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ORIGINAL PAPER

Impact of climate variability and anthropogenic activity on streamflow in the Three Rivers Headwater Region, Tibetan Plateau, China

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Abstract Under the impacts of climate variability and human activities, there is violent fluctuation for streamflow in the large basins in China. Therefore, it is crucial to separate the impacts of climate variability and human activities on streamflow fluctuation for better water resources planning and management. In this study, the Three Rivers Headwater Region (TRHR) was chosen as the study area. Long-term hydrological data for the TRHR were collected in order to investigate the changes in annual runoff during the period of 1956–2012. The nonparametric Mann–Kendall test, moving *t* test, Pettitt test, Mann–Kendall–Sneyers test, and the cumulative anomaly curve were used to identify trends and change points in the hydro-meteorological variables. Change point in runoff was identified in the three basins, which respectively occurred around the years 1989 and 1993, dividing the long-term runoff series into a natural period and a human-induced period. Then, the hydrologic sensitivity analysis method was employed to evaluate the effects of climate variability and human activities on mean annual runoff for the human-induced period based on precipitation and potential evapotranspiration. In the human-induced period, climate variability was the main factor that increased (reduced) runoff in LRB and YARB (YRB) with contribution of more than 90 %, while the increasing (decreasing) percentage due to human activities only accounted for less than 10 %, showing that runoff in the TRHR is more sensitive to climate variability than human activities. The intra-annual distribution of runoff shifted gradually from a double peak pattern to a single peak pattern, which was mainly influenced by atmospheric circulation in the summer and autumn. The inter-annual variation in runoff was jointly controlled by the East Asian monsoon, the westerly, and Tibetan Plateau monsoons.

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1 Introduction

The hydrological cycle of a basin is a complex process influenced by climate, the physical characteristics of the basin, and by anthropogenic activity. With the worsening of water shortage problems and the increasing number of water-related disasters globally, the effects of climate variability and anthropogenic activity on water resources have long been a focus of global hydrological research (Ren et al. 2002; IPCC 2007; Scanlon et al. 2007; Jiang et al. 2015). Climate variability is believed to have resulted in global warming and changing patterns of precipitation, while anthropogenic activities have changed the temporal and spatial distribution of water resources (Govinda 1995; Ye et al. 2003; Milly et al. 2005). In arid and semiarid regions, the effects of climate variability and anthropogenic activities on runoff are significantly more sensitive and, therefore, resulted in significant changes in water yield (Brown et al. 2005; Ma et al. 2008; Jiang et al. 2011;

Jiang and Zhang 2015; Jiang and Wang 2016). Quantitatively evaluating these effects is important for regional water resource assessment and management.

The effects of anthropogenic activities on runoff in northern China have traditionally been estimated by computing their impact on each component of the water balance equation (Ren et al. 2002). This method, however, is limited because of the challenges involved in computing the direct effect of anthropogenic activity on each component for complex and rapidly changing water supply and utilization scenarios. New attempts, including regression analysis (Ye et al. 2003; Huo et al. 2008; Tian et al. 2009), sensitivity analysis (Dooge et al. 1999; Milly and Dunne 2002; Ma et al. 2008; Jiang et al. 2011), and hydrologic model simulation methods (Jones et al. 2006; Wang et al. 2008; Liu et al. 2010), have recently been used to address this problem. The sensitivity analysis method is a widely used framework to estimate the sensitivity of annual runoff to precipitation and potential evaporation (Dooge et al. 1999; Milly and Dunne 2002). Li et al. (2007), Ma et al. (2008), and Jiang et al. (2011) used the sensitivity analysis method to separate the effects of climate variability and anthropogenic activity on runoff in the Wuding River Basin, Shiyang River Basin, and Laohahe River Basin, respectively, and showed that the impacts of climate variability and anthropogenic activity on river discharge were more significant in arid and semiarid areas than that in more humid regions.

The Three Rivers Headwater Region (TRHR) of China plays an important role in East Asian river systems (Liu et al. 2008; Fang 2012; Jiang et al. 2016a). The Lancang River (known as the Mekong River in its downstream portions) is an international river and is the 'mother river' of South Asian countries. Therefore, it occupies an extremely important position in terms of international geopolitical relationships and politics (Fang 2012). The TRHR has an altitude ranging between 3335 and 6564 m and includes widely distributed glaciers, permafrost, and wetlands, making its river systems unique, which has been a key factor in studies of cryosphere hydrology and water resources on the Tibetan Plateau (Liu et al. 2008; Tong et al. 2014). In the last several decades, studies have shown that the regional climate is becoming warmer and wetter (Shi et al. 2007; Piao et al. 2010) and this trend is likely to continue into the future (Shi et al. 2007). Therefore, it is important to understand the hydrological responses to these changes in order to develop sustainable basin management strategies.

The present study made use of observational data from 3 hydrological stations and 21 meteorological stations across the TRHR in order to pursue the following objectives: (1) to investigate long-term streamflow and climate variability in the major regional rivers, focusing on the analysis of intra-annual, inter-annual, and inter-decadal patterns of change; (2) to determine changes in streamflow and estimate the effects of climate variability and anthropogenic activity on streamflow; and (3) to investigate the possible relationship between climate and intra-annual and inter-annual distribution of streamflow. The primary goal of this study was to evaluate the impact of climate variability and anthropogenic activity on streamflow and to provide basic information for modelling, forecasting water resources, and for local governmental planning and decision-making with regard to water resources.

2 Study area

The TRHR is located in the hinterland of the Tibetan Plateau, southern Qinghai Province, and contains the headwaters of the Yellow River, Yangtze River, and Lancang River (Jiang et al. 2016b; see Fig. 1). Geographically, it is located between 31° 29′ N and 36° 12′ N and between 89° 45′ E and 102° 23′ E. It has a total area of ~39.6×10^4 km^2 (Fang 2012). The topography of the TRHR is mainly mountainous, with altitude ranging between 3335 and 6564 m, and high mountains with altitudes of 4000–5800 m forming the main skeleton of the topography. The dominant ranges within the area include the Kunlun Mountains (and their branch range, the Aemye Machhen Range), the Bayan Har Mountains, and the Tanggula Mountains. Among these, the Bayan Har Mountains form the watershed between the Yangtze River and the Yellow River (Tong et al. 2014). The TRHR has a typical plateau continental climate, with alternating hot and cold seasons, distinct wet and dry seasons, annual mean temperature of −5.6 to −3.8 °C, and annual precipitation between 262.2 and 772.8 mm from west to southeast (Liu et al. 2014; Tong et al. 2014; Jiang and Zhang 2016a, b). Regarding water resources, the annual mean sreamflow is 374.3×10^8 m^3 in total, which is respectively 201.0×10^8 m^3 in YRB, 129.0×10^8 m^3 in YARB, and 44.3×10^8 m^3 in LRB. TRHR is the world's largest alpine wetland ecosystem. As one of the regions with the world's highest elevation and the most extensive and concentrated distribution of wetlands, the TRHR also belongs to one of the most sensitive and fragile regions in the world (Tong et al. 2014).

