Research Article

Reference Evapotranspiration Changes: Sensitivities to and Contributions of Meteorological Factors in the Heihe River Basin of Northwestern China (1961–2014)

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This paper investigates reference evapotranspiration (ET_0) changes, sensitivities to and contributions of meteorological factors in the Heihe River Basin (arid and inland region). Results show that annual ET_0 over the whole basin has increasing trend ($2.01 \text{ mm} \cdot 10 \text{ yr}^{-2}$) and there are significant increasing spatial variations from the upper (753 mm yr^{-1}) to the lower (1553 mm yr^{-1}) regions. Sensitivity analysis indicates that relative humidity is the most sensitive factor for seasonal and annual ET_0 change, and the influence is negative. The sensitivity of minimum temperature is the weakest and negative. Contribution analysis shows that the main contributors to ET_0 changes are aerodynamic factors rather than radiative factors. This study could be helpful to understand the response of ecoenvironment to the meteorological factors changes in the Heihe River Basin.

1. Introduction

Evapotranspiration is an excellent indicator of hydroclimatic change and the response of water management, food security, and ecoenvironment [1]. Among different evapotranspiration terms, such as actual evapotranspiration (ET_a), potential evapotranspiration (ET_p), pan evaporation (E_{pan}), and reference evapotranspiration (ET_0), E_{pan} and ET_0 are often used as surrogates of ET_p to reflect the evaporation capability in a specific region. Because ET_p and ET_0 are dependent only on meteorological condition not underlying surface and are measurable or calculable, they are important hydroclimatic indicators for reflecting regional water-energy balance changes and the effect of climate change. Spatiotemporal variations of ET_0 in different climatic regions have been globally reported over the past decades [2, 3]. Many regions have experienced significant decreasing trends of E_{pan} or ET_0 , such as the US [4], China [5], Canada [6], Australia [7], India [8], Japan [9], and Romania [10]. However, ET_0 changes with significant positive trends have been reported

in other regions, such as the Mediterranean region [11], Iran [12], Spain [13], and Serbia [14]. Moreover, the interannual fluctuations of ET_0 for some regions are very significant; ET_0 may increase during one period but decrease during the next period [5, 7]. Therefore, the temporal variations of ET_0 are complex and diverse in different climatic zones. Reasons for the different temporal variations of ET_0 in different climatic zones need to be explored in further detail.

The causes of ET_0 changes in many regions have been studied. First, the effects of different methods on ET_0 changes have been discussed in different climatic zones. Popular methods for ET_0 calculation mainly include FAO P-M, Priestley-Taylor, Hargreaves, Makkink, Blaney-Criddle, and Samani-Hargreaves methods [15]. Comparisons showed that FAO P-M performs better among the different methods due to having the clear physical meaning and is recommended as a standard method for the ET_0 calculation [16, 17]. Second, a sensitivity coefficient was used to investigate the effects of meteorological factors on ET_0 change [18–20]. Most studies showed that aerodynamic factors are the major factors in

different regions. For example, air temperature, wind speed, and relative humidity have stronger effects on ET_0 change in Spain [20]. Air temperature and wind speed are the dominant variables influencing ET_0 in Iran [12]. Air temperature is the most sensitive variable to ET_0 change in India [21]. Similar results have also been found in some regions of China, in which wind speed, air temperature, and vapor pressure deficit are the major sensitive factors for ET_0 change in such areas as the Loess Plateau Region [22], the Liaohe delta [23], the Tibet Plateau [24], the Changjiang River Basin [19], and the Haihe River Basin [25]. Some studies proposed a close agreement between changes in ET_0 and solar energy in Greece [26], Korea [27], and the Yellow River Basin [28]. Sensitivity analysis could only describe the responses of ET_0 to changes in individual factor. However, it cannot determine how much the impact of each meteorological factor on ET_0 change is.

The Heihe River basin (HRB), the second largest inland river basin in northwestern China, consists of three regions with different landscapes and climate conditions, where the upper mountainous region is semiarid and natural with little human interference, the middle region is dry and intensively irrigated plain, and the lower region is an extremely dry Gobi desert plain. The spatial variation of ET_0 in such basin may supply more information of regional response to the climate. The previous studies only reported the spatiotemporal variations of ET_0 [29, 30] at a given period, but there is no common understanding of ET_0 change so far due to different data time series. The aim of this paper is to clarify the effect of meteorological factors on ET_0 change by comprehensively analyzing the sensitivity of ET_0 change and contributions of meteorological factors in the HRB using reliable and complete daily meteorological data from 16 stations for the period 1961–2014. This paper will determine (1) the spatial pattern and temporal trends of ET_0 for the HRB, (2) the sensitivity of ET_0 to meteorological factors, and (3) the contributions of the meteorological factors to ET_0 change.

