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Runoff processes in the Qinghai Lake Basin, Northeast Qinghai-Tibet Plateau, China: Insights from stable isotope and hydrochemistry

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ABSTRACT

Qinghai Lake is one of China's national nature reserves and supports the ecological security of the NE Qinghai-Tibet Plateau. More than half of the rivers flowing into Qinghai Lake are currently dry due to climate change and human activity. This study was designed to learn more about the causes of environmental problems in the basin, using stable isotopes and hydrochemistry of Qinghai Lake Basin river water to explore runoff processes and their relationship with climate change. Results indicated that the river water was mainly fed by precipitation in the basin, which has undergone weak evaporation. River discharges were generated mainly from the middle and upper basin, due to high precipitation, low evapotranspiration, and alpine swamp land use/cover in those areas. River water in tributaries would experience relatively stronger evaporation than in the main stream. Main hydrochemical types of river waters were Ca^{2+}-Mg^{2+}-HCO_3^-, and river water chemistry was mainly controlled by carbonate weathering in the Qinghai Lake Basin. The effects of human activity on water chemistry were relatively mild in the basin. The interaction between water and rocks was slighter in the river water than in the groundwater. River runoff was more sensitive to precipitation than to temperature. Lake level rises were closely related to increases in river runoff and precipitation. Conversely, Lake level declines were closely related to declines in river runoff and precipitation. Lake level could rise due to increasing precipitation and runoff in the future.

1. Introduction

Qinghai Lake (36°32′–37°15′N, 99°36′–100°47′E) is the largest inland lake in China. It has a surface area of 4400 km^2, a water volume of 71.6×10^9 m^3, and lies at an altitude of 3194 m above sea level. The lake lies in the cold and semiarid region of the NE Qinghai-Tibet Plateau, China (Fig. 1). It is a closed basin, with a watershed area of approximately 29,661 km^2 that has no surface water outflow. The area 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 recent decades, more than half of the rivers flowing into Qinghai Lake have dried up due to climate change and human activity (LZBCAS, 1994; Li et al., 2007). The lake level declined from 3196.55 m in 1959–3192.84 m in 2003, an average decreasing rate of 8.4 cm year^{-1} over those 44 years; the lake level rised from 3192.87 m in 2004–3194.08 m in 2012, an average rising rate of 15.1 cm year^{-1} over these 8 years (Li et al., 2007, 2012; Jin et al., 2013). These hydrological processes in part have led to some of the environmental problems in the basin, such as decreases in water supplies, deterioration of water quality, desertification, and the loss of grazing grassland (Qin and Huang, 1998; Zhang et al., 2003; Hao, 2008).

Understanding runoff processes and the water cycle of the river system in the Qinghai Lake Basin should support an understanding of the causes for environmental problems in the basin. However, due to the basin's large area, complex geological and geomorphic conditions (mountain area accounting for 68.6% of the basin area; river valleys and plain accounting for only 17.4%), and few meteorological and hydrological stations (Fig. 1), traditional hydrological research methods face more difficulty and uncertainty in the study of runoff process and water cycle in the basin. Previous published studies concerning river runoff focused on investigating the hydrochemistry compositions of river water, runoff variation, or land use impacts on runoff (Qin and Huang, 1998; Yan and Jia, 2003; Li et al., 2005, 2007,

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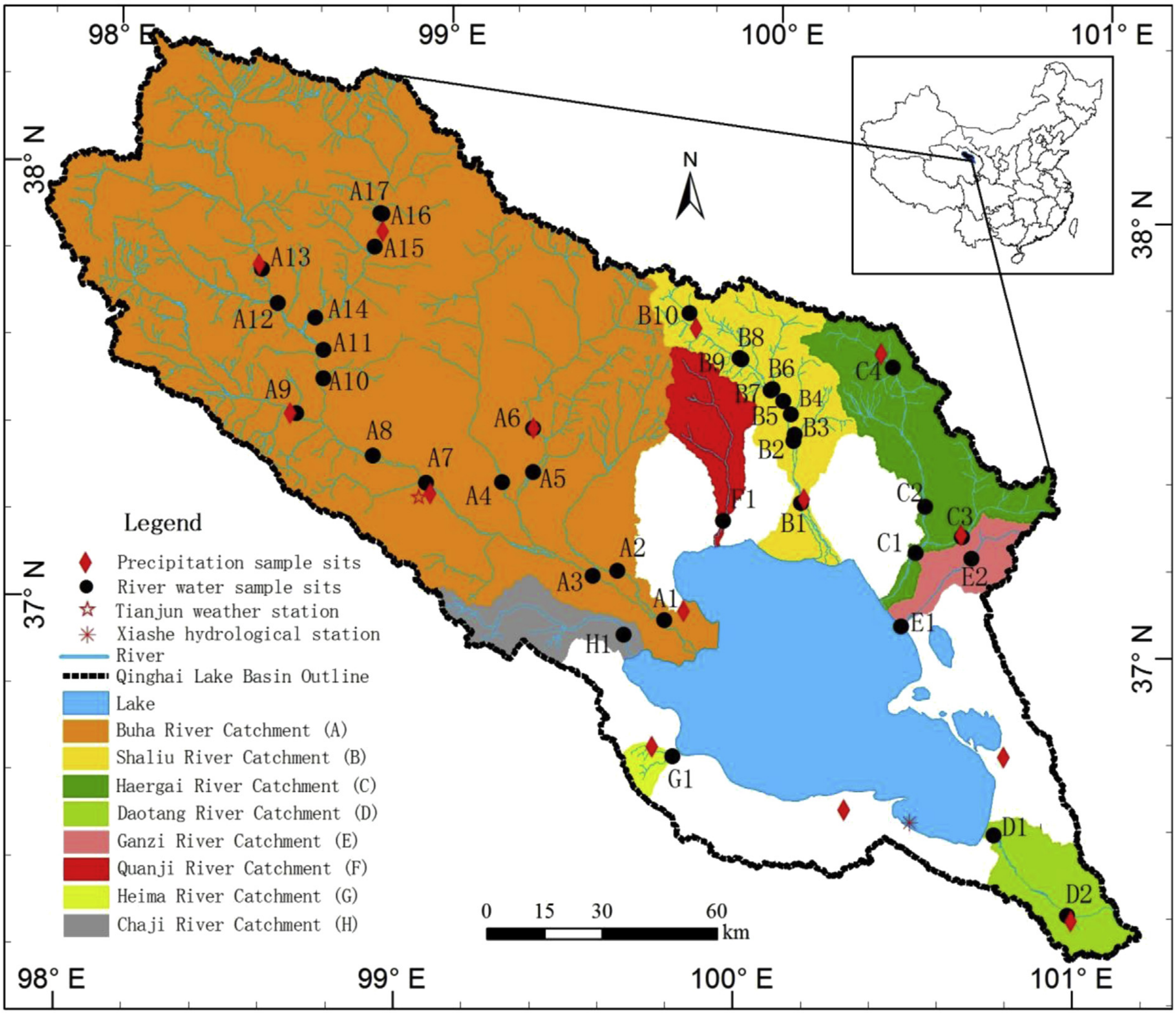

