Postprint: Evapotranspiration Estimation and Drought Characteristics of Shallow Mountain Watersheds in Hexi Inland Rivers

Authors: Cheng Wenju, Xi Haiyang, Jianhua Si, Li Ailin, Xi Haiyang

Date: 2020-12-17T00:00:00+00:00

Abstract

To reveal the evapotranspiration and drought characteristics of small watersheds in the shallow mountainous areas of inland rivers in Hexi, taking the Xitugou watershed in Dunhuang as an example, conventional meteorological data were used to calculate the potential evapotranspiration at 8 observation stations in the Xitugou watershed through the Penman-Monteith formula. Based on this, the crop coefficients at each observation station were calculated using the dual crop coefficient method, from which the actual evapotranspiration was calculated, and the SPEI (Standardized Precipitation Evapotranspiration Index) of the watershed was calculated. The results show that: (1) The annual potential evapotranspiration and actual evapotranspiration in the Xitugou watershed are 978 mm and 258 mm, respectively; from upstream to downstream, as altitude decreases, both potential evapotranspiration and actual evapotranspiration show varying degrees of increasing trends; (2) The downstream area of the watershed experiences longer drought duration, with droughts of varying degrees occurring throughout the entire growing season, and autumn is the main season for drought occurrence; (3) Precipitation in the upstream area of the watershed is greater than actual evapotranspiration, while in the middle and downstream areas, actual evapotranspiration exceeds precipitation; (4) The crop coefficient Kc values calculated using NDVI values show good applicability in the estimation of actual evapotranspiration, and in drought assessment of small watersheds in the shallow mountainous areas of inland rivers, the drought index SPEI has greater advantages than the SPI (Standardized Precipitation Index) and PDSI (Palmer Drought Severity Index) indices. This study calculated the potential and actual evapotranspiration of the Xitugou watershed, and evaluated the drought characteristics at monthly and seasonal scales in the watershed, providing guidance for production and domestic water use in the watershed, especially in the middle and downstream areas, and obtained universal patterns applicable to ungauged rivers throughout the shallow mountainous areas of inland rivers in Hexi.

Full Text

Estimation of Evapotranspiration and Drought Characteristics in Low Mountain Watersheds of Hexi Inland Rivers

CHENG Wenju[1],[2], **XI Haiyang**[1], **SI Jianhua**[1], **LI Ailin**[3]

[1]Northwest Institute of Eco-Environment and Resources, Chinese Academy of Sciences; Key Laboratory of Ecohydrology of Inland River Basin, Chinese Academy of Sciences; Alxa Desert Eco-Hydrological Experimental Research Station, Lanzhou 730000, Gansu, China

[2]University of Chinese Academy of Sciences, Beijing 100049, China

[3]Grassland Workstation of Alxa Right Banner, Alxa 750306, Inner Mongolia, China

Abstract

To reveal the evapotranspiration and drought characteristics of small watersheds in the low mountain areas of Hexi inland rivers, this study takes the Xitugou watershed in Dunhuang as a case study. Using conventional meteorological data, the Penman-Monteith formula was applied to calculate potential evapotranspiration at eight observation sites across the Xitugou watershed. Building upon this, the dual crop coefficient method was employed to compute crop coefficients at each site, which were then used to calculate actual evapotranspiration. The Standardized Precipitation Evapotranspiration Index (SPEI) was subsequently calculated for the watershed. The results indicate that: (1) The annual potential evapotranspiration and actual evapotranspiration in the Xitugou watershed are 978 mm and 258 mm, respectively. From upstream to downstream, with decreasing elevation, both potential and actual evapotranspiration exhibit varying degrees of increase. (2) The lower reaches of the watershed experience prolonged drought duration, with different degrees of drought occurring throughout the entire growing season. Autumn is the primary season for drought occurrence. (3) Precipitation in the upper reaches exceeds actual evapotranspiration, whereas in the middle and lower reaches, actual evapotranspiration exceeds precipitation. (4) The crop coefficient Kc calculated using NDVI demonstrates good applicability for actual evapotranspiration estimation. For drought assessment in inland river basins of arid northwestern China, the SPEI index offers greater advantages than the Standardized Precipitation Index (SPI) and Palmer Drought Severity Index (PDSI). This study quantifies potential and actual evapotranspiration in the Xitugou watershed and evaluates drought characteristics at monthly and seasonal scales, providing guidance for water resource management for production and domestic use, particularly in the middle and lower reaches. The findings yield generalizable principles applicable to ungauged rivers throughout the low mountain regions of Hexi inland rivers.

