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Effect of climate change on reference evapotranspiration and aridity index in arid region of China

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SUMMARY

Temporal variations in reference evapotranspiration (ET_0) and aridity index (AI) were comprehensively investigated for 23 meteorological stations during 1955–2008 in the northwest China. The quantitative contributions of the major meteorological variables to the ET_0 and AI trends were evaluated and the possible causes were also investigated. The results showed that in the past 50 years annual temperature, humidity and precipitation had significant increasing trends with time, and wind speed and radiation had decreasing trends. ET_0 had a significant decreasing trend with an averagely value of about 3 mm per year, and AI had also witnessed a decreasing trend; For ET_0, wind speed was the most sensitive meteorological variable, followed by relative humidity, temperature and radiation and for AI, precipitation was the most sensitive meteorological variable. The contribution of wind speed to the decrease of ET_0 is more than that of other meteorological variables. The increase of precipitation contributes more than the decrease of ET_0 to decrease of AI in past 50 years. This study provides an understanding of the effect of recent climate change on drought in arid northwest China.

1. Introduction

Human activities and climate changes have intensely influenced the ecohydrological pattern of many river basins all over the world and a series of problems about water resources have been induced in some regions. In arid or semi-arid regions, minor variations in precipitation and temperature easily induced significant changes in hydrological processes (Gan, 2000; Ma et al., 2004). Evapotranspiration is an important flux term in the water cycle that integrates atmospheric demands and surface conditions. Furthermore, evapotranspiration is also a factor determining climatic drought in arid and semi-arid regions. Any variations in meteorological variables induced by climate change will affect evapotranspiration or crop water requirements. Eventually, climate change will increase or decrease the dry conditions in the arid regions of the world by increasing or decreasing potential evapotranspiration. The prevalent understanding is that climate change increases drought and aggravates the process of desertification in conjunction with the ever-growing impact of humans and domestic animals on fragile and unstable ecosystems (Goyal, 2004). The trends in reference evapotranspiration (ET_0) and aridity index (AI) are of great significance for managing agricultural water resources.

Usually, actual crop evapotranspiration (ET_a) can be derived from reference evapotranspiration (ET_0) by means of appropriate crop and water stress coefficients. Consequently, understanding of the temporal and spatial variations of ET_0 is a vital component in regional hydrological studies. From the practical point of view, in semi-arid climates where water resources are limited and irrigation is the largest user of water, temporal variations of ET_0 and quantification of its trend can produce valuable reference data for regional studies, hydrological modeling, agricultural water requirement, irrigation planning and water resources management (Liang et al., 2010).

In recent years, extensive research efforts have examined the potential impact of climate change on reference evapotranspiration (ET_0). Analysis of long-term climate series indicate abrupt climate changes occurred in northwest China around 1975 (Chen et al., 1991). Thomas (2000) considered ET_0 time series in China during 1954–1993 and showed that ET_0 has decreased in all seasons especially in northwest and southeast China. Declining trends in ET_0 over the last 50 years were found for the Yangtze River basin, Yellow River basin, northern regions and Tibetan Plateau of China (Wang et al., 2007; Zhang et al., 2007; Zhang et al., 2010; Song et al., 2010), and over India (Bandyopadhyay et al., 2009) and

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USA (Irmak et al., 2012). However, ET_0 has increased in semi-arid Iran for the same period (Sabziparvar and Tabari, 2010; Tabari, 2010; Tabari and Marofi, 2010).

Sensitivity analyses have been used to determine the expected change of ET_0 in response to a known change of climatic variables. Results of these analyses make it possible to determine the accuracy required when measuring climatic variables used to estimate ET_0 (Irmak et al., 2006). McCuen (1974) was one of the first researchers to investigate the sensitivity of various evaporation models to meteorological variables. Since then more studies analyzed the sensitivity of evapotranspiration as the effect of global climate change on agricultural water requirement (Rana and Katerji, 1998; Hupet and Vanclooster, 2001). As influenced by the water and energy conditions, the responses of ET_0 to climate change present different characteristics in different regions. Liu et al. (2010) review recent studies about sensitivity analysis of ET_0 to key meteorological variables and found the sensitivity factors for ET_0 varied with location. For example, Goyal (2004) found ET_0 is sensitive to temperature and net solar radiation in arid Rajasthan in India. However, Gong et al. (2006) reported that relative humidity has the most influence on ET_0 in humid Yangtze River basin of China. It is noted that investigating the response of ET_0 to change in meteorological variables should be done separately for each region. Recently, several studies have begun to quantify the contribution of meteorological variables to the change in ET_0. Generally, the method is to detrend for each meteorological variable series and recalculate the ET_0 series with detrended meteorological variables. Xu et al. (2006) indicated that the decreasing trend in the net total radiation is the main cause of the decreasing trend found in the reference evapotranspiration, while the decreasing trend in wind speed contributed to a lesser degree. Liu et al. (2010) found that increasing air temperature is the main cause of increasing ET_0 in the Yellow River basin of China. Papaioannou et al. (2011) showed that the impact of temperature change on ET_0 is almost negligible during the dimming period in Greece. Some different results were obtained in the Mediterranean basin (Palutikof et al., 1994), Rasht City, Iran (Jahanbani et al., 2011), and Central Italy (Todisco and Vergni, 2008).

With global warming, aridity/humidity trends were predicted for some areas in some model scenarios which estimated that drought would persist in critical agricultural regions. Andreadis and Lettenmaier (2006) used a simulated data set of hydro-climatological variables to examine for 20th century trends in drought characteristics over the conterminous United States and found that, for the most part, droughts have become shorter, less frequent, and cover a smaller portion of the country over the last century. In previous studies, some significant evidence of extreme drought/wetness was found through extreme precipitation. For example, a decrease in drizzle in northern China is evidence of an atmospheric drought (Yan and Yang, 2000). Zhai et al. (1999) studied the trend of extreme precipitation in China. Wang et al. (2008) propose an aridity-wetness homogenized index and found there is greatly humidification trend in the whole northwest China. However, these studies only considered the impact of precipitation variation on drought, and could not show the influence of temperature on drought, especially under global warming.

Although trends of ET_0 and possible contribution of meteorological variables were investigated in many regions around the world, the effect of climate change (global warming) on ET_0 is unclear in the northwest region of China, one of the most arid areas all of world. Especially, understanding aridity trend and the main contribution is important to cope with the shortage of water resources, but no studies concentrating on northwest China have been published. So, the objectives of this study are, (1) to analyze seasonal and annual variation of meteorological variables, ET_0 and aridity; (2) to detect the sensitivity of ET_0 and aridity to key meteorological variables; (3) to quantify the contribution of key meteorological variables on ET_0 and aridity trends. This study will also improve our understanding of climate change and the accompanying effect on hydrology and climatic aridity, as ET_0 being not only a climatic variable but also an important hydrological process.

