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Responses of natural runoff to recent climatic changes in the Yellow River basin, China

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Abstract

The Yellow River, the second longest river in China, experienced frequent zero flow in the lower reaches of the mainstream in the 1990s. In recent years, the zero-flow phenomenon has almost disappeared. Besides engineering measures implemented to maintain ecological flows, the changes in natural runoff might have contributed to replenish the river. In this study, we used the Soil and Water Assessment Tool (SWAT) model and runoff elasticity analyses to assess the impacts of climatic changes on the natural streamflow at the Huayuankou station. The results show that there was little increase of precipitation but substantial recovery of natural runoff in the recent period (2003–2011) compared with the low flow period (1991–2002). The recent precipitation was slightly greater (~ 2 % of the mean annual precipitation in the baseline period of 1960–1990) than precipitation in the low flow period. However, the natural runoff in the recent period was much larger (~ 14 % baseline runoff) than runoff in the low flow period. The decreasing runoff in the low flow period was mainly caused by the decline in precipitation while the runoff recovery in the recent period was largely affected by the contributions from the climatic variables other than the precipitation. In the recent period, precipitation could account for a reduction of 21 % baseline runoff whereas the others – net radiation, wind speed, air temperature, and relative humidity – accounted for an increase of 7.5 % baseline runoff. The runoff reduction (~ 10.4 % baseline runoff) caused by the changes in temperature and relative humidity was offset by the contribution from the decreasing net radiation and wind speed which resulted in an increase of ~ 17.9 % baseline runoff.

1 Introduction

The Yellow River, the cradle of Chinese civilization, is a major source of freshwater for about 107 million people within the river basin, about 9 % of the total population in China (Wang et al., 2006). The upper and middle reaches of the river locate in semi-arid and arid regions, and the mean annual precipitation of this area is about 440 mm (Tang et al., 2008a). The Yellow River basin is one of the regions facing serious water shortages due to the dry climate and heavy water demands (Yang et al., 2004). After the completion of a few irrigation projects in the 1960s, the lower reaches have increasingly suffered from extreme low-flow conditions (Tang et al., 2008b). The Yellow River zero-flow phenomenon, i.e. zero flow in the lower reaches of the mainstream, has occurred since 1972 (Yang et al., 2004). The frequency of the zero-flow phenomenon increased rapidly in the 1990s. However, it seemed to disappear in the 2000s (Zhang et al., 2009). Numerous studies have investigated the hydrological cycle change in the Yellow River basin and tried to explain the causes of the zero-flow phenomenon (Yang et al., 2004; Liu and Zheng, 2004; Fu et al., 2004; Xu, 2005; Tang et al., 2006, 2007). The frequent zero-flow phenomenon in the 1990s was attributed to intensified human activities and climatic changes. As for the Huayuankou station, a hydrological gauge that controls most (approximating 97 % of the total) catchment area of the Yellow River basin, natural runoff has a significant decreasing trend during the period of 1952–1997 (Liu and Zheng, 2004). Climatic change is a dominant cause of the reduction in river flow above the Huayuankou station (Cong et al., 2009), accounting for about three quarters of annual streamflow changes (Tang et al., 2008b). In contrast, possible reasons for the disappearing zero-flow phenomenon in recent years (Zhang et al., 2009) have been less studied. With recognition of trade-off between human water use and eco-environmental water use and allocations of more water to maintain the ecological environment, engineering measures such as reservoir regulation might have contributed to prevent the zero-flow phenomenon (Yang et al., 2008; Hu et al., 2008; Cui et al., 2009). Recent climatic change in the river basin, which has a large impact on river flow, may also have contributed to the river replenishment.

Previous studies showed that the hydroclimatic changes in the Yellow River basin varied spatially. According to the river discharge records and the China Meteorological Administration (CMA) weather observations, precipitation in the source region of the Yellow River was low in the 1990s but returned to above normal after 2002 while

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discharge remained low (Zhou et al., 2012). In Hailiutu river basin, a small catchment (~2600 km^2) in the middle reaches of the Yellow River, the river discharge reached lowest in the 1990s and recovered in the 2000s (Yang et al., 2012). At the river mouth, annual streamflow decreased severely from 1997 to 2002 but increased thereafter as the direct beneficiary of the environment-friendly water resource allocation projects (Cui et al., 2009; Yu et al., 2011). Although reservoir regulations may help to increase the low flow in the river channel, there must be enough water in the river systems to enable the allocation for eco-environmental water use. Understanding the changes in natural runoff is essential to explain the observed streamflow change at the lower reaches in the recent decade and is informative for future water resources management in the Yellow River basin. Since the catchment area between Huayuankou station and river mouth is small (about 3 % of the total catchment area) and the river flow between Huayuankou station and river mouth is largely withdrawn for irrigation (Tang et al., 2008b), the natural runoff above the Huayuankou station is of special interest.