Keywords: Hexi inland rivers; evapotranspiration; drought index; Xitugou watershed

1. Introduction

The low mountain areas of Hexi inland rivers cover extensive regions, accounting for approximately 11.2% of the total mountainous area. These areas contribute 0.2197$×10^{3}$ of surface water resources and hold significant importance for the entire Hexi region. While some larger rivers in this area have measured meteorological and hydrological data, most smaller rivers lack such observations. Although these rivers have limited discharge, they possess independent water supply significance, nurturing different oases downstream and exerting important influences on local natural environments and human production and living conditions. Therefore, this study selected a typical ungauged watershed in the low mountain area—the Xitugou watershed in Dunhuang—to investigate precipitation, evapotranspiration, and drought characteristics, thereby revealing the watershed' s water balance and deriving universal patterns applicable to ungauged rivers throughout the low mountain regions of Hexi inland rivers.

Evapotranspiration is a critical parameter in surface water and heat balance, representing an essential component of both water and energy cycles that influences regional and global ecology, climate, and agricultural production. Potential evapotranspiration refers to the maximum possible evapotranspiration from a land surface when water supply is unlimited under given meteorological conditions. As a key parameter in farmland irrigation management, crop water requirement estimation, and water balance studies in data-scarce regions, its spatiotemporal distribution and variation patterns are crucial for research. Actual evapotranspiration refers to the actual water vapor flux achieved by the underlying surface under specific meteorological conditions and water supply limitations, comprising two main components: evaporation from soil surfaces or water bodies, and transpiration from vegetation surfaces. Calculating evapotranspiration typically requires including both aspects to reflect local actual conditions.

Previous research on evapotranspiration has primarily focused on several areas: spatiotemporal evolution of potential evapotranspiration (ET_0) using methods such as Penman-Monteith, Hargreaves, and Priestley-Taylor; estimation of actual evapotranspiration using approaches like the FAO-recommended crop coefficient method (Kc), eddy covariance techniques, remote sensing models based on empirical statistics or water balance methods, and lysimeter measurements; investigation of crop evapotranspiration driving mechanisms and the impacts of climate change and human activities; and evaluation of regional applicability of evapotranspiration calculation methods. In China' s arid regions, actual evapotranspiration is significantly smaller than potential evapotranspiration due to soil moisture limitations and differences in vegetation growth conditions. Using potential evapotranspiration as the water expenditure component in watershed water balance calculations for Hexi inland rivers would yield substantial errors and fail to guide human water use. Therefore, accurate estimation of actual evapotranspiration is particularly important, though challenging. Consequently, previous studies have predominantly focused on potential rather than actual evapotranspiration, with even fewer investigations at the watershed scale. This study aims to address this gap by examining low mountain watersheds of Hexi inland rivers, estimating potential and actual evapotranspiration at different elevations and watershed positions from upstream to downstream, and deriving universal evapotranspiration patterns applicable to ungauged rivers in these regions.

2. Study Area and Methods

2.1 Study Area Overview The Xitugou watershed originates from the northern foothills of the Altun Mountains, with a maximum elevation of 5536 m. It comprises several perennial water channels including Hongliugou, Aksaigou, and Heigou (Figure 1). The upstream area is characterized by mid-high mountain terrain with modern glaciers above 4100 m. The middle reaches feature alluvial fans as the primary landform, while the downstream area has relatively complex geomorphology including deserts, Gobi, saline-alkali land, barren grassland, wetlands, and cultivated land. The watershed covers an area of 1125 km^2 with an average annual surface water runoff of 0.1104$×10^{3}$. Vegetation shows zonal distribution: from high to low elevations in the upper reaches, sub-nival zones, alpine shrubland, steppe desert, and mountain grassland appear sequentially. The middle reaches flow primarily through the Aksai Basin with small shrubs as the main vegetation and desert-Gobi as the natural landscape, with vegetation coverage of 20%-40%. The lower reaches are dominated by desert shrubland, desert grassland, and farmland. Located in the Gobi desert region, the area has flat terrain, arid climate with low precipitation, low vegetation coverage, and typical continental climate characteristics: hot summers, cold winters, large diurnal temperature variations, low rainfall, high evaporation, dry air, high accumulated temperature, long sunshine hours, and low cloud cover. The average annual sunshine duration is 3240 h, mean annual temperature is 9.4°C, annual temperature range is 34.1°C, absolute frost-free period is 145 days, and the absolute minimum temperature is -9.2°C. The multi-year average precipitation is 40.1 mm, with large intra-annual variation concentrated in June-September, accounting for approximately 100% of annual precipitation, mostly occurring as heavy rainstorms. The multi-year average evaporation measured by evaporation pan is as high as 239.8 mm, about 50 times the precipitation.

