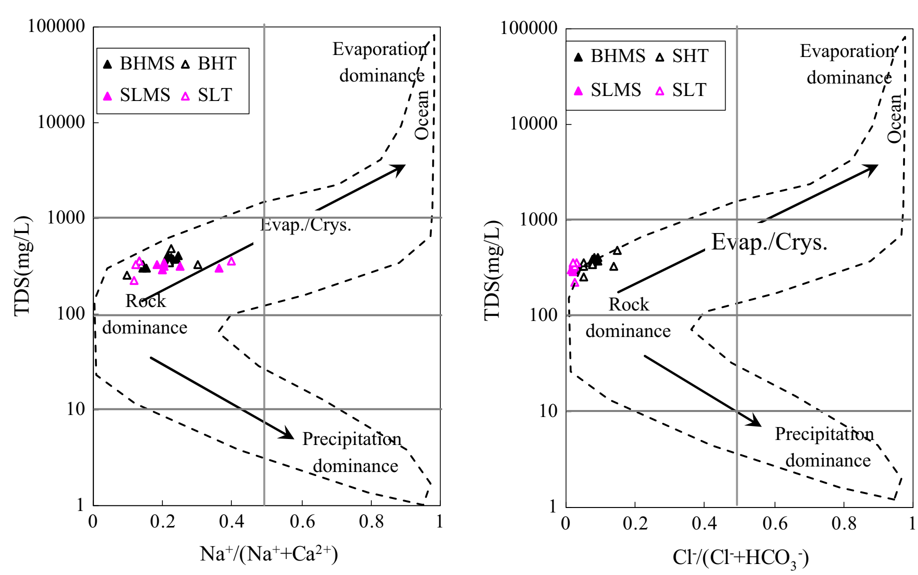

Fig. 7 Plots of the major ions within the Gibbs boomerang model for river water (*BHMS* Buha River mainstream, *BHT* Buha River tributary, *SLMS* Shaliu River mainstream, *SLT* Shaliu River tributary)

Runoff characteristics

By ignoring the evaporation in the channels and assuming that the discharges in locations A1 and B2 were the total discharges of the Buha and Shaliu Rivers, respectively, the contribution rates of mainstream and tributary were computed by hydrograph separation analysis using the two and three-component methods (Tables 4, 5). The drainage area (5,891.4 km^2) above location A11 occupied approximately 40.3 % of the Buha River Catchment (14,611.2 km^2, above location A1), but runoff from that area generated over half of the discharge for that river (57.9 %, Table 4). By comparison, the drainage area (517.8 km^2) above location B7 occupied approximately 41.3 % of the Shaliu River Catchment (1,252.7 km^2, above location B2); and the runoff also generated more than half of the discharge for

Table 4 Runoff percentage of the Buha River main stream (M) and tributaries (T)

Sites	A1	A2	A3	A4	A5	A7	A9	A11	A12	A14	A16	A17
Channel type	M	T	M	T	T	M	T	M	M	M	T	M
Runoff percentage (%)	100.0	7.2	92.8	14.0	11.2	67.6	9.8	57.9	32.9	25.0	3.1	21.8
Controlling area (km^2)	14,611.2	994.5	13,304.0	1,116.3	1,832.6	8,658.6	1,309.3	5,891.4	3,309.2	2,179.4	338.8	1,023.4
Area percentage (%)	100.0	6.8	91.1	7.6	12.5	59.3	9.0	40.3	22.6	14.9	2.3	7.0

Table 5 Runoff percentage of the Shaliu River main stream (M) and tributaries

Sites	B2	B3	B4	B5	B6	B7	B8	B9
Channel type	M	T	T	M	T	M	T	M
Runoff percentage (%)	100.0	5.7	14.9	79.4	28.1	51.2	15.0	36.2
Controlling area (km^2)	1,252.7	138.3	136.2	917.9	351.2	517.8	108.1	297.9
Area percentage (%)	100.0	11.0	10.9	73.3	28.0	41.3	8.6	23.8