A hydrological model, the Soil and Water Assessment Tool (SWAT), was used to reproduce the natural runoff in the catchment above the Huayuankou station. SWAT is a hydrologic model developed to evaluate water resources in large agricultural basins (Arnold et al., 1998; Arnold and Fohrer, 2005). It has been used to assess water resource and nonpoint pollution problems at a wide range of scales across the globe. The SWAT model has been used in many hydrological applications and climatic change studies in the Yellow River basin and its sub-basins (Li et al., 2009; Xu, et al., 2009, 2011; Liu et al., 2011). The climate elasticity of runoff derived by Yang and Yang (2011) was used to further attribute the changes in natural runoff to changes in different climatic variables. Climate elasticity of runoff was defined by the proportional change in runoff to the proportional change in a climatic variable such as precipitation (Schaake, 1990). Climate elasticity of runoff provides a measure of the sensitivity of runoff to the changes in the climatic variables and is widely used in impact assessment of climatic changes on hydrology (Sankarasubramanian et al., 2001; Chiew, 2006; Fu et al., 2007; Zheng et al., 2009; Tang and Lettenmaier, 2012). The streamflow sensitivities

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to the changes in the climatic variables have been analytically explored in some previous studies (Liu and Cui, 2011; Liu and McVicar, 2012). This paper compared the analytical estimates with the runoff simulations from SWAT model, and investigated the possible climatic factors contributed to the recent natural streamflow change at the Huayuankou station. The paper concluded with the recent natural streamflow change and the contributions from the changes in different climatic variables.

2 Study area and data

The Yellow River, the second longest river in China, originates in the Tibetan Plateau, flows through the Loess Plateau and North China Plain, and discharges into the Bohai Gulf (Fig. 1). The study area is the catchment above the Huayuankou station with a mainstream length of 4696 km and an area of 730 000 km^2 (~97 % of the total area of the Yellow River basin). The mean annual natural runoff in the study area accounts for ~98 % of that in the whole Yellow River basin (Liu et al., 2011). The study area is largely in the semi-arid and arid regions where the mean annual precipitation ranges from 300 to 700 mm.

Meteorological data from 50 weather stations inside and close to the study area were obtained from CMA. The dataset includes daily precipitation (P), mean air temperature (T), maximum temperature (T_{max}), minimum temperature (T_{min}), surface relative humidity (RH), wind speed at 10 m height (U_{10}), and sunshine duration (n) from 1955 to 2011. The monthly naturalized streamflow data from 1960 to 2000 were obtained from the Yellow River Conservancy Committee (YRCC) while the recent data (from 2001 to 2011) were unavailable. The naturalized streamflow data are measured streamflow data that have been adjusted to remove anthropogenic effects of both water management and use. The naturalized flow was directly comparable with the model simulated natural streamflow. The digital elevation model (DEM) with a spatial resolution of 1 km × 1 km was generated from the International Center for Tropical Agriculture (CIAT) product (Reuter et al., 2007) archived at the Computer Network Information

Center, Chinese Academy of Sciences (http://datamirror.csdb.cn). The land cover/use map of the 1980s was taken from the Institute of Geographical Sciences and Natural Resources Research (IGSNRR), Chinese Academy of Sciences (CAS) (Liu et al., 2003). The effects of land use change on runoff generation are not part of direct hydrological response to climatic changes. Furthermore, the previous studies suggested that comparing with climatic changes, land cover change might be a less significant factor to natural runoff change in the Yellow River basin (Cong et al., 2009). The fixed land cover map was used throughout the study period. The soil parameters were estimated by the Soil Water Characteristics application of Soil-Plant-Air-Water (SPAW) model (Saxton and Rawls, 2006), based on the soil texture and organic matter data provided in the China Soil Scientific Database (http://www.soil.csdb.cn).

