Contribution of cryosphere to runoff in the transition zone between the Tibetan Plateau and arid region based on environmental isotopes

Juan Gui [1,2], Zongxing Li [1*], Qi Feng [1], Qiao Cui [1], Jian Xue [1]

1. Key Laboratory of Ecohydrology of Inland River Basin/Gansu Qilian Mountains Eco-Environment Research Center, Northwest Institute of Eco-Environment and Resources, Chinese Academy of Sciences, Lanzhou 730000, China
2.University of Chinese Academy of Sciences，Beijing 100049，China

*Corresponding author: Tel: 86+13919887317, E-mail: lizxhhs@163.com (Li Zongxing).

Abstract: As the transition zone between the Tibetan Plateau and arid region, the Qilian Mountains are important ecological barriers and source regions of inland rivers in northwest China. In recent decades, drastic changes in the cryosphere have had a considerable impact on the formation process of water resources in the Qilian Mountains. In this study, 2,164 environmental isotope samples were used to quantify the runoff components of 11 major rivers in the Qilian Mountains and investigate the influence of cryosphere changes on mountain runoff. The results show that the mountain runoff primarily originates in the cryosphere belt, which contributes approximately 82%, 71%, and 80% to the Hexi inland water system, upper stream of the yellow river system, and Qinghai inland river system, respectively. The maximum contribution ratio of glacier and snow meltwater to runoff occurred in May. The contribution of supra-permafrost water to runoff gradually increased from May and reached approximately 40% in some rivers in October. Cryosphere degradation in the Qilian Mountains after the 1990s has caused a rapid increase in runoff, changes in the peak runoff time, and an increase in winter runoff. These

changes in the hydrological processes bring opportunities and challenges to managing inland river water resources, and various adaptive measures to seek advantages and avoid disadvantages have been proposed. The findings from the environmental isotope analysis provide insights into understanding water resources and realizing the nexus of life, agriculture, industry, and ecological water use.

Key Words: Runoff components; Stable isotope; Cryosphere degradation; Qilian mountains

1. Introduction

Characterizing the storage and streamflow-release mechanisms of precipitation in watersheds provides important hydrological information of a specific region (Miller et al., 2021) Particularly, in the current context, global freshwater resources are under increasing stress due to climate change and growing populations(Florke et al., 2018). Mountains act as natural reservoirs of freshwater in the form of snow and glaciers and are therefore the principal components of the global water supply. Numerous studies have shown that runoff from glacier discharge provides a valuable resource to downstream populations for domestic, agricultural, and other allied activities (Kriegel et al., 2013). Shortages and low use efficiencies of water resources have resulted in water-related issues in China, which will be further exacerbated by the continuous loss of glaciers and wetlands. Glacierized catchments in mountainous regions are generally headwater catchments, which are of great interest due to their complex runoff-generation processes and supply of water to downstream regions (Immerzeel et al., 2010). In various headwater catchments, seasonal water availability is strongly dependent on cryospheric processes; which become even more relevant in a changing climate (IPCC, 2013). In the last several decades, the

temperature of the permafrost in the Qinghai–Tibetan Plateau has significantly increased, and the region has experienced widespread degradation. Snow-covered areas have also been decreasing in the past 50 years (Yao et al., 2013), as a result, the runoff process has undergone significant changes in seasonal flow, flood peak discharge, and total runoff (Yang et al., 2000). Therefore, it is imperative to understand the source, processes, and mechanisms of runoff, particularly in seasonally arid mountainous areas, such as in Western China, where snowmelt is vital for downstream human activities (Milly and Dunne, 2020) and managing and sustaining water resources in a changing environment. However, runoff generation processes are complex and difficult to quantify, particularly, before an investigation is conducted (Uhlenbrook et al., 2002).

Environmental tracers are commonly used to investigate hydrological processes. Stable isotopes in water (^{2}H and ^{18}O) are important components of natural water bodies and powerful tools for investigating the water cycle and hydrological processes. Although their natural abundance is low, they are highly sensitive to changes in their environment and can indicate the source, migration, and transformation of water and are therefore ideal tracers for components of the water cycle (Gat et al., 1996; Bowen et al., 2019; Xianfang et al., 2007). The stable isotope ratios of hydrogen and oxygen in water samples can provide essential information about water dynamics within a given watershed (Rock and Mayer, 2007). For inland river basins with widely distributed glaciers and permafrost areas, stable isotope tracing is particularly useful, as direct, continuous field observations of hydrological processes are extremely difficult owing to the harsh environment(Cui & Li, 2015;Li et al., 2015). In the past few decades, isotopic tracers such as oxygen and deuterium isotopes have increasingly been utilized in conjunction with geochemical tracers and

hydrometric measurements to separate flow pathways and provide more information regarding temporal and geographic sources of runoff in temperate, humid, and arid environments(Hooper & Shoemaker, 1986; Li, et al., a, b). McDonnel et al., (1991) found that stream water in New Zealand was partially supplied by subsurface flow in humid zones. Mortatti et al., (1997) reported that the average surface runoff and baseflow (pre-event) contributions were 30.3% and 69.7%, respectively, in the Amazon River. This method has been used to estimate the changes in glacier meltwater in many different regions, for example, the contribution of glacier and snowmelt water to runoff in spring was found to be as high as 82% in a cold area in Colorado, USA (Liu et al., 2004). These studies found that glacier changes have a significant impact on the runoff of important rivers in cold regions. Climate warming has also significantly impacted on the cold regions of Western China where the annual average temperature has increased by 0.28 °C/10 a during the period of 1961–2016 (Li et al., 2019), which has caused the glaciers in the study area to melt rapidly.

The Qilian Mountains are important ecological barriers and environmentally functional areas in northwest China (Jia, 2012), which are not only the source region of the Hexi inland river system (HIRS), including the Shiyang, Heihe, and Shule River Basins, from southeast to northwest, but also the source regions of the Qinghai inland river system (QIRS), including the Qinghai and Hala Lake Basins and upper stream of the Yellow River system (USYR). In the Qilian Mountains, the glacier area was 2,017.81 km^2 in 1956, 1,761.3 km^2 in 1990, and 1,597.1 km^2 in 2010 (Liu et al., 2003; Wang et al., 2011; Sun et al., 2015). The permafrost area has continuously decreased in the Tibetan Plateau over the past 50 years (Cheng et al., 2012). Drastic changes in the cryosphere are bound to have a

significant impact on the quantity and formation process of water resources in the inland rivers of the Qilian Mountains, which are fed by glacial meltwater.

Against the background of climate warming and the continuous retreat of the cryosphere, it is urgency to understand the runoff replenishment amount of the cryosphere and runoff formation process in the study region for decision-making purposes regarding the rational development and utilization of water resources in river basins. However, due to the lack of data and observation and sampling related challenges in cold regions, previous studies have only focused on the statistical analysis of runoff and its associated influencing factors; hence, there is a lack of in-depth knowledge of the mechanism of the temporal and spatial variations in runoff components from a microscopic point of view, while the quantification of the impacts of the cryosphere meltwater on runoff in the Qilian Mountains remains ambiguous. The Danghe, Changma, Taolai, Heihe, Xiying, Nanying, Zamu, Jinqiang, Datong, Huangshui, and Buha Rivers were selected as the research objects in this study. The objectives of this study were as follows: (a) To understand the spatial and temporal differences of stable isotopes in various water bodies in the Qilian Mountains; (b) To quantify the runoff components of major rivers using $\delta^{18}O$ and deuterium excess (d-excess) as a proxy in the Qilian Mountains; and (c) To determine the impact of cryosphere degradation on water resources in inland river basins. The results of this study provide insights into the sources of streamflow and stream water fractions and can provide a more effective understanding of the internal hydrological processes in a depopulated alpine zone under the impacts of climate change.

2. Materials and methods

2.1 Study region

The Qilian Mountains are located on the northeastern edge of the Tibetan Plateau and consist of a series of NW trending parallel mountains and valleys with latitudes ranging from 36–43°N and longitudes of 92–107°E (Fig. 1). The average annual temperature in the Qilian Mountains is between -5.25~10.75 °C. The average annual precipitation in the Qilian Mountain ranges from 34.23 to 493.97 mm and increases gradually from west to east (Lv et al., 2019). Rivers are widely distributed with radial drainage characteristics in the Qilian Mountains, and the Leng Longling Range divides all rivers into internal flow and outflow systems. The outflow rivers mainly include the Huangshui and Datong Rivers, which are located in the USYR. The inner flow is divided into three parts: the Qaidam, Qinghai Lake, and Hexi Corridor Basins. Among these rivers, the Danghe, Taolai, Heihe, Xiying, Nanying, and Zamu Rivers belong to the HIRS, and the Buha River belongs to the QIRS. The southern rivers of the Qilian Mountains primarily flow into the Qaidam Basin, while the northern rivers flow into the Hexi Corridor (Deng Shaofu, 2013). Generally, the annual distribution of surface runoff is consistent with the precipitation process and high-temperature season in the Qilian Mountains. Runoff and precipitation are concentrated in the warm season in the area, and it is mainly recharged by ice and snow meltwater and groundwater in spring and precipitation in summer. The annual variation in runoff shows an evident periodic trend as detailed by Wang Jinye (2006). The rivers in the Qilian Mountains are mainly fed by glacial meltwater. Water originates in the cryosphere belt, which contains glaciers, snow, and permafrost (Zongxing et al., 2019). According to the second glacier inventory of China and permafrost dataset of the Qinghai-Tibet Plateau, there are 2,859 glaciers in the Qilian Mountains with a total area of 1,597 km^2; the total area of glacier reserve and permafrost are 84.48×10^8 m^3 (sun et al., 2015) and 9.39×10^4

140 km^2 (Zhou et al., 2000), respectively. The inter-annual variation in runoff is small, but there
141 are significant seasonal and daily variations. River runoff in the western and middle
142 sections of the Qilian Mountains has increased significantly since 1980, whereas the
143 eastern section has shown a slight decreasing trend (Zhang et al., 2018).

144 **2.2. Sample collection and analysis**

145 In addition to river water samples, precipitation, glacier and snow meltwater, super-
146 permafrost water, groundwater, and outlet river water were continuously collected in the
147 Qilian Mountains at the sampling sites shown in Fig. 1. The details of the samples are as
148 follows:

149 **River water (338)**: A sampling network of 11 river water sampling stations was
150 established in the Qilian Mountains. The Danghe, Xiying, Taolai, Heihe, Nanying, Zamu,
151 and Changma Rivers belong to the HIRS; three samples were collected every month at the
152 river outlets, and a total of 252 samples were collected from September 2018 to August
153 2019. The Datong, Huangshui, and Jinqiang Rivers belong to the USYR; water from the
154 river outlets was collected twice a month, and a total of 62 samples were collected from
155 July 2017 to July 2018 (no samples were collected from the Jinqiang River in January,
156 February, March, and December). Similarly, from August 2020 to August 2021, a total of
157 24 river water samples were collected from the Buha River in the QIRS.

158 **Precipitation (1310):** An observation network of precipitation stable isotope
159 including 19 sampling stations in the Qilian Mountains was established (Fig. 1), and 1310
160 groups of precipitation samples were collected during 2012–2018 in Menyuan, Huangyuan,
161 Gulang, Qilian, Yeniugou, Tuole, Minle, Gangcha, Tianjun, Jiayuguan ,Jiutiaoling, Xidahe,
162 Changma, Daiqian, Anyuan, Suli, Ebao ,Hulugou and Binggou . All the precipitation events

were collected according to the definition of precipitation event stipulated by meteorological observation, that is, all the precipitation from 20:00 of the same day to 20:00 of the next day was defined as a precipitation event, and a sample was collected. A total of 1,310 groups of precipitation samples have been collected during the period of 2012–2018. For more information, please refer to Juan et al., (2020)

Glacier and snow meltwater (96): The hydrochemical characteristics of glacier and snow meltwater in the cryosphere were analyzed. Glacier samples were collected every half a month from May to October. The snow covering the glaciers begins to melt in March and April as the weather warms, and by May, the snow is almost entirely melted. By June, the snow melts and flows into the channel as glacial meltwater. Essentially, snow cover at the end of the glacier has little effect on runoff segmentation.

The samples were taken at 14:00 every day, and the sampling location was in the hydrological section at the end of the glacier. During the sampling period, a total of 84 glacier meltwater samples were collected from the end of the Bayi, Shiyi, and source glaciers of the Taolai, Shule, Danghe, Shiyang, and Datong Rivers. Twelve snow meltwater samples were also collected in the Buha and Huangshui Rivers.

Supra-permafrost water (108): Supra-permafrost water, the most widely distributed type of groundwater, is mainly stored in the permafrost active layer. To determine the hydrochemical characteristics of supra-permafrost water in the study area, water samples were collected by comprehensive sampling from May to October in 2016 and 2018; sampling was performed manually. A 2 m deep profile of the permafrost active layer was dug at each sampling point. The water samples were then immediately filtered through a 0.45 μm Millipore filtration membrane and poured into clean polyethylene bottles. During

this period, a total of 108 samples were collected from the Danghe, Changma, Heihe, Taolai, Shiyang, Datong, and Buha Rivers.

Groundwater (312): A total of 240 samples were collected weekly from five wells in the Danghe, Changma, Taolai, Heihe, and Shiyang Rivers of the HIRS within one hydrological year. Similarly, another 48 samples were collected twice a month from the Datong and Huangshui Rivers in the USYR. Lastly, 24 samples were collected from the Buha River in the QIRS.

Meteorological and hydrological data: Meteorological and hydrological data for the Dang, Changma, Taolai, Heihe, Xiying, Nanying, Zamu, Datong, and Jinqiang Rivers were obtained from the Hydrology and Water Resources Bureau of Gansu Province (HWRBGS). Runoff data for the Huangshui River were obtained from Tiaofeng et al., (2014), and Buha River runoff data were obtained from Liu et al. (2020). The details are provided in Table 1.

2.3 Methods

2.3.1 Sample testing

Water sample analyses were carried out in the Key Laboratory of Eco-hydrology of Inland River Basins, Chinese Academy of Sciences to determine the presence of hydrogen and oxygen stable isotopes. For the analysis of the stable isotopes of soil water, water was extracted from the soil using a cryogenic freezing vacuum extraction system (LI-2000, Beijing Liga United Technology Co., Ltd., China), which can achieve complete extraction with high precision. The stable isotopes in the water samples were measured using a liquid water stable isotope analyzer (Model DLT-100, Los Gatos Research, Inc., Mountain View, CA, USA). The accuracies of $^{18}O/^{16}O$ and D/H were 0.2‰ and 0.5‰, respectively, which conform to the rule of valid digits for stable isotope analysis. The results were calibrated

using the Vienna Standard Mean Ocean Water (V-SMOW) and laboratory working standards. The final results were expressed in the form of micro-differences relative to V-SMOW:

$$\delta^{18}O(or\delta D) = \left[\frac{R_{Sample}}{R_{V\text{-}Smow}} - 1\right] \times ‰ \quad (1)$$

where R_{Sample} is the precipitation sample and $R_{V\text{-}Smow}$ is the ratio of the oxygen or hydrogen stable isotope in V-SMOW.