3 Data collection and methodology

3.1 Data collection and quality control

Monthly mean discharge data (1956–2012) from three hydrological gauging stations on the mainstreams of the YRB, YARB, and LRB were obtained from the Hydrological Year Book of Qinghai Province, issued by the Ministry of Water Resources of the People's Republic of China (PRC). In instances when runoff data were missing, we used runoff data from similar rainfall conditions at other times as a replacement. Monthly meteorological data (1956–2012) from 21 meteorological stations within the TRHR and surrounding areas

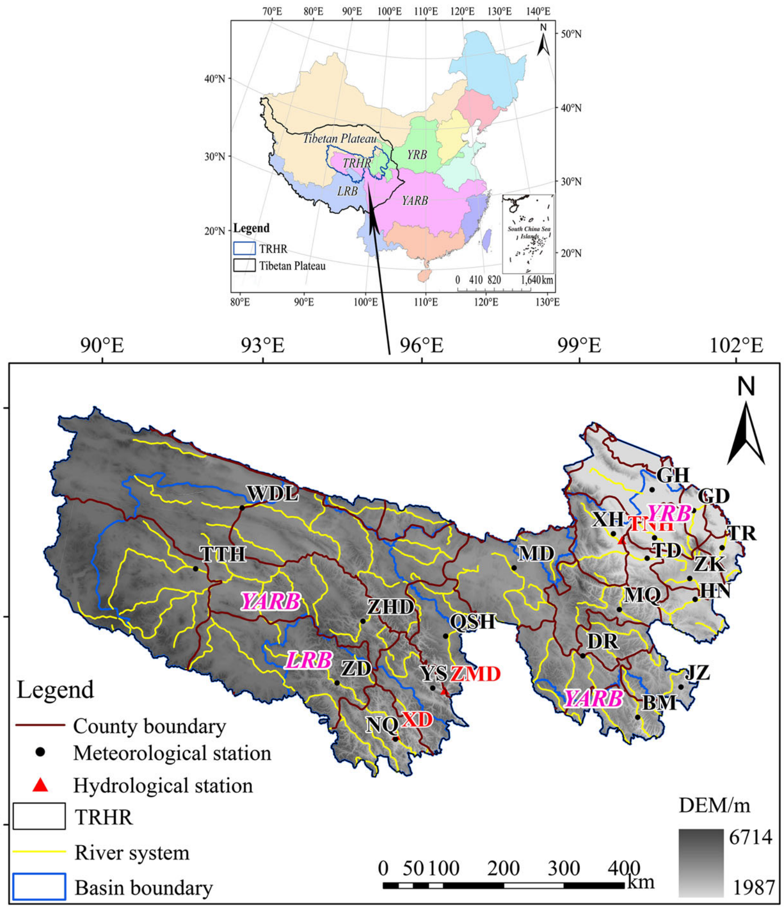

Fig. 1 Three Rivers Headwater Region (TRHR), northwest China, with the gauge stations (*red triangles*) and meteorological stations (*black dots*) used in this study denoted

were provided by the National Meteorological Information Centre (NMIC). To guarantee the accuracy of the results, the data was preprocessed before the analysis. The observational data of missing data years of more than 5 years (including 5 years) were excluded. The time series data of partial relocation stations were unified, and the remaining missing observation data were completed with a linear regression method and adjacent station interpolation to ensure the integrity of the time series. The division of the year is based on conventional meteorological seasons: December to February (DJF), March to May (MAM), June to August (JJA), and September to November (SON). The regional averages of meteorological variables are calculated from observation records of gauging stations, within and neighbouring area of each river basin, by using the Thiessen Polygon method.

Daily maximum and minimum air temperature, relative humidity, sunshine hours, and wind speed data were used to calculate potential evapotranspiration (*PET*) via the Penman-Monteith equation, as recommended by the United Nations Food and Agriculture Organization (FAO) (Allen et al. 1998; Jiang et al. 2016c). Spatial distribution and detailed information on the hydrological and meteorological gauging stations are provided in Fig. 1 and Tables 1 and 2. The Digital Elevation Model (DEM) and drainage map data were provided by the Data-Sharing Network of China Earth System Science (www.geodata.cn). The monthly reanalysis data of 100–600 hPa geopotential height from 1979 to 2012 was derived from National Centers for Environmental Prediction (NCEP). ArcGIS 10.0 software was used to process the mapping data. All measured data used in this study were checked by the corresponding agencies and rated as good quality.

3.2 Methodology

3.2.1 Nonparametric Mann-Kendall trend analysis

The nonparametric Mann–Kendall (MK) trend test (Mann 1945; Kendall 1975) was used in the flow, precipitation, and air temperature trend analysis. This test is widely used for trend detection because of its robustness for non-normally distributed and censored data, such as the hydroclimatic time series data used in this study (e.g., Yue et al. 2002a, b; Gao et al. 2011). The results of the trend tests were used to determine whether or not the observed time series of hydrological variables exhibited a statistically significant trend. For this, it was necessary to first test for, and then remove, any data with serial correlation before conducting the Mann–Kendall trend test (Miao et al. 2010). In this study, the trend-free pre-

Table 1 Selected meteorological stations in the research area, including ID number, station name, latitude, and longitude

ID	Name (abbreviation)	Latitude (°N)	Longitude (°E)	ID	Name (abbreviation)	Latitude (°N)	Longitude (°E)
1	Jiuzhi (JZ)	33.43	101.48	12	Wudaoliang (WDL)	35.22	93.08
2	Banma (BM)	32.93	100.75	13	Xinghai (XH)	35.58	99.98
3	Dari (DR)	33.75	99.65	14	Yushu (YS)	33.02	97.02
4	Gonghe (GH)	36.28	100.62	15	Zaduo (ZD)	32.90	95.30
5	Guide (GD)	36.03	101.43	16	Guinan (GN)	35.58	100.75
6	Henan (HN)	34.73	101.60	17	Tongde (TD)	35.27	100.65
7	Maduo (MD)	34.92	98.22	18	Zeku (ZK)	35.03	101.47
8	Nangqian (NQ)	32.20	96.48	19	Zhiduo (ZHD)	33.85	95.60
9	Qingshuihe (QSH)	33.80	97.13	20	Tongren (TR)	35.52	102.02
10	Qumalai (QML)	34.13	95.78	21	Maqin (MQ)	34.48	100.23
11	Tuotuohe (TTH)	34.22	92.43				

whitening (TFPW) method of Yue et al. (2002a) was used to analyse data trends. A Z statistic was obtained from the Mann-Kendall test on the whitened series by using the TFPW method. A negative value of Z indicated a downward trend, and vice versa.

In the Mann-Kendall test, the slope estimated using the Theil–Sen's estimator (Theil 1950; Sen 1968) is usually considered to represent the monotonic trend and indicates the variable quantity in the unit time. It is a robust estimate of the magnitude of a trend and has been widely used to identify the slope of trend lines in hydrological or climatic time series (Jhajharia et al. 2011). The estimator is given as:

$$\beta = \text{Median}\left(\frac{x_j - x_l}{j - l}\right) \quad \forall 1 < l < j \tag{1}$$

where $1 < l < j < n$, β is the median of all combinations of record pairs for the entire data set and is resistant to the effects of extreme observations. A positive β denotes an increasing trend, while a negative $\boldsymbol{\beta}$ indicates a decreasing trend.