that river (51.2 %, Table 5). These results suggested that the river discharge was generated mainly from the Qinghai Lake Basin's middle and upper catchments. This finding was similar to the mountainous runoff in the Heihe River in northwest China, where > 50 % of the total runoff at the catchment outlet was generated in the middle and upper mountain zones (Wang et al. 2009; Qin et al. 2013). High precipitation and low evapotranspiration led the upper catchment to be a major area of the mountainous runoff generation in the alpine zone (Kahn et al. 2008). Moreover, different types of grasslands led to different runoff generation regimes in alpine grassland vegetation cover regions (Wang et al. 2012; Qin et al. 2013). There were two main types of grasslands, alpine swamp and alpine meadow, in the Qinghai Lake Basin and the entire Qinghai-Tibet Plateau (Wang et al. 2007; Wang et al. 2013). The alpine meadow vegetation canopy had a higher maximum interception ratio and saturation precipitation than the alpine swamp vegetation (Wang et al. 2007), i.e., the runoff coefficient of the alpine swamp cover was higher than that of alpine meadow cover. Because the alpine swamp distributed mainly in the middle-upper catchment in the Qinghai Lake Basin (Cui et al. 2011), the runoff coefficient in the middle-upper catchment was relatively higher. These all indicated that the land use/cover would be the other reason for why the discharge of the rivers was generated mainly from the Qinghai Lake Basin's middle and upper catchments.

Glaciers occupy $\sim$13.29 km^2 in the Qinghai Lake Basin, mainly in the upper reaches of Buha River. The glacial meltwater runoff is approximately 0.1 $\times$ 10^8 m^3/year (LZBCAS 1994), accounting for only 1.24 % of the Buha River's annual runoff. These data suggested that the influence of glaciers on river runoff was negligible, and the river runoff was mostly controlled by precipitation. Therefore, the runoff coefficient [the portion of precipitation that becomes direct runoff, K (%) = runoff/precipitation] and runoff depth [the depth of the precipitation for direct runoff over catchment area, D (mm) = runoff/area] in the study area could be acquired by using the precipitation and hydrological observation data. Based on the runoff observation data of the Buha River (13.29 $\times$ 10^8 m^3) and Shaliu River (4.10 $\times$ 10^8 m^3) during the study period from July 2009 to June 2010, the drainage area above locations A1 and B1 (Fig. 1), the runoff depths of the Buha and Shaliu River Catchments were calculated (91.0 and 283.4 mm, respectively). To explore the reason for the large difference in the runoff depth between the two catchments, the catchment's amount of precipitation and runoff coefficient were calculated.

According to the observed precipitation data at the eight locations (Fig. 1), it was found that the annual precipitation increased with the altitude in the study area, with an altitudinal gradient of 19.6 mm/100 m (Fig. 8). The gradient was similar to that (17.1 mm/100 m) of the mountainous area of the Heihe River in northwest China (Wang et al. 2009). Therefore, based on the altitudinal precipitation gradient and drainage distribution at different altitudes in the study area, the mean precipitations of the Buha and Shaliu River Catchments were calculated as 555.9 and 533.9 mm, respectively. Consequently, the runoff coefficients of the Buha and Shaliu River Catchments were 0.164 and 0.531, respectively. The runoff coefficient of the Buha River Catchment was similar to that of the inland river basins of northwest China ($\sim$0.165; ECPGCCAS 1981) and that of the montane vegetation zone of the Heihe River in northwest China ($\sim$0.198; Wang et al. 2009), but much smaller than that of the Shaliu River Catchment (0.531) and the upper Heihe River watershed ($\sim$0.37; Qin et al. 2013). This was because the surface water flowed slowly and largely infiltrated in the large river catchment due to its relatively longer channel, lager drainage area, and smaller

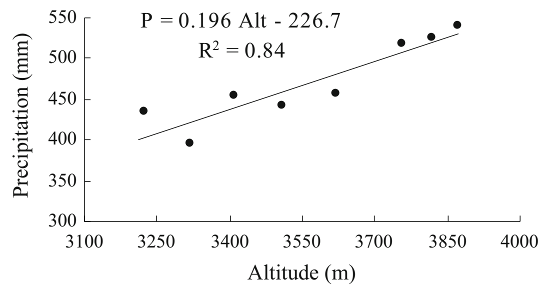

Fig. 8 Changes of precipitation with altitude in the study area

gradient (Carey and Woo 2001; Gibson et al. 2005; Buda 2013). The river channel gradient and the average slope of the Shaliu River Catchment (20 ‰ and 12.3°, respectively) were larger than those of the Buha River Catchment (10 ‰ and 11.5°, respectively); and the Shaliu River's catchment area and channel length were smaller/shorter than those of the Buha River Catchment (Table 1). Therefore, the runoff depth and runoff coefficient of the Buha River Catchment were smaller than those of the Shaliu River Catchment.