2. Materials and methods

2.1. Study area and data description

Arid inland regions of northwest China include Xinjiang province, Hexi Corridor of Gansu province, part of Qinghai province and the western part of Inner Mongolia district (Fig. 1). This region covers an area of 219.05×10^4 km^2 which is 22.82% of China. The region exhibits typical arid climatic features, in which the annual precipitation ranges between 16 and 267 mm and the annual reference evapotranspiration is 719–1375 mm. Since the foundation of the People's Republic, northwest China has become one of the most intensely cultivated areas of China. Rapid population growth and migration to the marginal semi-arid and arid areas exacerbated the land degradation and water depletion. Agriculture is the biggest water user in the basin. In 2000, irrigation accounted for more than 90% of the total water consumption of different sectors, equal to 28.5×10^8 m^3.

In this study, monthly meteorological data were obtained from 23 stations from January 1955 to December 2008 in the arid-inland basin of northwest China. The data have been provided by the National Climatic Centre (NCC) of CMA (the China Meteorological Administration). Six monthly meteorological variables were recorded including (1) mean minimum air temperature (T_{min}, °C); (2) mean maximum air temperature (T_{max}, °C); (3) mean relative humidity (Rh); (4) mean wind speed at 2 m (U_2, m/s); (5) bright sunshine hours (N, h/d); and (6) precipitation (P, mm/month). The information about the stations was presented in Table 1, and the geographical location of the stations was shown in Fig. 1. Mean monthly air temperature was calculated by taking the average of the maximum and minimum air temperature.

2.2. Reference evapotranspiration (ET_0)

The FAO Penman–Monteith method (Allen et al., 1998) was developed by defining the reference crop as a hypothetical crop with an assumed height of 0.12 m having a surface resistance of 70 s/m and an albedo of 0.23, closely resembling the evaporation of an extensive surface of green grass of uniform height, actively

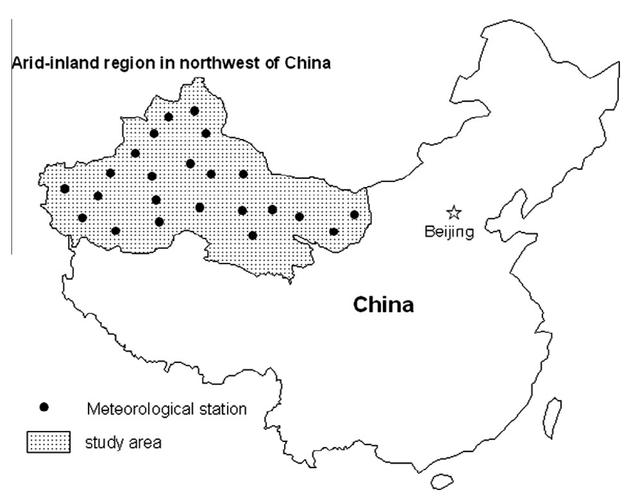

Fig. 1. Location of the arid-inland basin of China and the meteorological stations used in this study.

Table 1
Geographic characteristics of the stations used in the study.

Stations	Latitude (dec.deg.)	Longitude (dec. deg.)	Elevation (m.a.s.l.)	T (°C)	P (mm/year)	ET_0 (mm/year)
Aletai	47.44	88.05	735.3	4.56	195.3	769.5
Fuyun	46.59	89.31	807.5	3.63	186.4	719.0
Hebusaike	46.47	85.43	1291.6	4.19	144.1	752.5
Kelamayi	45.37	84.51	449.5	8.85	112.6	1374.9
Jinghe	44.37	82.54	320.1	8.16	101.6	755.4
Qitai	44.01	89.34	793.5	5.73	187.4	848.3
Yining	43.57	81.2	662.5	9.39	274.7	695.4
Wulumuqi	43.47	87.39	935.0	7.56	267.0	927.0
Tulufan	42.56	89.12	34.5	14.94	16.0	1031.0
Kuke	41.43	83.04	1081.9	11.64	67.5	1068.2
Kashi	39.28	75.59	1288.7	12.12	64.9	972.3
Bachu	39.48	78.34	1116.5	12.47	54.7	924.7
Tielikegan	40.38	87.42	846.0	11.47	34.1	992.5
Ruoqiang	39.02	88.1	888.3	12.18	27.5	1232.5
Shache	38.26	77.16	1231.2	12.15	51.5	851.6
Hetian	37.08	79.56	1374.5	13.08	37.7	1043.2
Hami	42.49	93.31	737.2	10.49	38.6	999.7
Donghuang	40.09	94.41	1139.0	9.99	38.1	860.4
Yumen	40.16	97.02	1526.0	7.54	63.8	1071.3
Jiuquan	39.46	98.29	1477.2	7.97	84.5	859.7
Minqin	38.38	103.1	1367.0	8.73	114.0	969.6
Jilantai	39.47	105.5	1031.8	9.46	109.1	1184.2
Geermu	36.25	94.54	2807.6	5.56	41.4	980.0

growing and adequately watered. The PM56 equation recommended for daily reference evapotranspiration ET_0 (mm/d) estimation (Allen et al., 1998) may be written as:

$$ET_0 = \frac{0.408\Delta(R_n - G) + \gamma(900/(T+273))U_2(e_s - e_a)}{\Delta + \gamma(1 + 0.34U_2)} \tag{1}$$

where R_n is the net radiation at the crop surface (MJ/m^2 d), G the soil heat flux density (MJ/m^2 d), T the air temperature at 2 m height (°C), U_2 the wind speed at 2 m height (m/s), e_s the vapor pressure of the air at saturation (kPa), e_a the actual vapor pressure (kPa), Δ the slope of the vapor pressure curve (kPa/°C) and γ the psychrometric constant (kPa/°C). A complete set of equations was proposed by Allen et al. (1998) to compute the parameters in Eq. (1) according to the available weather data and the time step, which constitute the PM56 method.

The key radiation part of the PM56 model can be calculated using the following equations:

$$R_n = R_{ns} - R_{nl} \tag{2}$$

$$R_{ns} = (1 - \alpha)R_s \tag{3}$$

$$R_s = \left(a + b\frac{n}{N}\right)R_a \tag{4}$$

$$R_{nl} = \sigma\left(\frac{T_{xk}^4 + T_{nk}^4}{2}\right)(c + d\sqrt{e_a})\left(e\left(\frac{R_s}{R_{so}}\right) + f\right) \tag{5}$$

where R_n is the net solar radiation (MJ m^{-2} d^{-1}), R_{ns} is the net shortwave radiation, R_{nl} is the net longwave radiation, α is the albedo, R_s is the solar radiation, n is the actual sunshine duration (h), N is the maximum possible sunshine duration, n/N is the relative sunshine duration, R_a is the extraterrestrial radiation, σ is the Stefan–Boltzmann constant (4.903×10^{-9} MJ K^{-4} m^{-2} d^{-1}), T_{xk} is the maximum absolute temperature during the 24-h period, T_{nk} is the minimum absolute temperature during the 24 h period, R_{so} is the clear-sky solar radiation, a–f are empirical coefficients. N, R_a and R_{so} were calculated by solar constant, latitude, elevation and the number of the day in the year according to the FAO56 report. The FAO56 recommended the coefficients of $a = 0.25$, $b = 0.50$, $c = 0.34$, $d = -0.14$, $e = 1.35$ and $f = -0.35$ to be used in regions where no actual R_s data are available and no calibration has been carried out (Allen et al., 1998).