Fig. 1. Location of the Qinghai Lake Basin and sampling sites.

2009; Xu et al., 2010; Yi et al., 2010; Jin et al., 2010; Xiao et al., 2012). Runoff processes in the basin are not fully understood.

The integrated use of isotopic and hydrochemical tracers has emerged as an effective approach to investigate the 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., 2008b; Meredith et al., 2009). Yuan and Miyamoto (2008) analyzed the oxygen and hydrogen isotopic compositions, major ion concentrations, and other physical variables of Mexico's Pecos River water to assess its physical features. The study found that up to 85% of stream flow was derived from local freshwater sources (mainly from Mexican monsoonal rainfall) in the lower valley, and that up to 33% of stream water was lost through evaporation from stream channels and middle basin fields. Halder et al. (2013) investigated the mixing of Rhône River water in Lake Geneva (Switzerland–France) using stable hydrogen and oxygen isotope tracers, and found that the fraction of Rhône River water within the Lake Geneva interflow was estimated to be up to 37% in summer. Fan et al. (2014) analyzed the isotopic composition of river water, precipitation, and ice-snowmelt water of the Tizinafu River, originating in the northern slope of the Kunlun, to investigate the distribution of stable isotopes and sources of river water. This study showed the mean contribution of ice-snowmelt water was 43%, meaning that ice-snowmelt water was a key water source for the Tizinafu River. All these studies demonstrated how isotopes and hydrochemical ions could be used in hydrologic studies.

These studies suggested that investigating the stable isotope and hydrochemistry of the Qinghai Lake Basin's river water could significantly contribute to knowledge of the basin's runoff processes and water cycle. As such, this study's objectives were to: (1) investigate the characteristics of stable isotopes and hydrochemistry of the Qinghai Lake Basin's river waters, (2) study the mechanisms of confluence and segmentation of the basin's river, and (3) evaluate the impact of runoff on the lake level. The study was designed to provide general insights into the hydrological and geochemical processes of cold and alpine rivers, as well as inform water resource management in the basin and the northeastern Qinghai-Tibet Plateau.

2. Methods

2.1. Background about rivers in the Qinghai Lake Basin

There are more than 50 rivers or streams flowing into Qinghai Lake (LZBCAS, 1994). The rivers are mainly located north and northwest of the lake, resulting in an asymmetric distribution in

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Table 1
Characteristics of the main rivers in the Qinghai Lake Basin.

No.	River name	Catchment area (km^2)	Main stream length (km)	Mean annual runoff (10^8 m^3/yr)	Percent of all basin discharge (%)
1	Buha River	14,932	286.0	8.09	45.5
2	Shaliu River	1645	105.8	3.12	17.5
3	Haergai River	1572	109.5	2.42	13.6
4	Quanji River	599	63.4	0.54	3.0
5	Heima River	123	17.2	0.11	0.6
	Total	18,872	581.9	14.28	80.3

the basin (Fig. 1; Yan and Jia, 2003). Because more than 77% of precipitation occurs during the monsoon season (June to October) (Cui, 2011), most rivers are seasonal; 85% of the annual discharge occurs between June and September. The main rivers are Buha River, Shaliu River, Haergai River, Quanji River and Heima River (LZBCAS, 1994; Li et al., 2007), collectively discharging 1.43×10^9 m^3/y (annual average from 1960 to 2010) and making up more than 80% of total water volume flowing into the Qinghai Lake (Table 1). Around Qinghai Lake, the average annual air temperature is ~1.2 °C. The average annual precipitation is 357 mm, with more than 65% occurring in summer; but evaporation is 3–4 times higher than precipitation (Li et al., 2007). The climate in the lake region is influenced by two air masses: Southeast Asian Monsoon (SAM) and Westerly Circulation (WC). The SAM reaches this region in summer, while the WC dominates in winter, resulting in a clear seasonality of precipitation (An et al., 2012).

For this study, the Buha River was chosen as the representative river to distinguish the origin of the water coming from different tributaries in the basin. The Buha River is the largest river in the Qinghai Lake Basin (LZBCAS, 1994; Li et al., 2007), with a catchment area of approximately 14,932 km^2, which is about 50% of the Qinghai Lake Basin. Its discharge is 8.09×10^8 m^3/y, which accounts for approximately 50% of the total volume of surface water flowing into Qinghai Lake (Table 1). Furthermore, from a study methods perspective, only two rivers (Buha River and Shaliu River) have hydrological observation stations at their estuaries (Locations A1 and B1, respectively (Fig. 1)), needed to quantitatively estimate runoff. Based on these factors, the Buha River was used as a representative river for the basin to study the sources of river water and runoff process.