2.2 Data Sources Eight small automatic weather stations (model: WPH1/HOBO-U30-GSM) were installed in the upper, middle, and lower reaches of the Xitugou watershed, with additional soil moisture observation probes at each station buried at depths of 10 cm, 20 cm, 40 cm, 60 cm, and 80 cm. The stations are named Dunhuanggong (DHG), Baishuitai (BST), Qingyaizishan (QYZS), Chengliulin (CLL), Jiulianhu (JLH), Qingyaiziding (QYZD), Qingyaizigou (QYZG), and Akesai (AKS) (Figure 1). Meteorological data including daily maximum temperature, daily minimum temperature, average temperature, relative humidity, wind speed at 2 m height, sunshine duration, and air pressure were obtained from these stations. Soil moisture data were measured by the five-layer moisture probes at 30-minute intervals. After outlier removal, data were processed into daily values covering the period from 2018 to 2019. MOD13A1 NDVI data were obtained from https://ladsweb.modaps.eosdis.nasa.gov/ with a spatial resolution of 250 m.

2.3 Methods

2.3.1 Penman-Monteith Formula The Penman-Monteith equation recommended by the Food and Agriculture Organization (FAO) in 1998 is a widely used method for evapotranspiration estimation, with reference underlying surface being well-watered grass with a height of 0.12 m. The formula is:

$$ET_0 = \frac{\Delta(R_n - G) + \gamma \frac{900}{T+273} u_2 (e_s - e_a)}{\Delta + \gamma(1 + 0.34 u_2)}$$

where ET_0 is potential evapotranspiration ($\mathrm{mm \cdot d^{-1}}$); G is soil heat flux ($\mathrm{MJ \cdot m^{-2} \cdot d^{-1}}$); T is mean daily temperature at 2 m height (°C); u_2 is wind speed at 2 m height ($\mathrm{m \cdot s^{-1}}$); e_s is saturation vapor pressure (kPa); e_a is actual vapor pressure (kPa); Δ is the slope of the vapor pressure curve ($\mathrm{kPa \cdot {}^{\circ}C^{-1}}$); γ is the psychrometric constant ($\mathrm{kPa \cdot {}^{\circ}C^{-1}}$); and R_n is net radiation ($\mathrm{MJ \cdot m^{-2} \cdot d^{-1}}$), calculated as the difference between net shortwave radiation (R_{ns}) and net longwave radiation (R_{nl}).

Since the small weather stations did not measure sunshine duration, this study used the Hargreaves radiation formula to estimate solar shortwave radiation R_s:

$$R_s = (1 - \alpha) R_{so} - (0.34 - 0.14\sqrt{e_a}) \sigma T^4$$

where R_{so} is clear-sky solar radiation ($\mathrm{MJ \cdot m^{-2} \cdot d^{-1}}$); α is albedo; σ is the Stefan-Boltzmann constant; and T is temperature.

The Hargreaves formula is expressed as:

$$R_s = K_{Rs} \sqrt{T_{max} - T_{min}} R_a$$

where T_{max} and T_{min} are daily maximum and minimum temperatures; R_a is extraterrestrial radiation ($\mathrm{MJ \cdot m^{-2} \cdot d^{-1}}$); and K_{Rs} is an empirical adjustment coefficient ranging from 0.16 to 0.19 (0.16 for inland areas, 0.19 for coastal areas).

Net shortwave radiation is calculated as:

$$R_{ns} = (1 - \alpha)R_s$$

Net longwave radiation is calculated as:

$$R_{nl} = \sigma \left(\frac{T_{max}^4 + T_{min}^4}{2} \right) (0.34 - 0.14\sqrt{e_a})(1.35\frac{R_s}{R_{so}} - 0.35)$$

Soil heat flux G is calculated from the five-layer moisture probe measurements at each station.