2.3. Aridity index

There are many formulas to describe an aridity index (AI) (Mannocchi et al., 2004). UNESCO (1979) calculated AI as the mean annual precipitation (P) over the mean annual ET_0. In arid area, generally the agricultural aridity is induced mainly by evapotranspiration exceeding precipitation. ET_0 means the potential evapotranspiration determined by climate factors which can be observed easily. Thornthwaite (1948) expressed AI as dividing the difference between precipitation and potential evapotranspiration by potential evapotranspiration. This definition can express the arid degree in arid area or season. In this study, similarly the AI was calculated following as,

$$AI = (ET_0 - P)/ET_0 \tag{6}$$

This equation defines AI as the ratio of difference between ET_0 and P to ET_0 and can be understood as the water requirement to satisfy reference evapotranspiration. If there is no precipitation, AI will be 1 and mean aridity is the highest. In contrast, if the precipitation be equal to or higher than reference evapotranspiration, the AI will equal to zero or be negative. In the study area, AI was a value between 0 and 1 because the ET_0 are far larger than the P.

2.4. Statistical tests for trend analysis

2.4.1. Mann–Kendall test

The Mann–Kendall test (M–K test), also called Kendall's tau test due to Mann (1945) and Kendall (1975), is a rank based non-parametric test for assessing the significance of a trend, and has been widely used in hydrological trend detection studies. The null hypothesis H_0 is that a sample of data $\{x_i, i = 1, 2, \ldots, n\}$, x_i is independent and identically distributed. The alternative hypothesis H_1 is that a monotonic trend exists in X. The statistic S of Kendall's tau is defined as follows:

$$S = \sum_{i=1}^{n-1}\sum_{j=i+1}^{n}\operatorname{sgn}(x_j - x_i) \tag{7}$$

where the x_j are the sequential data values, n is the length of the data set, and

$$\operatorname{sgn}(\theta) = \begin{cases} 1 & \theta > 0 \\ 0 & \theta = 0 \\ -1 & \theta < 0 \end{cases} \tag{8}$$

Mann (1945) and Kendall (1975) have documented that when $n \geqslant 8$, the statistic S is approximately normally distributed with the mean and the variance as follows:

$$E(S) = 0$$
$$Var(S) = \frac{n(n-1)(2n+5) - \sum_{m=1}^{n} t_m(m-1)(2m+5)}{18} \tag{9}$$

where t_m is the number of ties of extent m. The standardized test statistic Z_c is computed by

$$Z_c = \begin{cases} \frac{S-1}{\sqrt{\mathrm{var}(S)}} & S > 0 \\ 0 & S = 0 \\ \frac{S+1}{\sqrt{\mathrm{var}(S)}} & S < 0 \end{cases} \tag{10}$$

where Z_c is the test statistics. When $|Z_c| > Z_{1-\alpha/2}$, in which $Z_{1-\alpha/2}$ are the standard normal deviates and α is the significance level for the test. In addition, simple linear regression was used to establish trend of five meteorological variables.

2.4.2. Change-point analysis

The non-parametric approach developed by Pettitt (1979) was used in this study. This approach detects a significant change in the mean of a time series when the exact time of the change is unknown. The test uses a version of the Mann–Whitney statistic $U_{t,N}$, which verifies whether two samples $x_1, \ldots, x_t$ and $x_{t+1}, \ldots, x_N$ are from the same population or not. The test statistic $U_{t,N}$ is given by

$$U_{t,N} = U_{t-1,N} + \sum_{j=1}^{N} \mathrm{sgn}(x_t - x_j) \quad \text{for } t = 2, \ldots, N \tag{11}$$

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 Pettitt's test is the absence of a changing point. Its statistic $k(t)$ and the associated probabilities used in significance testing are given as

$$k(t) = Max_{1 \leqslant t \geqslant N} |U_{t,N}| \tag{12}$$

and

$$p \cong 2\exp\{-6(K_N)^2/(N^3 + N^2)\} \tag{13}$$

2.5. Sensitivity analyses and sensitivity coefficients

For multi-variable models (e.g., the PM method), different variables have different dimensions and different ranges of values, which makes it difficult to compare the sensitivity by partial derivatives. Because of the different approaches used in parameterizing ET_0 models, there are different definitions of the sensitivity coefficients and different ways to carry out the sensitivity analyses on ET equations (Gong et al., 2006). Literature reviews of previous studies revealed that there is no standard or common procedure for computing sensitivity coefficients for climate variables (Irmak et al., 2006). The non-dimensional relative sensitivity coefficients used in this study were calculated following McCuen (1974):

$$Se_{Vi} = \lim_{\Delta v_i \to 0} \left(\frac{\Delta ET_0/ET_0}{\Delta V_i/V_i} \right) = \frac{\partial ET_0}{\partial V_i} \cdot \frac{V_i}{ET_0} \tag{14}$$

Se_{Vi} is the sensitivity coefficient and V_i is the ith variable. The transformation gives the non-dimensional relative sensitivity coefficient. Basically, a positive/negative sensitivity coefficient of a variable indicates that ET_0 will increase/decrease as the variable increases. The larger the sensitivity coefficient, the larger relative effect a given variable has on ET_0. In graphical form, the sensitivity coefficient is the slope of the tangent at the origin of the sensitivity curve. In practice, the coefficient is accurate enough to represent the slope of the sensitivity curve within a certain "linear range" around the origin. The width of the range depends on the degree of non-linearity of the sensitivity curve. If a sensitivity curve is linear, the sensitivity coefficient is able to represent the change in ET_0 caused by any perturbation of the variable concerned. A sensitivity coefficient of 0.2 for a variable would in this case mean that a 10% increase of that variable, while all other variables are held constant, may increase ET_0 by 2%. If the sensitivity curve is significantly non-linear, the predictive power of the sensitivity coefficient will be limited to small perturbations only.

2.6. Quantitative estimation of the influence of climate variables change on ET_0

In order to quantify the contributions of key meteorological variables to the change trend of evapotranspiration and aridity, similar to Xu et al. (2006), the following steps are performed: (1) removing the change trend in temperature, wind speed, humidity, net total radiation to convert them as stationary time series; (2) recalculating the reference evapotranspiration using the detrended data series for wind speed or net total radiation and using the original data for other variables; and (3) comparing the recalculated reference evapotranspiration with the original value. Any difference is considered to be due to the influence of that variable on the trend.