3.2.2 Change point analysis

Identifying change points is one of the most important statistical techniques for runoff data analysis to study the effects of climate variability and human activities. In this study, the Mann–Kendall–Sneyers test (Mann 1945; Kendall 1975; Sneyers 1975), Pettitt test (Pettitt 1979), moving t test (Wei 2007), cumulative anomaly curve (CAC) (Wei 2007), and precipitation-runoff double cumulative curve (DCC) (Matouškov and Kliment 2009; Mu et al. 2010) were used to explore (abrupt) changes within hydro-meteorological series of the TRHR. The Mann–Kendall–Sneyers test, which was originally devised by Mann (1945) as a nonparametric test for detecting trends and was then used for the distribution of the test statistic by Kendall (1975), was used to test nonlinear trends as well as turning points. This test has the advantage of not assuming any distribution form for the data and has similar power to its parametric counterparts. Therefore, it is highly recommended for general use by the WMO and is widely used in the literature to analyse abrupt change in

Table 2 Summary of gauging stations and hydrological characteristics of the Three Rivers Headwater Region (TRHR), northwest China. DA denotes the drainage area

ID	River	Gauge station (abbreviation)	DA (km^2)	Location		Time series
				Longitude (°E)	Latitude (°N)	
1	Yellow River	Tangnaihai (TNH)	121,972	100.15	35.50	1956–2012
2	Yangze River	Zhimenda (ZMD)	137,704	97.24	33.01	1956–2012
3	Lancang River	Xiangda (XD)	17,907	96.48	32.25	1956–2012

climate data. The detailed theory of the Mann-Kendall test has been described by many previous researches (e.g., Miao et al. 2010).

We also used the nonparametric approach developed by Pettitt (1979) to detect change points in streamflow time series. This method detects a significant change in the mean of a time series when the exact time of the change is unknown (Pettitt 1979). The test uses a version of the Mann–Whitney statistic $U_{t,N}$ that tests whether two sample sets $x_1, \ldots x_t$ and $x_{t+1}, \ldots x_N$ are from the same population. The test statistic $U_{t,N}$ is given by:

$$U_{t,N} = U_{t-1,N} + \sum_{j=1}^{N} \mathrm{sgn}\left(X_t - X_j\right) \ \textit{for} \ \ t = 2, \ldots, N \tag{2}$$

and

$$\begin{array}{l} \textit{if} \ \left(X_t - X_j\right) \ > \ 0, \ \ \textit{sgn} \ \left(X_t - X_j\right) \ = \ 1 \\ \textit{if} \ \left(X_t - X_j\right) \ = \ 0, \ \ \textit{sgn} \ \left(X_t - X_j\right) \ = \ 0 \\ \textit{if} \ \left(X_t - X_j\right) \ < \ 0, \ \ \textit{sgn} \ \left(X_t - X_j\right) \ = -1 \end{array} \tag{3}$$

The test statistic counts the number of times a member of the first sample exceeds a member of the second sample. The null hypothesis of the Pettitt's test is the absence of a change point. The test statistic K_N and the associated probability (P) used in the test are given as:

$$K_N = {}^N \max_{1 \le t \le N} \left|U_{t,N}\right| \tag{4}$$

$$P \cong 2\exp\left\{-6(K_N)^2 \Big/ \left(N^3 + N^2\right)\right\} \tag{5}$$

In addition, the moving *t* test (Wei 2007) was used in order to enhance confidence in the (abrupt) change results. A general discussion of this method is available in Wei (2007).

Finally, the cumulative anomaly curve (CAC) (Wei 2007) and precipitation-runoff double cumulative curve (DCC) (Matouškov and Kliment 2009; Mu et al. 2010) were generated to identify change points of the runoff series. Precipitation-runoff DCC analysis provides a visual representation of the consistency of the precipitation and runoff data (Matouškov and Kliment 2009). The DCC should be a straight line if two variables are proportional, and the slope of this line will present the ratio between the two variables. Changes in the gradient of the curve may indicate that the characteristics of the precipitation or runoff have changed (Mu et al. 2010). In this study, DCC between precipitation and runoff was used as an auxiliary confirmation of the change points when anthropogenic activity influenced the river.

3.2.3 Hydrologic sensitivity analysis method

The hydrologic sensitivity analysis method can be described as the percentage change in mean annual runoff in response to changes in mean annual precipitation and *PET* (Jones et al. 2006; Li et al. 2007). The water balance for a basin can be written as:

$$P = \mathrm{AET} + Q + \Delta S \tag{6}$$

where P is precipitation, AET is the actual evapotranspiration, Q is runoff, and ΔS is the change in basin water storage. Over a long period of time (i.e., 10 years or more), it is reasonable to assume that $\Delta S = 0$.

Annual mean *AET* can be estimated from precipitation and *PET* (Zhang et al. 2001) as:

$$\frac{AET}{P} = \frac{1 + \omega\left(PET/P\right)}{1 + \omega\left(PET/P\right) + \left(P/PET\right)} \tag{7}$$

where ω is the plant-available water coefficient related to vegetation type (Zhang et al. 2001). Details of the relationship can be found in Zhang et al. (2004). In this study, we calibrated parameter ω by comparison with the long-term annual *AET*, as calculated from Eqs. (6) and (7).

A change in mean annual runoff can be calculated as:

$$\Delta Q_{\mathrm{tot}} = Q_{\mathrm{obs2}} - Q_{\mathrm{obs1}} \tag{8}$$

where ΔQ_{tot} indicates the observed change in mean annual runoff between two different periods, Q_{obs1} is the average annual runoff during the reference period, and Q_{obs2} is the average annual runoff during other periods.

As a first-order approximation, the total change in mean annual runoff can be estimated as follows:

$$\Delta Q_{\mathrm{tot}} = \Delta Q_{\mathrm{clim}} + \Delta Q_{\mathrm{hum}} \tag{9}$$

where ΔQ_{clim} is the change in mean annual runoff due to climate variability and ΔQ_{hum} represents the change in mean annual runoff due to various anthropogenic activities.

Perturbations in both precipitation and *PET* can lead to changes in the water balance (Dooge et al. 1999). Based on the hydrologic sensitivity relationship, the change in mean annual runoff due to climate variability can be approximated as (Koster and Suarez 1999; Milly and Dunne 2002):

$$\Delta Q_{clim} = \frac{\partial Q}{\partial P}\Delta P + \frac{\partial Q}{\partial PET}\Delta PET \tag{10}$$

where ΔP and ΔPET denote changes in precipitation and *PET*, respectively, $\frac{\partial Q}{\partial P}$ and $\frac{\partial Q}{\partial PET}$ are the coefficients of sensitivity of runoff to precipitation and *PET*, respectively. They can be expressed as follows (Li et al. 2007):

$$\frac{\partial Q}{\partial P} = (1 + 2x + 3\omega x)\Big/\left(1 + x + \omega x^2\right)^2 \tag{11}$$

$$\frac{\partial Q}{\partial PET} = -(1 + 2\omega x)\Big/\left(1 + x + \omega x^2\right)^2 \tag{12}$$

where x is the index of dryness and is equal to PET/P, and ω is same as in Eq. (7).

Once ΔQ_{clim} is obtained, ΔQ_{hum} can be derived using Eq. (9). The relative contributions of climate variation and anthropogenic activity to runoff can be then expressed by a percentage, calculated as:

$$\eta_{\text{clim}} = \frac{\Delta Q_{\text{clim}}}{|\Delta Q_{\text{obs}}|} \times 100\% \tag{13}$$

$$\eta_{\text{hum}} = \frac{\Delta Q_{\text{hum}}}{|\Delta Q_{\text{obs}}|} \times 100\% \tag{14}$$

where η_{clim} and η_{hum} are the percentage of the climate-variation-induced impact and human-activity-inducted impact on runoff, respectively.