2.3.2 Crop Coefficient K_c The dual crop coefficient method from FAO-56 divides the crop coefficient into plant transpiration coefficient (K_{cb}) and soil evaporation coefficient (K_e):

$$K_c = K_{cb} + K_e$$

where K_{cb} is the basal crop coefficient representing crop transpiration effects, and K_e is the soil evaporation coefficient representing soil evaporation effects.

The basal crop coefficient is calculated using NDVI:

$$K_{cb} = 1.07 \times \frac{NDVI - NDVI_{min}}{NDVI_{max} - NDVI_{min}} + 0.54$$

where $NDVI_{max}$ and $NDVI_{min}$ represent NDVI values for bare soil and dense vegetation, respectively.

The soil evaporation coefficient is:

$$K_e = K_r \times (1 - f_c) \times K_{c,max}$$

where K_r is a reduction factor; f_c is the effective fraction of soil surface covered by vegetation; and $K_{c,max}$ is the maximum crop coefficient.

The water stress coefficient K_s characterizes the impact of insufficient root zone soil moisture on crop transpiration and is calculated using a piecewise function based on field capacity and wilting coefficient for different soil types as specified in FAO-56.

2.3.3 Standardized Precipitation Evapotranspiration Index (SPEI) The SPEI, based on the Standardized Precipitation Index (SPI) and grounded in water cycle theory, considers the effects of precipitation and potential evapotranspiration under changing environments. The calculation process involves: (1) computing potential evapotranspiration using the improved Penman-Monteith method; (2) using the difference between monthly precipitation and monthly potential evapotranspiration as input data for the SPI; and (3) calculating the SPEI from the frequency values of the three-parameter (scale parameter, shape parameter, and location parameter) Logistic distribution.

The Logistic distribution calculation is:

$$F(x) = \left[1 + \exp\left(\frac{x - \xi}{\alpha}\right)\right]^{-\beta}$$

where ξ, α, and β are location, scale, and shape parameters, respectively.

The SPEI calculation uses the approximation:

$$SPEI = W - \frac{c_0 + c_{1W} + c_{2W}^2}{1 + d_{1W} + d_{2W}^2 + d_{3W}^3}$$

where $W = \sqrt{-2\ln(P)}$ for $P \leq 0.5$, $W = \sqrt{-2\ln(1-P)}$ for $P > 0.5$; P is cumulative probability; and constants are $c_0 = 2.515517$, $c_1 = 0.802853$, $c_2 = 0.010328$, $d_1 = 1.432788$, $d_2 = 0.189269$, $d_3 = 0.001308$.

The drought classification based on SPEI values is shown in Table 2.

3. Results

3.1 Daily Variation of Potential and Actual Evapotranspiration As shown in Figure 2, potential evapotranspiration (ET_0) begins to increase rapidly in mid-to-late May, reaches its maximum in July with daily maximum potential evapotranspiration of 11 $\mathrm{mm} \cdot \mathrm{d}^{-1}$, then gradually decreases with slowing rate after October, showing overall symmetrical distribution centered on July. Different observation points exhibit substantial variation in daily potential evapotranspiration amplitude, with greater variation at lower-elevation downstream sites and smaller variation at higher-elevation upstream sites. Actual evapotranspiration (ET) shows smaller daily variation amplitude than potential evapotranspiration at all sites except DHG. The ET at the DHG observation point shows the largest variation amplitude, primarily because this site is located in farmland in the lower reaches of the Xitugou watershed, significantly influenced by irrigation and vegetation coverage.

[Figure 2: see original paper]

3.2 Annual Potential and Actual Evapotranspiration As shown in Figure 3, annual potential evapotranspiration in the Xitugou watershed ranges from a maximum of 1478 mm to a minimum of 426 mm. Generally, upstream areas at higher elevations have smaller potential evapotranspiration, while downstream areas have larger values. Actual evapotranspiration is significantly smaller than potential evapotranspiration, with the maximum annual actual evapotranspiration of approximately 400 mm occurring at the downstream DHG site, and the minimum of 110 mm at the upstream QYZ4 site. Averaging data from all eight observation sites across the watershed yields mean annual potential evapotranspiration of 978 mm and mean annual actual evapotranspiration of 258 mm.