3 Method

The potential evaporation (E_0) was estimated using Penman equation (Penman, 1948):

$$E_0 = \frac{\Delta}{\Delta+\gamma}(R_n - G)/\lambda + \frac{\gamma}{\Delta+\gamma}6.43(1+0.536U_2)(1-\mathrm{RH})e_s/\lambda, \quad (1)$$

where Δ is the slope of the saturated vapor pressure versus air temperature curve ($\mathrm{kPa\,^\circ C^{-1}}$), γ is psychrometric constant ($\mathrm{kPa\,^\circ C^{-1}}$), λ is the latent heat of vaporization ($2.45\,\mathrm{MJ\,kg^{-1}}$), R_n and G are the net radiation and soil heat flux ($\mathrm{MJ\,m^{-2}\,d^{-1}}$) respectively, e_s is the saturated vapor pressure (kPa), RH is the relative humidity (%), and U_2 is the wind speed at a height of 2 m ($\mathrm{m\,s^{-1}}$). The observed wind speed at 10 m height was adjusted to the standard height of 2 m (U_2, $\mathrm{m\,s^{-1}}$) (Allen et al., 1998):

$$U_2 = U_z\frac{4.87}{\ln(67.8z-5.42)} = 0.75U_{10}, \quad (2)$$

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where U_z measured wind speed at z meters above ground surface ($\mathrm{m\,s^{-1}}$), z is the height of measurement above ground surface (m). The daily net radiation R_n ($\mathrm{MJ\,m^{-2}\,day^{-1}}$) was estimated as:

$$R_\mathrm{n} = (1-\alpha)R_\mathrm{s} - \sigma\left[\frac{(T_\mathrm{max}+273.2)^4+(T_\mathrm{min}+273.2)^4}{2}\right]\left(0.1+0.9\frac{n}{N}\right) \times \left(0.34-0.14\sqrt{\frac{\mathrm{RH}}{100}e_\mathrm{s}}\right) \tag{3}$$

where α is albedo or canopy reflection coefficient (dimensionless), R_s is solar or shortwave radiation ($\mathrm{MJ\,m^{-2}\,d^{-1}}$), σ is Stefan-Boltzmann constant ($4.903\times10^{-9}\,\mathrm{MJ\,K^{-4}\,m^{-2}\,d^{-1}}$), T_max is daily maximum air temperature (°C), T_min is daily minimum air temperature (°C), n is daily actual sunshine duration (h), N is daily maximum possible duration of sunshine (h), RH is daily relative humidity (%). Albedo (α) was here set as 0.23 for the hypothetical grass reference crop considering that grass was the main land use type in the Yellow River Basin (Wang et al., 2004). R_s is calculated using the Angström formula relating solar radiation to extraterrestrial radiation and relative sunshine duration (Angström, 1924). e_s is estimated as:

$$e_\mathrm{s} = 0.3054\left[\exp\left(\frac{17.27T_\mathrm{max}}{T_\mathrm{max}+237.3}\right)+\exp\left(\frac{17.27T_\mathrm{min}}{T_\mathrm{min}+237.3}\right)\right]. \tag{4}$$

The linear trends of the mean annual basin-averaged climatic variables were calculated. The statistical significance of the annual trend was tested by the two-tailed Student's t test. The mean annual value during the historical period of 1960–1990 was used as the baseline. The period of 1991–2002 was the low flow period (Cui et al., 2009; Yu et al., 2011; Zhou et al., 2012). The relative changes of the climatic and hydrological variables to the baseline were computed for the low flow (1991–2002) and recent (2003–2011) periods, respectively.

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The SWAT model was set up to reproduce the natural streamflow at the Huayuankou station. The catchment above that station was divided into 76 sub-basins, ranging from 32 to 40 194 km^2 (Fig. 1). The SWAT model ran at daily time step from 1955 to 2000. The first five years (1955–1959) served as a warm-up period to general initial conditions for the model experiments. The simulated natural runoff was manually calibrated against the monthly naturalized streamflow in the period of 1960–1979 and validated in the period of 1980–2000. The Relative Error (E_r), Nash–Sutcliffe efficiency (E_{NS}), and Coefficient of Determination (R^2) were used to evaluate the model performance:

$$E_r = \frac{S_i - O_i}{O_i} \times 100\,\%, \tag{5}$$

$$E_{NS} = 1 - \frac{\sum_{i=1}^{N} (O_i - S_i)^2}{\sum_{i=1}^{N} (O_i - \overline{O})^2}, \tag{6}$$