4 Results and analysis

4.1 Temporal and trend analysis of hydro-meteorological factors

4.1.1 Temporal and trend analysis of meteorological factors

The results show that over the past 57 years, precipitation and temperature systematically increased across the entire TRHR (Fig. 2a, b), with the slopes of the trend lines being 1.06 mm year^{-1} ($P<0.01$) and 0.03 °C year^{-1} ($P<0.001$), respectively. Annual precipitation in TRHR averaged 417.3 mm and ranged from 428.0 mm (1962) to 562.8 mm (1989; Fig. 2a). Spatially, the mean annual precipitation varied from 381.2 mm in the YARB to 405.5 mm in the YRB. Precipitation in LRB (485.5 mm) was higher than in YRB and YARB (Table 3). The annual mean temperature in the TRHR averaged 0.2 °C and ranged from −1.0 °C (1957) to 1.8 °C (2009; Fig. 2b). Spatially, the annual mean temperature varied from −2.8 °C in the YARB to 1.1 °C in the YRB. The mean annual temperature in LRB (2.3 °C) was higher than that in the YRB and YARB (Table 3); *PET* for the entire TRHR showed an obvious increasing trend with the slope of 1.43 mm year^{-1} ($P<0.001$; see Fig. 2c). *PET* had a mean value of 827.9 mm; the estimated *PET* in the TRHR varied from 698.2 mm (1956) to 914.7 mm (2009; Fig. 2c). The results suggest that the humid index (ratio of precipitation to *PET*) remained relatively stable across the 57-year study period. However, due to dramatic increases in rainfall and temperature since 2000, the humid index also presented a significant increasing trend since 2000, which had a rate of 0.01 year^{-1} ($P<0.01$).

4.1.2 Temporal and trend analysis of streamflow

Annual runoff depth in the Tangnaihai station of YRB showed an average of 146.8 mm for the 57-year study period (1956–2012), fluctuating widely from 86.6 mm in the driest year (2002) to 268.7 mm in the wettest year (1989; Fig. 3). Runoff depth in the Tangnaihai station (YRB) declined slowly at a rate of 1.6 mm decade^{-1} ($P>0.1$). During the same period, runoff in the Zhimenda station (YARB) and Xiangda station (LRB) showed slightly increasing trends with rates of 2.7 mm decade^{-1} ($P<0.1$) and 6.1 mm decade^{-1} ($P<0.1$), respectively (Fig. 3).

Fig. 2 Temporal variations of meteorological factors in the Three Rivers Headwater Region (TRHR) between 1956 and 2012, including: **a** annual rainfall, **b** annual mean temperature, **c** annual potential evapotranspiration (*PET*), and **d** annual mean humid index. The change trends of meteorological variables reached different significance level, for example, $P=0.001$, $P=0.01$, $P=0.05$, and $P=0.1$. When the $P>0.1$, we consider the trend is not significant. The same as follows

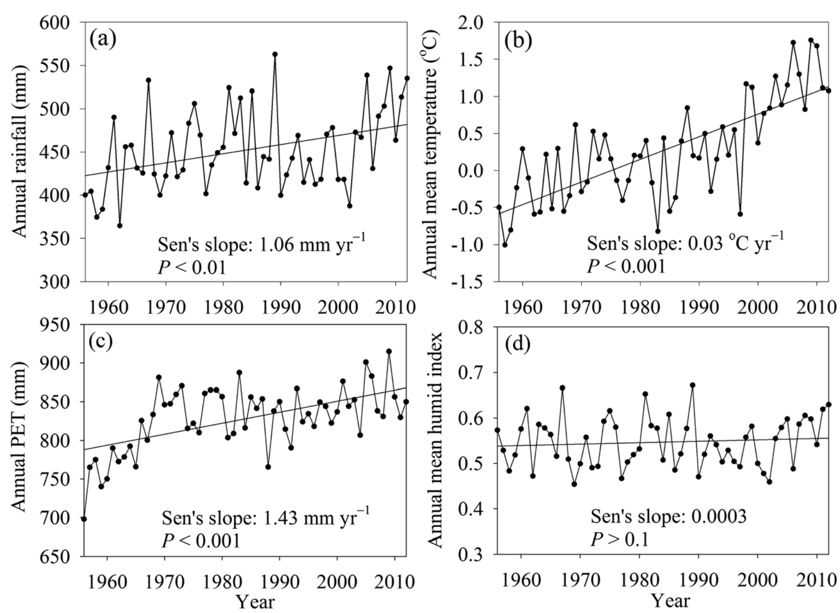

Table 3 Decadal means of precipitation and temperature from 1956 to 2012 in the Three-Rivers Headwater Region (TRHR), northwest China. YRB, YARB, and LRB represent the Yellow River Basin, Yangtze River Basin, and Lancang River Basin, respectively. $\overline{P}$ and $\overline{T}$ are the mean values of precipitation and temperature, respectively

Basin	Precipitation (mm)							Temperature (°C)						
	1956–1959	1960s	1970s	1980s	1990s	2000–2012	$\overline{P}$	1956–1959	1960s	1970s	1980s	1990s	2000–2012	$\overline{T}$
YARB	317.1	388.3	400.5	381.6	405.6	394.2	381.2	−4.3	−3.0	−2.8	−2.8	−2.4	−1.6	−2.8
LRB	401.3	470.5	454.5	494.4	540.4	552.0	485.5	1.3	1.9	2.1	2.3	2.5	3.4	2.3
YRB	329.8	375.6	424.1	425.5	430.1	447.8	405.5	0.05	0.9	1.1	1.1	1.4	2.1	1.1
TRHR	390.5	398.5	419.9	417.4	435.5	442.0	417.3	−0.6	−0.1	0.04	0.1	0.4	1.1	0.2

4.2 Change point analysis of hydro-meteorological factors

We used several methods to detect change points of temperature, precipitation and streamflow, as shown in Table 4. For the Pettitt test and Mann–Kendall–Sneyers test, no change points were detected in the series of precipitation and streamflow, and the change points for temperature were not same by using these two methods. Therefore, we used the CAC to detect change (turning) points (see Fig. 4 and Table 4). In Fig. 4a–c, the change points for streamflow of YRB and YARB occurred in 1989, while it was 1993 for LRB. The same as streamflow, the change points of precipitation occurred at the same pace as streamflow in each basin (Fig. 4d–f), which indicated that precipitation fluctuation had a direct impact on streamflow change and that they varied at the same pace. In order to valid the change point detected by CAC, we used moving t test to detect (see Table 4). The change points generated by moving t test were almost same as those of CAC. For the temperature, the change points for three rivers occurred in 1986; since then, the CAC turned from downward trend to upward trend (see Fig. 4g–i). The result of moving t test for temperature reached the $P = 0.05$ significance level.

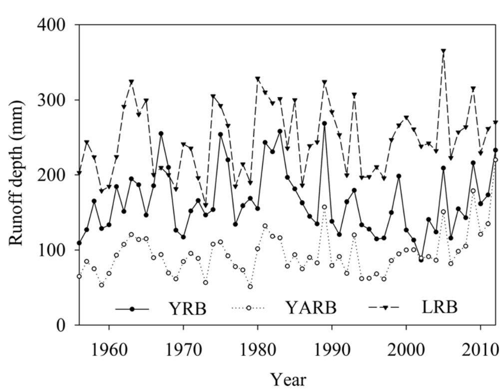

Fig. 3 Inter-annual variability of runoff depth in the three basins of the Three Rivers Headwater Region (TRHR) between 1956 and 2012

The relationship between cumulative annual precipitation and runoff depth (Fig. 5) demonstrated that in all the three basins (LRB, YRB, and YARB), precipitation and runoff were relatively uniform before 1989 (1993) and changed thereafter. Combined analysis of CAC and precipitation-runoff DCC indicated that 1989 and 1993 could be the change points reflecting the effect of evaporation and anthropogenic activity on runoff in YRB, YARB, and LRB. Therefore, the 57-year study period was divided into two time series: 1956–1989 (1956–1993), where minimal anthropogenic activity and a slow warming trend took place, and 1990–2012 (1994–2012), where land use change and river engineering projects were undertaken, accompanied by a dramatic warming trend.