The d-excess can be defined as follows:

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

The value of d-excess is equivalent to the intercept when the slope of local meteoric water is 8, which represents the imbalance degree of the evaporation process.

2.3.2 End-member mixing analysis

Hooper (2003) introduced the end-member mixing analysis (EMMA) using chemical/isotopic compositions in waters. The techniques involve graphical analyses, in which chemical and isotopic parameters are used to represent the designated end members. Tracer concentrations are constant in space and time. Essentially, the change in the composition of the water can be attributed to the intersections it makes during its passage through each landscape zone. Tracers can be used to determine both sources and flow paths. The EMMA tracer approach is a conventional method for analyzing potential water sources contributing to streamflow (Li, et al., 2016 ; Zongxing et al., 2014). In this study, a three end-member mass-balance mixing model was employed to calculate the contribution of three water sources in stream water according to Eq. 3a and b below:

$$X_S = F_1X_1 + F_2X_2 + F_3X_3 \quad (3a)$$

$$Y_S = F_1Y_1 + F_2Y_2 + F_3Y_3 \quad (3b),$$

where X and Y represent the concentrations of two types of different tracers. In this study,

232 the concentrations of $\delta^{18}O$ and deuterium excess were chosen for comparison; subscripts 1,
233 2, and 3 represent the water from the respective contributions of source waters 1, 2, and 3
234 (end members) to stream water; F denotes the fraction of each end-member; and the
235 solutions for F_1, F_2, and F_3 in regards to tracer concentrations in Eq. (3) can be given as:

236 $$F_1 = [(X_3-X_S)/(X_3-X_2)-(Y_3-Y_S)/(Y_3-Y_2)]/[(Y_1-Y_3)/(Y_3-Y_2)-(X_1-X_3)/(X_3-X_2)] \quad (4a)$$

237 $$F_2 = [(X_3-X_S)/(X_3-X_1)-(Y_3-Y_S)/(Y_3-Y_1)]/[(Y_2-Y_3)/(Y_3-Y_1)-(X_2-X_3)/(X_3-X_1)] \quad (4b)$$

238 $$F_3 = 1-F_1-F_2 \quad (4c)$$

239 This method has been used in previous studies (Zongxing et al., 2014; Li et al., 2015)
240 to evaluate the contribution of possible sources to the river water.

241 **2.3.3 Uncertainty in hydrograph separation**

242 The uncertainty of tracer-based hydrograph separations can be calculated using the error
243 propagation technique (Genereux, 1998;Klaus & McDonnell, 2013); this approach considers
244 errors of all separation equation variables. Assuming that the contribution of a specific
245 streamflow component to streamflow is a function of several variables c1, c2, …, cn and
246 the uncertainty in each variable is independent of the uncertainty in the others, the
247 uncertainty in the target variable (e.g., the contribution of a specific streamflow component)
248 is estimated using the following equation (Uhlenbrook et al., 2002):

249 $$W_{fx} = \sqrt{\left(\frac{\partial z}{\partial c_1} W_{c1}\right)^2 + \left(\frac{\partial z}{\partial c_2} W_{c2}\right)^2 + \cdots + \left(\frac{\partial z}{\partial c_n} W_{cn}\right)^2}, \quad (5)$$

250 where W represents the uncertainty in the variable specified in the subscript; and f_x is the
251 contribution of a specific streamflow component, x, to streamflow. The software package
252 MATLAB was used to apply equation 3 to the different hydrograph separations in this study.
253

3. Results

3.1 Characteristics of stable isotopes in different waters

3.1.1 Precipitation

The $\delta^{18}O$ in the precipitation was characterized by pronounced seasonal variations, with maximum values in summer and minimum values in winter. The monthly average values varied from -21.62‰ to -5.15 ‰, with an annual average of -12.11‰. From January to December, $\delta^{18}O$ showed an initial increasing trend, after which it decreased, with the maximum value occurring in June. Owing to the combined influence of the water vapor source, temperature, precipitation, and other factors, the stable isotopes of the precipitation fluctuated greatly during the year (Fig. 2a).

The HIRS is located on the northern slope of the Qilian Mountains, and the monthly average $\delta^{18}O$ in the precipitation ranged from -21.26‰ to -5.03‰, with an annual average value of -12.21‰ (Fig. 2b). In the USYR and QIRS, located on the southern slope, the range of monthly average $\delta^{18}O$ values were -17.72–-3.85‰ and -27.32–-5.51‰, respectively, with annual average values of -10.53‰ and -11.57‰, respectively (Fig. 2). Compared with the HIRS and QIRS, the average value of $\delta^{18}O$ in the USYR was more positive. Different water vapor sources and climatic and topographic conditions resulted in evident spatial and temporal differences in stable isotopes in the precipitation in the study area.

3.1.2 Glacier and snow meltwater

Glacier and snow meltwater were collected from seven basins in the study area from May to October. As shown in Fig. 3, the mean $\delta^{18}O$ in glacier and snow meltwater of the Qilian Mountains was -9.61‰, which is significantly lower than the corresponding river

277 water. The seasonal fluctuation of glacier and snow meltwater was very low; although there
278 were differences among several river systems, they were not significant. During the
279 sampling period, the $\delta^{18}O$ in glacier and snow meltwater in the HIRS varied from -13.07‰
280 to -7.97‰, with an average value of -9.69‰. The $\delta^{18}O$ values of the glacier and snow
281 meltwater in the Danghe River, Changma River, Shiyi Glacier, Bayi Glacier, Taolai River,
282 and Shiyang River were -10.64‰, -9.31‰, -9.86‰, -9.77‰, -9.29‰, and -9.25‰,
283 respectively. The mean $\delta^{18}O$ values of glacier and snow meltwater increased from west to
284 east. In the USYR, the $\delta^{18}O$ in glacier and snow meltwater in the Datong River ranged from
285 -10.24‰ to -8.06‰, with a mean value of -9.54‰. In general, from May to October,
286 fluctuations in $\delta^{18}O$ were low, and the overall trend first increased and then decreased, with
287 the maximum value appearing in August and minimum value in May. In the QIRS, the $\delta^{18}O$
288 in glacier and snow meltwater in the Buha River ranged from -11.69‰ to -8.26‰, with a
289 mean value of -9.24‰. These $\delta^{18}O$ values in the QIRS were higher than those in the HIRS
290 and USYR system and fluctuated greatly during the study period.

291 The $\delta^{18}O$ values in glacier and snow meltwater were relatively stable during the
292 sampling period and did not show any notable temporal variations. The maximum $\delta^{18}O$
293 values in the HIRS, USYR system, and QIRS occurred in May, August, and July, while
294 the minimum values were in June, May, and August. The time that the maximum and
295 minimum values of $\delta^{18}O$ occurred in the three river systems was not consistent, which may
296 be attributed to the start time of glacier and snow melting and storage. The d-excess of
297 glacier and snow meltwater also showed no evident seasonal characteristics during the
298 sampling period. The d-excess in the HIRS, USYR system, and QIRS ranged from 19.33‰
299 to 22.06‰, 14.17‰ to 22.90‰, and 11.38‰ to 11.57‰, respectively, with mean values of

300 21.53‰, 17.42‰, and 11.47‰, respectively.

301 **3.1.3 Supra-permafrost water**

302 Supra-permafrost water is groundwater that exists mainly in the permafrost active
303 layer, which forms an important part of the runoff in cold regions (Li et al., 2016a). The
304 $\delta^{18}O$ value in the supra-permafrost water was significantly higher than those of glacier and
305 snow meltwater (Fig. 4); this was noted because during the study period, the supra-
306 permafrost water was subjected to higher temperatures, which resulted in intense
307 evaporation. In the HIRS, the mean $\delta^{18}O$ values in the supra-permafrost water ranged from
308 -7.16‰ to -6.49‰, with an average of -6.69‰. The $\delta^{18}O$ in the supra-permafrost water in
309 the Heihe River was the most stable, with almost no fluctuations from May to October. The
310 Danghe River showed slight fluctuations, with the maximum value occurring in September.
311 In the Shiyang River Basin, $\delta^{18}O$ values were negative from May to June, positive from
312 July to August, and negative from September to October, whereas in the Changma River,
313 they were positive, with a maximum value of -5.12‰ in June and minimum of -7.15 ‰ in
314 October. In the USYR system, the $\delta^{18}O$ in the supra-permafrost water ranged from -7.61‰
315 to -6.87‰, with an average value of -7.34‰. The values in the Datong and Huangshui
316 Rivers of the USYR system were significantly different; the values of the Datong River
317 were relatively positive, with an average value of -6.34‰, while those of the Huangshui
318 River were relatively negative, with an average value of -8.34‰. In the Buha River of the
319 QIRS, $\delta^{18}O$ values were relatively stable, and the weak temporal variation showed peaks
320 in June and September. In conclusion, the $\delta^{18}O$ in the supra-permafrost water of the three
321 river systems was generally positive and did not show notable temporal variations. This
322 phenomenon can be explained as follows: Firstly, supra-permafrost water is mainly stored

in the active layer of the permafrost, and under strong evaporation (noted during the study period), the stable isotope concentration becomes unbalanced through the influence of dynamic fractionation. Secondly, the supra-permafrost water is then replenished by a mixture of precipitation and glacier and snow meltwater, which results in random fluctuations in the stable isotope concentrations. The range of the d-excess values for the supra-permafrost water for the corresponding river systems were 10.03~17.69‰, 13.56~15.73‰, and 11.43~14.87‰, respectively, with mean values of 13.36‰, 14.29‰ and 12.93‰, respectively (Fig. 4).

3.1.4 River water

As shown in Fig. 5, the $\delta^{18}O$ in the river water in all basins in the study area was relatively stable, and had significant seasonal variation characteristics, which differed from that of precipitation. The $\delta^{18}O$ values in the different river systems in the Qilian Mountains were relatively stable and did not show evident seasonal variations, although there were some differences. The range of $\delta^{18}O$ values were -8.75–-8.27‰, -8.97–8.53‰, -8.15–7.26‰, and -9.47–7.96‰ for the Qilian Mountains, HIRS, USYR system, and QIRS, respectively, with corresponding mean values of -8.49‰, -8.72‰, -7.79‰, and -8.49‰. Although the $\delta^{18}O$ value in rivers was relatively stable, there were slight variations from the annual mean value among the different river systems: USYR > QIRS > HIRS (Fig. 5a).

The maximum monthly mean values of $\delta^{18}O$ in the HIRS, USYR, and QIRS were in September, February, and October, respectively, while the minimum monthly mean values were in May, July, and June, respectively. Owing to the significant distance from west to east in the HIRS, the $\delta^{18}O$ values in the different rivers varied greatly; for example, in the Danghe River, in May and October, there was a higher depletion of $\delta^{18}O$ (-11‰), while in

February, it was more enriched (-9‰). The $\delta^{18}O$ in the Shule River water in the western segment fluctuated greatly compared with that of the three rivers in the Shiyang River Basin in the eastern segment. The Heihe River Basin in the middle reaches had the highest average value of $\delta^{18}O$, followed by the Shiyang River Basin in the eastern part, while that of the Shule River Basin in the western part was significantly negative, with an average value of -9.97‰. The $\delta^{18}O$ value in the river water of the Heihe River, located in the middle of the Qilian Mountains, was positive, with an annual average of -7.75‰ and -6.79‰ in April.

In the USYR system, except in the Jinqiang River, the river water $\delta^{18}O$ values of the Huangshui and Datong Rivers were relatively positive, with mean values of -7.76‰ and -7.58‰, respectively. The $\delta^{18}O$ values of the river water in the Jinqiang River were relatively negative, with large fluctuations. In the QIRS, because of the large extent of uninhabited areas, sampling was conducted only in the Buha River, in which the $\delta^{18}O$ values ranged from -9.47‰ to -7.96‰, with a mean of -8.59‰. In terms of its weak seasonal variation, the maximum $\delta^{18}O$ occurred in October, and the minimum occurred in June.

The mean d-excess values in the Qilian Mountains, HIRS, USYR system, and QIRS were 13.63‰, 13.78‰, 13.56‰, and 20.20‰, respectively. According to the monthly mean values, the following was noted: USYR < HIRS < QIRS. In HIRS, the d-excess value ranged from 8.52‰ to 16.15‰, except in the Dang River, where the annual variations in the other rivers were low. In the USYR, the d-excess value in the Jinqiang River was more positive than those of the Datong and Huangshui Rivers (Fig. 6).

The differences in the stable isotopes in river water reflect the differences in the

recharge sources and recharge processes in different river systems. If river water is only supplied by precipitation, its stable isotope composition should be closer to the stable isotope variation in precipitation. However, in this study, there was a significant difference between $\delta^{18}O$ in the river water and precipitation, which indicates that the potential recharge source of these three river systems is not only precipitation.

3.1.5 Groundwater

Variations in the stable isotopes in groundwater were highly consistent with those of river water, the annual fluctuation was low, and there were no evident time variations. The $\delta^{18}O$ in the groundwater in the Qilian Mountains ranged from -8.99‰ to -8.50‰, with a mean value of -8.76‰. In the HIRS, USYR system, and QIRS, the $\delta^{18}O$ variation ranged from -9.01–8.43‰, -9.37–8.43‰, and -9.12–7.60‰, respectively, with corresponding mean values of -8.78‰, -8.83‰, and -8.55‰ (Fig. 7). The temporal variations in the $\delta^{18}O$ values in river water and groundwater in the study area were highly consistent, which confirms that precipitation and supra-permafrost water may be transformed into groundwater runoff and then recharge. The $\delta^{18}O$ concentration in the groundwater was more negative than that of the river water, which indicates that the evaporation effect of groundwater was relatively weak compared with that of river water. The mutual transformation relationship and frequent exchange processes between groundwater and river water were confirmed. In alpine desert zones, sparse vegetation, large soil particles, and high permeability are conducive to the infiltration of precipitation, snow, and ice meltwater, which can be converted into groundwater thus replenishing runoff.