2.2. Sampling and data

Precipitation samples were collected monthly from July 2009 to June 2010 at 14 locations distributed evenly throughout the Qinghai Lake Basin (Fig. 1). A total of 124 precipitation samples, including 75 rainwater samples and 49 samples of snow or sleet, were collected during the observation period. The volumes of the precipitation samples were measured, and then stored in 100 ml high-density polyethylene square bottles for isotopic analyses. River water samples were collected along the main stream and tributary in July 2009 at 38 locations (Fig. 1). A total of 38 river water samples were collected. In order to determine the monthly change of stable isotope in the river water, 12 samples of river water were collected monthly in the lower Buha River (location A1) within the year. All river water samples were hand-dipped approximately 0.2 m below the water surface and 5–10 m from the river's bank from clean and flowing water. These water samples were collected in 100 ml high-density polyethylene (HDPE) square bottles for isotopic analyses and in 500 ml HDPE bottles for chemical analyses. The river water's electrical conductivity (EC) was measured in situ using a handheld meter with probe.

Monthly records of river runoff and lake level from 1961 to 2012 came from the Buha River hydrological station (Fig. 1, Location A1) and Xiashe hydrological station (Fig. 1); monthly precipitation and temperature records for the same period came from Tianjun weather station (Fig. 1). Annual data were derived from the monthly data.

2.3. Analysis methods

Isotopic values were expressed using delta notation (δ) relative to the V-SMOW standard (Vienna Standard Mean Ocean Water). The $\delta^{18}O$ and δ^2H contents of the precipitation and river water samples were measured using a Picarro L1102-i water isotope analyzer in the Stable Isotope Laboratory, Institute of Geology and Geophysics, Chinese Academy of Sciences. Overall analytical precision was ±0.5‰ and ±0.1‰ for δ^2H and $\delta^{18}O$, respectively. The chemical analyses of the water samples were performed at the Analytical Laboratory, Beijing Research Institute of Uranium Geology. Major cations were determined using a Dionex-600 ion chromatograph; anions were measured using a Dionex-500 ion chromatograph.

Two and three-component isotope hydrograph separation was used to distinguish the origin of the water coming from different tributaries (Pearce et al., 1986; DeWalle et al., 1988; Ryu, 2007). The $\delta^{18}O$ and total dissolved solids (TDS) were used as tracers in this study due to their significant differences in recharging sources (Laudon and Slaymaker, 1997; Gibson et al., 2005; Kong and Pang, 2012). The boundaries, altitudes, and catchment or sub-watershed areas were plotted using the software ArcGIS9.0 from the Environmental Systems Research Institute (ESRI).

3. Results and discussion

3.1. Stable isotope in precipitation

The $\delta^{18}O$ values in precipitation ranged from −24.40 to −2.80‰ (mean −11.72‰); the δ^2H values ranged from −180.80 to −11.54‰ (mean −77.15‰; Fig. 2). These values are within previously reported ranges for Chinese precipitation: −35.5‰ to +2.5‰ for $\delta^{18}O$ and −280.0‰ to +24.0‰ for δ^2H (Tian et al., 2001). The Local Meteoric Water Line (LMWL) was simulated using the relationship

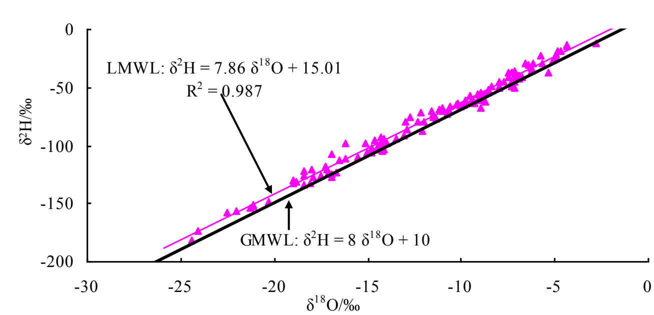

Fig. 2. Relationship between the δ^2H and $\delta^{18}O$ of precipitation in the Qinghai Lake Basin. LMWL, Local Meteoric Water Line; GMWL, Global Meteoric Water Line.

between $\delta^{18}O$ and δ^2H contents of precipitation in the Qinghai Lake Basin (Fig. 2):

$$\delta^2H = 7.86\delta^{18}O + 15.01, \quad VSMOW(n = 124, \ R = 0.99) \qquad (1)$$

The LMWL slope was similar to the slopes of meteoric water lines for northwestern China (7.05, Liu et al., 2008a) and western China (7.56, Ma et al., 2009). All slopes were less than 8, indicating that some non-equilibrium evaporation processes occurred as raindrops fell below the cloud base (Friedman et al., 1962; Dansgaard, 1964; Araguás-Araguás et al., 1998).

The altitude effect of $\delta^{18}O$ in annual precipitation was statistically significant; the best-fit equation was:

$$\delta^{18}O = -0.002 \text{ Alt} - 1.025 \quad (n = 14, \ P < 0.01, \ R = -0.66) \qquad (2)$$

The equation indicates that the precipitation's $\delta^{18}O$ was depleted at a rate of −0.2‰ for every 100 m in elevation. This was similar to findings for Tianshan Mountain precipitation (Pang et al., 2011) and for global precipitation (Bowen and Wilkinson, 2002).

3.2. Stable isotope in river waters

The $\delta^{18}O$ and δ^2H values of the river waters collected in July 2009 ranged from −9.78‰ to −3.9‰ and from −63.62‰ to −32.58‰, respectively (Table 2, Fig. 3). The lowest $\delta^{18}O$ and δ^2H values were observed in location A13 (upper part of Buha River); the highest values were observed in location D1 (lower part of Daotang River). Comparing the isotope values of river water samples with the LMWL assists in investigating water sources in regional hydrology (Clark and Frjtz, 1997). In Fig. 3, all isotopic data points lay close to the LMWL; the local evaporation line's slope (LEL: $\delta^2H = 5.7 \ \delta^{18}O$ - 4.05 (n = 38; r = 0.93)) of the river water samples was smaller than the LMWL slope (7.86). These indicated that the river water mainly came from the basin precipitation that had undergone variable degrees of evaporation (Friedman et al., 1962; Paul and Wanielista, 2000).