$$R^2 = \frac{\left(\sum_{i=1}^{N} (O_i - \overline{O})(S_i - \overline{S})\right)^2}{\sum_{i=1}^{N} (O_i - \overline{O})^2 \sum_{i=1}^{N} (S_i - \overline{S})^2} \tag{7}$$

where, O_i is the observed naturalized streamflow, S_i is simulated natural runoff; $\overline{O}$ is the mean observed value, $\overline{S}$ is the mean simulated value, and N is the total number of paired values, i.e. the number of years in the evaluated period. E_r gives the percent difference between the simulated and observed natural runoff over the evaluated period, thus is of special interest in this study. A E_{NS} value of 1 is a perfect match of observed and simulated data. Generally model performance is very good if $E_{NS} > 0.75$, satisfactory if $0.36 < E_{NS} < 0.75$, and unsatisfactory if $E_{NS} < 0.36$ (Nash and Sutcliffe, 1970; Krause et al., 2005; Moriasi et al., 2007). R^2 is the square of the correlation coefficient

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between the observed and simulated data values. The validated model continued to simulate the natural streamflow into 2001–2011 when the observed streamflow data were unavailable.

The climate elasticity of runoff (ε) was used to attribute the changes in natural runoff to changes in different climatic variables for the low flow and recent periods. The runoff elasticities to precipitation (P), net radiation (R_n), mean air temperature (T), wind speed (U_2), and relative humidity (RH) were derived using the mean annual climatic variables in the baseline period following the derivation described in Yang and Yang (2011). Runoff (R) change was expressed as:

$$\frac{\mathrm{d}R}{\overline{R}} = \varepsilon_P \frac{\mathrm{d}P}{\overline{P}} + \varepsilon_{R_n} \frac{\mathrm{d}R_n}{\overline{R_n}} + \varepsilon_T \mathrm{d}T + \varepsilon_{U_2} \frac{\mathrm{d}U_2}{\overline{U_2}} + \varepsilon_{\mathrm{RH}} \frac{\mathrm{dRH}}{\overline{\mathrm{RH}}} \tag{8}$$

where $\overline{R}$, $\overline{P}$, $\overline{U_2}$, $\overline{R_n}$, $\overline{\mathrm{RH}}$ are the mean annual values in the baseline period, ε_P, ε_{R_n}, ε_T, ε_{U_2} and $\varepsilon_{\mathrm{RH}}$ are the runoff elasticities. The runoff elasticity to temperature (ε_T) implies that 1 °C increase in T could lead to ε_T% change in runoff, and the elasticity to the other climatic variables (i.e. P, R_n, U_2, and RH) implies that 1 % changes in the climatic variables could induce ε% change in runoff. Once the runoff elasticities were estimated, relative runoff changes in the low flow and recent periods to the baseline period could be derived from the changes of climatic variables according to Eq. (8). The derived runoff changes were compared with the SWAT model estimates and the naturalized streamflow.

4 Results

Figure 2 shows the changes in the climatic variables and potential evaporation during the period of 1960–2011 in the study area. There are a significant increase trend ($p < 0.001$) in T and significant decrease trends in U_2, R_n and RH. The warming trend of the Yellow River basin has been reported in the previous studies (Fu et al., 2004; Tang

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et al., 2008a) and is consistent with the generally increase in surface air temperature over global land surface (Hansen et al., 2006). The decline in wind speed has been documented over China (Jiang et al., 2010; Fu et al., 2011; Lin et al., 2012) and seems to be a part of widespread terrestrial stilling across the globe (McVicar et al., 2012). The decreasing trend of R_n is consistent with the reported declines in solar radiation across China (Tang et al., 2011) and R_n decrease at the adjacent Yangtze River basin (Xu, et al., 2006). The decreasing RH is line with previous studies which reported large decease in relative humidity in many parts of China (Wang et al., 2012; Liu et al., 2010; Song et al., 2012). P showed a decrease trend although the trend was not statistically significant during the period of 1960–2011 (Fu et al., 2004). A decreasing trend of E_0 is consistent with that in many previous studies (Ma et al., 2012; Liu and McVicar, 2012).