3.2 Relationships of stable isotopes between river water and various water sources

The isotopic relationship between river water and various water sources is shown in

Fig. 8. The local meteoric water line (LMWL) of the Qilian Mountains was $\delta D = 7.99\delta^{18}O + 14.57$ ($R^2 = 0.96$), and the $\delta^{18}O$ concentrations was clearly higher in summer and autumn but lower in winter and spring (Gui et al., 2020). For glacier and snow meltwater, the stable isotope was distributed in the center of the LMWL, while that of the supra-permafrost water was mainly distributed below the LMWL. The stable isotopic compositions of the river water were close to the LMWL, and between those of precipitation, glacier and snow meltwater, and supra-permafrost water, indicating that the river water was fed by all these sources of water. At the same time, the distribution of river water and groundwater showed a hydraulic connection. The local evaporation lines (LEL) of the Qilian Mountains, HIRS, USYR system, and QIRS were $\delta D = 5.19\delta^{18}O - 8.35$ ($R^2 = 0.55$, $P < 0.01$), $\delta D = 5.96\delta^{18}O - 2.77$ ($R^2 = 0.72$, $P < 0.01$), $\delta D = 5.44\delta^{18}O - 6.37$ ($R^2 = 0.79$, $P < 0.01$), and $\delta D = 3.72\delta^{18}O - 15.67$ ($R^2 = 0.58$, $P < 0.01$), respectively. The corresponding $\delta^{18}O$ and δD values at the intersection of the LEL and LWML were (-8.19‰, -50.84‰), (-8.54‰, -53.68), (-8.21‰, -51.04‰), and (-7.08‰, -42.01‰) for the Qilian Mountains, HIRS, USYR, and QIRS, respectively, which were very similar to the corresponding mean values of $\delta^{18}O$ and δD in groundwater (-8.76‰,-52.10‰), (-8.78‰,-52.54‰), (-8.83‰,-53.02‰), and (-8.55‰,-48.07‰). This indicates that various types of water in the study area are converted into groundwater and then resupplied by river water.

3.3 Components of outlet runoff

EMMA was used to determine the contribution ratio of the different water sources to runoff. The results of the above analysis show that there were significant spatio-temporal variations in d-excess and $\delta^{18}O$ concentrations in the river and its supplementary components; therefore, $\delta^{18}O$ and d-excess were selected as tracers for analysis because this

415 combination provides a reasonable separation of sources. In the three river systems of the
416 Qilian Mountains and the Jinqiang and Huangshui Rivers in the QIRS, there were no
417 glaciers in the upper reaches; therefore, the binary segmentation model was adopted to
418 calculate the supply source (Fig. 9h and 9j). For the remaining rivers, precipitation, supra-
419 permafrost water, and glacier and snow meltwater were taken as three end-members, and
420 the recharge sources were then calculated (Fig. 9). The results show that the precipitation
421 was the main recharge source of seven rivers in the HIRS (Fig. 9a–g), and its contribution
422 ratios to the Danghe, Changma, Taolai, Heihe, Xiying, Nanying, and Zamu Rivers were
423 65%, 51%, 69%, 59%, 75%, 80%, and 79%, respectively. More than half of the runoff in
424 the HIRS was converted from precipitation; the contribution ratio increased from west to
425 east and was as high as 80% in the Nanying and Zamu Rivers in the eastern part of the
426 Qilian Mountains. This is consistent with the spatial distribution characteristics of higher
427 precipitation in the east and lower precipitation in the west. In addition to precipitation,
428 supra-permafrost water was an important recharge source for the HIRS. The contribution
429 of supra-permafrost water to the Danghe, Changma, Taolai, Heihe, Xiying, Nanying, and
430 Zamu Rivers was approximately 21%, 33%, 20%, 33%, 19%, 15%, and 16%, respectively,
431 as the third end-member; the corresponding glacier and snow meltwater contributed
432 approximately 14%, 16%, 11%, 8%, 6%, 5%, and 5%, respectively, to runoff. Contrary to
433 the spatial distribution trend of the precipitation contribution, the contribution ratio of
434 glacier and snow meltwater showed a decreasing trend from west to east. This was mainly
435 related to the storage and distribution of glaciers in the Qilian Mountains, in terms of glacier
436 area and ice storage: Shule River Basin > Heihe River Basin > Shiyang River Basin (Sun
437 et al., 2015).

Compared with the HIRS, the contribution of glacier and snow meltwater to the runoff in the USYR was significantly lower, and among the three rivers, only the Datong River was weakly replenished by glacier and snow meltwater. The contribution ratios of the precipitation, supra-permafrost, and glacier and snow meltwater to the Datong River were 63%, 35%, and 2%, respectively. The Jinqiang River was mainly replenished by precipitation and groundwater, which contributed 30% and 70%, respectively, while the Huangshui River was mainly replenished by precipitation and supra-permafrost water, which contributed 83% and 17%, respectively. Located in the QIRS, the Buha River was mainly replenished by precipitation, supra-permafrost, and glacier and snow meltwater, with the contributions of the three end-members to the runoff being 58%, 40%, and 2%, respectively.

Studies have shown that runoff in inland river basins in China is mainly derived from precipitation in mountainous areas, supra-permafrost, and glacier and snow meltwater Zhou et al., (2000) and Li, et al., (2016b) determined that the elevation of mountain runoff is approximately 3,500 m, according to the elevation effect of $\delta^{18}O$ in the precipitation of the Qilian Mountains, which is 0.18‰/100 m. These facts show that the water resources of the Qilian Mountains mainly originate from the upper reaches of the mountain area. As the permafrost boundary is 3600 m in the Qilian Mountains (Zhou et al., 2000), and the altitude at which the river flows from the mountains in the USYR and QIRS is higher, with 3700 m as the boundary, the mountainous area was divided into the cryosphere and vegetation belt. EMMA was then used to calculate the contributions of the cryosphere and vegetation belts to mountain runoff from the three major water systems in the Qilian Mountains.

For HIRS, the $\delta^{18}O$ values in the precipitation of the cryosphere and vegetation belts

were -9.02‰ and -7.32‰, respectively. Calculation using a binary mixed segmentation model showed that the contribution ratio of the cryosphere belt to HIRS reached 82%, consisting of precipitation, glacier and snow meltwater, and supra-permafrost water at 50%, 9%, and 23%, respectively; the remaining 18% was contributed by precipitation in the vegetation belt below 3600 m.

Similarly, the $\delta^{18}O$ values in the precipitation in the cryosphere and vegetation belts in the USYR were -10.74‰ and -8.42‰, respectively. The contribution ratio of the cryosphere belt was 71% of runoff (comprising precipitation, glacier and snow meltwater, and supra-permafrost water at 43.5%, 24%, and 3.5%, respectively), and that of the vegetation belt was 29%. In the QIRS, the $\delta^{18}O$ values in the precipitation of the cryosphere and vegetation belts were -12.29‰ and -8.95‰, respectively. Calculations using the binary mixed segmentation model showed that the contribution of the cryosphere belt to the runoff was 80% (comprising precipitation, glacier and snow meltwater, and supra-permafrost water at 38%, 2%, and 40%, respectively), and that of the vegetation belt was 20% (Fig. 10, 11).

Using the approach shown in Eq. (5), the uncertainty originating from the variation in the tracers of components and measurement methods can be calculated separately (Pu et al., 2013). The uncertainty was estimated to be 0.04 for the three-component mixing model in the study region. The uncertainty terms for the supra-permafrost water accounted for more than 47% of the total uncertainty, indicating that the $\delta^{18}O$ variations in the supra-permafrost water accounted for majority of the uncertainty. Although there is some uncertainty in hydrograph separation, isotope-based hydrograph separations are still valuable tools for evaluating the contribution of meltwater to water resources and

particularly helpful for improving the understanding of hydrological processes in cold regions, where there is a lack of observational data.

3.4 Hydrological processes

The runoff process in the Qilian Mountains occurs mainly from May to October and is influenced by glacier and snow meltwater, precipitation, and freezing and thawing processes in different ways and months; therefore, the monthly runoff components were segmented according to the daily isotope values from May to October (Fig. 12). In May, the temperature and precipitation gradually increased, causing the glacier, snow, and permafrost active layer to melt gradually. Snowmelt runoff increased with increasing temperature and, together with precipitation, replenished the river. Snow meltwater and supra-permafrost water in the surface layer are generally blocked by the frozen layer, which directly merges into river recharge runoff and becomes an important part of the spring runoff; this is also the main reason for spring floods in inland rivers. As shown in Fig. 12, the contribution of glacier and snow meltwater to river runoff in all three river systems was the largest in May. For this period, the contributions of glacier and snow meltwater to runoff in the HIRS, USYR system, and QIRS were approximately 16%, 3%, and 7%, respectively, which were significantly higher than those of the average level of the entire growing season. As the temperatures continue to rise and precipitation continues to increase in June, the snow melts rapidly, but reserves continue to decline. With the thawing of the soil at the top of the permafrost active layer, the contribution of supra-permafrost water to runoff is further enhanced, but precipitation still plays a leading role. Hence, in June, the contribution of precipitation to the runoff from mountains in the HIRS, USYR system, and QIRS was 73%, 64%, and 72%, respectively.

The temperature and precipitation in July and August reached the highest value in study period with the melting depth of the active layer for permafrost further increases and the water absorption is enhanced. Thus, the precipitation and surface water can be quickly transformed into groundwater in the active layer of permafrost. In steep terrains, it rapidly replenishes runoff in the form of spring water, whereas in relatively flat terrains, it slowly replenishes runoff in the form of groundwater, which becomes an important part of runoff in cold regions; hence, the contribution of supra-permafrost water to runoff further increased. However, in the warmer months, the snow completely melted, but the massive increase in rainfall, instead, had a leading role in runoff, particularly in August when the contribution ratio of precipitation to the Huangshui and Zamu Rivers was as high as 90%. Although the glacier and snow meltwater flows remained unchanged, the total flow at this time was larger than that in May and June, resulting in a lower contribution ratio of glacier meltwater.

From September to October, the temperature and precipitation began to decline, and the saturated active layer of the permafrost began to release water; the contribution of supra-permafrost water to runoff was at its highest level in a year, particularly in October, with more than half of the runoff in many rivers, including the Changma (57%), Heihe (59%), Datong (57%), and Buha Rivers (65%) (Fig. 12), originating from the supra-permafrost water. However, the contribution of glacier and snow meltwater remained high owing to the significant decrease in the overall runoff and relative decrease in the precipitation. In conclusion, from May to October, the runoff from the Qilian Mountain is dominated by precipitation, and it plays an imperative role from July to August. However, with the thickening of the permafrost active layer, the contribution ratio of the permafrost

active layer to runoff from May to October increased. The contribution ratio of glacier and snow meltwater to runoff in the Qilian Mountains was relatively low overall, and the highest value occurred in May, whereas the contribution ratio was relatively low in July–August when the temperature was relatively high.

It can be seen from the above analysis that precipitation has an absolute replenishment effect on rivers in the Qilian Mountains. In some rivers, precipitation contributed more than 80% to the runoff, except for the Jinqiang River, which mainly relies on groundwater; more than half of the recharge of the other rivers are from precipitation. In the HIRS, the contribution of precipitation to runoff increased from west to east. In the western section of the Qilian Mountains, the contribution of precipitation to the Shule River was between 50% and 65%. For the Heihe River, located in the middle of the Qilian Mountains, the contribution of precipitation was approximately 60–70%, and in the Shiyang River, in the eastern part of the Qilian Mountains, the contribution of precipitation was approximately 75–80%. This regional difference was mainly caused by the spatial distribution differences in precipitation in the Qilian Mountains, which gradually increases from west to east. The contribution of precipitation to runoff was slightly lower in the USYR system and QIRS than in the HIRS.

4. Discussion

4.1 Contribution of the cryosphere to runoff under changing environments

4.1.1 Contributions of glacier and snow meltwater to runoff

Glacier and snow meltwater are an important part of runoff in alpine mountains in western China. Different types and sizes of glaciers have different sensitivities to climate change that lead to significant differences in their processes of runoff yield and

concentration. The contribution of glacier meltwater to mountain runoff showed significant spatial differences. Glacial meltwater and its contribution to runoff in cold basins is controlled by the number, size, area ratio, and storage capacity of glaciers in the basin. In general, the larger the scale of the glacier, the higher the ice reserves, more glacier and snow melting water can be provided. Secondly, climate conditions, especially temperature changes, have a great influence on the amount of glacier and snow melting water. As a result, the contribution of meltwater to runoff varies from river to river because of differences in the glaciers themselves and differences in climate conditions and sampling times. For Bai Shui River in the Yulong Mountains, the two-component mixing model showed that an average of 53.4% of runoff came from glacier and snow meltwater during the wet season (Pu et al., 2013). Maurya (2011) et al., found that the average contribution of meltwater to runoff is 32% in typical glacial basins on the southern slope of the Himalayas, while in the western Himalayan region, the glacier and snow meltwater contribution has been estimated to be 35% to 50% of the river discharge(Laskar et al., 2018). The results suggested that the fraction of glacier and snow meltwater input over the total stream flow ranged from 84.50 to 86.52% in Hailuogou watershed on the eastern slope of Mount Gongga, China (Meng and Liu, 2013). In the Hengduan Mountains, hydrograph separation analysis showed that the contribution to runoff from glacier and snow meltwater varied from 63.8% to 92.6% (Liu et al., 2008). The contribution of glacial melt to the annual runoff varied greatly among the basins, with approximately 22.3%, 25.2%, 34.6%, and 79% for the source regions of the Mekong, Salween, Brahmaputra, and Indus, respectively(Zhang et al., 2013). In the source regions of the Yangtze River, meltwater accounted for 23% of river water at the TTH station from June 2016 to May 2018, while

the corresponding value at the ZMD station was 17% (Li et al., 2020). In the USYR, the contribution of snow and ice meltwater to runoff was approximately 23%. In the Kunlun Mountains, the average contribution of meltwater to runoff in the Tizinafu River was 43% (Fan et al., 2015), and in Urumqi River, it was 14.7%, while in Kumalak River, the contribution was more than 57% (Sun et al., 2015; Kong & Pang, 2012) (Fig. 13).

As solid reservoirs, glacier retreat will inevitably lead to a reduction in total water resources. Glacier meltwater runoff increases when the glacier is degraded and then tends to decrease as the glacier area decreases (Weiming et al., 2012). During 1960-2019, the Qilian Mountains showed an overall warming trend with an average annual temperature rise of 0.319°C/a (Ye et al., 2022). Under the influence of global warming, glaciers have become shorter, narrower, and thinner, and statistical results showed that the glacier area of the Qilian Mountains in 1987, 1991, 1997, 2001, 2007, 2013, and 2018 was 2080.39 km^2, 1939.12 km^2, 1805.65 km^2, 1691.13 km^2, 1619.26 km^2, 1531.21 km^2, and 1,442.09 km^2, respectively. During 1987-2018, the glacier area was in continuous retreat, at an average annual rate of 1.34% during 1987 - 2001 and 0.87% during 2001 - 2018 (Yingzheng et al., 2020). The massive retreat of glaciers will inevitably have a large impact on glacial meltwater runoff.

4.1.2 Contribution of permafrost to runoff

In cold watersheds, winter precipitation is solid and cannot directly recharge rivers, so winter runoff mainly comes from groundwater. In the long term, the active layer of permafrost is also a solid water source, and permafrost degradation will lead to thickening of its active layer, which in turn will cause an increase in the permafrost water storage capacity and groundwater volume, ultimately causing changes in the hydrological

processes and seasonal structure of runoff in cold regions. Permafrost is widely distributed in the Qilian Mountains, which are mainly characterized by steep topography, sparse vegetation, and cold climate, and it plays an important role in the exchange of surface and groundwater within the basin as well as in the intra-annual distribution (Cheng and Wu, 2007). On the one hand, the increased depth of the active layer of permafrost reduces the depth of the water barrier, thereby reducing direct runoff, and on the other hand, as the active layer deepens, the frozen water stored in the active layer will be released, thereby recharging runoff. The effect of temperature on runoff is a complex interaction of various factors.