[Figure 3: see original paper]

3.3 Elevation Gradient Analysis To analyze elevation effects, observation sites were divided into five elevation gradients: downstream (1258 m average, 1236-1274 m), middle reaches (1686 m), upstream (3100 m, 3370 m, and 4100 m). The downstream sites have small elevation differences and similar underlying surfaces, so they were grouped together. The QYZ3 and QYZG sites at 3100 m and 3370 m have significantly different vegetation, so they were analyzed separately to fully reflect elevation impacts. QYZ4 at 4100 m is above the snow line.

As shown in Figure 4, potential evapotranspiration decreases with increasing elevation, dropping rapidly from 1258 m to 1686 m, then decreasing only slightly from 1686 m to 3100 m, with a larger decrease from 3370 m to 4100 m. Actual evapotranspiration also decreases with elevation, with the maximum of 356 mm at 1258 m and minimum of 110 mm at 4100 m.

[Figure 4: see original paper]

3.4 Monthly-Scale SPEI Variation Monthly SPEI values were calculated for all sites. As shown in Figure 5, the annual average SPEI values for each site follow the order: BST < JLH < CLL < DHG < 0 < AKS < QYZ3 < QYZG < QYZ4. Downstream sites show negative SPEI values throughout the year, with moderate drought in June at DHG, extreme drought in April at CLL, and severe drought in September at BST. Middle reaches show moderate drought in April at AKS. Upstream sites have positive SPEI values, indicating relatively wet conditions. In summary, autumn is the main drought season, followed by summer, with drought duration and severity in the lower reaches significantly greater than in the middle and upper reaches.

[Figure 5: see original paper]

3.5 Seasonal-Scale SPEI Variation Seasonal SPEI was calculated using 3-month scales for each season. As shown in Figure 6, all sites show positive

SPEI in spring. In summer, only downstream sites have negative SPEI. In autumn, all sites except QYZ4 show negative SPEI values less than -1, indicating severe drought. Upstream precipitation is characterized by high-intensity, low-frequency events that easily form floods flowing to middle and lower reaches. The middle reaches consist of alluvial fans with high infiltration rates, where precipitation and upstream floodwater infiltrate to form subsurface flow. The lower reaches have complex underlying surfaces including desert, barren grassland, shrubland, and farmland, where soil evaporation occurs mainly at desert margins and barren grasslands, while vegetation transpiration occurs mainly in farmland and shrub areas. Floodwater infiltrating in the middle reaches emerges in the lower reaches as surface flow for irrigation and domestic use. Precipitation and groundwater are water sources for evapotranspiration in the lower reaches, making it the primary water consumption zone of the entire watershed.

[Figure 6: see original paper]

4. Discussion

4.1 Estimation of Actual Evapotranspiration and Watershed Water Consumption As shown in Figure 7, precipitation gradually decreases from upstream to downstream, with average annual precipitation of 383.7 mm in the upper reaches, 203 mm in the middle reaches, and only 58.6 mm at the DHG site in the lower reaches. Actual evapotranspiration increases significantly from upstream to downstream, averaging 123.6 mm upstream, 239.8 mm in the middle reaches, and 356 mm downstream. Only the upper reaches generate runoff where precipitation exceeds actual evapotranspiration. The middle reaches have high infiltration rates that convert surface runoff to subsurface flow. The lower reaches have minimal precipitation but abundant water supply from emerging groundwater for human use and evapotranspiration.

[Figure 7: see original paper]

Actual evapotranspiration observation is difficult, so current estimates are primarily model-based. Common methods include watershed water balance, Penman-Monteith with dual and single crop coefficients, sap flow, eddy covariance, and lysimetry. Due to data scarcity, few studies have estimated actual evapotranspiration in northwestern China’ s arid inland river basins, particularly in desert areas. This study used 2018-2019 meteorological and soil moisture data from small automatic weather stations and soil moisture observation equipment installed across the watershed to quantitatively investigate water cycle components including precipitation and evapotranspiration, providing a reference for similar small watersheds in northwestern inland river regions.

This research employed the dual crop coefficient method because it adequately considers soil evaporation effects on total evapotranspiration, which is crucial in desert areas where vegetation transpiration is minimal. The calculated annual actual evapotranspiration of 258 mm is consistent with previous studies: Tian et al. reported 248 mm for the Manas River basin, Li Xiucang found 200-250 mm for the Tarim River basin, and Li Xiucang reported 278 mm for the Tarim River basin. The consistency suggests that the NDVI-based Kc estimation method is applicable in arid inland river basins.