The isotopic values of $\delta^{18}O$ and δ^2H and the sampling site distribution (Figs. 1 and 3 and Table 2) showed that the isotope values of the river water decreased consistently from lower to upper part of the river; the lowest $\delta^{18}O$ and δ^2H values were generally observed in the upper part of the river, and the highest $\delta^{18}O$ and δ^2H values were generally observed in the lower part of the river. This trend suggested that the isotope altitude effect in precipitation (equation (2)) caused the river water to originate from a state of depleted precipitation or snow melt in the upper parts of the river, and then transition into a relatively enriched precipitation state in the lower parts (Rank et al., 1998; Miljević et al., 2008; Ogrinc et al., 2008). Furthermore, the $\delta^{18}O$ and δ^2H values were especially high in the lower parts of Daotang River, Ganzi River and Quanji River (Fig. 3). These three rivers were all in the lower parts of the Qinghai Lake Basin, with relatively small drainage areas (Fig. 1, Table 2). The water in these rivers would come from precipitation with an enriched level of stable isotopes (caused by the altitude effect of isotopes in precipitation). The water flowed slowly and evaporated heavily in a small drainage area with relatively small gradient; this was followed by extensive evaporation in the lower parts of the basin (Gibson et al., 2005; Miljević et al., 2008; Ogrinc et al., 2008; Cui and Li, 2014a). According to the $\delta^{18}O$ values of Buha River water (Table 2, Fig. 4), the values in tributary were higher than that in main stream at the same altitude. The evaporation line slope of river water in tributary was smaller than that in main streams (Fig. 5). These all indicated that the tributary water undergoes stronger evaporation than the main stream water.

Comparing the monthly $\delta^{18}O$ in the Buha River water (Location A1) with those in precipitation and groundwater (Fig. 6), the $\delta^{18}O$ of river water fluctuated slightly within the year. The fluctuation of $\delta^{18}O$ of river water was smaller than that of precipitation, especially in the dry season (between November and April), and larger than that of groundwater (Cui and Li, 2014b). These all indicated that the river water came mainly from rain during the wet season (June to September), and from interflow in the soils and groundwater during the dry season (between October and April). This interflow would originate from the rain in the wet season (Liu et al., 2008b), given the lack of precipitation during the dry season (the average precipitation was 22.5 mm in the dry season from 1961 to 2010). The lowest $\delta^{18}O$ values in the river water appeared in May and June, indicating that river water came mainly from snow melt upstream. The $\delta^{18}O$ of the river water were higher than the $\delta^{18}O$ of the groundwater, indicating a relatively high evaporation in river water.

3.3. River water hydrochemistry

Most of the total dissolved solids (TDS) values of river waters ranged from 226.27 mg/L to 488.13 mg/L, with the average of 341.79 mg/L (Table 2). Based on the ternary diagrams in Fig. 7, the hydrochemical type of river water was Ca^{2+}–Mg^{2+}-HCO_3^-. Ca^{2+} and Mg^{2+} concentrations accounted for more than 77% of the cations (Table 2), indicating that river water chemistry was mainly controlled by carbonate weathering in the Qinghai Lake Basin (Cui and Li, 2014b). The river water TDS was lower than groundwater TDS and higher than precipitation TDS (the average TDS of groundwater and precipitation were539.05 mg/L and 68.10 mg/L, respectively; Hou et al., 2009; Cui and Li, 2014b), indicating that the

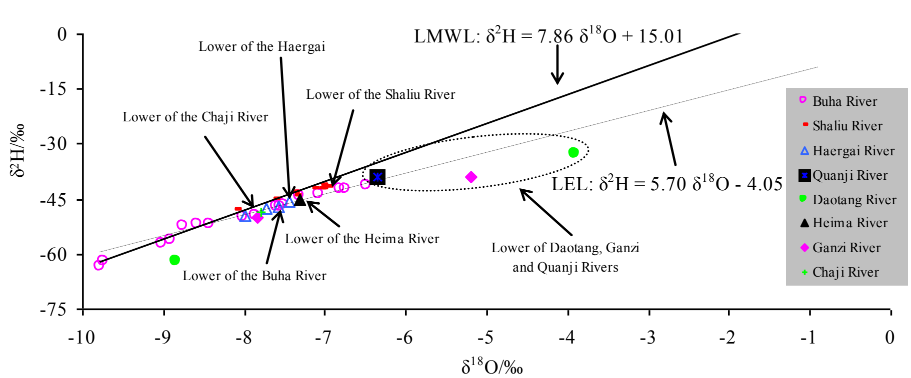

Fig. 3. Characteristics of the stable isotopes in river waters with the LMWL in the Qinghai Lake Basin. (LMWL: local meteoric water line; LEL: local evaporation line of river water).