The spatial and temporal distributions and hydrothermal characteristics of different types of permafrost differ to some extent, which leads to significant differences in hydrological processes. The above analysis confirms that supra-permafrost water is also an important part of the runoff from the Qilian Mountains. In some months, its contribution to runoff from the mountains can reach more than 60%. The contribution of supra-permafrost water to runoff from mountains shows obvious spatiotemporal characteristics. In terms of time, the contribution ratio gradually increases from May to October, and the largest contribution ratio appears in October. In terms of spatial distribution characteristics, the contribution is significantly higher in the Datong River of the USYR system and the Buha River of the QIRS than in the HIRS. This is mainly related to the spatial distribution and thickness of the permafrost. In the Qilian Mountains, the average contribution of supra-permafrost water to runoff from May to October is 10%-40%, while in the source region of the Yangtze River in China, due to the widespread distribution of permafrost, the contribution is generally over 40% (Li et al., 2020).

4.2 Hydrological effects of cryospheric change

4.2.1 Runoff changes

Numerous studies have shown that under global warming, glacier degradation and precipitation have continued to increase in the study area, resulting in a significant increase in runoff in the Qilian Mountains since 1990 (Li et al., 2019a, b; Cao et al., 2010). Therefore, the runoff variation after 1990 in the study area was analyzed. As shown in Fig. 14, from 1990 to 2020, the seven rivers in the HIRS all showed an increasing trend. The increasing rates of runoff in the Danghe, Changma, Taolai, Heihe, Xiying, Nanying, and Zamu Rivers were $0.16 \times 10^8\ m^3/10a$, $2.7 \times 10^8\ m^3/10a$, $0.48 \times 10^8\ m^3/10a$, $2.6 \times 10^8\ m^3/10a$, $0.36 \times 10^8\ m^3/10a$, $0.04 \times 10^8\ m^3/10a$, and $0.01 \times 10^8 m^3/10a$, respectively. Accordingly, the peak runoff of the seven rivers appeared in 2019 ($5.057 \times 10^8\ m^3$), 2017 ($17.43 \times 10^8\ m^3$), 2018 ($7.82 \times 10^8\ m^3$), 2017 ($23.31 \times 10^8\ m^3$), 2019 ($4.416 \times 10^8\ m^3$), 1993 ($1.738 \times 10^8\ m^3$), and 2003 ($3.542 \times 10^8\ m^3$), respectively (Fig. 14).

In the USYR system, the runoff of the Datong and Huangshui Rivers still showed an increasing trend, with rates of $1.3 \times 10^8\ m^3/10a$ and $1.55 \times 10^8\ m^3/10a$, respectively. The runoff of the Jinqiang River showed a decreasing trend, but the rate was very slow ($-0.08 \times 10^8\ m^3/10a$). In summary, after 1990, runoff from the Qilian Mountains generally showed an increasing trend, and similar studies have shown that global warming has increased runoff from rivers that are heavily affected by glacier recharge. Based on the annual runoff data from 1951 to 2000, the results show that the runoff of most rivers in western China has been increasing (Baisheng et al., 2006), particulary since 1980. The runoff from mountains in Xinjiang has increased significantly, with a maximum increase of 40% (Ding et al., 2020).

4.2.2 Seasonal structure of runoff

The seasonal patterns of runoff have also changed over the past 30 years, for example, the time of maximum monthly runoff in some rivers is changing. As shown in Fig. 15, except for the Danghe and Changma Rivers, the runoff peaks of the other five rivers have all changed in the HIRS. From 1990 to 2020, the time of peak runoff in the Heihe River was delayed from July to August, but in the other six rivers, it advanced from August to July. This was closely related to the mass melting of the glaciers. Glacier meltwater runoff is mainly influenced by two factors: temperature and the size of the glacial reserves and area. In the context of global warming, the climate of northwest China was warm and dry from the end of the Little Ice Age to the 1980s and began to change to warm and humid around 1990 (Yafeng et al., 2003). The results show that the warming of the annual mean temperature in the Qilian Mountains from 1991 to 2016 was twice as much as that in the period of 1961–1990, making the ablation period in the Qilian Mountains longer (Li et al., 2019). This caused the glacier snowpack to start melting earlier and combine with heavy precipitation to form a flood season. Regardless of whether the peak runoff was advanced or retarded, it was still concentrated in July and August in the HIRS. The peak runoff of the Jinqiang River in the USYR moved from August to September (Fig. 15).

4.2.3 Winter runoff increases

Global warming has caused a temperature rise and active layer thickening of the permafrost in the Qinghai-Tibet Plateau of China (Cheng and Wu, 2007). Monitoring results show that the permafrost has been continuously degraded in the past few decades, and statistical model estimations show that the thickness of the active layer along the Qinghai-Tibet Highway has increased significantly from 1981 to 2018, with an average change rate of 19.5 cm/10 a; much of the underground ice that had been trapped near the

upper boundary of the permafrost has melted (Lin et al., 2010; Ren et al., 2012). The distribution of permafrost in Qilian Mountains in the 1960s, 1970s, 1980s, 1990s, the first decade of the 21st century, and 2010–2015 was 0.61×10^4 km^2, 0.58×104 km^2, 0.57×10^4 km^2, 0.50×10^4 km^2, 0.42×10^4 km^2, and 0.43×10^4 km^2, respectively (Chen et al., 2019). Permafrost degradation increases the infiltration rate of the soil, resulting in the weakening or even loss of the water barrier effect of the permafrost layer. In summer, an increase in the depth of permafrost thaw increases the recharge of groundwater from precipitation, while some underground ice melts and the area of the thaw zone expands, thus increasing the recharge of winter runoff (Clark et al., 2001).

The winter runoff (total runoff in January, February, and December) of some rivers in the Qilian Mountains increased after 1990. For example, the winter runoff of the Changma River in the HIRS was 0.759×10^8 m^3 in 1990–2000, 1.175×10^8 m^3 in 2001–2010, and 1.250×10^8 m^3 in 2011–2020, representing an increase of approximately 25%. The winter runoff of the Heihe River was 1.114×10^8m^3 in 1990–2000, 1.225×10^8 m^3 in 2001–2010 and 1.320×10^8 m^3 in 2011–2020, representing an increase of approximately 18.5%. Similarly, winter runoff increased by approximately 6% and 57% in the Nanying and Zamu Rivers, respectively. A similar situation occurred in the USYR (Fig. 15); the winter runoff of the Datong River was 1.432×10^8 m^3 in 1990–2000, 1.629×10^8 m^3 in 2001–2010, and 2.280×10^8 m^3 in 2011–2020, representing an increase of approximately 59%. This is mainly because permafrost degradation has led to further weakening of the permafrost water barrier, thereby changing the groundwater reservoir storage capacity and drainage paths. At the same time, glacier melt has accelerated, resulting in mixing of glacier meltwater and rainfall that recharges groundwater and subsequent monthly runoff in the

691 form of baseflow (Chen et al., 2019). Similar results have been reported for other cold
692 regions of the world. For example, analysis and simulation of runoff changes showed that,
693 owing to the decline of permafrost and change in the thawing process, the winter runoff in
694 the European parts of Russia has increased significantly by as much as 50–120%
695 (Kalyuzhnyi and Lavrov, 2012). In a typical basin of the Northern Slope of the Himalayas
696 during the cold period (from November to March of the following year), the runoff
697 increased by different amounts, particularly in January, and increased by 67% in the last 10
698 years compared with that of the previous 10 years (Fei et al., 2006). Studies on rivers in
699 the Qinghai-Tibet Plateau, Tianshan Mountains, Northeast China, and other regions of
700 China have found that permafrost degradation leads to an increase in winter runoff, which
701 is consistent with the trend of frozen soil degradation over time (Liu et al., 2006; Gong et
702 al., 2006; Liu et al., 2003)

703 **4.3 Implications for water resources management**

704 Water resources are a key factor limiting the development of the arid zone in
705 Northwest China. In these regions, water resources mainly come from the surrounding
706 mountains; therefore, changes in mountain runoff will directly affect socioeconomic
707 activities (Ye et al., 2008). According to the results in this study, 82%, 71%, and 80% of
708 the runoff in the HIRS, USYR system, and QIRS, respectively, come from the cryosphere
709 belt above 3600/3700 m altitude in the Qilian Mountains. Kang et al. (2008) also confirmed
710 that the contribution ratio of the cryosphere belt to mountain runoff in the HIRS was 83%,
711 based on the water balance.

712 These facts indicate that the water resource security of inland river basins in China is
713 highly dependent on the stability of the cryosphere belt. The severe retreat of the cryosphere

after 90 years has changed the characteristics of runoff, such as a rapid increase in volume, change in the peak time, and increase during winter. More seriously, as the glacier shrinks and its area decreases, the amount of glacial meltwater is bound to decrease at some point in the future, which is called the inflection point of the glacial meltwater increase to decrease, also known as the peak of glacial meltwater (Ding et al., 2020). These changes will pose challenges to water security in the future. Model predictions also indicate future glacial meltwater reductions of approximately 34%, 62%, and 74% by the end of the 21st century under the RCP2.6 (low), RCP4.5 (medium), and RCP8.5 (high) discharge scenarios, respectively (Yingsong et al., 2012; Zhao et al., 2019). Against the background of changing water resources, the Gansu and Qinghai provinces, which depend on the Qilian Mountains for water supply, will also actively adjust their water-use strategies. From 2010 to 2019, the total utilization of water resources in these provinces showed a downward trend. Among them, the agricultural irrigation and industrial water consumption showed a downward trend, while the water consumption for forestry, husbandry, fishery, livestock, comprehensive living, and ecological environment showed an upward trend. The structure of water use in Gansu and Qinghai is still dominated by agricultural water, which accounts for more than 70% of the total water consumption, whereas ecological water use remains less than 10% (Hou et al., 2021).

In conclusion, runoff from the Qilian Mountains has shown a pronounced increasing trend in recent decades, and the runoff composition has changed to some extent. However, it is predicted that the contribution of meltwater to runoff will decrease significantly in the future as glaciers melt substantially. This will lead to reduced runoff in basins where glacial meltwater contributes more to runoff.

Changes in runoff and its composition have resulted in many opportunities. For example, increased runoff means that there are more water resources to use, and in winter the uneven distribution of runoff during the year can be effectively alleviated. More importantly, it provides good conditions for ecological restoration and protection in arid areas. However, sudden increases in runoff and changes in runoff peaks indicate an increased risk of flooding. At the same time, the increase in runoff in winter does not match the main period of use for local production and living, and the utilization efficiency of this part of the water resource is not high; an increase in runoff may not be sustainable. In the future, with the continuous decrease in glacier area and numbers, the amount of runoff from glacier and snow meltwater will decrease, which is likely to lead to a decrease in runoff from mountains, bringing great loss to local economic and social development. Based on the current situation of water resources, the following suggestions are proposed (Fig. 16):

(1) Strengthen the capacity for water storage and release by designing water supply and flood control projects. Owing to the restrictions on economic and social conditions, development ideas, and water resource conditions at that time, water conservancy projects, such as reservoirs built in the early years have been unable, to some extent, to meet and cope with the current situation of increasing mountain runoff. Therefore, it is necessary to continually improve regional water supply and flood control capacity. It is necessary to consider storing more water for use in the dry season, and it is also necessary to strengthen the awareness and ability to cope with flood disasters and flexibly respond to changes in water resources.

(2) Redistribute water resources over time according to actual water use. In the oasis areas that depend on the water resources of the Qilian Mountains, the water use structure

is mainly agricultural; however, agricultural activities have relatively fixed amounts and times, and their requirements often cannot completely adapt to the situation of natural water supply, especially in the current context of increased winter runoff. On the one hand, the increased runoff in winter can be stored for use in the dry season through water conservancy projects and other measures. On the other hand, agricultural irrigation methods can be adjusted appropriately, such as replacing winter irrigation with spring irrigation.

(3) Strengthen regional communication and cooperation to reallocate water resources. Although the runoff from mountains has generally shown an increasing trend in recent decades in the study area, this is not uniform in space, and some regions even show negative growth, which leads to a more uneven distribution of water resources in space. To better adapt to this situation, each region should build and improve the inter-basin water transfer project according to the actual situation, adjusting the remaining water resources efficiently to realize the optimal allocation of water resources.

(4) Accelerate the reform of water-saving agriculture and actively address the impact of glacier and snow meltwater on agriculture. It is predicted that with the continuous decrease in glacier reserves, the contribution from glacier and snow meltwater to runoff from mountains will decrease in the future, which will significantly reduce runoff where glacier and snow meltwater is the main recharge source. To deal with the impact of water resource reduction on agriculture, on the one hand, we should vigorously promote water-saving irrigation; on the other hand, we should promote drought-resistant tillage and cultivation methods. By applying various measures to seek advantages and avoid disadvantages, the harmony of life, agriculture, industry, and ecological water use can be realized.

5. Conclusions

Based on the isotopic data of 1,310 precipitation, 338 river water, 96 glacier and snow meltwater, 108 supra-permafrost water, and 312 groundwater samples, this study quantified the runoff components of 11 major rivers in the Qilian Mountains and investigated the influence of cryosphere changes on runoff from mountains. It was found that the stable isotopes of river water and groundwater in the study area were relatively invariable, unlike that of precipitation, which showed significant seasonal variations. The annual mean values of $\delta^{18}O$ in the river and groundwater in the Qilian Mountains were -8.49‰ and -8.76‰, respectively.

The stable isotope relationships of various waters showed that the river water was fed by precipitation, glacier and snow meltwater, and supra-permafrost water. EMMA was used to determine the contribution ratios of different water bodies to runoff. The calculations showed that precipitation was the main recharge source of seven rivers in the HIRS; the contribution ratios to the Danghe, Changma, Qiaolai, Heihe, Xiying, Nanying, and Zamu Rivers being 65%, 51%, 69%, 59%, 75%, 80%, and 79%, respectively. Supra-permafrost water was also an important recharge source for the HIRS. The contribution of supra-permafrost water to the Dang, Changma, Taolai, Heihe, Xiying, Nanying, and Zamu Rivers was approximately 21%, 33%, 20%, 33%, 19%, 15%, and 16%, respectively. As the third end-member, the corresponding glacier and snow meltwater contributed approximately 14%, 16%, 11%, 8%, 6%, 5%, and 5% to runoff, respectively. In the USYR system, the contribution of glacier and snow meltwater to the runoff was notably low. The contribution ratios of precipitation, supra-permafrost, and glacier and snow meltwater to the Datong River was 63%, 35%, and 2%, respectively. The Jinqiang River was mainly replenished by

precipitation and groundwater, which contributed 30% and 70%, respectively, while the Huangshui River was mainly replenished by precipitation and supra-permafrost water, which contributed 83% and 17%, respectively. Located in the QIRS, the Buha River was mainly replenished by precipitation, supra-permafrost, and glacier and snow meltwater, with the contributions of these three end-members to the runoff being 58%, 40%, and 2%, respectively.