Mean annual precipitation in the low flow period (1991–2002) was 47 mm (10.5 %) less than that in the baseline period (1960–1990) (Table 1). The precipitation in the recent period (2003–2011) remained the low level of the low flow period although there was a little rebound (about 2 %). The precipitation in the recent period was 36 mm (8.1 %) below the baseline period. The temperatures in both the low flow and the recent periods were higher than that in the baseline period while there was little temperature difference between the low flow and recent periods. The relative humidity in the recent period dropped about 9 % from the baseline (Table 1). The increase in temperature and decrease in relative humidity favor an increase in potential evaporation. In the recent period, net radiation was about 10 % below that in the baseline period and 2 m wind speed reduced about 18 % from the baseline wind speed. The decreases in net radiation and wind speed would reduce the potential evaporation (Liu and McVicar, 2012). Furthermore, the net radiation and wind speed in the recent period were the lowest among the three periods and thus have greater impacts on the potential evaporation than in the low flow period. Overall, the potential evaporation in the low flow period is about the same as that in the baseline period while the potential evaporation in the recent period is 5.3 % lower than that in the baseline period (Table 1). The potential evaporation reflects the energy condition which affects the partition of precipitation into

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runoff and actual evaporation (Budyko, 1974; Roderick and Farquhar, 2011; Liu and McVicar, 2012). The reduction of potential evaporation might have affected runoff in the recent period.

Figure 3 shows the monthly comparisons between the SWAT simulated streamflow and observed naturalized streamflow in the calibration (1960–1979) and validation (1980–2000) periods at the Huayuankou station. Table 2 gives the evaluation scores of the SWAT performance in the calibration and validation periods. The SWAT simulated streamflow agrees favorably with the observed naturalized streamflow. The E_{NS} is greater than 0.5 in both the calibration and validation periods, suggesting a satisfactory model performance (Krause et al., 2005; Moriasi et al., 2007). The relative error is small (less than 4 %) in either the calibration or validation period. These indicate that the SWAT simulations can capture the temporal variations of streamflow reasonably well in the study area.

Figure 4 shows the SWAT simulated annual natural streamflow at the Huayuankou station from 1960 to 2011. The mean annual natural streamflow is the largest in the baseline period and smallest in the low flow period. In the recent period when precipitation slightly rebounded (Table 1), the natural streamflow has substantially recovered (Fig. 4). The recent recovery of the natural streamflow enabled a greater amount of the available water resources for reservoir regulations and might contribute to the disappearance of the Yellow River zero-flow phenomenon after 2002.

Table 3 shows the mean annual streamflows estimated from SWAT model and derived from the runoff elasticities in the baseline, low flow, and recent periods. The mean annual streamflow derived from the runoff elasticities method was identical as the observed naturalized flow in the baseline period because the runoff elasticities were calculated using the data in that period. The mean annual streamflows derived from the runoff elasticities and estimated from SWAT model match well with the observed naturalized streamflow in the low flow period, with about 4 % relative errors (Table 3). The observed naturalized streamflow in the low flow period was 27 % below the streamflow in the baseline period. Both the SWAT model simulation and runoff elasticities

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estimates captured the low flow, showing 26 and 24 % reduction of natural streamflow respectively. In the recent period, the SWAT model showed a 12 % reduction and the runoff elasticities method estimated a 14 % reduction from the natural streamflow in the baseline period. The natural streamflow in the recent period was lower than that in the baseline period but higher than that in the low flow period. Both the SWAT model simulations and runoff elasticity estimates show that about half reduction amount in the low flow period has recovered in the recent period. The natural streamflow recovery amount is about 13 % of the mean annual steamflow in the baseline period (Table 3), which may have helped to replenish the drying river in the recent period.