Impact of climate changes on runoff and lake level

The precipitation and temperature observed at the Tianjun weather station and the Buha River runoff were used to analyze the impact of climate change on the runoff in the Qinghai Lake Basin, because climatic factors (e.g., precipitation and temperature) observed at the Gangcha weather station were consistent with those observed at the Tianjun weather station; and the runoff of the rivers in the Qinghai Lake Basin had the same interannual variation (Li et al. 2007; Yi et al. 2010; Zhang et al. 2011a; Jin et al. 2013).

During the period between 1961 and 2010, the Buha River's runoff fluctuated significantly and ranged from 1.99×10^8 to 19.47×10^8 m^3 (Fig. 9). It had an increasing trend in the 1980s and 2000s, and a decreasing trend in the 1970s and 1990s (Table 6). Precipitation showed the same fluctuation as runoff (Fig. 9), with an increasing trend in the 1980s and 2000s and a decreasing trend in the 1970s and 1990s (Table 6), indicating that the runoff variations corresponded well with the change in precipitation. The temperature increased between 1961 and 2010 (Fig. 9), with a remarkable warming trend in the 1990s and 2000s. The temperature was 0.4 °C higher in the 1970s and 1980s than in the 1960s, 0.8 °C higher in the 1990s than in the 1960s, and 1.7 °C higher in the 2000s than in the 1960s. The annual temperature increased at a rate of 0.4 °C/decade. The increasing trend of temperature in the Qinghai Lake Basin was similar to that in the "Three-River Headwaters" regions (the Yangtze River, the Yellow River, and the Lancang River) and the Tibetan Plateau (Wang et al. 2008; Lan et al. 2010; Zhang et al. 2011b).

However, the influences of temperature and precipitation on runoff were different, because the precipitation changes affected the generation of surface runoff directly, whereas the temperature changes affected the evaporation of surface runoff. As shown in Fig. 9, the runoff peaks corresponded to the highest precipitation but to the lowest

Fig. 9 Time series of the Buha River runoff (**a**), precipitation (**b**) and temperature (**c**) at the Tianjun weather station between 1961 and 2010