4.3 Intra-annual variation of streamflow

Streamflow variation in all three basins was similar, and so we here discuss YRB as a representative example. Table 5 shows the decadal mean streamflow, anomalous percentage of streamflow, and coefficient of variation at Tangnaihai Station (in the YRB) from 1956 to 2012. As shown in Table 5, C_v was the largest in the 1980s, followed by the 1960s and 1970s. Accordingly, streamflow in 1980s was the highest, at 21.7 % greater than the mean level, while the anomalies in the 1960s and 1970s were 9.1 and 2.8 % higher than the mean level, respectively. The smallest C_v appeared in the 1950s and 1990s, when flow was 16.8 and 10.6 % lower than mean level, respectively; however, C_v between 2000 and 2012 was similar to the mean level, with the flow just 6.0 % less.

As shown in Table 5 and Fig. 6, the flow in autumn maintained the same pace as the annual flow. When the flow in autumn presented positive anomalies, a second peak appeared; thus, the annual flow increased. The flow in the 1980s was unevenly distributed and presented an obvious bimodal distribution (Fig. 6b). Intra-annual distribution in the 1960s and 1970s was also bimodal, with two peak values in

Table 4 The change point detection of hydro-meteorological variables by using different methods in the TRHR

Hydro-meteorological variables		Moving t–test	Pettitt test	Mann–Kendall–Sneyers test	Cumulative anomaly curve
Annual mean temperature	YRB	–	–	–	1986
	YARB	1986	1986	1997	1986
	LRB	1986	1983	1997	1986
Annual precipitation	YRB	1989	–	–	1989
	YARB	1989	–	–	1989
	LRB	—	–	–	1993
Annual streamflow	YRB	1989	–	–	1989
	YARB	1989	–	–	1989
	LRB	1993	–	–	1993

– ' denotes no significant change point

July and September; however, these peak values were smaller than those in the 1980s. The smallest C_v appeared in the 1990s, when the intra-annual distribution showed a single peak pattern, with the peak value, which was the smallest in several decades, appearing in July (Fig. 6b). The flow in the period 1956–1960 also presented a single peak pattern and had the smallest observed mean flow values. Flow since 2000 was still relatively low and had an intra-annual distribution similar to the bimodal pattern but with a lower flow in the second peak (September).

4.4 Calibration and validation of the hydrologic sensitivity analysis method

The effect of climate variability on runoff was estimated using the hydrologic sensitivity analysis method. During the baseline period, the anthropogenic activity did not cause significant perturbations in runoff within the TRHR, and so we can assume that anthropogenic activity did not affect runoff during the baseline period. Thus, the baseline period of 1956–1989 (1956–1993) was used to estimate the effects of climate

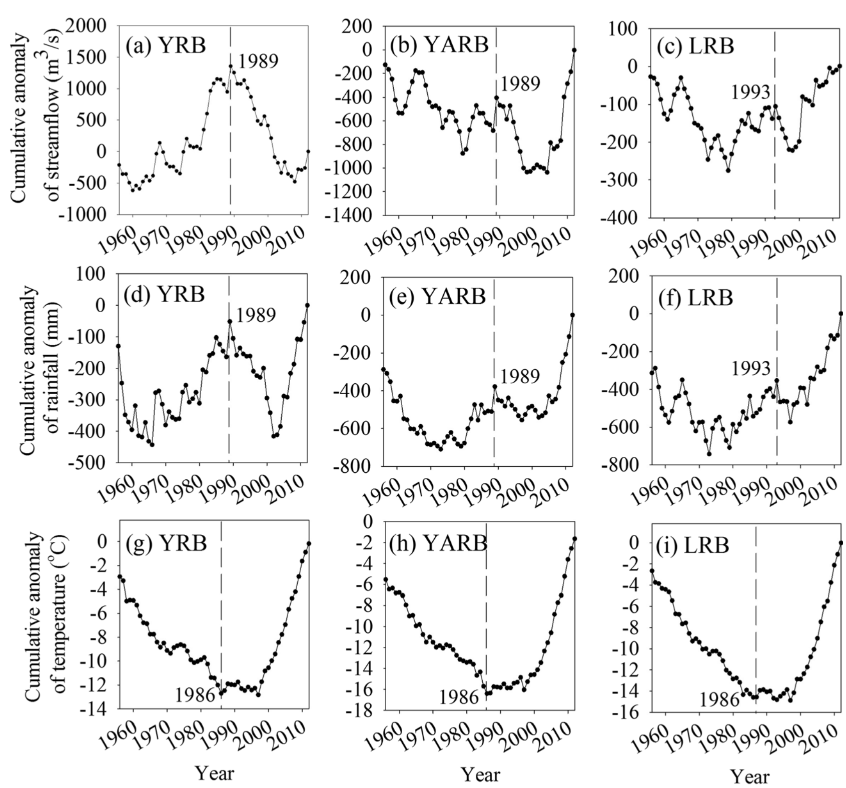

Fig. 4 Cumulative anomaly curve (CAC) of annual mean temperature (**a**–**c**), annual precipitation (**d**–**f**), and annual streamflow (**g**–**i**) in the Three Rivers Headwater Region (TRHR) between 1956 and 2012

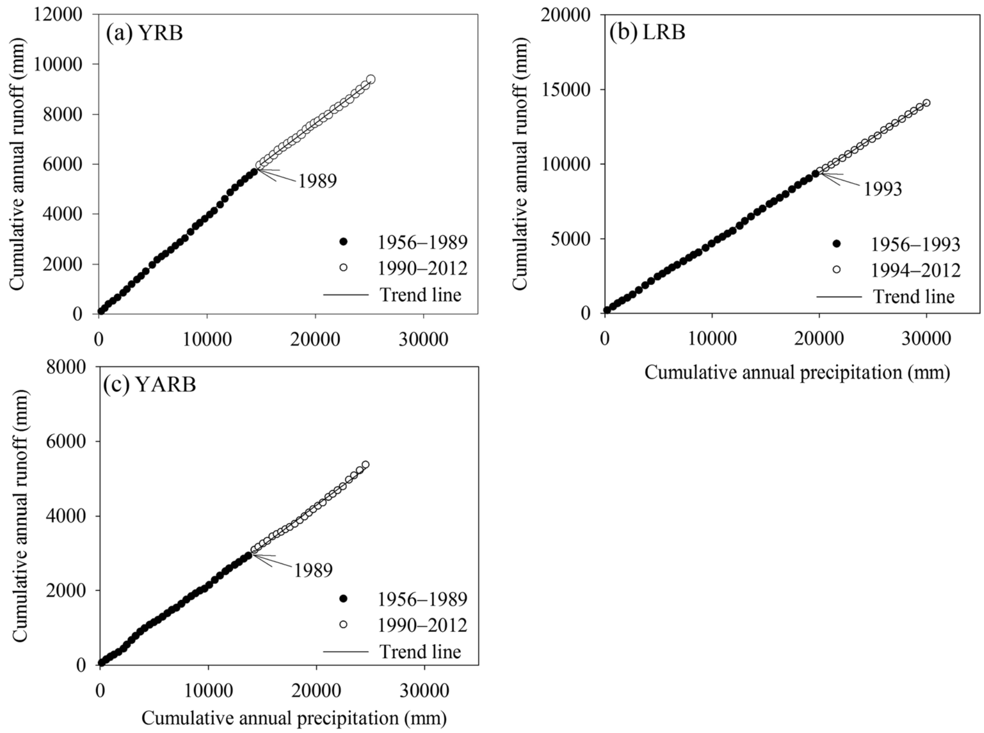

Fig. 5 Double cumulative curves (DCC) of annual precipitation and runoff depth in the baseline period of 1956–1989 (1956–1993) and change period of 1990–2012 (1994–2012) for: **a** the Yellow River Basin (YRB), **b** the Lancang River Basin (LRB), and **c** the Yangtze River Basin (YARB)

variability and anthropogenic activity on runoff in change period of 1990–2012 (1994–2012) using the hydrologic sensitivity analysis method.