Table 4 shows the runoff elasticities and the changes in climatic variables in the low flow (1991–2002) and recent (2003–2011) periods to the baseline period (1960–1990), and the contributions of climatic variable changes to runoff change. The runoff elasticity to precipitation (ε_P) is 2.6, indicating that a 10 % change in mean annual precipitation results in a 26 % change in mean annual runoff (Table 4). The precipitation in the low flow period was 10.5 % below that in the baseline period and led to a 27.3 % reduction in runoff. The runoff reduction caused by precipitation change was close to the decrease in the observed naturalized streamflow (27 % baseline runoff) in the low flow period (Table 3). In the low flow period, the total contribution from the climatic variables other than precipitation to runoff change was relative small (3.2 % baseline runoff). In the recent period, the precipitation was 8.1 % below the baseline which would lead to a 21 % reduction in runoff. However, the reduction in natural streamflow was about 13 % baseline runoff (Table 3). This suggests that climatic variables other than precipitation may have affected runoff in the recent period. The increase in temperature (0.71 °C) had relative small effects on runoff change, responsible for a reduction of only 3.3 % baseline runoff. The decrease in relative humidity (−9.1 %) was responsible for a decrease of 7.1 % baseline runoff. Contrastively, the decrease in R_n (−9.5 %) and wind speed (−18.1 %) contributed an increase of 7.2 and 10.7 % baseline runoff, respectively. The contributions of R_n and wind speed offset the runoff reduction caused by temperature increase and relative humidity decrease in the recent period, resulting

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in a total contribution of 7.5 % of baseline runoff increase from the climatic variables other than precipitation. The contribution from the climatic variables other than precipitation formed the major part of the natural streamflow recovery amount which was about 13 % of the mean annual steamflow in the baseline period (Table 3). The large positive contribution from the climatic variables other than precipitation is consistent with the decrease in potential evaporation in the recent period (Table 1).

5 Conclusions

The Yellow River experienced frequent zero-flow phenomenon in the 1990s. The river drying was largely attributed to the decrease in natural runoff in the upper and middle reaches (above Huayuankou station) and the increase in water withdrawals in the lower reaches (from Huayuankou station to the river mouth). In the recent years, the zero-flow phenomenon has almost disappeared. We used a hydrological model together with runoff elasticity analyses to investigate the recent change in natural streamflow at the Huayuankou station and the possible contributions of climatic factors to the natural streamflow change.

Our results show that there was little rebound of precipitation but substantial recovery of natural runoff in the recent period (2003–2011) compared with the low flow period (1991–2002). The precipitation in the recent period was slightly greater than precipitation in the low flow period by 2 % of the mean annual precipitation in the baseline period (1960–1990). However, the natural runoff in the recent period estimated by the model and runoff elasticity analyses was much larger than runoff in the low flow period ($\sim$ 14 % of the mean annual runoff in the baseline period). Although the natural runoff in the recent period was still 12 % less than the baseline runoff, the substantial runoff recovery may have contributed to replenish the drying river.

The runoff elasticity analyses show that the decrease in runoff in the low flow period was mainly caused by the decrease in precipitation whereas decreasing R_n and wind speed were largely responsible for recent runoff recovery. In the low flow period,

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precipitation was responsible for a runoff reduction of 27.3 % baseline runoff while the climatic variables other than precipitation accounted for a small runoff increase (3.2 % baseline runoff). In the recent period, precipitation accounted for a runoff reduction of 21 % baseline runoff and the climatic variables other than precipitation accounted for a runoff increase of 7.5 % baseline runoff. The changes in temperature and relative humidity have caused a reduction in runoff of 3.3 and 7.1 % baseline runoff, respectively. The runoff reductions were largely offset by the contribution from the decreasing net radiation and wind speed which resulted in an increase in runoff of 7.2 and 10.7 % baseline runoff, respectively.

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Table 1. Mean annual climatic variables and potential evaporation of the study area in the baseline (1960–1990), low flow (1991–2002), and recent (2003–2011) periods.

	P (mm yr^{-1})	R_n (MJ m^{-2} yr^{-1})	T (°C)	U_2 (m s^{-1})	RH (%)	E_0 (mm yr^{-1})
Baseline	449	2599	7.48	1.735	58.67	1010
Low flow	402	2543	8.24	1.539	57.37	1002
Recent	413	2353	8.2	1.421	53.31	956
Low flow change relative to baseline	−10.5 %	−2.2 %	0.76 °C	−11.3 %	−2.2 %	−0.8 %
Recent change relative to baseline	−8.1 %	−9.5 %	0.71 °C	−18.1 %	−9.1 %	−5.3 %

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Table 2. Performance of SWAT model in the calibration (1960–1979) and validation (1980–2000) periods.

	E_{NS}	E_r	R^2
Calibration	0.53	−1.0 %	0.69
Validation	0.68	−3.5 %	0.74

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Table 3. Mean annual streamflows estimated from SWAT model and derived from the runoff elasticities in the baseline (1960–1990), low flow (1991–2002), and recent (2003–2011) periods.