2. Study Area and Data

2.1. Study Area. As shown in Figure 1, the drainage map and the basin border of the HRB are extracted using a 90 m resolution digital elevation model (DEM) data from the Shuttle Radar Topography Mission (SRTM) website of the NASA (<http://srtm.csi.cgiar.org/SELECTION/inputCoord.asp>) (basin length: 820 km; total area: 143,000 km²; elevation: 870–5545 m).

The HRB is divided into three regions according to basin characteristics, shown in Figure 1. The upper mountainous region belongs to the cold and semiarid mountain zone with an elevation from 2000 to 5000 m, annual mean temperature of less than 2°C, pan evaporation of 700 mm yr⁻¹, and precipitation of 350–400 mm yr⁻¹. The middle region is the main irrigation zone and residential area with more than 90% of the total population of the basin; it has a precipitation of 100–250 mm yr⁻¹ and pan evaporation of 2000 mm yr⁻¹. The lower region is covered by the arid Gobi

FIGURE 1: Location of the HRB and spatial distribution of the meteorological and radiation stations. U, M, and L represent the upper, middle, and lower regions in the basin, respectively.

desert in the north of the basin with an elevation of 870–1500 m and is characterized by an extremely arid climate, with pan evaporation of 3500 mm yr⁻¹ and precipitation of 10–50 mm yr⁻¹.

2.2. Data. In this study, daily meteorological data of 16 stations from 1961 to 2014 in and around the HRB are available from the National Climatic Centre of the China Meteorological Administration. The three solar radiation stations correspond to the upper, the middle, and the lower region (Figure 1). The data set includes daily observations of atmospheric pressure, maximum and minimum air temperatures at 2 m height ($T_{\max}$, $T_{\min}$), relative humidity at 2 m height (RH), daily sunshine duration, pan evaporation measured using a metal pan, 20 cm in diameter and 10 cm high, installed 70 cm above the ground, and wind speed measured at 10 m height which was transformed to wind speed at 2 m height (WS) by the wind profile relationship from Chapter 3 of the FAO paper 56 [16]. In addition, the three radiation stations were used to calibrate the Ångström parameters of extraterrestrial radiation reaching the earth on clear days in the FAO P-M equation. The spatial patterns of the meteorological factors, ET_0 , and sensitivity coefficients were obtained by the inverse distance weight (IDW) interpolation method.

In this study, the four seasons of the HRB are defined as spring (from March to May), summer (from June to August), autumn (from September to November), and winter (from December to February).

3. Methodology

3.1. FAO Penman-Monteith Method. The Penman-Monteith method can be used globally to estimate potential evapotranspiration. Allen et al. simplified the Penman-Monteith equation and defined the hypothetical reference grass with an assumed height of 0.12 m, a fixed surface resistance of 70 s m^{-1} , and an albedo of 0.23 [16]. This method can provide good and reliable results for ET_0 because it is physically based and explicitly incorporates both physiological and aerodynamic parameters and has been accepted as a standard to compare evapotranspiration capability for various climatic regions [31]. Moreover, this method has been successfully applied across the whole of China [32, 33]. The FAO P-M for calculating daily ET_0 is described as

$$ET_0 = \frac{0.408\Delta(R_n - G) + \gamma(900/(T_{\text{mean}} + 273))u_2(e_s - e_a)}{\Delta + \gamma(1 + 0.34u_2)}, \quad (1)$$

where ET_0 is the reference evapotranspiration (mm day^{-1}), R_n is the net radiation at the crop surface ($\text{MJ m}^{-2} \text{ day}^{-1}$), G is the soil heat flux density ($\text{MJ m}^{-2} \text{ day}^{-1}$), T_{mean} is the mean daily air temperature at 2 m height ($^{\circ}\text{C}$), u_2 is the wind speed at 2 m height (m s^{-1}), e_s is the saturation vapor pressure (kPa), e_a is the actual vapor pressure (kPa), Δ is the slope vapor pressure curve ($\text{kPa}^{\circ}\text{C}^{-1}$), γ is the psychrometric constant ($\text{kPa}^{\circ}\text{C}^{-1}$); the atmospheric pressure used in this study is the measured value. More details regarding the data processing in (1) can be found in FAO paper 56.