In the Xitugou watershed, vegetation coverage is very low (0.01-0.47), making vegetation transpiration significantly smaller than soil evaporation. Soil evaporation is primarily controlled by precipitation and irrigation. In such low-precipitation, low-coverage areas, limited soil evaporation results in relatively small crop coefficients. With small interannual variations in meteorological factors, both potential and actual evapotranspiration show small interannual variability. Despite having only one year of data, the results likely represent evapotranspiration and drought characteristics of the watershed and similar low mountain watersheds in Hexi inland rivers.

4.2 Mechanism and Applicability of SPEI in Arid Inland River Basins

Drought indices are controlled by precipitation and potential evapotranspiration. From upstream mountainous areas to downstream desert regions, significant changes in precipitation, temperature, wind speed, and air pressure create regional differences in potential evapotranspiration, resulting in spatial variation of drought indices. In the Xitugou watershed, the upstream area on the eastern edge of the Altun Mountains has higher elevation, lower temperature, lower pressure, and greater precipitation than the middle and lower reaches. Low temperature and low pressure are the main reasons for relatively low potential evapotranspiration in the upstream mountainous area. In the downstream area with precipitation below 50 mm, high temperature and high wind speed create large potential evapotranspiration. These differences produce varying precipitation-potential evapotranspiration differences: smaller differences upstream, larger differences downstream.

The SPEI method reveals these patterns: average SPEI is -0.5 downstream (light drought), 0.2 in the middle reaches (normal), and 0.5 upstream (slightly wet). The drought index decreases with decreasing elevation, becoming negative downstream. Downstream drought is primarily controlled by potential evapotranspiration, while upstream drought is influenced by both precipitation and potential evapotranspiration. Lower SPEI values indicate greater water expenditure and less water income.

In arid inland river basins with precipitation mostly below 50 mm, considering only precipitation (as in SPI) yields low accuracy. SPEI, which incorporates temperature effects, provides more reasonable drought assessment. Studies have shown that SPEI calculated using Penman-Monteith potential evapotranspiration better reflects actual conditions than methods using Hargreaves or Thornthwaite. The consistency between this study’ s results and previous research demonstrates that one year of data can adequately represent watershed characteristics.

5. Conclusions

1. **Intra-annual variation**: Potential and actual evapotranspiration in the Xitugou watershed show symmetrical variation centered on July, which is the main period of water consumption. Annual potential evapotranspiration is 978 mm and actual evapotranspiration is 258 mm. From upstream to downstream, both potential and actual evapotranspiration show gradually increasing regional patterns.
2. **Drought characteristics**: The annual average SPEI values for the eight observation sites follow the order: BST < JLH < CLL < DHG < 0 < AKS < QYZ3 < QYZG < QYZ4. Drought occurs across the watershed with varying severity, becoming more severe from upstream to downstream. Autumn is the primary drought season, followed by summer. Drought duration and severity in the lower reaches significantly exceed those in the middle and upper reaches.
3. **Water balance**: Precipitation exceeds actual evapotranspiration in the upper reaches (the main runoff generation zone), while actual evapotranspiration exceeds precipitation in the middle and lower reaches. The middle reaches have high infiltration rates that convert surface runoff to subsurface flow. The lower reaches have minimal precipitation but receive water from emerging groundwater for human use and evapotranspiration. Overall watershed evapotranspiration significantly exceeds precipitation.
4. **Method applicability**: This study evaluated the applicability of NDVI-based crop coefficient Kc for actual evapotranspiration estimation in Hexi inland river basins and found it produces reliable results. SPEI calculated using the Penman-Monteith formula provides more accurate and effective drought assessment than SPI or PDSI. As a representative small watershed in the low mountain area of Hexi inland rivers, the Xitugou watershed' s evapotranspiration patterns and drought characteristics are applicable to other ungauged small watersheds in this region.

References

[1] Wang Yanxin, Li Ruiping, Li Xiazi. Estimation and variability of evapotranspiration for different land types during the growing season in the Hetao Irrigation District[J]. Arid Zone Research, 2020, 37(2): 364-373.

[2] Gu L, Hu Z, Yao J, et al. Actual and reference evapotranspiration in a cornfield in the Zhangye Oasis, Northwestern China[J]. Water, 2017, 9(7): 1-16.