Runoff in the inland rivers of the Qilian Mountains is mainly derived from the cryosphere belt. Calculations using a binary mixed segmentation model revealed that the contribution ratios of the cryosphere belt to mountain runoff in the HIRS, USYR system, and QIRS were 82%, 71%, and 80%, respectively. Cryospheric changes have impacted hydrological processes in the Qilian Mountains. After the 1990s, the runoff from the Qilian Mountains generally increased rapidly, the peak time of runoff changed, and runoff showed an increasing trend in winter. These changes in hydrological processes provide both opportunities and challenges, and require various measures to exploit advantages and avoid disadvantages to achieve harmony in ecological, living, and production water use.

Code/Data availability

The raw/processed data required to reproduce these findings cannot be shared at this time as the data also forms part of an ongoing study. We will not share our data until all relevant results are completed.

Author Contributions

Juan Gui led the write-up of the manuscript with significant contribution. Zongxing Li developed the research and designed the experiments. Qi Feng collected the water

samples and analysed the data, Qiao Cui and Jian Xue Modifyed the language. All authors discussed the results and contributed to the preparation of the manuscript.

Competing interests

This manuscript has not been published or presented elsewhere in part or in entirety and is not under consideration by another journal. We have read and understood your journal's policies, and we believe that neither the manuscript nor the study violates any of these. There are no conflicts of interest to declare.

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Tables:

Table 1 Hydrological data of various basins in Qilian Mountains

Table 1

Data type			glacier data		Hydrological data	
Drainage	River	River basin	Glacier area/km^2	Source	Period	Source
HIRS	Danghe River	Shule river	203.77	Sun et al.,2015	1990-2020	HWRBGS
	Changma River		509.87	Sun et al.,2015	1990-2020	HWRBGS
	Taolai River	Heihe river basin	78.33	Sun et al.,2015	1990-2020	HWRBGS
	Heihe River				1990-2020	HWRBGS
	Xiying River	Shiyang River Basin	39.94	Sun et al.,2015	1990-2020	HWRBGS
	Nanying River				1990-2020	HWRBGS
	Zamu River				1990-2020	HWRBGS
	Datong River	Datong River Basin	20.83	Sun et al.,2015	1990-2020	HWRBGS
USYR	Jinqiang River	Jinqiang River basin	---	---	1990-2020	HWRBGS
	Huangshui River	Huangshui Rive basin r	----	------	1990-2010	Zhang et al,.2014
QIRS	Buha River	Buha River basin	10.27	Sun et al.,2015	1990-2016	Liu et al,.2020

Figures:

Fig.1 Location of (a): study region (1. Danghe river; 2. Changma river; 3. Taolai river; 4. Heihe river; 5. Xiying river; 6. Nanying river; 7. Zamu river; 8. Datong river; 9. Huangshui river;10. Jinqiang river; 11.Buha river) and (b): sampling site

Fig.2 Temporal variation of $\delta^{18}O$ in precipitation in Qilian Mountains

Fig.3 Temporal variation of stable isotopes in glacier and snow meltwater in Qilian Mountains

Fig.4 Temporal variation of stable isotopes in supra-permafrost water in Qilian Mountains

Fig.5 Temporal variation of $\delta^{18}O$ in the outflow river in the Qilian Mountains

Fig.6 Temporal variation of d-excess in outlet river in Qilian Mountains

Fig.7 Stable isotope characteristics of groundwater in Qilian Mountains

Fig. 8 Relationships of $\delta^{18}O$ and δD between river water and various water in (a) Qilian mountains, (b) HIRS, (c) USYR, (d) QIRS

Fig. 9 Mixing diagram using the mean values of $\delta^{18}O$ and d-excess for river water Qilian mountains

Fig. 10 Mixing diagram using the mean $\delta^{18}O$ and d-excess values for the outlet runoff in (a): HIRS;(b): USYR and (c): QIRS

Fig. 11 Conceptual model of the contribution of cryosphere belt and vegetation belt to runoff in (a): Hexi inland river system;(b): Upper stream of yellow river system;(c) Qinghai inland river system

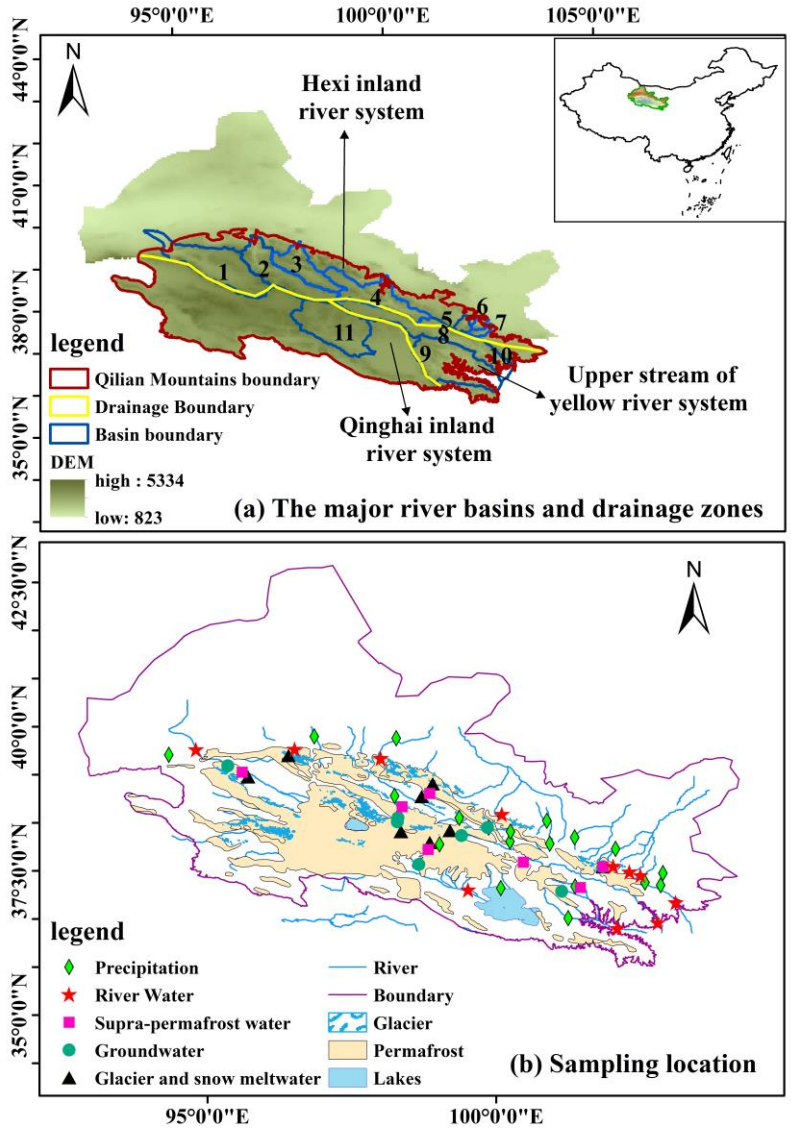
Fig. 12 Contribution rate from runoff components to monthly runoff

Fig. 13 Contribution of glacier and snow meltwater to runoff in alpine regions of China

Fig. 14 Annual variation of runoff after 1990 in Qilian mountains

Fig. 15 Seasonal variation of runoff after 1990 in hexi inland river system: (a) Danghe River; (b) Changmahe River; (c) Taolai River; (d) Heihe River; (e) Xiying River; (f) Nanying River and; (g)Zamu River

Fig. 16 Conceptual model of runoff change, water resource effect and countermeasures under global warming in the Qilian Mountains



1146

1147 Fig.1

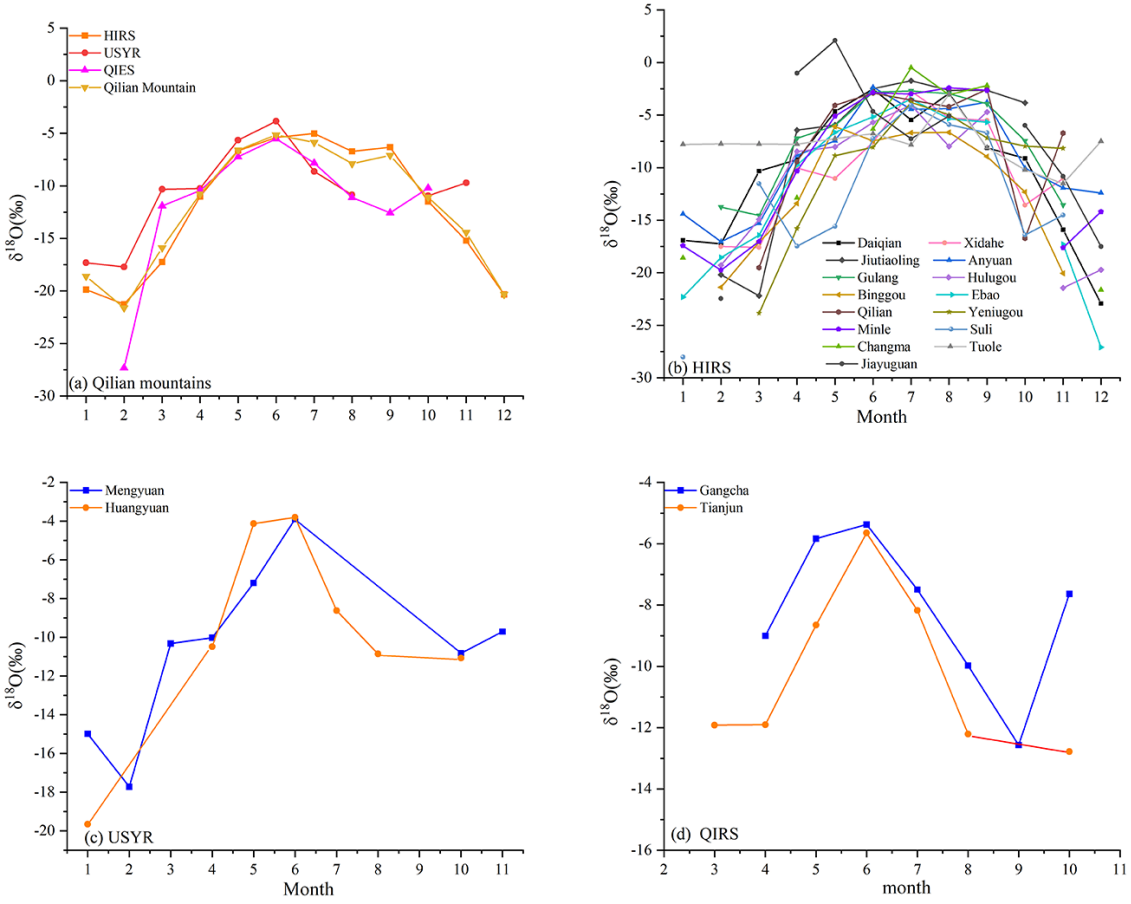

HIRS
USYR
QIES
Qilian Mountain
(a) Qilian mountains
δ18O(‰)
(b) HIRS
Daiqian
Xidahe
Jiutiaoling
Anyuan
Gulang
Hulugou
Binggou
Ebao
Qilian
Yeniugou
Minle
Suli
Changma
Tuole
Jiayuguan
Month
Mengyuan
Huangyuan
(c) USYR
Gangcha
Tianjun
(d) QIRS
month

Fig.2

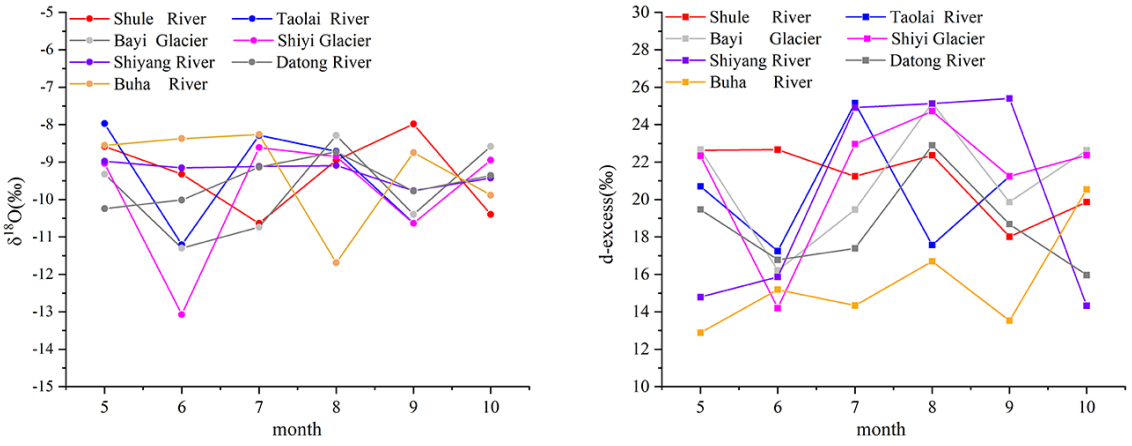

Shule River
Taolai River
Bayi Glacier
Shiyi Glacier
Shiyang River
Datong River
Buha River
δ18O(‰)
d-excess(‰)
month

Fig.3

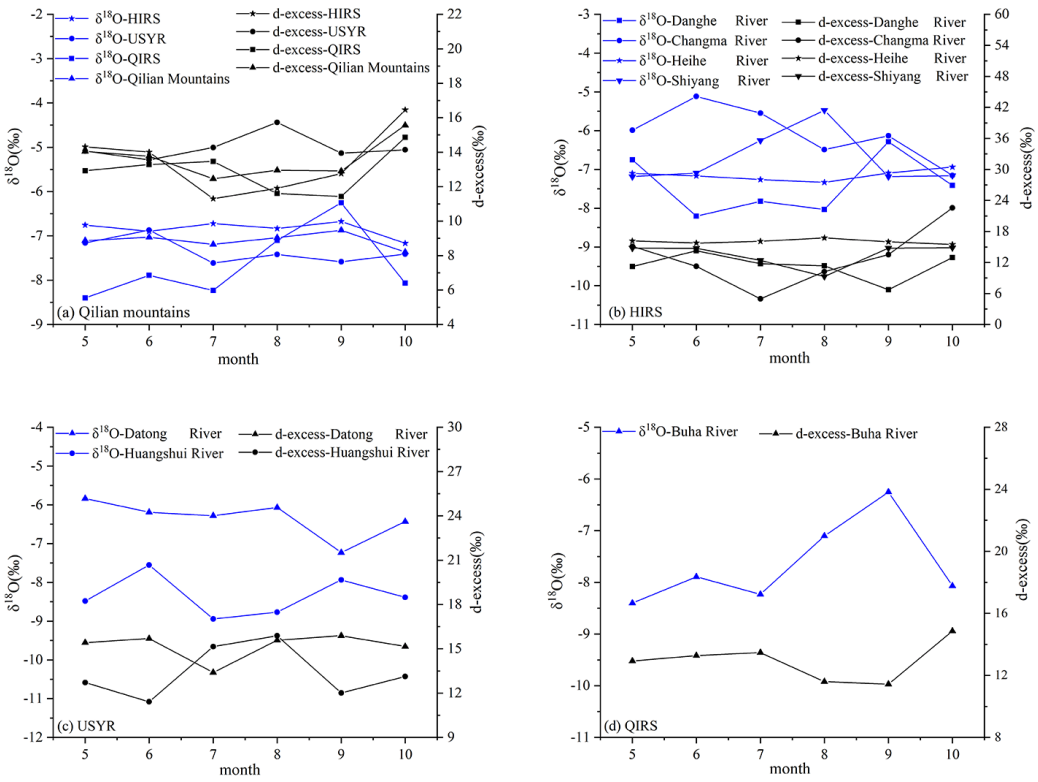

δ18O-HIRS
δ18O-USYR
δ18O-QIRS
δ18O-Qilian Mountains
d-excess-HIRS
d-excess-USYR
d-excess-QIRS
d-excess-Qilian Mountains
(a) Qilian mountains
(b) HIRS
δ18O-Danghe River
δ18O-Changma River
δ18O-Heihe River
δ18O-Shiyang River
d-excess-Danghe River
d-excess-Changma River
d-excess-Heihe River
d-excess-Shiyang River
(c) USYR
δ18O-Datong River
δ18O-Huangshui River
d-excess-Datong River
d-excess-Huangshui River
(d) QIRS
δ18O-Buha River
d-excess-Buha River
δ18O(‰)
d-excess(‰)
month