3. Results and discussions

3.1. Change trends of climate variables

Seasonal trends of several key climate variables from 1955 to 2008 are presented in Table 2. Linear regression revealed that T, Rh and P had an increasing trend for four seasons. For spring and winter T respectively increased by 0.043 and 0.032 °C per year in the past 50 years. An increasing trend of 0.029 and 0.010 °C per year was also observed respectively in summer and autumn. The increase in T was lower in warmer seasons than in cooler seasons. According to M–K test, the increasing trend of T was significant ($p < 0.01$) for three seasons. Furthermore, the Pettit test indicated that T series had a change point in 1996 for warmer seasons. For Rh, the increasing trend was significant in cooler season ($p < 0.05$), while for P only in winter ($p < 0.05$).

It is noted that U_2 had a significant declining trend ($p < 0.01$) in all four seasons. In spring and winter seasons, U_2 reduced by 0.015 m/s per year, and 0.021–0.024 m/s per year in summer and autumn. The decreasing trend was stronger in warmer seasons than in cooler seasons. R_n had a decreasing trend in all four seasons, and the trends were significant in spring, autumn ($p < 0.05$) and winter ($p < 0.01$).

It was noted that annual average T and Rh significantly increased by 0.028 °C and 0.054 respectively in the past 50 years ($p < 0.01$). However, annual average U_2 and R_n had significant decreasing trends with a reduction of 0.019 m/s and 0.011 MJ/m^2 d, respectively. For the annual P, a significant increasing trend ($p < 0.01$) was observed which was 0.522 mm/year. Furthermore, the Pettit test showed that there were change points in 1986, 1987, 1983, 1972 and 1987 respectively for the five climate factors.

According to the IPCC (2007), climatic variability combined with human-induced emission of greenhouse gases has resulted in an increase of surface temperature of the Earth by about 0.13 °C per decade over the past 50 years. This study revealed that the effect of global climate change on the increase in air temperature was more remarkable in northwest China. Although there was a consistent conclusion that air temperature had been increasing in the past several decades, the trends of other meteorological variables are different in different regions. Xu et al. (2006) detected that R_a and U_2 had decreasing trends in the past 50 years, while there was no significant trend for Rh in the Yangtze River basin. Liu et al. (2010) also reported a similar conclusion in the Yellow River basin. In this study, we found that not only U_2 and R_a decreased remarkably but also Rh enhanced significantly in arid northwest China. P is the most important factor to assess the climate aridity and the trend was different in each region. A declining trend was found in the Yellow River basin in the past 50 years (Liu et al., 2008). It is noteworthy that annual P presented an increasing trend and, although the trend was not significant for spring, summer and autumn, it had an important influence on reducing climatic drought in the arid region. In fact, different change trends of climate variables were found in different regions over world (Jhajharia et al., 2009; Jhajharia and Singh, 2011).

Table 2
Trend analysis results of climate variable with linear regression, M–K test.

	Linear regression		M–K analysis		Pettit test
	Slope	Std	Z_c	Significance level	Change point (year)
Spring					
T (°C)	0.043	1.869	2.512	–*	–
Rh (%)	0.011	4.253	0.420	–	–
U (ms^{-1})	−0.015	0.182	−6.130	–**	1983
R_a ($MJ\,m^{-2}\,d^{-1}$)	−0.024	1.154	−2.042	–*	–
P (mm)	0.013	2.512	0.634	–	–
Summer					
T (°C)	0.029	1.029	2.596	–**	1996
Rh (%)	0.028	3.524	0.828	–	–
U (ms^{-1})	−0.024	0.228	−6.662	–**	1983
R_a ($MJ\,m^{-2}\,d^{-1}$)	−0.018	1.194	−0.316	–	–
P (mm)	0.035	4.897	0.594	–	–
Autumn					
T (°C)	0.010	0.720	1.610	–	1996
Rh (%)	0.107	3.322	3.235	–**	1987
U (ms^{-1})	−0.021	0.202	−6.369	–**	1983
R_a ($MJ\,m^{-2}\,d^{-1}$)	−0.028	1.153	−2.191	–*	–
P (mm)	0.072	5.341	1.012	–	–
Winter					
T (°C)	0.032	1.662	2.176	–*	–
Rh (%)	0.069	3.562	2.111	–*	–
U (ms^{-1})	−0.015	0.180	−6.080	–**	1983
R_a ($MJ\,m^{-2}\,d^{-1}$)	−0.032	1.168	−2.907	–**	1981
P (mm)	0.054	2.688	2.186	–*	–
Whole year					
T (°C)	0.028	0.650	5.304	–**	1986
Rh (%)	0.054	1.942	2.992	–**	1987
U (ms^{-1})	−0.019	0.334	−6.87	–**	1983
R_a ($MJ\,m^{-2}\,d^{-1}$)	−0.011	0.360	−3.56	–**	1972
P (mm)	0.522	21.150	2.887	–**	1987

Notes: Std mean standard error of linear regression. The positive value in the table means upward trend and negative value means downward trend. – Means significance level exceeds 0.05.
* Indicate significance levels of 0.05.
** Indicate significance levels of 0.01.

3.2. Change trend of ET_0 and AI

3.2.1. Seasonal cycles of regional ET_0

As expected, ET_0 exhibited strong seasonal fluctuations since it are affected by many meteorological variables (Fig. 2). Its seasonal cycles were similar to P, although ET_0 is much higher than P in all months. ET_0 was greater than 80 mm/month during April–September and not high during December and January. In July, ET_0 reached a maximum with a value of 150 mm. The ET_0 during April–September accounted for 75% of the annual total ET_0, and the value was about 60% of the growing season.

Fig. 2 showed that *AI* also had a significant seasonal change. In the colder December and January, *AI* was lower than 0.52, while *AI* was higher than 0.7 from February to November. In particular, *AI* was higher than 0.85 from March to October indicating an arid climate. This means ET_0 was far higher than P in these seasons. The highest *AI* appeared in May resulting in extreme drought at crop seeding stage in this region.

3.2.2. Annual change of ET_0 and AI series

The K–M test showed that annual ET_0 and *AI* at 16 stations have a decreasing trend (Table 3). Linear regression showed that average ET_0 and *AI* over the study area have witnessed a declining trend for every month. In order to test the significance of the trend, M–K test was performed and the results are presented in Figs. 3 and 4. It is seen that the trend shows an obvious seasonal change. Fig. 4 had clearly shows that the decreasing trend of ET_0 was significant at the 5% level from May to October, even at 1% from June to October.