In (1), the solar radiation (R_s) is obtained with the following Ångström formula:

$$R_s = \left(a + b \frac{n}{N}\right) R_a, \quad (2)$$

where R_s is the solar radiation ($\text{MJ m}^{-2} \text{ day}^{-1}$), n is the actual sunshine duration (hours), N is the maximum possible sunshine duration or daylight hours (hours), R_a is the extraterrestrial radiation ($\text{MJ m}^{-2} \text{ day}^{-1}$), and a and b are regression constants.

Because of the effects of the atmospheric conditions (humidity, dust) and solar declination (latitude and month) as well as the elevation variations, the Ångström values a and b in the HRB were calibrated using the observed radiation data at the three solar radiation stations (Figure 2).

3.2. Trend Analysis. The long-term trends and changes of ET_0 and meteorological factors are detected using the linear fitted method:

$$\hat{y} = \hat{a}t + \hat{b}, \quad (3)$$

where $\hat{y}$ is the fitted trend during a given period and $\hat{a}$ and $\hat{b}$ are the estimated regression slope and the regression constant, respectively. Positive slope indicates an increasing trend and negative slope indicates a decreasing trend.

For data sets without seasonality, the significance of a trend is described using the Mann-Kendall (MK) test method

[34, 35], which is to statistically assess if there is a monotonic trend of the variable of interest over time [36], whilst the Seasonal Kendall (SK) test is extension of the MK test and is suitable for trend applicable to data sets with seasonality, missing values, and serial correlation over time [37, 38]. The SK test begins by computing the MK test separately for each month or season and then summing the statistic S_i and variance $\text{Var}(S_i)$. Following Hirsch et al. [37], the entire sample X is made up of subsamples X_1 through X_{12} (one for each month), and each subsample X_i contains the n_i annual values from month i :

$$X = (X_1, X_2, \dots, X_{12}), \quad (4)$$

$$X_i = (X_{i1}, X_{i2}, \dots, X_{i12}).$$

The null hypothesis H_0 for the SK test is that the X is a sample of independent random variables (x_{ij}) and that each X_1 is a subsample of independent and identically distributed random variables over years. The alternative hypothesis H_1 is that for one or more months the subsample is not distributed identically over years.

According to the MK test, the statistic S_i is defined by

$$S_i = \sum_{k=1}^{n_i-1} \sum_{j=k+1}^{n_i} \text{sign}(x_{ij} - x_{ik}), \quad (5)$$

where

$$\text{sign}(\theta) = \begin{cases} 1, & \theta > 0, \\ 0, & \theta = 0, \\ -1, & \theta < 0. \end{cases} \quad (6)$$

Now, the subsample X_i satisfies the null hypothesis of Mann's test. Therefore relying on Mann and Kendall we have

$$E(S_i) = 0 \quad (7a)$$

$$\text{Var}(S_i)$$

$$= \frac{1}{18} \left[n_i(n_i - 1)(2n_i + 5) - \sum_{t_i}^{g_i} t_i(t_i - 1)(2t_i + 5) \right], \quad (7b)$$

where g_i is the number of tied groups for the i th month and t_{ip} is the number of data in the p th group for the i th month. S_i is normal in the limit as $n_i \rightarrow \infty$. The SK test statistic S is given by

$$S = \sum_{i=1}^m S_i, \quad (8)$$

where m is the number of months for which data have been obtained over years. The expectation and variance can be derived as follows:

$$E(S) = \sum_{i=1}^m E(S_i), \quad (9a)$$

$$\text{Var}(S) = \sum_{i=1}^m \text{Var}(S_i) + \sum_{i=1}^m \sum_{j=1}^m \text{cov}(S_i S_j), \quad (9b)$$

FIGURE 2: Calibration of the Ångström coefficients for the three radiation stations: (a) Gangcha station in the upper region, (b) Jiuquan station in the middle region, and (c) Ejina station in the lower region.

where S_i and S_j ($i \neq j$) are function of independent random variables, so $\text{cov}(S_i S_j) = 0$.