In this method, ω is the main model parameter. We calibrated ω by comparison with the long-term annual *AET*, estimated by using Eq. (7), and the water balance estimated from Eq. (6) for the baseline period. We calculated the $\frac{\partial Q}{\partial P}$ and $\frac{\partial Q}{\partial PET}$ using Eqs. (11) and (12). With a value of $\omega_{\mathrm{YRB}}=-0.041$, $\omega_{\mathrm{YARB}}=-0.043$, and $\omega_{\mathrm{LRB}}=-0.045$, the results of annual *AET* estimated by using Eq. (7) are realistic and acceptable and have a value of 275.9, 285.2, and 290.1 mm in YRB, YARB, and LRB, respectively. Thus, we set $\omega_{\mathrm{YRB}}=-0.041$, $\omega_{\mathrm{YARB}}=-0.043$, and $\omega_{\mathrm{LRB}}=-0.045$. The terms of $\frac{\partial Q}{\partial P}$ and $\frac{\partial Q}{\partial PET}$ in Eqs. (11) and (12) can be considered as the sensitivity coefficients of runoff to changes in precipitation and *PET*, respectively. When $\omega_{\mathrm{YARB}}=-0.043$, the values of these

Table 5 Decadal mean streamflow, anomalous percentages of streamflow, and coefficients of variation (C_v) at Tangnaihai Station (Yellow River Basin; YRB) from 1956 to 2012

Items	Mean	1950s (1956–1959)	1960s (1960–1969)	1970s (1970–1979)	1980s (1980–1989)	1990s (1990–1999)	2000–2012
C_v	0.68	0.65	0.79	0.73	0.82	0.64	0.67
Annual flow (m^3/s)	636	521	683	644	762	560	588
Anomaly of annual flow (%)	0	−16.8	9.1	2.8	21.7	−10.6	−6.0
Spring flow (m^3/s)	376	326	398	403	404	396	329
Anomaly of spring flow (%)	0	−13.2	5.9	7.2	7.4	5.4	−12.5
Summer flow (m^3/s)	1098	1020	1119	1059	1347	989	1053
Anomaly of summer flow (%)	0	−7.1	1.9	−3.6	22.7	−9.9	−4.1
Autumn flow (m^3/s)	867	707	1019	926	1089	677	780
Anomaly of autumn flow (%)	0	−18.4	17.6	6.8	25.6	−21.9	−10.0
Winter flow (m^3/s)/	186	161	196	186	207	197	190
Anomaly of winter flow (%)	0	−13.3	5.1	−0.1	11.3	−4.0	2.4

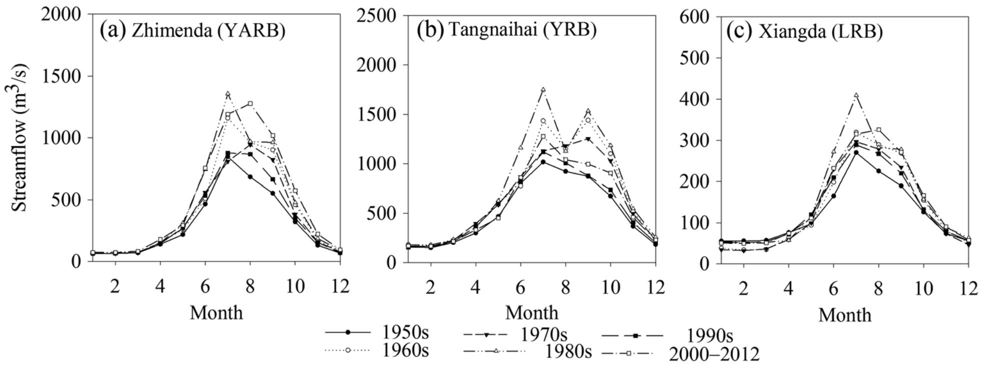

Fig. 6 Intra-annual variability of streamflow in the three basins of the Three-Rivers Headwater Region (TRHR) between 1956 and 2012: **a** Zhimenda Station in the Yangtze River Basin (YARB), **b** Tangnaihai Station in the Yellow River Basin (YRB), and **c** Xiangda Station in the Lancang River Basin (LRB)

sensitivity coefficients were 0.60 and 0.12, respectively, revealing that the change in runoff was more sensitive to precipitation than to *PET*.

The sensitivity coefficient to precipitation was higher for lower ω values and decreased with increases in the dryness index, and changes in precipitation led to greater changes in runoff for grassed basins than for forested basins as forested basins generally had larger ω values (Zhang et al. 2004; Ma et al. 2008). The magnitudes of $\frac{\partial Q}{\partial P}$ and $\frac{\partial Q}{\partial PET}$ approach zero under very arid conditions (e.g., large E/P ratios), suggesting that basins in humid regions respond more strongly to changes in precipitation and *PET* than basins in arid regions (Ma et al. 2008). However, the magnitude of the change in runoff depends on both the sensitivity coefficients and on changes in precipitation and *PET*.

4.5 Quantitative assessment of the impact of climate variation and anthropogenic activity on streamflow

The results indicated that in the TRHR and sub-basins, the proportional change in annual runoff due to climate variability accounted for >90 % of the observed change, while anthropogenic activity was responsible for ~10 % (Table 6). The contribution of anthropogenic activity in the YRB and LRB was 9.5 and 9.2 %, respectively, which were a little higher than that in YARB (6.1 %), due to the higher population densities and greater anthropogenic activity. During the change period, the effects of climate variability and anthropogenic activity on runoff showed a significant difference. It can be inferred that the increase in runoff during the change period was mainly due to climate variability. Fortunately, adopting exclusion measures for preventing grassland degradation have increased vegetation recover to a certain extent, which has had a positive effect on runoff.

5 Discussion

5.1 The connection between atmospheric circulation and intra-annual distribution of flow

The intra-annual distribution of flow in the YRB presented an obvious bimodal pattern; however, bimodal pattern distributions in the YARB and LRB were not clearly observed. Therefore, we chose the YRB as a case study for attempting to find a possible climatic attribution. Peak values of the bimodal pattern distribution in the YRB appeared in summer and autumn, and the flow was directly influenced by atmospheric circulation. Therefore, from the atmospheric circulation perspective, we discuss the relationships between flow in summer and autumn with the 600 hPa geopotential height in summer, so as to explore the possible causes of the bimodal pattern distribution. Something to note is that 100, 200, …,

Table 6 Quantifying the effects of climate variability and anthropogenic activities on runoff depth

Basin	Change period	ΔQ_{tot} (mm)	Runoff depth change (mm)		Proportional change in annual runoff depth due to	
			ΔQ_{clim}	ΔQ_{hum}	Climate variability (%)	Anthropogenic activities (%)
YRB	1990 − 2012	−25.5	−23.1	−2.4	90.5	9.5
YARB	1990 − 2012	11.2	10.5	0.7	93.9	6.1
LRB	1994 − 2012	4.3	3.9	0.4	90.8	9.2

and 600 hPa geopotential height are commonly used in related analysis. Actually, we plotted the contour map of 100, 200, …, and 600 hPa geopotential height vs. streamflow, but the correlation in other maps was not as good as 600 hPa geopotential height; therefore, we only showed this figure. Regarding the contour map, we calculated the correlation coefficient of streamflow and height field at 600 hPa in each grid; thus, the contour map can be plotted.