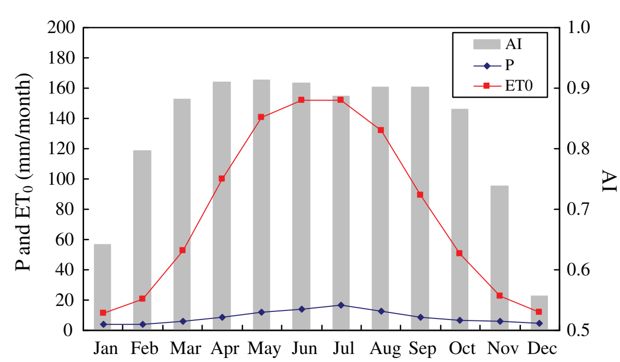

Fig. 2. Seasonal cycles of ET_0, P and AI averaged in 1922–2008.

Linear regression results indicated that the ET_0 averagely reduced 0.226–0.454 mm per month from May to October in the past 54 years. But from January to April and November to December, the decrease trends were not significant according to the M–K test. McVicar et al. (2012) reviewed the global study of trend in ET_0 in past several decades and declining trends were found in lots regions. In detail, Jhajharia et al. (2012) found ET_0 decreased significantly at annual and seasonal time scales during the last 22 years in NE India and NE India as a whole. However, the change trends of ET_0 are different among different stations over Iran (Dinpashoh et al., 2011; Hossein et al., 2011).

Similarly, monthly *AI* had also a decreasing trend in the past 54 years. However, the M–K test showed that the trend was

Table 3
Test results of trend and change point for ET_0 and AI.

	ET_0			AI		
	M–K test		Change point (year)	M–K test		Change point (year)
	Z_c	Significance level		Z_c	Significance level	
Aletai	−3.05	–**	1981	−2.082	–*	1982*
Fuyun	3.23	–**	1970	−3.723	–**	1983**
Hebusaike	−3.59	–**	1986	−1.35	–	–
Kelamayi	−3.2	–**	1986	−1.35	–	1986*
Jinghe	−3.22	–**	1986	−3.081	–**	1979**
Qitai	−4.33	–**	1986	−2.231	–*	1983*
Yining	−1.16		–	−2.455	–*	1984*
Wulumuqi	−0.9		1978	−2.648	–**	1980**
Tulufan	−6.39	–**	1984	−1.977	–*	–
Kuke	−7.53	–**	1986	−4.379	–**	1985*
Kashi	−2.22	–*	1979	−1.619	–	–
Bachu	−1.38		–	−1.992	–*	–
Tielikegan	−3.87	–**	1986	−2.141	–*	–
Ruoqiang	−4.93	–**	1986	−4.379	–**	1980**
Shache	−4.57	–**	1984	−1.44	–	–
Hetian	1.01		–	−0.754	–	–
Hami	−7.14	–**	1985	−4.126	–**	1983**
Donghuang	−2.07	–*	1981	−2.32	–*	1968*
Yumen	−6.15	–**	1985	−2.917	–**	–
Jiuquan	−0.92		–	−1.231	–	–
Minqin	−0.4		–	−0.828	–	–
Jilantai	−3.04	–**	1989	−0.619	–	–
Geermu	−1.26		1995	−2.141	–*	–

Notes: The positive value in the table means upward trend and negative value means downward trend. – Means significance level exceeds 0.05.
* Indicate significance levels of 0.05.
** Indicate significance levels of 0.01.

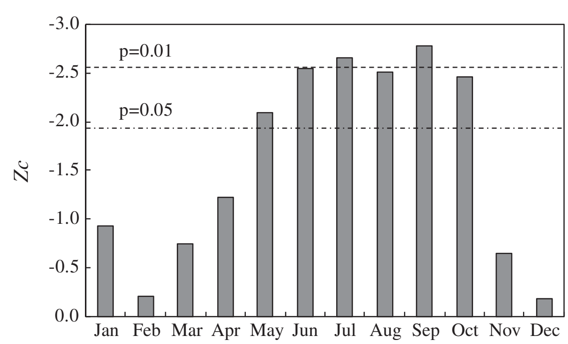

Fig. 3. M–K test results of monthly ET_0.

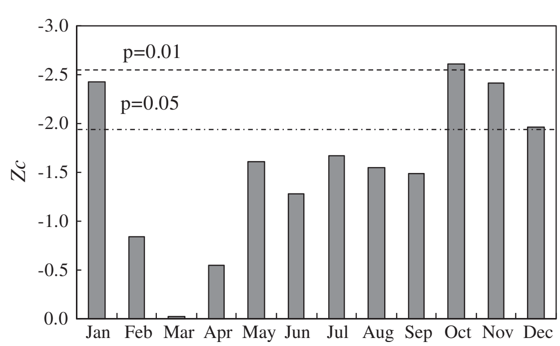

Fig. 4. M–K test result for monthly *AI*.

significant ($p < 0.05$) only in January, October, November and December (Fig. 4). It was found that *AI* declined by an average 0.0015–0.0052 per year in the 4 months and the decline was much higher in January and December than in the other 2 months.

In addition, temporal trends of ET_0 and *AI* in four seasons were analyzed with linear regression, M–K test and Pettit test method (Tables 4 and 5). The results showed ET_0 in all four seasons had decreasing trends. The linear regressions indicate that the reducing trends in ET_0 in warmer seasons (summer and autumn) were higher than in cooler seasons (spring and winter). The average decrease in ET_0 was 0.409 and 0.539 mm per year in summer and autumn, and 0.056 and 0.088 mm in spring and winter, respectively. Furthermore, the M–K test indicated that the trends in all seasons were significant ($p < 0.01$). The Pettit test revealed that the trend of seasonal ET_0 series had a change point at 1983 except in spring.

As for *AI* in the four seasons, a decreasing trend also was found, but being different from the trend of ET_0, the reduction extent was slight in warmer seasons with 0.001 per year (Table 5). In contrast, the decrease value of *AI* in colder seasons was higher with 0.002–0.003 per year, which was attributed to the difference of change trend of *P* in different seasons. A M–K test revealed that reduction trend of *AI* was significant in spring and summer ($p < 0.05$) and in autumn and winter ($p < 0.01$). Change points of *AI* seasonal series also were found in 1985 for summer and autumn and in 1982 for winter.

Fig. 5 shows the trends in ET_0 and *AI* on an annual basis, linear regression indicated that both ET_0 and *AI* had a decreasing trend. On average, annual total ET_0 declined 3.091 mm per year in the past 54 years. The rate of decline for annual mean *AI* was 0.001 per year. M–K analysis proved the trend for annual ET_0 and *AI* were significant ($p < 0.01$), however, no change point was found in the annual ET_0 and *AI* series.