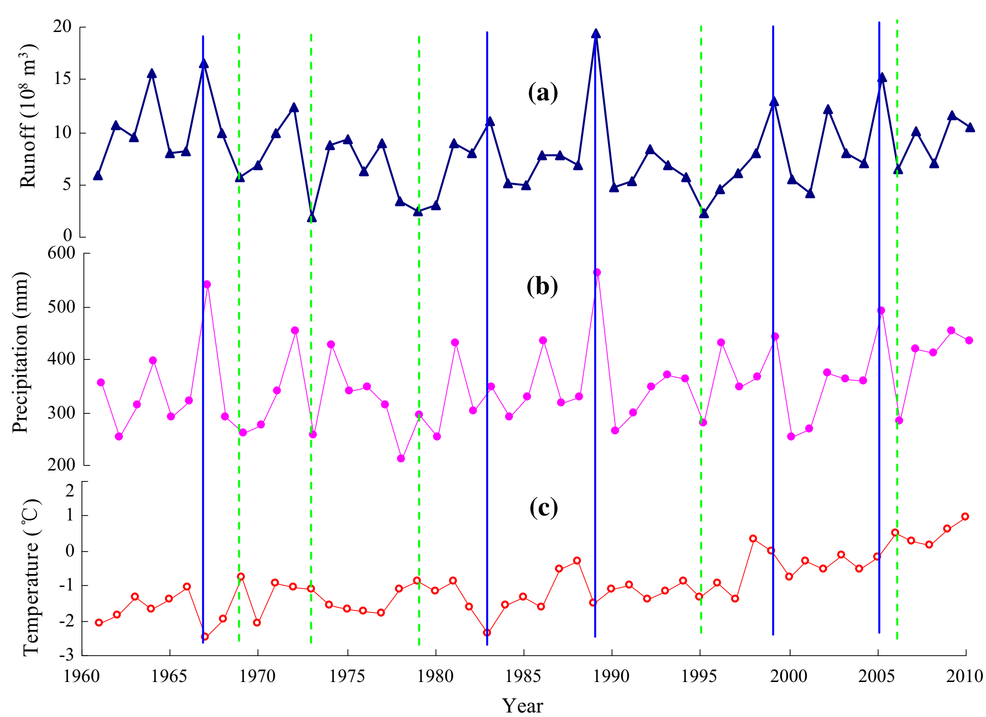

Table 6 Mean annual runoff of Buha River, mean annual precipitation and temperature of Tianjun weather station at different time periods

Period	1961–2010	1961–1970	1971–1980	1981–1990	1991–2000	2001–2010
Mean annual runoff (10^8 m^3/year)	8.13	9.71	6.67	8.49	6.56	9.23
Mean annual precipitation (mm)	349.75	329.5	323.47	361.22	349.27	385.31
Mean annual temperature (°C)	−1.00	−1.67	−1.30	−1.27	−0.86	0.08

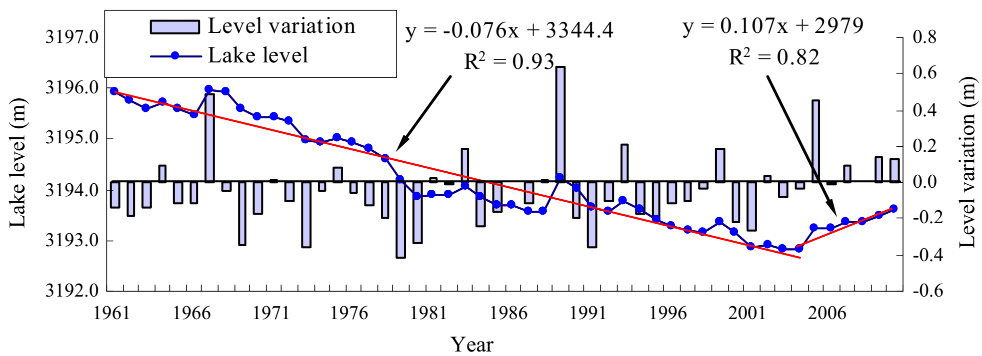

Fig. 10 Time series of lake level and annual variation of Qinghai Lake between 1961 and 2010

evaporation (e.g., 1967, 1983, 1989, 1999, and 2005); conversely, the peaks in the decline of runoff corresponded to the lowest precipitation but to the highest evaporation (e.g., 1969, 1973, 1979, 1995, and 2006). These results indicated that dry and relatively warm climate was closely related to the decline of river runoff, whereas wet and relatively cold climates were closely related to the increase of river runoff, i.e., the precipitation had a positive effect on the runoff, while the potential evaporation due to temperature played a negative role (Zhang et al. 2011a). Table 6 shows the Pearson correlation coefficients between the runoff and climatic factors (precipitation and temperature). The runoff variation was highly positively correlated to the precipitation in all time periods, with a correlation coefficient of 0.74 during the period between 1961 and 2010 (significant at the level of < 0.001). However, the correlation coefficients between the runoff and temperature were not significant at all the time periods. In general, these results indicated that the river runoff was more sensitive to precipitation than temperature, i.e., the river runoff in the Qinghai Lake Basin was primarily influenced by precipitation.