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	Ca^{2+}	Mg^{2+}	Na^{+}	K^{+}	HCO_3^{-}	Cl^{-}	SO_4^{2-}	Controlling area
A1	Main stream	−7.52	−46.69	375.65	69.9	14.6	18.5	1.55	219	18.1	34	14,611.15
A2	Tributary	−6.49	−41.29	346.55	59.7	12.7	16.9	1.35	218	17.6	20.3	994.54
A3	Main stream	−7.6	−47.27	355.53	61.4	13.6	17.4	1.63	209	17.7	34.8	13,303.95
A4	Tributary	−7.08	−43.56	327.99	64.1	11.4	10.8	1.16	209	9.63	21.9	1116.31
A5	Tributary	−6.74	−42.39	336.48	62.8	11.9	10.1	1.08	215	12.1	23.5	1832.59
A6	Tributary	−6.82	−42.21	355.05	68	12.2	10.2	0.95	225	11.9	26.8	1439.49
A7	Main stream	−7.85	−49.38	367.2	63	14.8	19.6	1.8	204	20.3	43.7	8658.61
A8	Main stream	−8.01	−50.08	380.45	70.7	16	21.3	1.75	205	19.2	46.5	8318.65
A9	Tributary	−7.5	−47.14	332.36	57.8	12.5	24.6	1.66	178	28.1	29.7	1309.28
A10	Tributary	−7.56	−47.07	488.13	92.2	22.1	26.3	2.03	229	40	76.5	276.30
A11	Main stream	−8.91	−56.05	389.91	73.9	15.5	21.3	2.21	215	17.7	44.3	5891.38
A12	Main stream	−9.78	−63.62	401.94	78.2	13	21.4	3.04	230	20.5	35.8	3309.21
A13	Main stream	−9.73	−62.21	401.82	74.7	13.5	24	3.22	224	22.6	39.8	2733.23
A14	Main stream	−8.43	−51.84	327.68	62	13.7	9.95	2.16	196	7.67	36.2	2179.43
A15	Main stream	−8.58	−51.83	308.61	60.4	12.7	9.99	1.65	187	5.17	31.7	1454.69
A16	Tributary	−7.32	−44.29	257.95	55.3	7.4	6.06	0.8	168	8.89	11.5	338.76
A17	Main stream	−8.76	−52.41	302.77	56.9	12.9	10.4	1.46	181	4.81	35.3	1023.44
B1	Main stream	−6.96	−41.97	344.73	57.4	12.5	14.8	1.4	232	4.93	21.7	1446.17
B2	Main stream	−7.04	−41.53	325.89	54.1	12.1	12	1.32	220	4.67	21.7	1252.66
B3	Tributary	−6.38	−38.65	356.08	63.7	10.5	9.63	2.08	250	7.27	12.9	138.25
B4	Tributary	−7.35	−44.41	226.27	40.9	6.5	5.46	1.13	157	3.78	11.5	136.17
B5	Main stream	−7.14	−42.35	321.43	52.1	12.5	13.4	1.28	218	4.65	19.5	917.93
B6	Tributary	−7.02	−42.36	329.11	51.9	15.2	7.35	1.29	229	5.87	18.5	351.24
B7	Main stream	−7.38	−43.08	314.75	49.1	10.5	16.5	1.27	211	3.98	22.4	517.84
B8	Tributary	−7.06	−42.13	359.29	45.8	11.1	30.5	1.82	236	4.87	29.2	108.14
B9	Main stream	−7.61	−45.05	291.7	48.3	10.1	12.1	1.09	197	3.61	19.5	297.92
B10	Main stream	−8.11	−47.96	303.05	39.3	11.4	22.5	1.64	196	3.41	28.8	104.91
C1	Main stream	−7.73	−47.71	378.51	65.40	22.10	10.80	1.83	248.00	5.98	24.40	1495.56
C2	Main stream	−7.57	−47.24	384.63	63.70	23.40	10.50	1.99	254.00	5.84	25.20	–
C3	Tributary	−8.00	−49.75	328.38	69.10	15.80	6.84	1.00	186.00	4.64	45.00	–
C4	Main stream	−7.44	−45.71	514.96	89.00	25.60	12.40	3.09	352.00	7.47	25.40	–
D1	Main stream	−3.90	−32.58	1014.11	39.30	53.00	225.00	5.81	353.00	239.00	99.00	740.07
D2	Main stream	−8.84	−62.07	527.74	69.20	26.00	45.90	3.14	327.00	29.00	27.50	–
E1	Main stream	−5.19	−38.99	441.72	63.80	30.40	19.50	0.92	286.00	20.80	20.30	429.94
E2	Main stream	−7.83	−49.89	314.1	61.40	14.40	9.54	1.26	201.00	6.00	20.50	–
F1	Main stream	−6.35	−38.84	342.06	64.90	10.30	8.98	1.73	230.00	8.35	17.80	589.08
G1	Main stream	−7.31	−44.98	437.76	91.70	15.50	21.30	3.26	254.00	28.20	23.80	120.13
H1	Main stream	−7.79	−48.57	443.33	87.20	15.50	24.80	2.63	260.00	27.00	26.20	562.62

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

chemical interaction between water and rocks was lower in river water than in groundwater. However, one sample collected from the lower part of the Daotang River (location D1) had a TDS of 1014.11 mg/L (Table 2), with a hydrochemical type of Na^{2+}-Cl^{-} (Fig. 5). This suggested that river water underwent relatively heavy evaporation in the lower part of the Daotang River Catchment, where water flowed more slowly.

The boomerang envelope model, developed by Gibbs (1970), describes and classifies natural water chemistry controls into three types: rock dominance, evaporation/crystallization dominance and precipitation dominance (Machender et al., 2014). As shown in Fig. 8, most samples fall within the mid-upper branch of the Gibbs boomerang envelope, suggesting that rock-weathering dominated the river water chemistry. Fig. 8 also showed that the water collected from location D1 (lower of the Daotang River) experienced heavy evaporation. Overall, the TDS values, hydrochemical type and boomerang envelope model all suggested that rock weathering, ion exchange and precipitation were the major geochemical processes responsible for the river water solutes within the Qinghai Lake Basin. As such, the effects of human

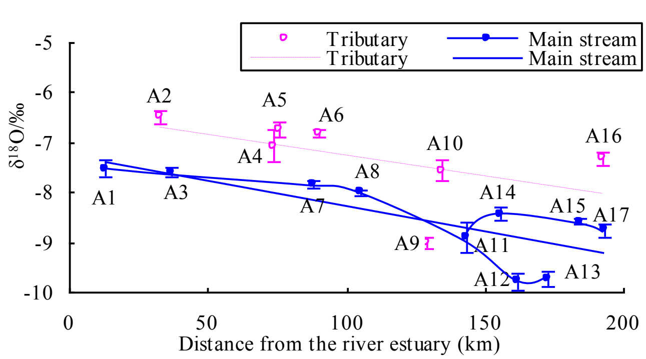

Fig. 4. Changes of stable isotope in main stream water and tributary water in Buha River.