The ET_0 variation was influenced by both of the radiometric and aerodynamic variables which reflected the combined effect of topography, soil, vegetation and climate (McVicar et al., 2007b). The trend in changes in ET_0 varied between regions. Liang et al. (2010) found ET_0 had an increasing trend in the upper reaches, but a decreasing trend in the lower reaches in humid Taoer River basin of Northeast China. Tabari (2010) revealed that annual average ET_0 increased by 2.3–11.28 mm in the arid western side of Iran.

Table 4
Temporal trend analysis of seasonal ET_0 with linear regression and M–K analysis.

	Linear regression		M–K analysis		Change point
	Slope (mm/year)	Std (mm/year)	Z_c	Significance level	
Spring	−0.056	2.638	−2.096	–*	–
Summer	−0.409	9.578	−5.394	–**	1983
Autumn	−0.539	10.658	−6.394	–**	1983
Winter	−0.088	2.818	−3.395	–**	1983
Annual	−3.091	64.720	−5.901	–**	–

Note: Std mean standard error of linear regression. The positive value in the table means upward trend and negative value means downward trend. – Means significance level exceeds 0.05.
* Indicate significance levels of 0.05.
** Indicate significance levels of 0.01.

Table 5
Temporal trend analysis of seasonal AI with linear regression and M–K analysis.

	Linear regression		M–K analysis		Change point
	Slope (year)	Std (year)	Z_c	Significance level	
Spring	−0.002	0.095	−2.216	–*	1985
Summer	−0.001	0.031	−1.873	–	–
Autumn	−0.001	0.033	−3.186	–**	1985
Winter	−0.003	0.106	−3.126	–**	1982
Annual	−0.001	0.029	−3.409	–**	–

Note: Std mean standard error of linear regression. The positive value in the table means upward trend and negative value means downward trend. – Means significance level exceeds 0.05.
* Indicate significance levels of 0.05.
** Indicate significance levels of 0.01.

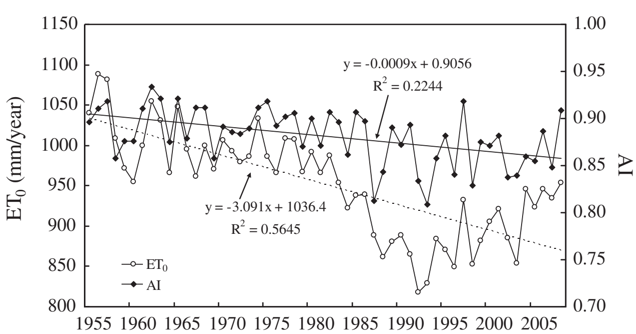

Fig. 5. Long term trend of ET_0 and AI in the study area.

Significant declining trends were found for ET_0 and AI in Northwest China in this study. As Liang et al. (2010) summarized, explanations for causes of temporal trends for ET_0 or pan evaporation mainly lie in: (1) declining global solar irradiance (Roderick and Farquhar, 2002; Roderick et al., 2007); and (2) decreasing wind speed (Xu et al., 2006; Rayner, 2007; McVicar et al., 2008). In the present study, T had a significant increasing trend, but U_2 and R_n declined in the past several decades. Moreover, annual mean Rh increased noticeably within the past 50 years although the trend was not significant in some seasons. The increase in ET_0 induced by rising T can be compensated by reduced ET_0 with decrease of U_2 and R_n and increase of Rh. As a result, ET_0 appeared to show a declining trend, which was consistent with the study in India (Chattopadhyay and Hulme, 1997) and in the Yangtze River basin in China (Xu et al., 2006).

3.3. Sensitivity of ET_0 and AI to key meteorological variables

Sensitivity coefficients for each meteorological variable and for the quarter periods are summarized in Table 6. These coefficients represent the average value for each meteorological variable and analyzed period. During summer months, higher temperatures have a larger effect on ET_0, while the effect is lower in spring and winter months. Rh has a negative effect on ET_0 and the effect is larger in cooler seasons. Furthermore, Rh has the strongest influence on ET_0 in spring for the study area ($Se = -0.509$). In the other months, U_2 has the most influence. Moreover, the sensitivity coefficient is higher in autumn months. This can be contributed to the higher wind speed at these months in the study area. As for the R_n, the sensitivity is lower relative to the other variables and the variations were small within the four seasons. On a full year basis, the variables with most influence on ET_0 are U_2 and Rh with T and R_n have some influence.

We analyzed the sensitivity of AI to meteorological variables (Table 7). The results revealed that relative influence of the four

Table 6
Sensitivity coefficients for T, Rh, U_2 and R_a to ET_0 in different seasons and full year.

	Spring (January, February, March)	Summer (April, May, June)	Autumn (July, August, September)	Winter (October, November, December)	Full year
T	0.381	0.513	0.386	0.339	0.405
Rh	−0.509	−0.478	−0.371	−0.508	−0.467
U_2	0.312	0.589	0.633	0.539	0.518
R_a	0.345	0.366	0.326	0.317	0.339

Table 7
Sensitivity coefficients for T, Rh, U_2, R_n, P and ET_0 to AI in different seasons and full year.

	Spring (January, February, March)	Summer (April, May, June)	Autumn (July, August, September)	Winter (October, November, December)	Full year
T	0.116	0.354	0.206	0.157	0.208
Rh	−0.470	−0.240	−0.200	−0.522	−0.358
U_2	0.476	0.500	0.558	0.544	0.520
R_n	0.134	0.219	0.282	0.104	0.185
P	−0.648	−0.818	−0.863	−0.684	−0.753