Fig.4

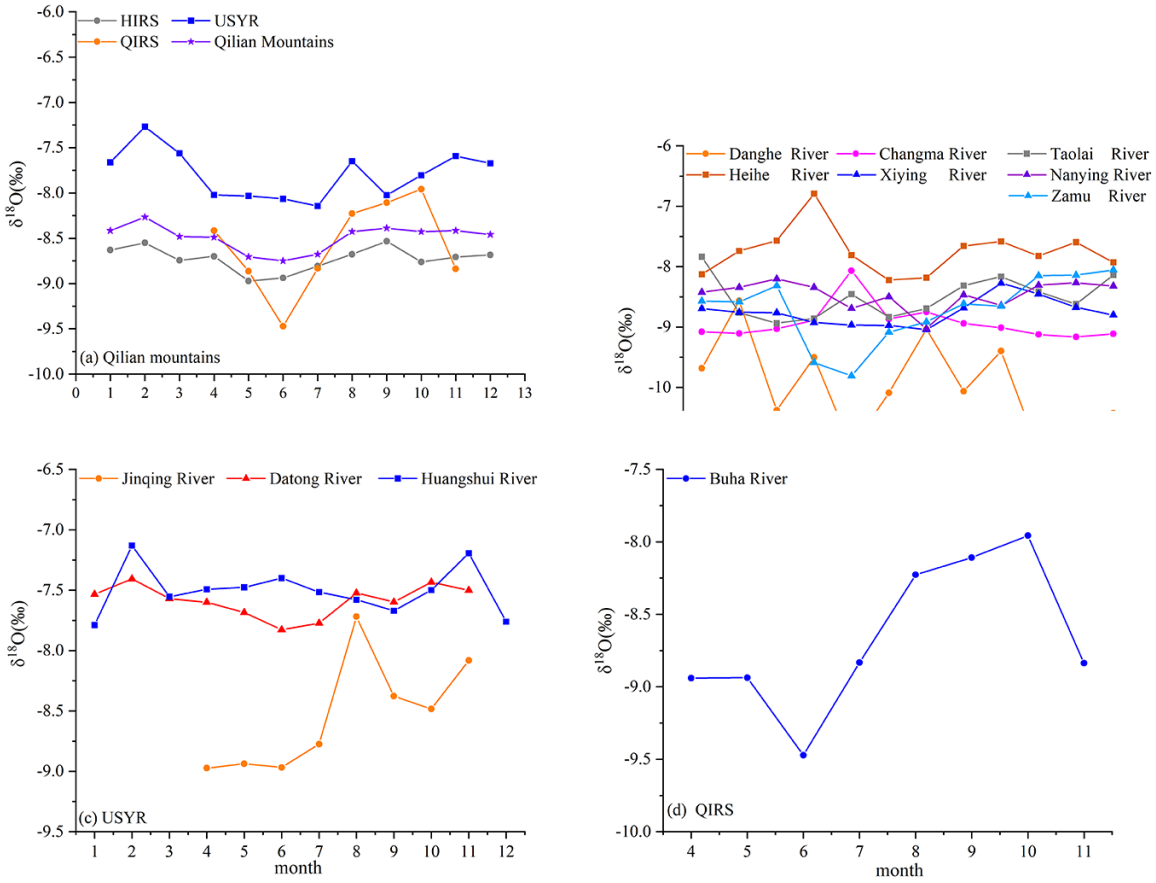

HIRS
USYR
QIRS
Qilian Mountains
(a) Qilian mountains
δ18O(‰)
Danghe River
Changma River
Taolai River
Heihe River
Xiying River
Nanying River
Zamu River
Jinqing River
Datong River
Huangshui River
(c) USYR
month
Buha River
(d) QIRS
month

Fig.5

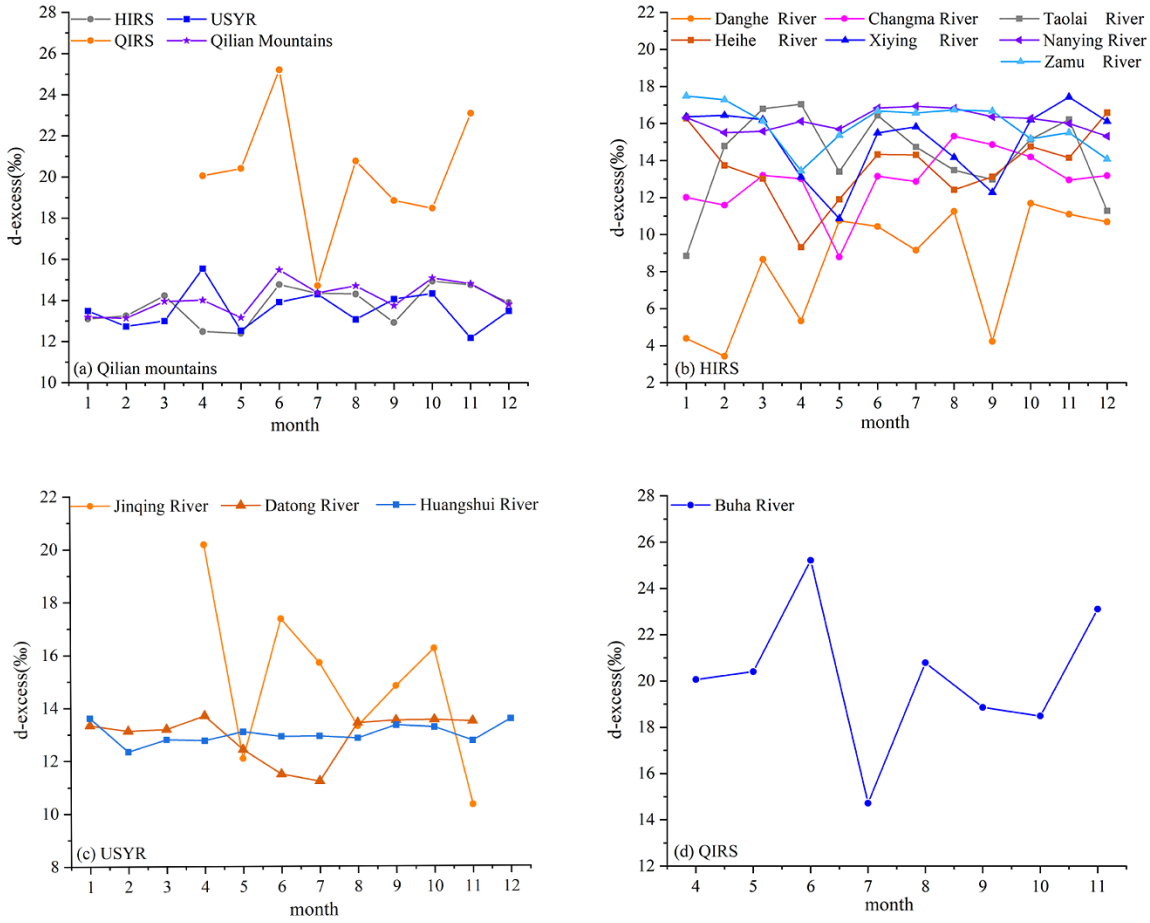

HIRS
USYR
QIRS
Qilian Mountains
d-excess(‰)
month
(a) Qilian mountains
Danghe River
Changma River
Taolai River
Heihe River
Xiying River
Nanying River
Zamu River
(b) HIRS
Jinqing River
Datong River
Huangshui River
(c) USYR
Buha River
(d) QIRS

Fig.6

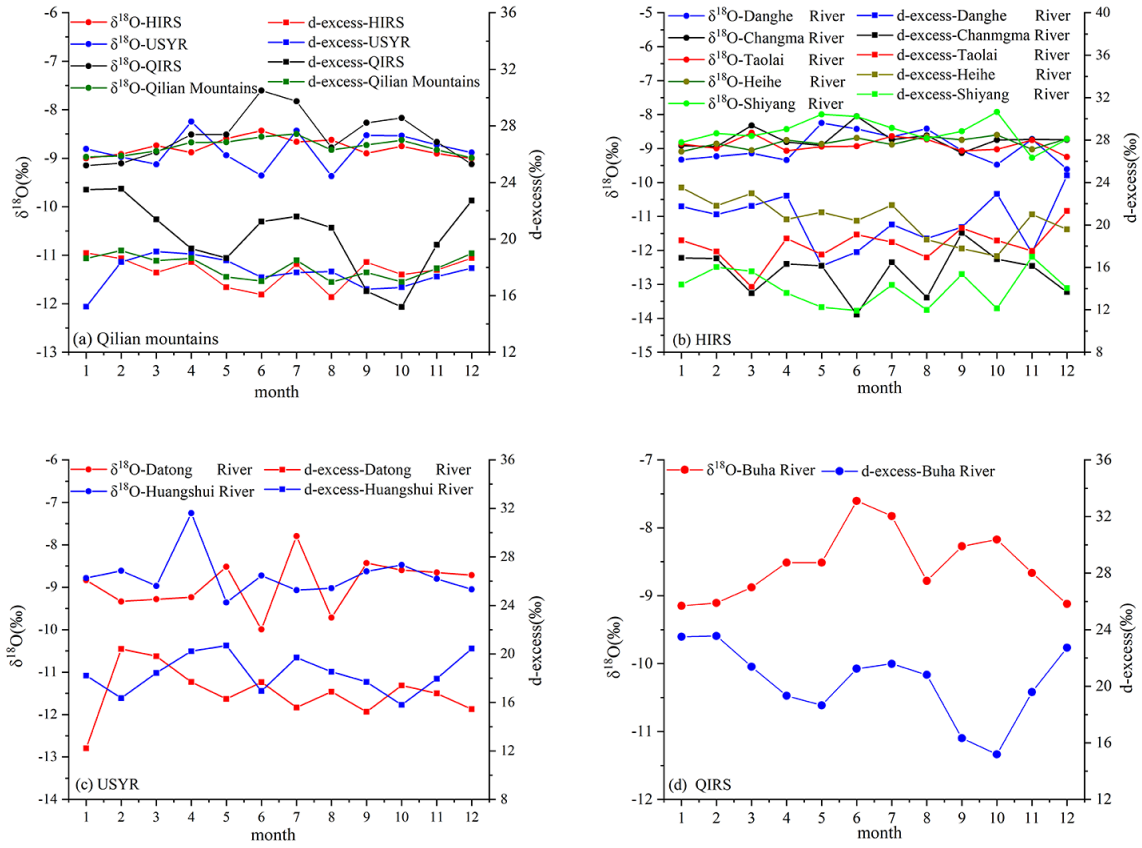

δ18O-HIRS
d-excess-HIRS
δ18O-USYR
d-excess-USYR
δ18O-QIRS
d-excess-QIRS
δ18O-Qilian Mountains
d-excess-Qilian Mountains
(a) Qilian mountains
δ18O(‰)
d-excess(‰)
month
δ18O-Danghe River
d-excess-Danghe River
δ18O-Changma River
d-excess-Chanmgma River
δ18O-Taolai River
d-excess-Taolai River
δ18O-Heihe River
d-excess-Heihe River
δ18O-Shiyang River
d-excess-Shiyang River
(b) HIRS
δ18O-Datong River
d-excess-Datong River
δ18O-Huangshui River
d-excess-Huangshui River
(c) USYR
δ18O-Buha River
d-excess-Buha River
(d) QIRS