[3] Kool D, Agam N, Lazarovitch N, et al. A review of approaches for evapotranspiration partitioning[J]. Agricultural and Forest Meteorology, 2014, 184: 56-70.

[4] Allen R G, Pereira L S, Raes D, et al. Crop evapotranspiration: Guidelines for computing crop water requirements[J]. FAO Irrigation and Drainage Paper

56, 1998.

[5] Han S, Zhang B. Advances of evapotranspiration research based on the Penman approach and complementary principle[J]. Journal of Hydraulic Engineering, 2018, 49(9): 1158-1168.

[6] Hargreaves G H, Samani Z A. Reference crop evapotranspiration from temperature[J]. Applied Engineering in Agriculture, 1985, 1(2): 96-99.

[7] Priestley C B, Taylor R J. On the assessment of surface heat flux and evaporation using large scale parameters[J]. Monthly Weather Review, 1972, 100(2): 81-92.

[8] Blaney H F. Determining Water Requirements in Irrigated Areas from Climatological and Irrigation Data[M]. United States: Soil Conservation Service, 1950.

[9] Rohwer C. Evaporation from Free Water Surfaces[M]. US Department of Agriculture, Washington, D.C. in cooperation with Colorado Agricultural Experiment Station, 1931.

[10] Zhao Liwen, Ji Xibin. Quantification of transpiration and evaporation over agricultural field using the FAO-56 dual crop coefficient approach: A case study of the maize field in an oasis in the middlestream of the Heihe River Basin in Northwest China[J]. Scientia Agricultura Sinica, 2010, 43(19): 4016-4026.

[11] Sun Xiaolong, Wu Rongsheng, Li Ping, et al. An evaluation of the Hargreaves method for estimating reference evapotranspiration in different grassland types in Inner Mongolia, China[J]. Acta Prataculturae Sinica, 2016, 25(5): 13-20.

[12] Sun Li, Song Changchun. Estimation of evapotranspiration from a typical marsh in Sanjiang Plain[J]. Chinese Journal of Applied Ecology, 2008, 19(9): 1925-1930.

[13] Zhao Linlin, Wang Zhonggen, Xia Jun, et al. Improved Priestley-Taylor method and its application in complementary relationship evapotranspiration model[J]. Progress in Geography, 2011, 30(7): 805-810.

[14] Liu Yu, Peng Zhigong. A review of monitoring and estimating methods for regional evapotranspiration[J]. Journal of China Institute of Water Resources and Hydropower Research, 2009, 7(2): 256-264.

[15] Li Tiansheng, Xia Jun, Kuang Yang, et al. The applicability of various potential evapotranspiration estimation methods in the middle and upper reaches of Hanjing River Basin[J]. South North Water Transfers and Water Science & Technology, 2017, 15(6): 1-10.

[16] Ren Qingfu, Yang Zhiyong, Li Chuanzhe, et al. Advances in the study of the crop evapotranspiration in changing environment[J]. Advances in Earth Science, 2013, 28(11): 1227-1238.

[17] Guo X, Cheng G. Advances in the application of remote sensing to evapotranspiration research[J]. Advance in Earth Sciences, 2004, 19(1): 107-114.

[18] Zhang Yang, Qiu Guoyu, Yan Chunhua, et al. Studies on the influence of altitudes on the trend of reference evapotranspiration in recent 50 years: A case study of Sichuan province[J]. Ecology and Environmental Sciences, 2018, 27(12): 2208-2216.

[19] Qiu Meijuan, Liu Buchun, Liu Yuan, et al. Temporal spatial variation characteristics of reference crop evapotranspiration and its influence factors in Jilin Province[J]. Journal of Arid Meteorology, 2019, 37(1): 119-126.

[20] Liu X, Zheng H, Zhang M, et al. Identification of dominant climate factor for pan evaporation trend in the Tibetan Plateau[J]. Journal of Geographical Sciences, 2011, 21(4): 594-608.

[21] Li Yuanfei, Zhang Lanxia, Cao Yongqiang, et al. Spatiotemporal variation characteristics of potential evapotranspiration and climate influencing factors in Hebei province[J]. South North Water Transfers and Water Science & Technology, 2019, 17(3): 67-78.

[22] Li Nian, Sun Weijun, Qin Xiang, et al. Estimation of evapotranspiration in an alpine meadow in zone of Laohugou in Qilian Mountains[J]. Journal of Arid Land Resources and Environment, 2016, 30(6): 173-178.