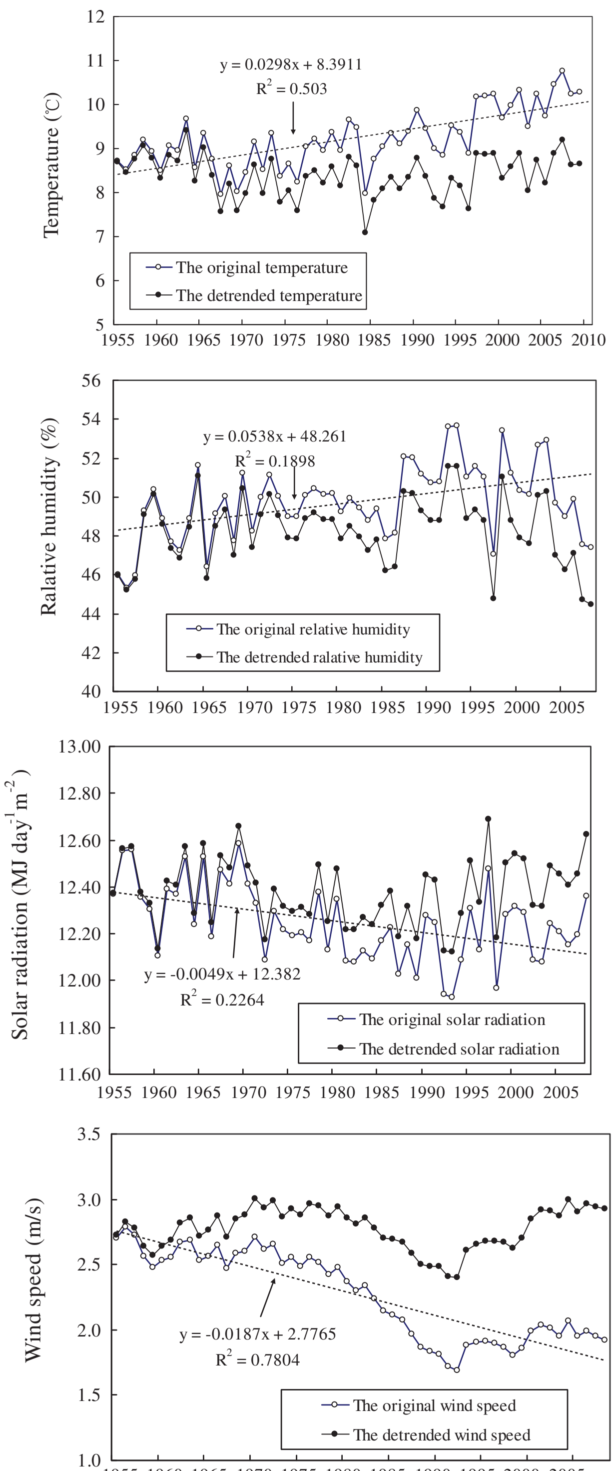

Fig. 6. The plot of the original annual values, the trend and the recovered stationary series for T, Rh, U and R_a.

meteorological variables on AI is the same as on ET_0, although the absolute values of the sensitivity coefficients are different from those for ET_0. On a full year basis, the most sensitive variable is U_2. This means a higher U_2 possibly is also a reason of climatic drought in the study area. As we noticed, T and R_n are the most influential climate factor affecting AI in summer and autumn but Rh is the most influential factor in spring and winter. The sensitivity of AI to P also was analyzed according the method used in this study. As expected, P has a higher effect on AI than other four meteorological variables. The sensitivity coefficients are larger in warmer seasons where there is much precipitation than in cooler seasons. For the four meteorological variables determining ET_0, U_2 has a greater effect on AI.

The sensitivity of ET_0 to meteorological variables varies with the climate characteristics in each of the study regions. Gong et al. (2006) found that lower sensitivity coefficients for U_2 throughout the year in humid Yangtze River basin of China. In the warmer seasons, but, the response of ET_0 to meteorological variables is different. Estévez et al. (2009) reported that ET_0 is sensitive to T and R_n in warmer season but U_2 is the sensitive factor in cooler seasons in southern Spain with a semi-arid climate. In the present study, we conclude that ET_0 is very sensitive to U_2 and Rh in arid northwest China. The greater impact of wind as compared to that of solar radiation on ET_0 is in agreement with results from Irmak et al. (2006), and it can be explained by the lower amount of water vapor carried by the wind in drier climates as compared to the higher humidity of the wind flow in humid climates.

3.4. Contribution of climate change to ET_0 and AI

The original and detrended data series for T, Rh, U_2, R_n over the study area are shown in Fig. 6. The obvious difference can be seen from original and detrended data series for the four meteorological variables. With adverse trends obtained for P, T and Rh, and negative trends for U_2 and R_n, the detrended data series were lower than the original data series for P, T and Rh with adverse trend obtained for R_n and U_2.

Detrended T, Rh, U_2, and R_n series respectively were used to compute the ET_0 series (Fig. 7). The results indicate that the difference between the original annual ET_0 series and the recalculated ET_0 series with the respectively detrended T, Rh, U_2, and R_n. The results indicated that recalculated ET_0 with detrended T was lower than that obtained from the original meteorological variables. Conversely, the recalculated ET_0 with detrended Rh, U_2, and R_n was larger than that calculated from original meteorological variables. Furthermore, larger differences in the increasing trend were found among recalculated ET_0 series with detrended U_2 than that with detrended R_n or Rh. In addition, a remarkable difference is obtained between the original ET_0 and the recalculated ET_0 with detrended T and the difference is larger than that between original ET_0 and recalculated ET_0 with detrended R_n or Rh. This means that the decline in the U_2 is the main cause of the decreasing ET_0 trend, whereas the increasing trend in Rh and the decreasing trend in R_a makes a much smaller contribution. From the results presented, it is obvious that the temporal trend of ET_0 reflects a combined effect of the meteorological variables. In our study, the increasing air

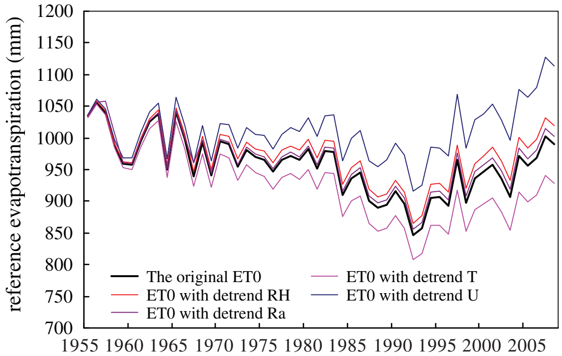

Fig. 7. The comparison of the original mean annual reference evapotranspiration with the recalculated evapotranspiration with respectively detrend *T*, *Rh*, *U* and R_a.

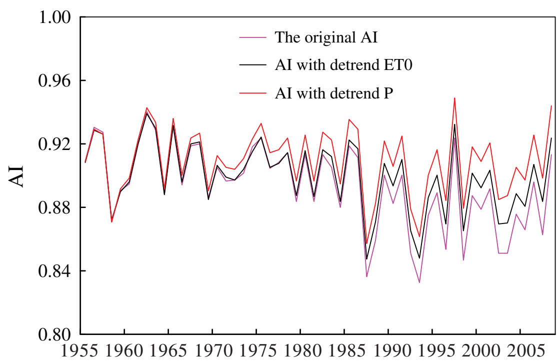

Fig. 8. The comparison of the original mean annual *AI* with the recalculated *AI* with respectively detrend *P* and ET_0.

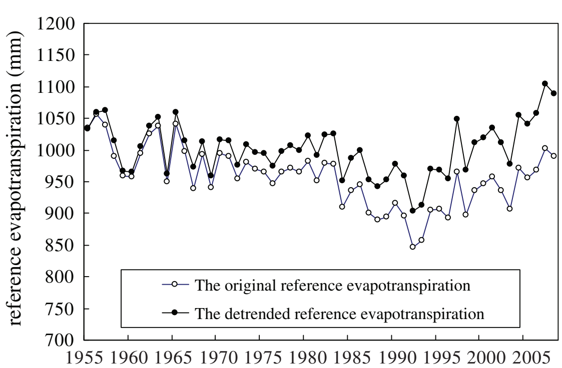

Fig. 9. The comparison of the original mean annual ET_0 with the recalculated *AI* with detrend *T*, *Rh*, *U* and R_a.

temperature or relative humidity was not the main cause of ET_0 change. Furthermore, this result also reveals that the decreasing trend in U_2 is the main cause of the decreasing trend of ET_0, while the increasing trend in *T* contributes to much a smaller extent.