	Observed ($m^3 s^{-1}$)	SWAT Simulated ($m^3 s^{-1}$)	Derivedfrom elasticities ($m^3 s^{-1}$)
Baseline	1899	1934	1899
Low flow	1382*	1441	1443
Recent		1710	1633
Low flow change relative to baseline	−27 %*	−26 %	−24 %
Recent change relative to baseline		−12 %	−14 %

* The mean annual observed streamflow in the period of 1991–2000 was used because of the naturalized streamflow data were unavailable after 2000.

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Table 4. Estimated runoff elasticities, climatic variable changes in the low flow (1991–2002) and recent (2003–2011) periods to the baseline period (1960–1990), and contributions of climatic variable changes to runoff (R) change.

Period		P	R_n	T	U_2	RH
Baseline	Runoff elasticities (ε)	2.6	−0.76	−0.046	−0.59	0.78
Low flow	Relative change to baseline	−10.5 %	−2.2 %	0.76 °C	−11.3 %	−2.2 %
	Contribution to R change (%)	−27.3	1.7	−3.5	6.7	−1.7
Recent	Relative change to baseline	−8.1 %	−9.5 %	0.71 °C	−18.1 %	−9.1 %
	Contribution to R change (%)	−21	7.2	−3.3	10.7	−7.1

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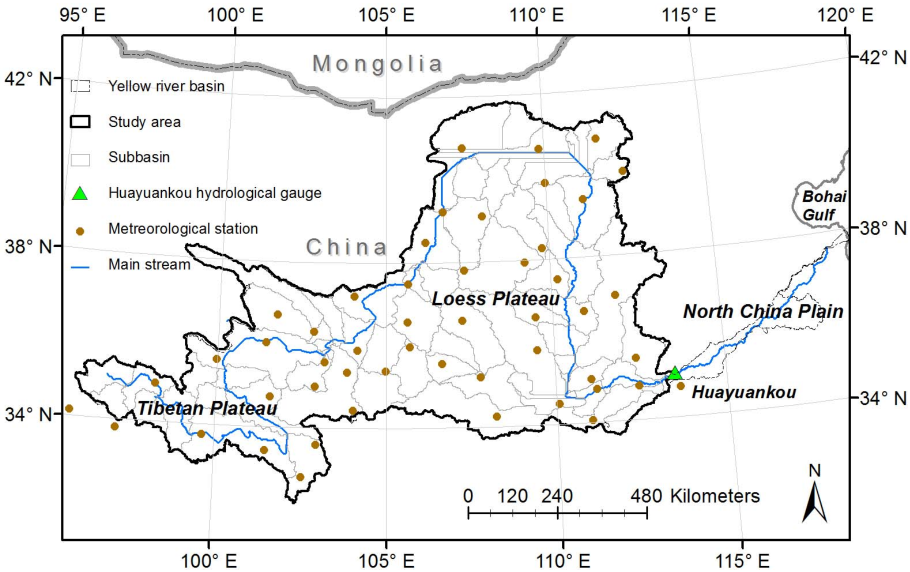

Fig. 1. The study area and sub-basins.

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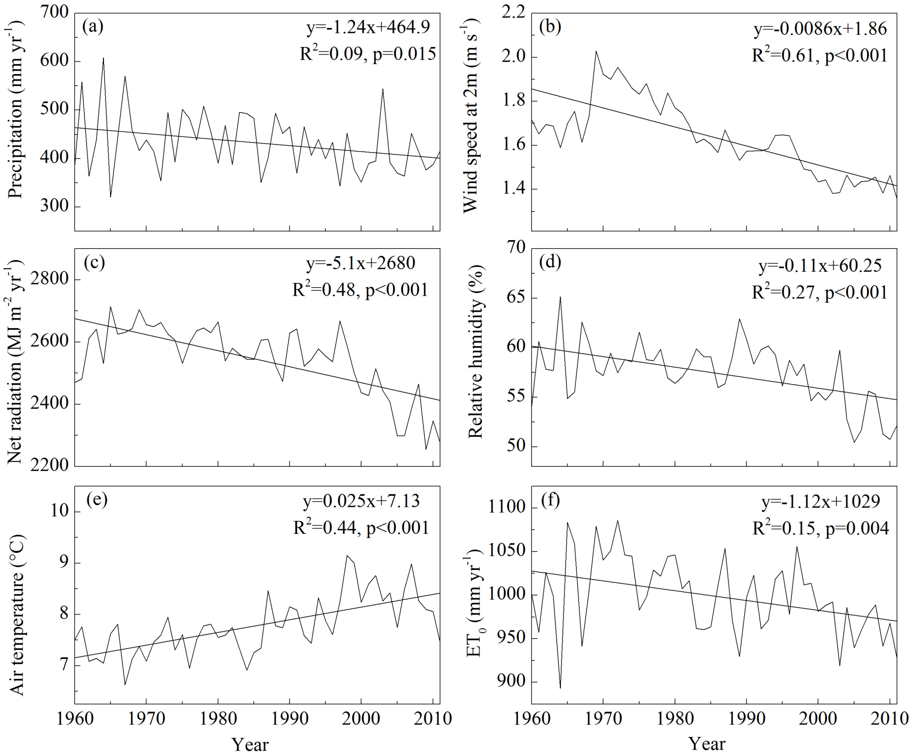