For $n_1 > 10$, the standard normal deviate Z is estimated by (10) to test the significance of trends:

$$Z = \begin{cases} \frac{(S-1)}{\sqrt{V(S)}}, & S > 0, \\ 0, & S = 0, \\ \frac{(S+1)}{\sqrt{V(S)}}, & S < 0. \end{cases} \quad (10)$$

For the SK test, the null hypothesis H_0 means that there is no monotonic trend over time; when $|Z| > Z_{1-\alpha/2}$, the original null hypothesis is rejected; this means that the trend of the time series is statistically significant. In this study, significance level of $\alpha = 0.1$ is employed.

3.3. Sensitivity Analysis. Saxton [18] and Smajstrla et al. [39] defined the sensitivity coefficient by drawing a curve of

the change of a dependent variable versus the changes of independent variables. For multifactor models (e.g., the FAO P-M), due to different dimensions and ranges of different factors, the ratios of ET_0 changes and factors changes cannot be compared. In addition, this approach could introduce errors to understand the response of model behaviors to the factors because of changing one of the factors but holding other factors stationary [27]. To avoid the above two disadvantages, the dimensionless sensitivity coefficient defined by the dimensionless partial derivative with respect to the independent factors is used in this study:

$$S(x_i) = \lim_{\Delta x_i \rightarrow 0} \left(\frac{\Delta ET_0 / ET_0}{\Delta x_i / x_i} \right) = \frac{\partial ET_0}{\partial x_i} \cdot \frac{x_i}{ET_0}, \quad (11)$$

where x_i is the i th meteorological factor and $S(x_i)$ is the dimensionless sensitivity coefficient of reference evapotranspiration related to x_i . Greve et al. [40] used this method to estimate the effects of variation in meteorological factors and

measurement error on evaporation change. If the sensitivity coefficient of a factor is positive (negative), ET_0 will increase (decrease) as the factor increases. The larger the absolute value of the sensitivity coefficient, the more ET_0 is sensitive to a factor.

In this study, the meteorological factors $T_{\max}$, $T_{\min}$, WS, RH, and R_s are chosen for sensitivity analysis. Sensitivity coefficients ($S_{T_{\max}}$, $S_{T_{\min}}$, S_{WS} , S_{RH} , and S_{R_s}) were calculated on a daily dataset. Monthly and annual average sensitivity coefficients were obtained by average daily values. Regional sensitivity coefficients were obtained by averaging station values.

3.4. Contribution Estimation. Although sensitivity coefficients can reflect the sensitivity of ET_0 change to the perturbation of a factor, it cannot describe the contribution of a factor change to ET_0 change. Because both of the sensitivity and changes in meteorological factors affected ET_0 change, an approach to integrating the sensitivity and changes of meteorological factors is proposed to quantify influence magnitude individual meteorological factors changes to the trends of ET_0 .

Mathematically, for the function $ET_0 = f(x_1, x_2, \dots, x_n)$, where $x_1, x_2, \dots, x_n$ are independent variables, the first order Taylor approximation of the dependent variable ET_0 in terms of the independent variables is expressed as

$$\Delta ET_0 = \sum \frac{\partial ET_0}{\partial x_i} \cdot \Delta x_i + \delta, \quad (12)$$

where ΔET_0 is the change of ET_0 during a period, x_i is the i th meteorological factor, Δx_i is the change of x_i during the same period, $\partial ET_0 / \partial x_i$ is the partial differential of ET_0 with respect to x_i , and δ is the Lagrange remainder.

If both sides of (12) are divided by $\overline{ET_0}$ (the average value of ET_0 during a period), (12) can be written as

$$\frac{\Delta ET_0}{\overline{ET_0}} = \sum \frac{\partial ET_0}{\partial x_i} \cdot \frac{\Delta x_i}{\overline{ET_0}} + \varepsilon, \quad (13)$$

where $\Delta ET_0 / \overline{ET_0}$ is the relative change of ET_0 during a given period; $\varepsilon = \delta / \overline{ET_0}$ is the error item, which can be neglected because of its small value.

The first term in the right side of equation is multiplied by $\bar{x}_i / \bar{x}_i$; (13) can be written as

$$\frac{\Delta ET_0}{\overline{ET_0}} = \sum \frac{\partial ET_0}{\partial x_i} \cdot \frac{\bar{x}_i}{\overline{ET_0}} \cdot \frac{\Delta x_i}{\bar{x}_i} + \varepsilon, \quad (14)$$

where $(\partial ET_0 / \partial x_i) \cdot (\bar{x}_i / \overline{ET_0})$ is the average sensitivity coefficient of factor x_i during a period, denoted as $\bar{S}_{xi}$. If we let $C(x_i) = \sum \bar{S}_{xi} \cdot (\Delta x_i / \bar{x}_i)$, (14) can be written as

$$\frac{\Delta ET_0}{\overline{ET_0}} \approx \sum C(x_i) \quad (15)$$

$C(x_i)$ is the relative change in ET_0 contributed by $T_{\max}$, $T_{\min}$, WS, RH, and R_s .