Fig.7

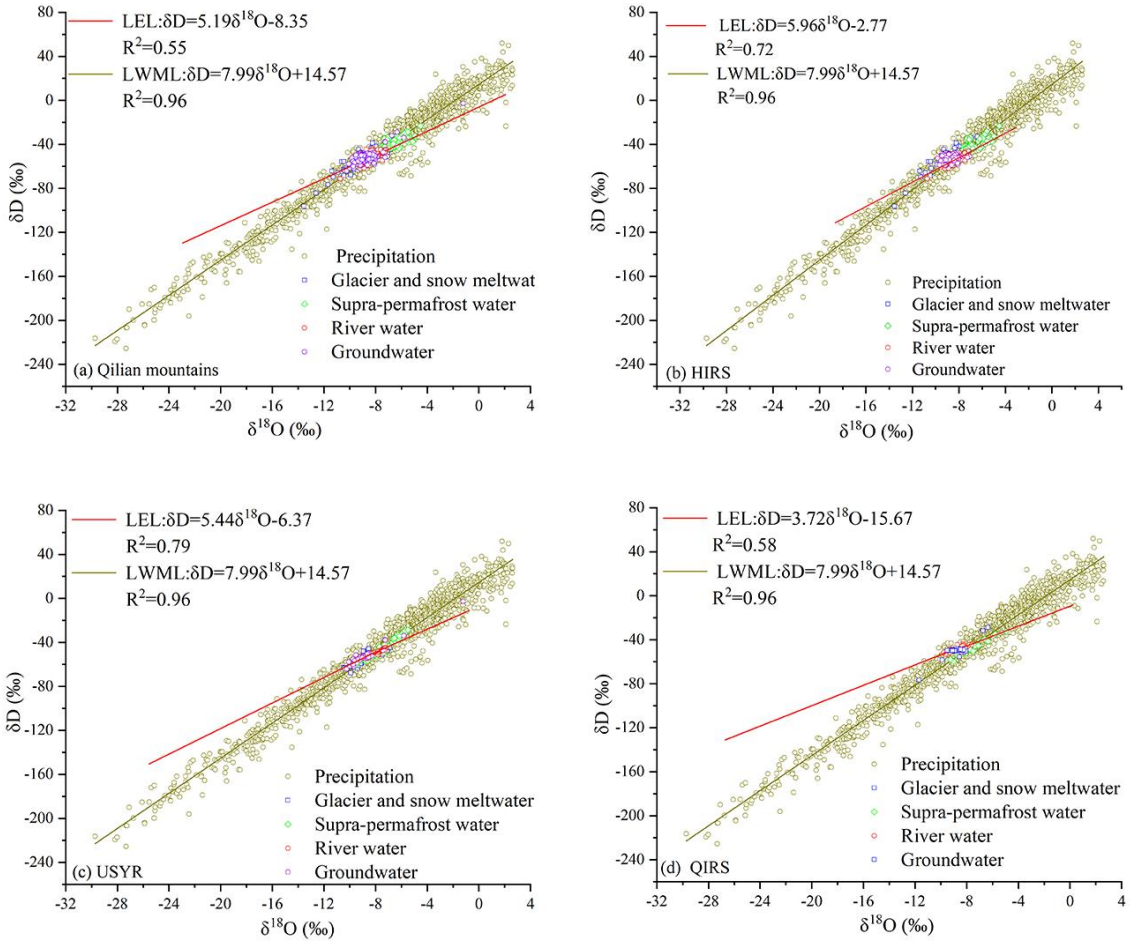

LEL:δD=5.19δ18O-8.35
R2=0.55
LWML:δD=7.99δ18O+14.57
R2=0.96
Precipitation
Glacier and snow meltwat
Supra-permafrost water
River water
Groundwater
(a) Qilian mountains
LEL:δD=5.96δ18O-2.77
R2=0.72
LWML:δD=7.99δ18O+14.57
R2=0.96
Precipitation
Glacier and snow meltwater
Supra-permafrost water
River water
Groundwater
(b) HIRS
LEL:δD=5.44δ18O-6.37
R2=0.79
LWML:δD=7.99δ18O+14.57
R2=0.96
Precipitation
Glacier and snow meltwater
Supra-permafrost water
River water
Groundwater
(c) USYR
LEL:δD=3.72δ18O-15.67
R2=0.58
LWML:δD=7.99δ18O+14.57
R2=0.96
Precipitation
Glacier and snow meltwater
Supra-permafrost water
River water
Groundwater
(d) QIRS
δD (‰)
δ18O (‰)

Fig. 8

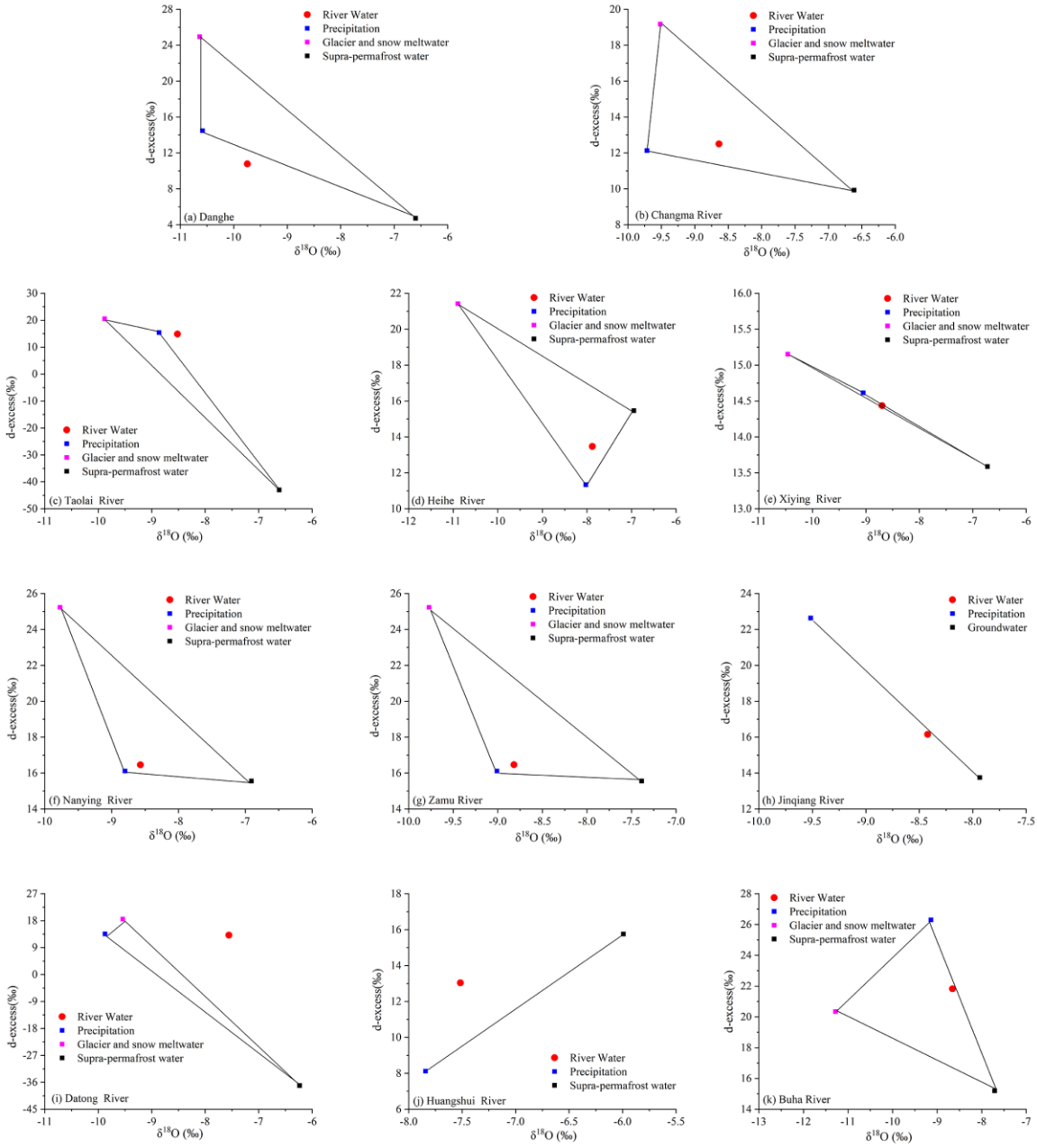
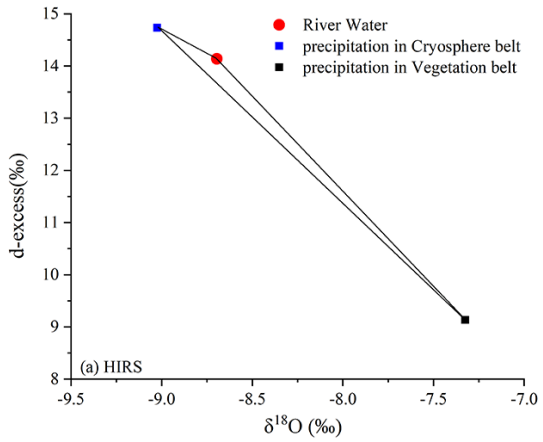
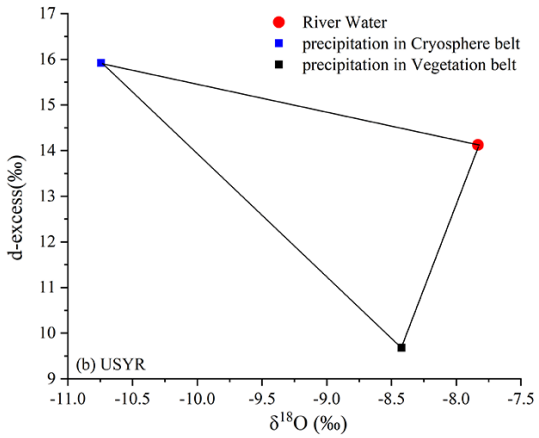

River Water
Precipitation
Glacier and snow meltwater
Supra-permafrost water
Groundwater
d-excess(‰)
δ18O (‰)
(a) Danghe
(b) Changma River
(c) Taolai River
(d) Heihe River
(e) Xiying River
(f) Nanying River
(g) Zamu River
(h) Jinqiang River
(i) Datong River
(j) Huangshui River
(k) Buha River

Fig. 9


River Water
precipitation in Cryosphere belt
precipitation in Vegetation belt
d-excess(‰)
δ18O (‰)
(a) HIRS


River Water
precipitation in Cryosphere belt
precipitation in Vegetation belt
d-excess(‰)
δ18O (‰)
(b) USYR

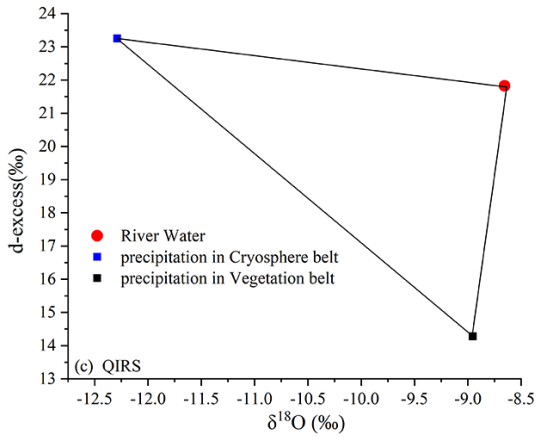
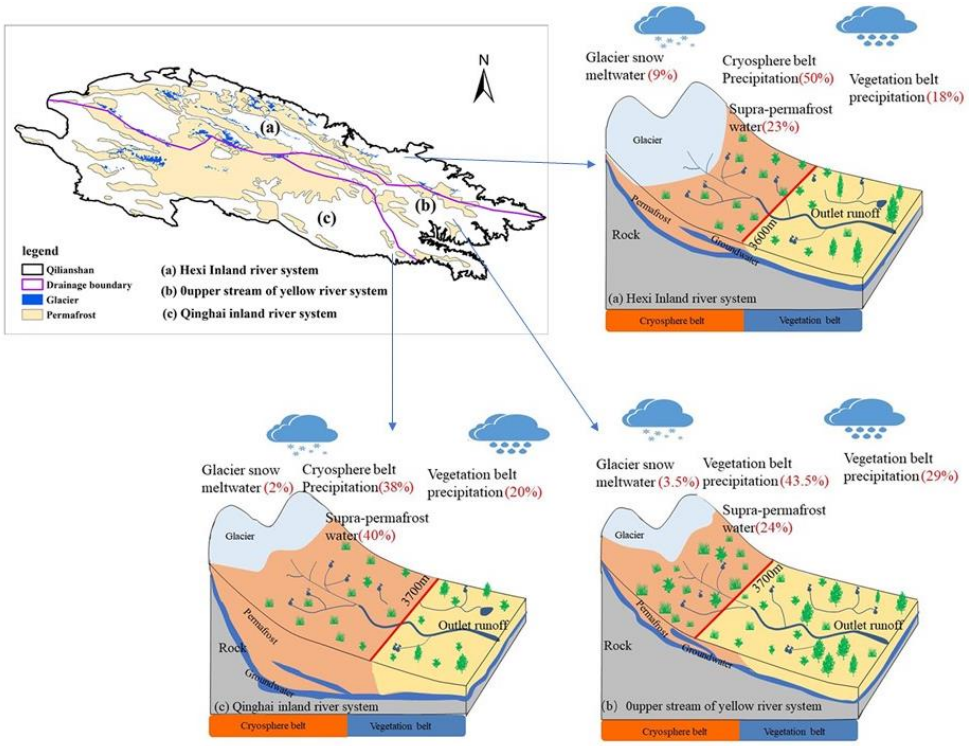

River Water
precipitation in Cryosphere belt
precipitation in Vegetation belt
d-excess(‰)
δ18O (‰)
(c) QIRS