The overall Qinghai Lake water level decreased from 1961 to 2010, from 3,195.93 m in 1961 to 3,193.62 m in 2010 (Fig. 10). Two phases of annual lake level were determined from Fig. 10: a decreasing phase during the period from 1961 to 2004 with a decreasing rate of −7.6 cm/year; and a rising phase during the period from 2004 to 2010 with a rising rate of 13.5 cm/year. All the long-term linear trends achieved a 0.001 significance level. The lake level declined in 33 years and showed signs of recovery in 15 years (Fig. 10), and the level declined by > 15 cm/year in 15 years and rose by > 15 cm/year in 6 years. The maximum decline of 0.41 m was recorded in 1979, and the maximum rise of 0.64 m was recorded in 1989. Compared to the Buha River's runoff (Figs. 9, 10), the lake level showed the same runoff fluctuation, with an increasing trend in the 1980s and 2000s and a decreasing trend in the 1970s and 1990s. Moreover, as shown in Fig. 11, the peaks in the rise of lake level corresponded to the highest runoff (e.g., 1967, 1983, 1989, 1999, and 2005); conversely, the peaks in the decline of lake level corresponded to the lowest runoff (e.g., 1969, 1973, 1979, 1995, and 2006). The Pearson correlation coefficients between the lake level variation and runoff was 0.84 during the period between 1961 and 2010 (significant at a level of < 0.001). Combined with the relationship between precipitation and runoff (Table 7; Fig. 9), these findings suggested that the decline in the lake level was closely related to the decline in the river runoff and precipitation, and the increase in the lake level was closely related to the increase in the river runoff and precipitation. In general, the level of the Qinghai Lake was primarily influenced by precipitation and runoff.

An annually resolved and absolutely dated tree ring-width chronology of the NE Tibetan Plateau spanning 4,500 years showed that any further large-scale climate warming might be associated with an even greater moisture supply in this region (Yang et al. 2014). If so, the lake's level would rise due to increasing precipitation and runoff

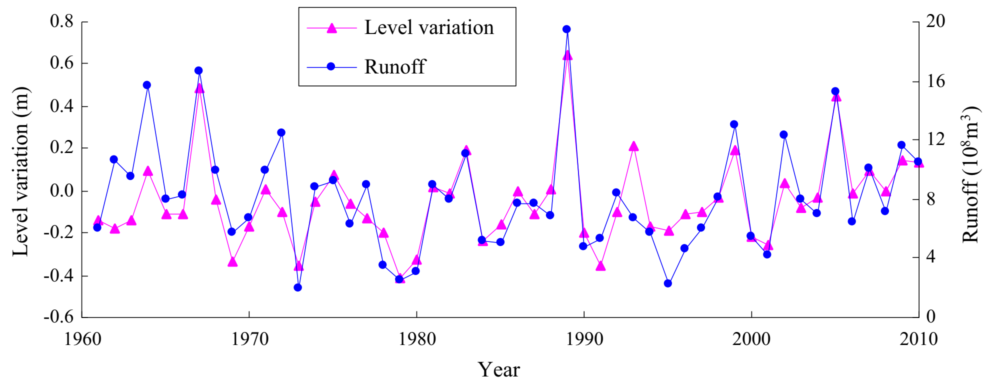

Fig. 11 Relationship between the Buha River runoff and Qinghai Lake water level variation

Table 7 Pearson correlation coefficients between runoff, precipitation, and temperature

Period	1961–2010	1961–1970	1971–1980	1981–1990	1991–2000	2001–2010
Precipitation	0.74***	0.76**	0.83**	0.85***	0.61*	0.83**
Temperature	−0.04	−0.46	−0.32	−0.28	0.57	0.09

*** Correlation is significant at the 0.001 level; ** correlation is significant at the 0.01 level; * correlation is significant at the 0.05 level

on the TP (Christensen et al. 2014; Song et al. 2014); i.e., the rising water level of the Qinghai Lake since 2004 may be part of a comprehensive effect of increasing precipitation, runoff, and changing rainfall patterns under an unfolding global warming scenario.

Conclusions

The river water assessed for this study originated primarily from precipitation, with slight evaporation; the river water in the tributaries evaporated more than that in the main stream. Chemically, Ca–Mg–HCO_3 was the primary substance in the river waters, driven by rock-weathering rather than evaporation.

The river discharge mainly originated from the middle and upper reaches in the Qinghai Lake Basin. The runoff depth and runoff coefficient in the large catchments were smaller than those in the small catchments, because in the large river catchment, the surface water flowed more slowly and infiltrated more due to its relatively longer channel, larger drainage area, and smaller gradient. River runoff was more sensitive to the precipitation than to temperature-driven evaporation in the Qinghai Lake Basin, i.e., the river runoff in the Qinghai Lake Basin was primarily influenced by precipitation.

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