5.1.1 Relationship between summer flow and atmospheric circulation

Figure 7a shows the correlation coefficient between the summer flow anomaly of Tangnaihai station in YRB and the 600 hPa geopotential height anomaly in summer. The correlation analysis was done on seasonal scale (summer flow anomaly vs. summer geopotential height anomaly). Both flow anomaly and geopotential height anomaly were calculated based on mean level from 1979 to 2012. The negative correlation relationship was relatively strong, with the highest correlation coefficient (−0.45) located in within 90°–110°E and 30°–40°N, with the value passing the significance test level of 0.01. The flow correlated positively with the 600 hPa geopotential height field east of the Ural Mountains (65°–85°E, 50°–60°N), while it correlated negatively in Baikal Lake. Figure 7a shows that, when the Ural ridge and Baikal groove strengthened, it was conducive for cold air to move southward. When the 600 hPa geopotential height of the Tibetan Plateau decreased and the plateau summer monsoon was enhanced, low-level convergence and high-level divergence occurred, which increased rainfall and flow, resulting in the first flow peak (July). On the contrary, when the 600 hPa geopotential height of Tibetan Plateau increased and the plateau summer monsoon weakened, low-level divergence and high-level convergence reduced rainfall and flow, resulting in a relatively small flow peak.

5.1.2 Relationship between autumn flow and atmospheric circulation

The correlation between autumn flow anomaly of Tangnaihai station in YRB and the 600 hPa geopotential height anomaly in autumn was similar to that of the summer (figure has been omitted). However, it is worth noting that the flow in autumn was negatively correlated with the geopotential height field in summer (Fig. 7b), with the highest correlation coefficient (−0.60) located within 80°–110°E and 30°–40°N, with the value passing the significance test value of 0.001. The autumn flow in the YRB negatively correlated with the geopotential height of nearby Baikal Lake and positively correlated east of Ural Mountains (60°–90°E, 55°–70°N). When the Ural ridge and Baikal groove strengthened, it was conducive for cold air to move southward. An increase in the low pressure and plateau summer monsoon was conducive to increasing the autumn flow, resulting in the formation of the second peak. In contrast, if the autumn flow fell, the second peak disappeared.

5.2 Impact of monsoon change on streamflow inter-annual variation

The ebb and flow of the monsoons determines the start and end of the rainy season, and the strength of the monsoon affects the amount of rainfall. The TRHR is located in the hinterland of the Tibetan Plateau and is directly affected by the monsoon of the Tibetan Plateau itself. However, as the area is located in a transition zone between prevailing westerlies and the East Asian summer monsoon, runoff in the region is affected by the joint action of a number of monsoons (Xu et al. 2006). To determine the possible causes of streamflow variation, we examined Asian monsoon indices, including the East Asian monsoon index (EAMI), Tibet–Qinghai Plateau monsoon index (TPMI), and westerly index (WI). The Asian monsoon indices were created by Li et al.

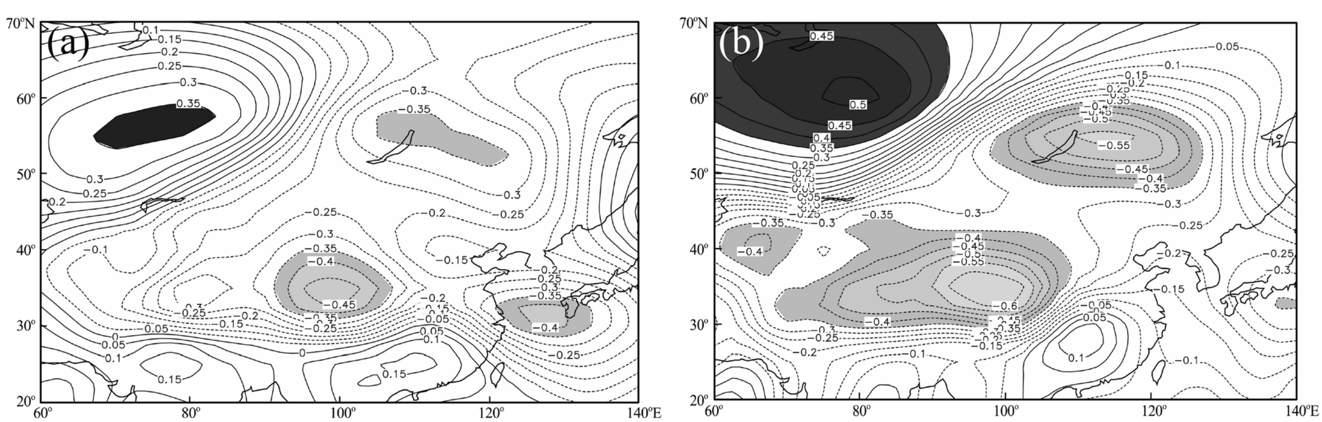

Fig. 7 Correlation coefficients between **a** summer runoff anomalies and **b** autumn runoff anomalies in the YRB and the summer anomaly of height field at 600 hPa between 1979 and 2012

(2010; 2011), and WI was derived from National Centers for Environmental Prediction (NCEP) reanalysis data. The TPMI was calculated from NCEP 600 mb height reanalysis data based on the method given by Wang et al. (1984).

Cumulative anomaly curves were drawn using the runoff vs. monsoon intensity indices at the Tangnaihai, Zhimenda, and Xiangda stations, in order to explore this relationship (Fig. 8). Runoff in the YRB (Tangnaihai) correlated positively with EAMI and negatively with WI. In years with a strong East Asian monsoon, runoff was high, but in years with a strong westerly, runoff was very low. In fact, these two monsoon systems have an inverse phase relationship in the Tibetan Plateau, such that when the East Asian monsoon is strong, the westerly is weak, and vice versa (Xu et al. 2006). It is worth noting that the correlation coefficient between runoff and the EAMI was low. It is therefore suggested that runoff is affected by a variety of monsoon systems, and so lacked direct correspondence with a specific index. In the YARB, the Tibetan Plateau monsoon plays leading roles, resulting in the correspondingly high correlation coefficients observed. This is especially the case with the TPMI, which had a correlation coefficient with runoff of 0.39 ($P<0.01$). The situation in the LRB was similar to that in the YRB, in that runoff was jointly controlled by the East Asian monsoon and westerlies; however, it had a higher correlation coefficient than the YRB.