Fig. 10



1171
1172 Fig. 11

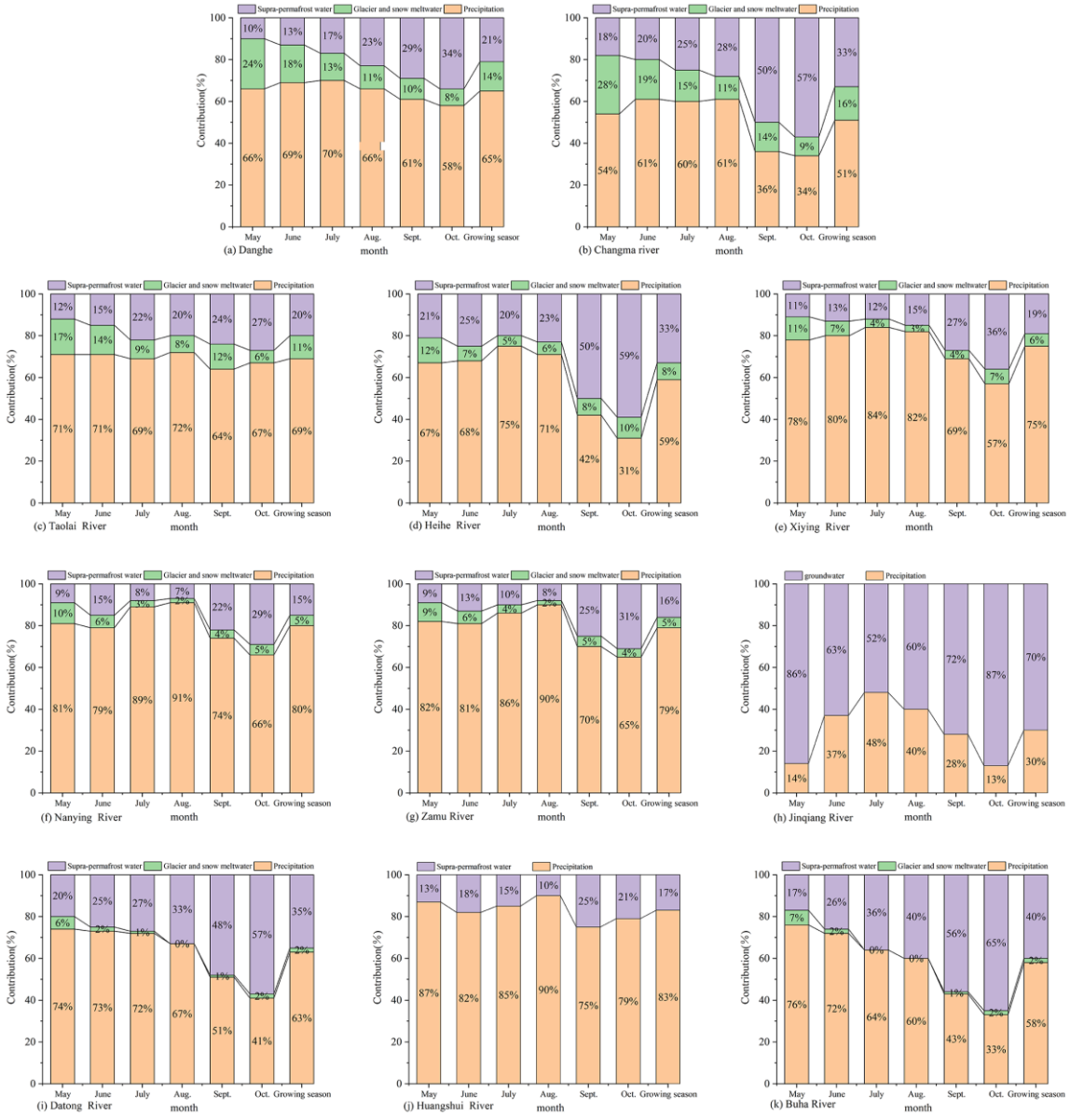
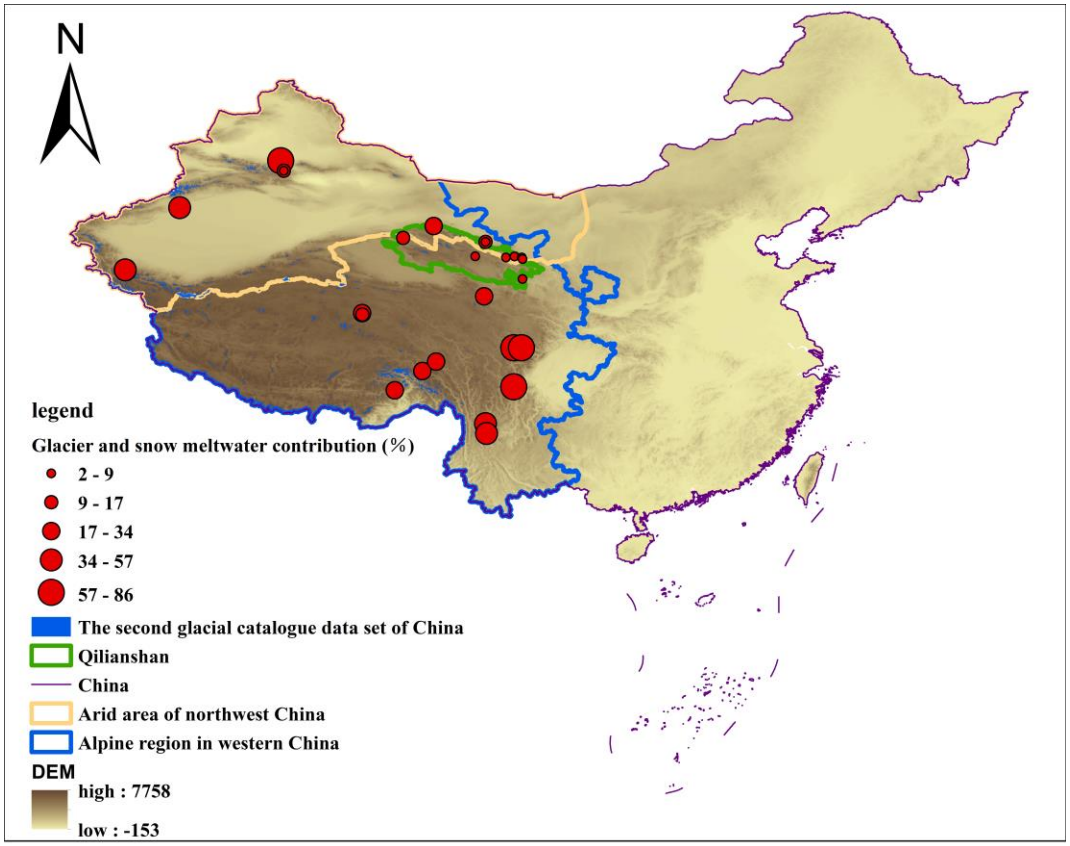

Supra-permafrost water
Glacier and snow meltwater
Precipitation
groundwater
Contribution(%)
May
June
July
Aug.
Sept.
Oct.
Growing season
month
(a) Danghe
(b) Changma river
(c) Taolai River
(d) Heihe River
(e) Xiying River
(f) Nanying River
(g) Zamu River
(h) Jinqiang River
(i) Datong River
(j) Huangshui River
(k) Buha River

Fig. 12



1175

1176 Fig. 13

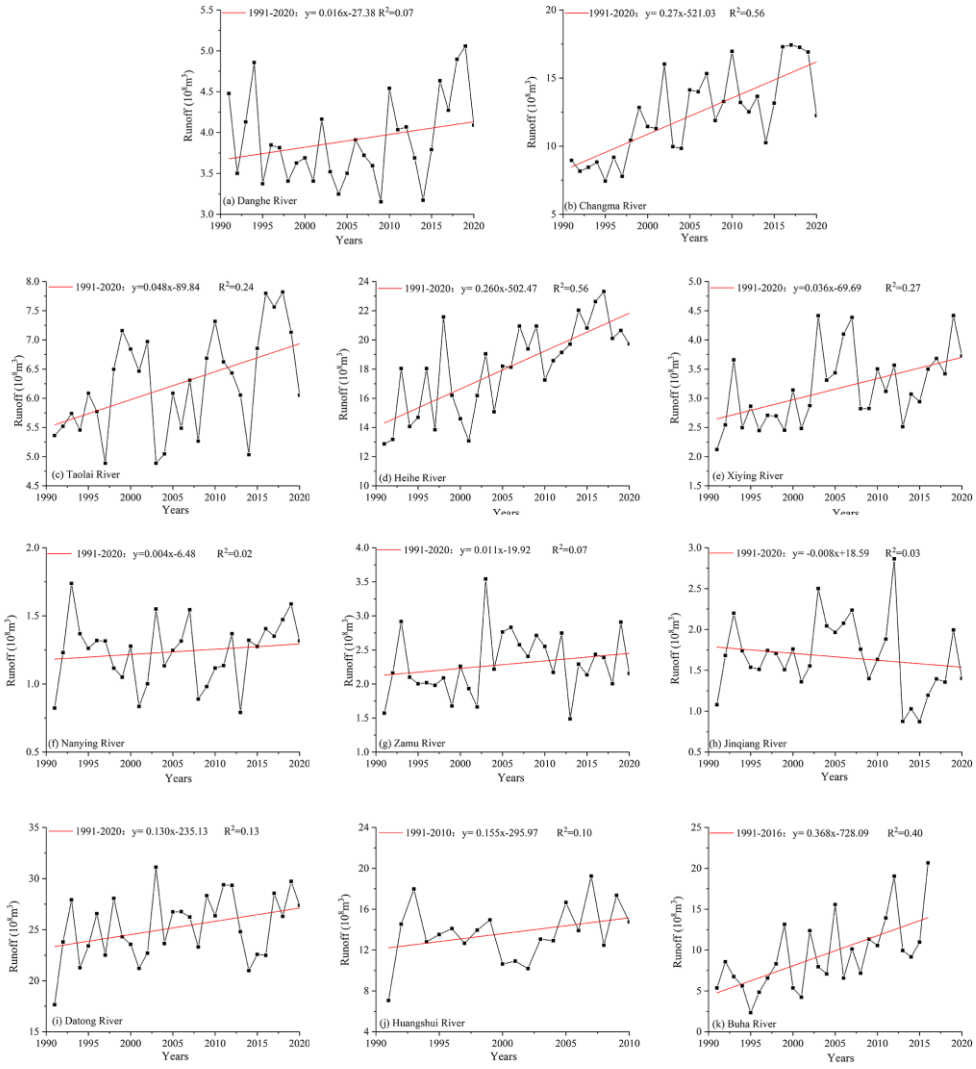

1991-2020：y= 0.016x-27.38 R²=0.07
Runoff (10^8m^3)
(a) Danghe River
Years
1991-2020：y= 0.27x-521.03 R²=0.56
(b) Changma River
1991-2020：y=0.048x-89.84 R²=0.24
(c) Taolai River
1991-2020：y= 0.260x-502.47 R²=0.56
(d) Heihe River
1991-2020：y=0.036x-69.69 R²=0.27
(e) Xiying River
1991-2020：y=0.004x-6.48 R²=0.02
(f) Nanying River
1991-2020：y= 0.011x-19.92 R²=0.07
(g) Zamu River
1991-2020：y= -0.008x+18.59 R²=0.03
(h) Jinqiang River
1991-2020：y= 0.130x-235.13 R²=0.13
(i) Datong River
1991-2010：y= 0.155x-295.97 R²=0.10
(j) Huangshui River
1991-2016：y= 0.368x-728.09 R²=0.40
(k) Buha River

Fig. 14

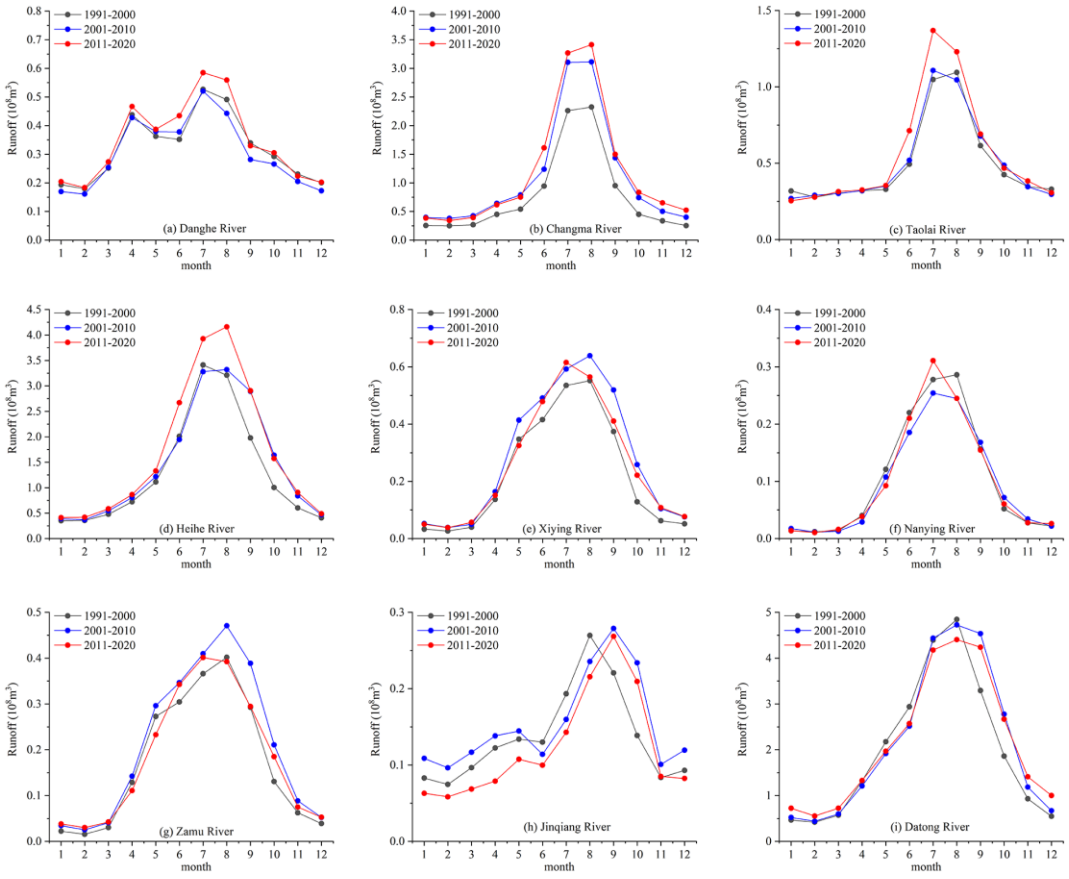

1991-2000
2001-2010
2011-2020
Runoff (10^8m^3)
month
(a) Danghe River
(b) Changma River
(c) Taolai River
(d) Heihe River
(e) Xiying River
(f) Nanying River
(g) Zamu River
(h) Jinqiang River
(i) Datong River

Fig. 15

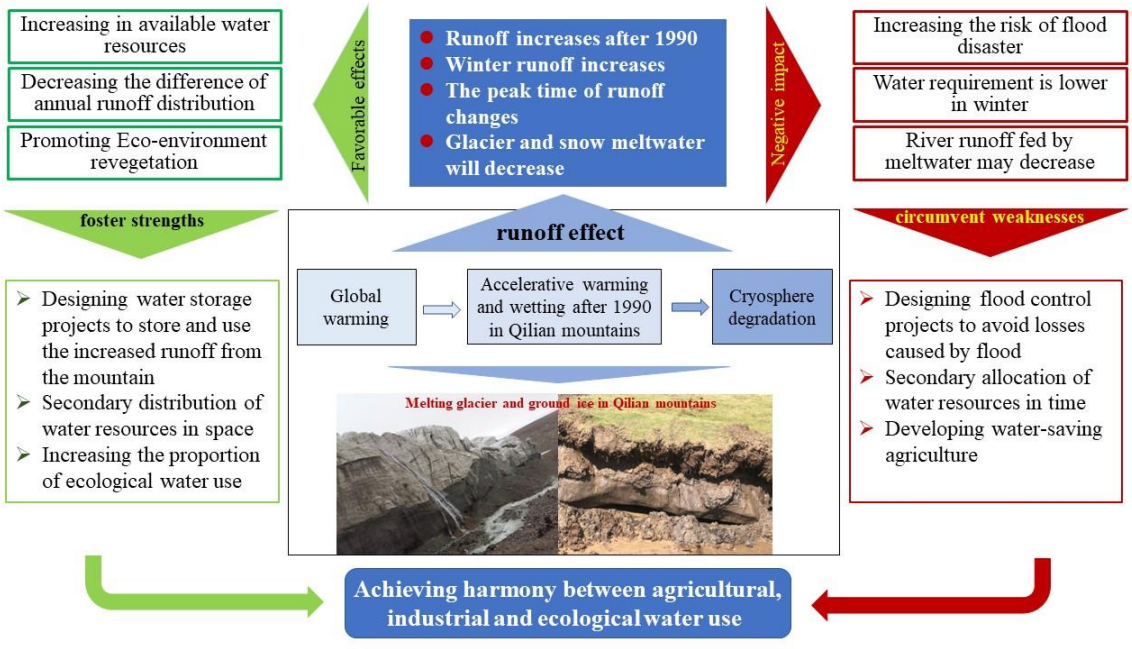

Increasing in available water resources
Decreasing the difference of annual runoff distribution
Promoting Eco-environment revegetation
Favorable effects
Runoff increases after 1990
Winter runoff increases
The peak time of runoff changes
Glacier and snow meltwater will decrease
Negative impact
Increasing the risk of flood disaster
Water requirement is lower in winter
River runoff fed by meltwater may decrease
foster strengths
runoff effect
circumvent weaknesses
Global warming
Accelerative warming and wetting after 1990 in Qilian mountains
Cryosphere degradation
Melting glacier and ground ice in Qilian mountains
Designing water storage projects to store and use the increased runoff from the mountain
Secondary distribution of water resources in space
Increasing the proportion of ecological water use
Designing flood control projects to avoid losses caused by flood
Secondary allocation of water resources in time
Developing water-saving agriculture
Achieving harmony between agricultural, industrial and ecological water use

Fig. 16