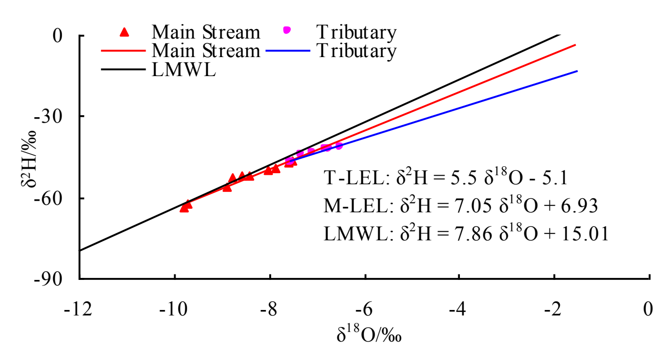

Fig. 5. LEL of main stream water and tributary water in Buha River. (LMWL: local meteoric water line; LEL: local evaporation line of river water).

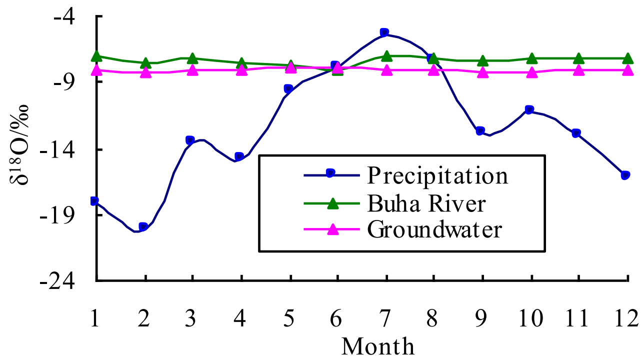

Fig. 6. Monthly $\delta^{18}O$ in river water (location A1), precipitation and groundwater around Qinghai Lake.

activity on water chemistry were relatively mild in the basin (Xu et al., 2010).

3.4. Sources of river water and runoff processes

Glaciers, covering an area of 13.29 km^2, lie mainly in the upper reaches of Buha River in the Qinghai Lake Basin. Glacial meltwater runoff is about 0.1×10^8 m^3/y\ (LZBCAS, 1994), accounting for 1.24% of the Buha River's annual runoff. This suggested that glacier influence on river runoff in the Qinghai Lake Basin should be negligible. Fig. 5 showed that the Buha River water mainly came from precipitation that had undergone weak evaporation. Therefore, to eliminate the influence of evaporation, the intersection between the LEL of river water in Buha River ($\delta^2H = 6.55\ \delta^{18}O + 2.53$ ($n = 17$, $R = 0.99$)) and the LMWL of precipitation ($\delta^2H = 7.86\ \delta^{18}O + 15.01$ ($n = 124$, $R = 0.99$)) were calculated. The δ^2H and $\delta^{18}O$ values on the intersection ($\delta^{18}O$: −9.53‰; δ^2H: −59.89‰) were the average initial recharge isotope values from the precipitation to the river water. The Buha River water's average recharge altitude was then calculated using the altitude effect of precipitation and Equation (2), and found to be approximately 4250 m.a.s.l., higher than the average altitude (approximately 3970 m.a.s.l.) of the Buha River Catchment.

It was assumed that the discharge in location A1 was the total discharge of Buha River. Ignoring channel evaporation, the contribution rates of each mainstream and tributary were computed using two and three-component hydrograph separation analysis (Kong and Pang, 2012). The drainage area above location A11 represented approximately 40.3% of the Buha River Catchment, However, runoff generated more than half of the Buha River's discharge (57.9%) (Table 3). This result mirrored the result obtained at the LEL and LMWL intersection (the average recharge altitude of Buha River was about 4250 m.a.s.l.); the Buha River's discharge came mainly from the middle and upper catchment. 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). As a similar region comparison, more than 50% of the total runoff of the Heihe River in northwest China was also generated in the middle and upper mountain zone (Wang et al., 2009; Qin et al., 2013).

In regions with alpine grassland vegetation, different grasslands types led to different runoff generation regimes (Wang et al., 2012; Qin et al., 2013). The main types of grasslands on the whole Qinghai-Tibet Plateau included alpine swamp and alpine meadow (Wang et al., 2007, 2013). Alpine swamp vegetation had a lower maximum interception ratio and saturation precipitation than alpine meadow vegetation (Wang et al., 2007), indicating that the runoff coefficient of alpine swamp cover was higher than alpine meadow cover. Due to the middle and upper parts of the Qinghai

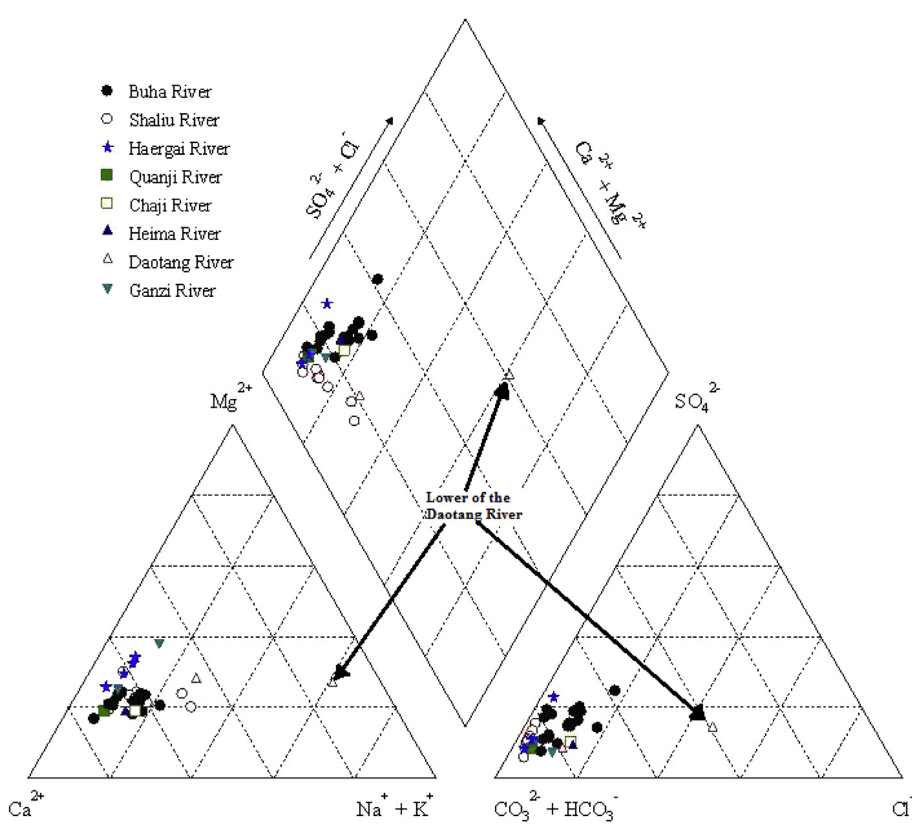

Fig. 7. Ternary plots of cations and anions of river waters in the Qinghai Lake Basin.