The contribution of meteorological variables to ET_0 change was determined by studying the sensitivity of ET_0 and its changing trend to these variables. Liu et al. (2010) suggested that increasing *T* is the most important cause of the increasing ET_0 in the Yellow River basin of China, because, not only was it one of the variables influencing ET_0, but it also it a variable with significant increasing trend in the whole basin. In their study, although ET_0 is more sensitive to *Rh* than to *T*, the contribution of *Rh* to increasing ET_0 is less than that of *T* as it has no significant decreasing trend. We also draw a similar conclusion in the present study. U_2 is the main variable contributing to decreasing ET_0 due to its sensitivity to ET_0 and significant change trend. McVicar et al. (2012) suggested that U_2 cannot be neglected when explaining the change trend of ET_0. However, the contribution of *Rh* to change of ET_0 are smaller than that of *T*, although the sensitivity coefficient of *Rh* is higher than that of *T*.

Although not investigated in this study, the contributions of *T*, *Rh*, U_2, and R_a change to decrease of *AI* are similar to their effects on decreasing ET_0 because of its simple relationship with *AI*. This means a decrease in U_2 has a higher contribution than change of *T*, *Rh* and R_a. To investigate combined effect of *T*, *Rh*, U_2 and R_n on *AI*, detrended annual *AI* were calculated with detrended ET_0 (Fig. 8). At the same time, the detrended *AI* with detrended *P* are also shown in the figure. It is noted that a decrease in ET_0 has an expected contribution to the decline in *AI*. However, the contribution of increasing *P* is larger. This can be attributed to *AI* being more sensitive to *P* than to ET_0 on a yearly basis. Although ET_0 also has significant declining trend, its effect on *AI* is smaller than that of *P* in the past 50 years in the arid northwest China.

To understand the combined effect of *T*, *Rh*, U_2 and R_n on ET_0 over the past 50 years, we calculated an ET_0 series using all four detrended meteorological variables (Fig. 9). This shows a remarkable increasing divergence between the original and recalculated ET_0. The magnitude of change in ET_0 induced by climate change in past 50 years was estimated by comparing the original and recalculated ET_0 (Table 8). In the period of 1955–1989, the difference between original and the detrended ET_0 is slight, with a range of −0.23% to 5.25%. This can be attributed to a negligible climate change effect in this period. Since 1990s, however, the difference between the original and the detrended ET_0 is greater. Annual ET_0 decreased on average by 100.45 mm and 124.37 mm from the corresponding value of the detrended series respectively in 1990s and 2000s. This means that ET_0 declined by 10.38–11.92% compared to the detrended value as climate change in recent 20 years. Similarly, annual *AI* was recalculated with detrended *T*, *Rh*, U_2, R_n and *P* to investigate the effect of climate change on *AI* (Fig. 10). As a result of the declining and the rising trend respectively for ET_0 and *P*, *AI* reduced by 0.019–0.05 from 1955 to 2008 and subsequently the reduction ratio with detrend *AI* is 2.07–5.44%. These reflect that there is an alleviation trend for aridity in the study area in the recent 54 years. This shows that climatic drought has an alleviative trend in arid northwest China in past

Table 8
Changes of ET_0 and *AI* between original and detrend values during different years.

	ET_0				*AI*			
	Original (mm/year)	Detrend (mm/year)	Changed (mm/year)	Change ratio (%)	Original	Detrend	Changed	Change ratio (%)
1955–1979	1004.23	1006.58	−2.35	−0.23	0.896	0.915	−0.019	−2.07
1980–1989	932.87	984.61	−51.74	−5.25	0.879	0.914	−0.035	−3.84
1990–1999	867.36	967.82	−100.45	−10.38	0.861	0.908	−0.047	−5.2
2000–2008	918.78	1043.15	−124.37	−11.92	0.866	0.916	−0.05	−5.44

Note: Change ratios were computed by changed value dividing original value.

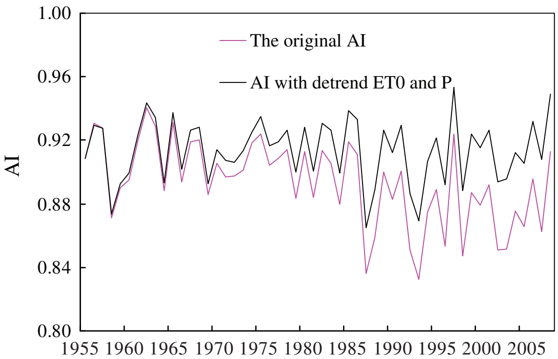

Fig. 10. The comparison of the original mean annual *AI* with the recalculated *AI* with detrend *P* and ET_0.

50 years although climate is becoming warmer and warmer. This is a good evidence for agriculture practice.

4. Conclusions

In this study, we presented a comprehensive analysis of the effect of climate change on ET_0 and *AI* in terms of their temporal variation during 1955–2008 in an arid inland region of northwest China. The possible causes and quantitative contributions of the major meteorological variables to the ET_0 and *AI* trends were also investigated. Annual temperature, humidity and precipitation have significant increasing trends and decreasing trends were found for wind speed and radiation although the seasonal trend varies. As a result, ET_0 has a significant declining trend with an average decrease of about 3 mm per year. *AI* has also a decreasing trend over the past 50 years. ET_0 is most sensitive to wind speed, followed by relative humidity, temperature and solar radiation. However, *AI* is more sensitive to *P* than is ET_0. Both ET_0 and *AI* decrease due to the combined effect of these meteorological variables. The contribution of the decline in wind speed to decreasing ET_0 is larger than for other meteorological variables. Increasing precipitation contributes more to decline in *AI* than it does to the decrease of ET_0 in the past 50 years.

In general, the results of this study showed climate drought has been alleviated in recent several decades in arid region of northwest China. These suggest a decrease in irrigation water demand for agricultural crops in northwest China. The finding is helpful for planning and efficient use of agricultural water resources. Results of this research need to be verified in other climatic conditions; especially in arid climates where ET_0 variations are crucial for water management. The authors regret that spatial variation of the contribution of climate change to ET_0 and climatic drought could not be investigated here because of the shortage of meteorological observation stations in the region. In future, we are excited to investigate this variation if data from more stations becomes available. Furthermore, this larger inland area can be divided into several sub-areas according to the climate character and ET_0 variation and its contribution can be analyzed for every sub-area.

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