[23] Tian Jing, Su Hongbo, Chen Shaohui, et al. Spatiotemporal variations of evapotranspiration in China mainland in recent 20 years[J]. Resources Science, 2012, 34(7): 1277-1286.

[24] Abid N, Bargaoui Z, Mannaerts C M. Remote sensing estimation of the water stress coefficient and comparison with drought evidence[J]. International Journal of Remote Sensing, 2018, 39(14): 4616-4640.

[25] Rocha J, Perdigao A, Melo R, et al. Managing water in agriculture through remote sensing applications[C]//Proceedings of 30th EARSeL Symposium on Remote Sensing for Science, Education, and Natural and Cultural Heritage, Paris, France, 2010, 31: 223-230.

[26] Chen Xuelin, Niu Zuirong, Huang Weidong, et al. Mode and effect of flood resources utilization in Xitugou Watershed of Dunhuang[J]. Journal of China Hydrology, 2017, 37(2): 73-77.

[27] Li Xiaoyuan, Yu Deyong. Progress on evapotranspiration estimation methods and driving forces in arid and semiarid regions[J]. Arid Zone Research, 2020, 37(1): 26-36.

[28] Wright J L. New Evapotranspiration crop coefficients[J]. Journal of the Irrigation & Drainage Division ASCE, 1982, 108: 57-74.

[29] Yang Qing, Li Mingxing, Zheng Ziyan, et al. Regional adaptability of 7 meteorological drought indexes in China[J]. Scientia Sinica Terrae, 2017, 47(3): 337-353.

[30] Zhang Leyuan, Wang Yi, Chen Yaning. Spatial and temporal distribution characteristics of drought in Central Asia based on SPEI index[J]. Arid Zone Research, 2020, 37(2): 331-340.

[31] Sun Yijie, Liu Xianfeng, Ren Zhiyuan, et al. Spatiotemporal variations of multi-scale drought and its influencing factors across the Loess Plateau from 1960 to 2016[J]. Geographical Research, 2019, 38(7): 1820-1832.

[32] Yin Wenjie, Zhang Menglin, Hu Litang. Spatiotemporal variation of drought in the Qaidam Basin[J]. Arid Zone Research, 2018, 35(2): 387-394.

[33] Jiang R, Xie J, He H, et al. Use of four drought indices for evaluating drought characteristics under climate change in Shaanxi, China: 1951-2012[J]. Natural Hazards, 2014, 75(3): 2885-2903.

[34] Guo Xuxin, Zhao Ying, Gao Zhiyong, et al. SPEI based drought characters and factors in Loess Hilly regions of Northern Shaanxi[J]. Journal of Northwest Forestry University, 2019, 34(1): 69-76.

[35] Wang Cui. Correlation analysis of evapotranspiration changes in Manas River Basin based on water saving measures[J]. Shaanxi Water Resources, 2019(7): 104-109.

[36] Li X, Sha J, Wang Z L. Comparison of drought indices in the analysis of spatial and temporal changes of climatic drought events in a basin[J]. Environmental Science Pollution Research, 2019, 26(11): 10695-10707.

[37] Li Xiucang. Spatiotemporal Variation of Actual Evapotranspiration in the Pearl, Haihe and Tarim River Basins of China[D]. Nanjing: Nanjing University of Information Science & Technology, 2013.

[38] Luo Nana, Bake Bateer, Wu Yanfeng. Correlation analysis of potential evapotranspiration and key climatic factors in Shihezi city[J]. Research of Soil and Water Conservation, 2016, 23(5): 251-255.

[39] Xu Junzeng, Peng Shizhang, Zhang Ruimei, et al. Reference evapotranspiration varied with latitude and altitude[C]//Chinese Society of Agricultural Engineering. Agricultural Engineering Technology Innovation and Construction of Modern Agriculture: Second Volume of Proceedings of the 2005 Annual Conference of the Chinese Society of Agricultural Engineering, 2005: 134-137.

[40] Jin X, Guo R, Xia W. Distribution of actual evapotranspiration over Qaidam Basin, an arid area in China[J]. Remote Sensing, 2013, 5(12): 6976-6996.

Note: Figure translations are in progress. See original paper for figures.

Source: ChinaXiv —Machine translation. Verify with original.