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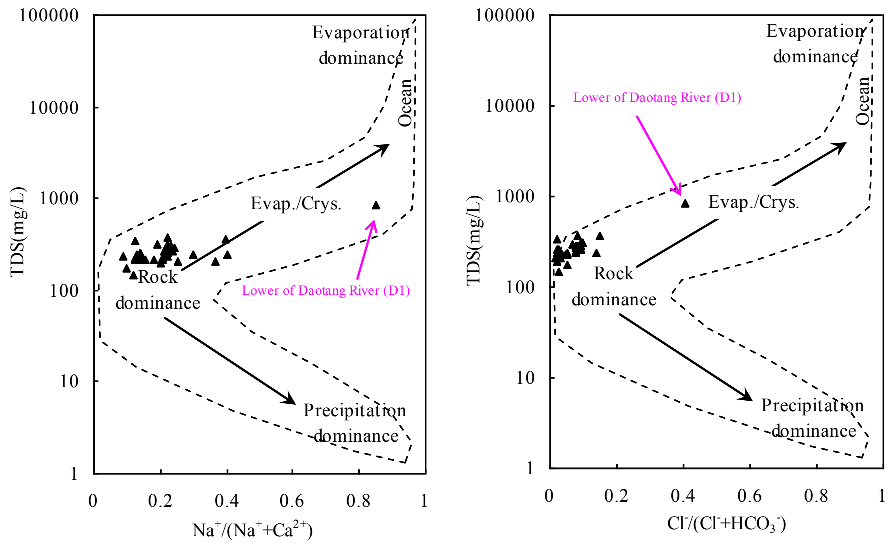

Fig. 8. Gibbs boomerang model for river waters in the Qinghai Lake Basin.

Table 3
Runoff percentage of the main stream (M) and tributaries (T) in Buha River.

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	1116.3	1832.6	8658.6	1309.3	5891.4	3309.2	2179.4	338.8	1023.4
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

Lake Basin covered mainly by the alpine swamp (Cui, 2011), the runoff coefficient in the middle-upper part would be higher. This land use/cover characteristic would be another reason why the river discharge came mainly from the middle and upper basin in the Qinghai Lake Basin.

3.5. Impact of runoff on lake level

The runoff of Buha River fluctuated, with no clear increasing or decreasing trend from 1961 to 2012 (Fig. 9). The runoff increased in the 1980s and 2000s and decreased in the 1970s and 1990s. The maximum and minimum amounts of annual runoff were 19.47 $\times$ 10^8 m^3 in 1989 and 1.99 $\times$ 10^8 m^3 in 1973, respectively (Fig. 9). Runoff changes were divided into three stages: fluctuating and decreasing stage (1961–1980), with a decreasing rate of 0.34 $\times$ 10^8 m^3/y; fluctuating stage (1981–1995); and fluctuating and increasing stage (1996–2012), with an increasing rate of 0.44 $\times$ 10^8 m^3/yr. The annual precipitation fluctuated similarly with runoff of Buha River (Fig. 9), with increasing peaks in 1967, 1989 and 2006; and decreasing peaks in 1969, 1990, 1995 and 2007. The

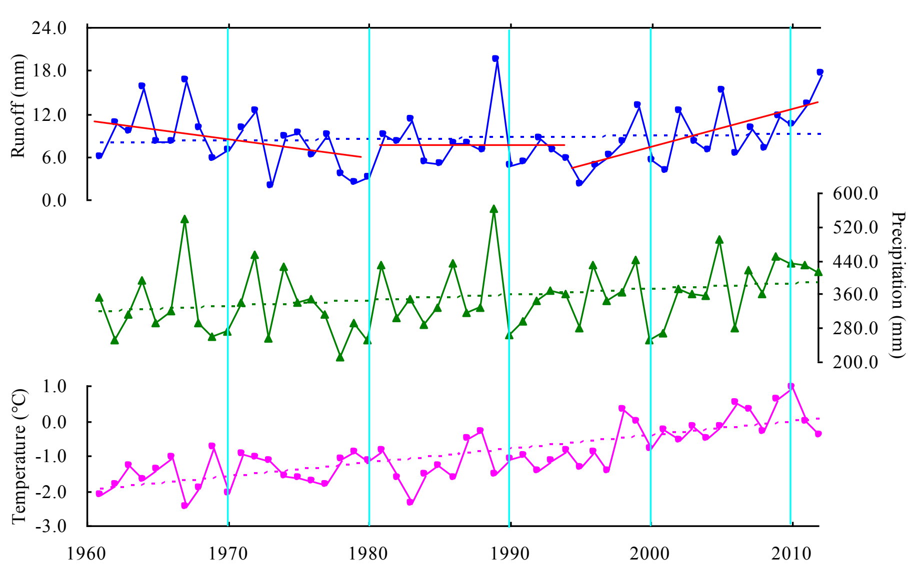

Fig. 9. Time series of the Buha River runoff, precipitation and temperature at the Tianjun weather station between 1961 and 2012.