Fig. 2. Changes in precipitation **(a)**, wind speed at 2 m above the ground **(b)**, net radiation **(c)**, relative humidity **(d)**, air temperature **(e)**, and potential evapotranspiration **(f)** from 1960 to 2011 in the study area.

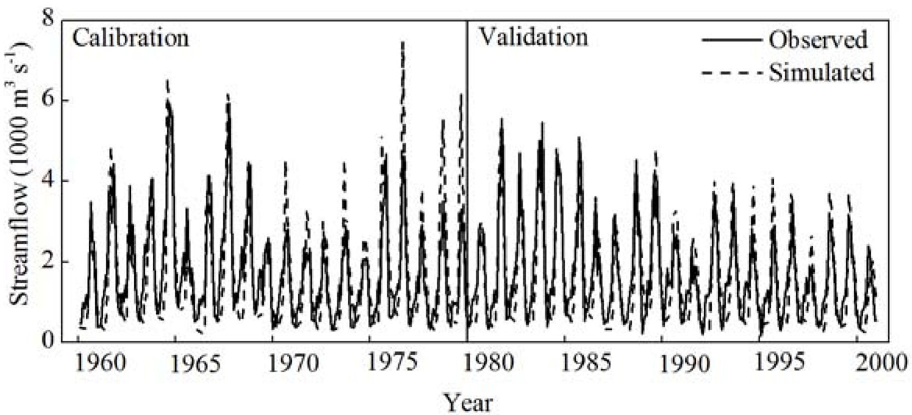

Fig. 3. Monthly comparisons between the SWAT simulated streamflow and observed naturalized streamflow in the calibration (1960–1979) and validation (1980–2000) periods at the Huayuankou station. The vertical line divides the calibration and validation periods.

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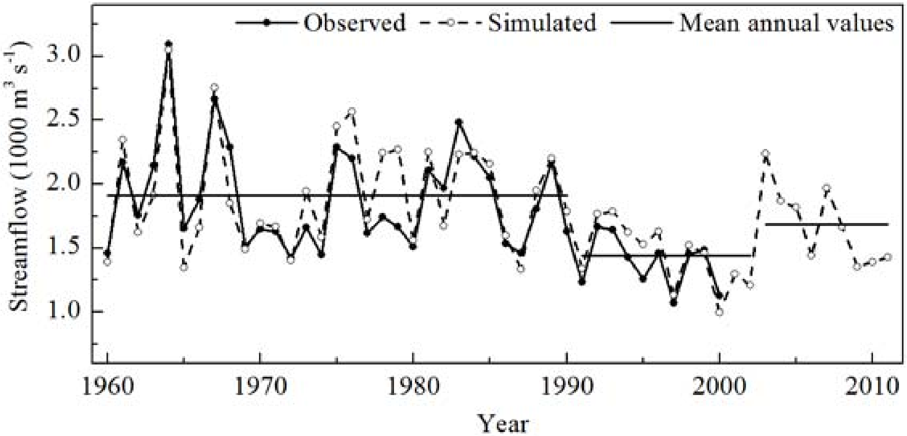

Fig. 4. SWAT simulated annual natural streamflow at the Huayuankou station from 1960 to 2011. The horizontal lines show the SWAT simulated mean annual streamflow in the baseline (1960–1990), low flow (1991–2002), and recent (2003–2011) periods.

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