5.3 Overview of the possible anthropogenic influences on water resources

The population in the TRHR is relatively small comparing with that in other regions due to the harsh natural environment. However, the population in this area presented

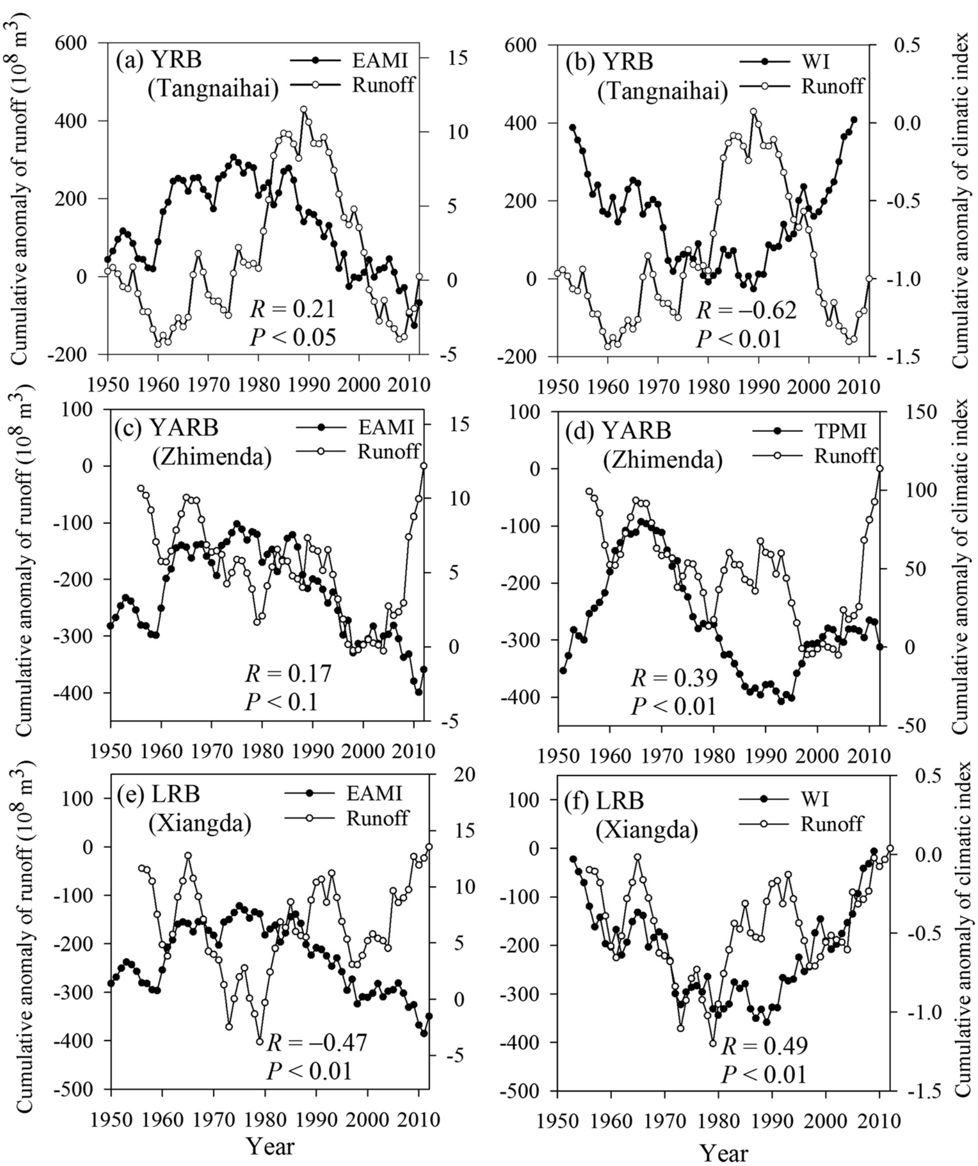

Fig. 8 Cumulative anomaly curves of runoff vs. the monsoon intensity index in the Three Rivers Headwater Region (TRHR). EAMI, WI, and TPMI represent the East Asian monsoon index, the westerly index, and the Tibetan Plateau monsoon index, respectively. **a**-**b**, **c**-**d**, and **e**-**f** are for YRB, YARB, and LRB, respectively. *R* and *P* represent the correlation coefficient and significance level, respectively

significant increasing trend, which led to more anthropogenic influences. The population in YRB increased from less than 0.1 million in the early 1980s to more than 0.2 million in 2007 (Shao et al. 2012). The condition was similar in YARB; the population increased from less than 0.2 million in the early 1980s to more than 0.3 million in 2007 (Shao et al. 2012). With the increase in population, the amount of production and living water consumption also increased significantly (Zhang et al. 2012). In the TRHR, grassland degradation was aggravated in the past two decades, which was partly induced by overgrazing (Liu et al. 2008). In 2008, the largest livestock overload capacity in TRHR occurred in Xinghai County at 142.2×10^4 sheep units (Shao et al. 2012). Overgrazing can induce grassland degradation by influencing soil water properties and further influence water yield (Shao et al. 2009).

Meanwhile, a series of ecological projects were implemented since 1990s to protect eco-environment. For instance, the reducing livestock project started from 2005, and the first phase last for 5 years (2005–2009). Since the implementation of project, livestock decreased from 1548.7×10^4 sheet unit in 2005 to 1482.3×10^4 sheet unit in 2009. The ecological projects effectively reduced pressure of livestock and contained the disorderly increase of grazing area (Liu et al. 2008; Shao et al. 2012). This has promoted vegetation restoration and ecological improvements, which will improve water conservation function of ecosystem.

5.4 Uncertainty in the hydrological sensitivity analysis

Some uncertainties may exist in the hydrologic sensitivity analysis method when assessing the effects of climate variation and anthropogenic activity on runoff. First, the method is based on long-term hydro-meteorological observation data. The limited number and distribution of hydro-meteorological stations may affect the simulation accuracy of the hydro-climatic variables such as *PET*, runoff, and *AET*. In addition, the method assumes that anthropogenic activity and climate variation are independent of each other (Zheng et al. 2009). However, climate variation may influence anthropogenic activity, for example, land use, while anthropogenic activity, for example, extensive urbanization and an expanded population may cause changes in climate. Thus, these two factors are inter-related, and due to the limited length of data combined with the insufficiency in change-point analysis, there were in fact some human disturbances during the baseline period. Uncertainty may also arise from the parameter ω. However, despite the uncertainties and limitations, our study sheds light on understanding the sensitivity of runoff to climate variation and anthropogenic activity in the TRHR. Additional work will be carried out on the estimation of these uncertainties in our future studies in order to improve the results of our quantification.

6 Conclusions

Climate variability and anthropogenic activity have significantly affected runoff in the TRHR. This study applied the hydrologic sensitivity analysis method to quantify the effects of climate variability and anthropogenic activity on runoff. On the basis of our results, the following conclusions can be drawn:

1. Annual runoff in the LRB and YARB showed an increasing trend, while runoff in the YRB fell slightly in the 57-year period. A change point reflecting the effect of climate variability and anthropogenic activity on runoff occurred in 1989 and 1993. Mean annual runoff in change period increased by 12.3 and 1.7 % in the YARB and LRB, respectively, as compared with the baseline period. Meanwhile, the annual runoff in change period decreased by 14.5 % in the YRB, as compared with the baseline period.
2. The hydrologic sensitivity analysis indicated that climate variability was the dominant factor, accounting for more than 90 % of the increase in runoff of YARB and LRB; the increase due to human activities was less than 10 %. We suggest that the change in runoff during 1990–2012 (1994–2012) was mainly due to climate variability; the contribution of anthropogenic activity was relatively small.
3. The intra-annual distribution of runoff shifted gradually from a double peak pattern to a single peak pattern, which was mainly influenced by atmospheric circulation in the summer and autumn. The inter-annual variation in the YRB and the LRB correlated positively with the East Asian monsoon index, but negatively with the westerly index. The runoff in the YARB was mainly controlled by the East Asian monsoon and the Tibetan Plateau monsoon.
4. Quantifying the effects of climate variability and human activities on runoff will contribute to regional water resources assessment and management. The TRHR is a producing flow area providing water resources for the socio-economic development of the western China and central and eastern China (downstream of three rivers). Climate variability has distinctly increased runoff since 2000, which will play a positive effect on the socio-economic development and ecological conservation of the TRHR and, meanwhile, suggests that the local government should take reasonable measures to improve the water use efficiency and reduce water resources waste.

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