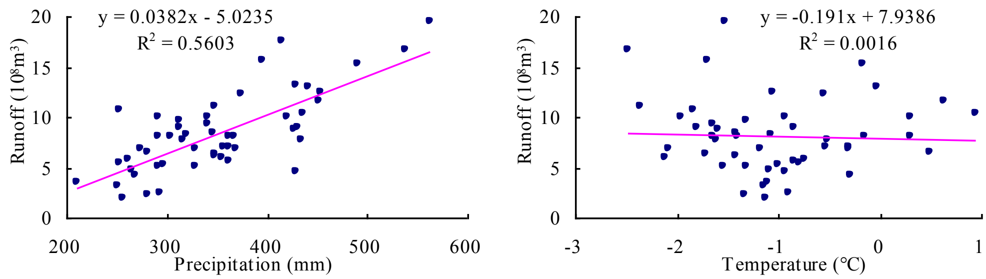

Fig. 10. Relationship between runoff of Buha River, precipitation and temperature.

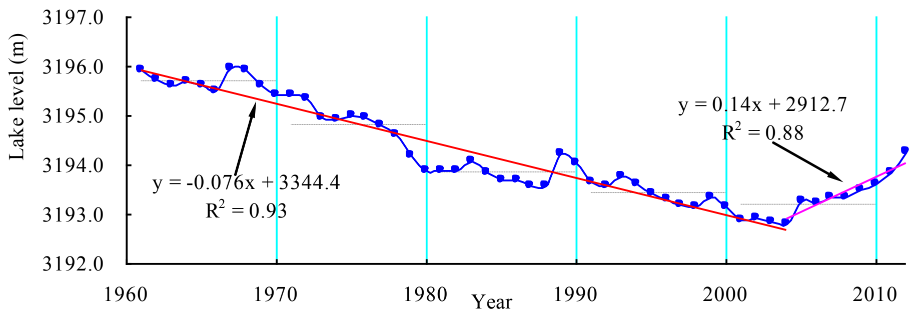

Fig. 11. Time series of lake level of Qinghai Lake between 1961 and 2012.

annual temperature for the same period, however, showed an overall increase, with a clear warming trend in the 1990s and 2000s (Fig. 9). This increasing trend was similar to the "Three-River Headwaters" regions (the Yangtze River, the Yellow River, the Lancang River) and the Tibetan Plateau (Wang et al., 2008; Lan et al., 2010; Zhang et al., 2011).

The similar fluctuations between runoff and precipitation, and different fluctuations between runoff and temperature indicated that the different influences that temperature and precipitation had on runoff. Precipitation changes directly affected the generation of surface runoff, whereas temperature changes affected evaporation of surface runoff (Zhang et al., 2011). Runoff variation was positively correlated with precipitation (correlation coefficient of 0.75) from 1961 to 2012, a statistically significant result ($p < 0.001$; Fig. 10). However, the correlation coefficient between runoff and temperature was not statistically significant. These results showed that river runoff was more sensitive to precipitation than to temperature in the Qinghai Lake Basin.

The Qinghai Lake level decreased from 1961 to 2012 (3195.93 m a.s.l. in 1961 and 3194.26 m a.s.l. in 2012) (Fig. 11). Lake level could be segmented into two phases: (1) a phase with a decreasing rate of −7.6 cm/y from 1961 to 2004, with severe declines in the 1970s and early 1990s); and (2) a phase with a rising rate of 14 cm/y from 2004 to 2012 (long-term linear trends at a $p < 0.001$ significance level). A maximum decline of 0.41 m was recorded in 1979; the maximum rise of 0.64 m was recorded in 1989 (Fig. 12). Comparing the variation of lake level with the runoff of Buha River (Fig. 12), the rising peaks of lake level corresponded to the highest runoff (e.g., 1967, 1983, 1989, 1999, 2005, 2012); vice versa, the decreasing peaks of lake level corresponded to the lowest runoff (e.g., 1969, 1973, 1979, 1995, 2001). The correlation coefficient between runoff and lake level variation was 0.84 ($p < 0.001$). When considered with the relationship between precipitation and runoff (Figs. 9 and 10), the change of the Qinghai Lake's level was sensitive to both precipitation and river runoff. In general, lake level increase was closely related to river runoff and precipitation increases, similarly, lake level decrease was closely related to declines in river runoff and precipitation. Previous studies suggested that any further large-scale climate warming would likely be associated with a greater moisture supply in the

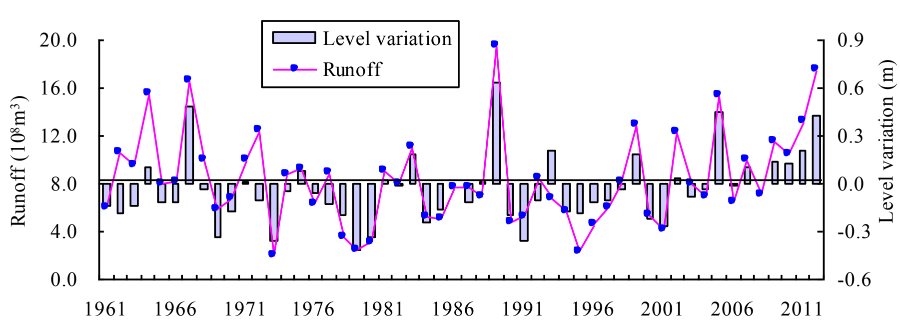

Fig. 12. Relationship between the Buha River runoff and Qinghai Lake water level variation.

northeastern Tibetan Plateau (Yang et al., 2014). If so, the river runoff would likely increase, and lake's level would rise due to increasing precipitation and runoff on the Tibetan Plateau (Christensen et al., 2014; Song et al., 2014).

4. Conclusions

The river water in the Qinghai Lake Basin came mainly from precipitation and evaporated only slightly. Tributary river water experienced stronger evaporation than the main stream. The river water's hydrochemical type was mainly Ca^{2+}–Mg^{2+}-HCO_3^-; river water chemistry was mainly controlled by carbonate weathering in the Qinghai Lake Basin.

River discharge was generated mainly from the Qinghai Lake Basin's middle and upper basin reaches, with high precipitation and low evapotranspiration, and covered by alpine swamp. River runoff was more sensitive to precipitation than to temperature. Qinghai Lake level increased with increases in river runoff and precipitation, and declined with decreases in river runoff and precipitation. The lake level could rise due to increasing precipitation and runoff in the future.

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