TABLE 1: Coefficient of determination of monthly E_{pan} and ET_0 for nine meteorological stations.

Station	U1	U2	U3	M1	M2	M3	M4	L1	L2
R^2	0.945	0.939	0.966	0.969	0.973	0.959	0.968	0.965	0.986

4. Results

4.1. Correlation of ET_0 and E_{pan} . The coefficients of determination R of monthly E_{pan} and ET_0 for different stations (Table 1) are between 0.939 and 0.986, which means that the monthly E_{pan} and ET_0 have a very close linear relationship in the HRB. Such a close linear relationship suggests that ET_0 can be a good estimation using the observed E_{pan} in the HRB if the regression coefficients are given. Moreover, Figures 3(a) and 3(b) show that monthly and annual E_{pan} and ET_0 both present good linearity. The monthly and annual R values are 0.967 and 0.906, respectively. And the correlation of monthly E_{pan} and ET_0 appears to be a strong seasonal characteristic and becomes less centralized from winter to summer.

4.2. Evolution and Spatial Pattern of ET_0 at Different Time Scales. Figure 4 shows the average monthly ET_0 change during a year for the whole basin during 1961 and 2014. The mean monthly ET_0 is 97.8 mm month⁻¹ over the whole basin in the last 50 years. Monthly ET_0 first increases and then decreases during a year. The peak value occurs in June and July, approximately 177 mm month⁻¹, whereas the bottom values occur during November and February and are less than 50 mm month⁻¹. This strong monthly variation has a similar shape feature to the natural change in temperature and solar radiation (Figures 4 and 7). In addition, ET_0 in summer months differs more dramatically than that in winter months. And the difference between the maximum and the minimum of ET_0 reaches 50 mm in July, whereas the difference in December is only 10 mm. The evaporation capability in summer months accounts for 44% of annual ET_0 .

Figure 5 shows the trends of annual and seasonal ET_0 for the whole basin from 1961 to 2014. The mean annual ET_0 is 1175 mm yr⁻¹. The increasing trend of annual ET_0 is 2.01 mm·10 yr⁻² over the 54 years and has no statistical significance. Annual ET_0 variations exhibit three different phases, which has a significant increasing trend during 1961–1974 and 1997–2014 but clearly decreases during 1975–1996 at 0.05 levels (Figure 5(e)). The 1961–2014 means of ET_0 from spring to winter are 363 mm yr⁻¹, 511 mm yr⁻¹, 220 mm yr⁻¹, and 81.4 mm yr⁻¹, respectively. The climatic trends of ET_0 in spring and winter are 2.07 mm·10 yr⁻² and 0.52 mm·10 yr⁻², respectively, whereas ET_0 changes in summer and winter have decreasing trends of -0.7 mm·10 yr⁻² and -0.06 mm·10 yr⁻².

Table 2 reports the mean values and trends of seasonal and annual ET_0 in the three subregions from 1961 to 2014. The seasonal and annual ET_0 have gradually increasing spatial gradients from the upper region to the lower region. The mean annual ET_0 of the upper, middle, and lower regions are 902 mm yr⁻¹, 1051 mm yr⁻¹, and 1289 mm yr⁻¹,

FIGURE 3: Relationship between E_{pan} and ET_0 at monthly (a) and annual (b) scales for nine meteorological stations from 1961 to 2001.

FIGURE 4: Box and whisker plots of monthly ET_0 of the HRB from 1961 to 2014. The line inside the boxes represents the median, and the upper and lower lines of the boxes indicate the 75th and 25th percentiles, respectively. The upper and lower parts of the whiskers indicate the maximum and the minimum of monthly ET_0 , respectively.

respectively. The ET_0 change in the upper region appears to be a statistically increasing trend at $6.61 \text{ mm} \cdot 10 \text{ yr}^{-2}$. The climatic trends of annual ET_0 in the middle and lower regions are $2.25 \text{ mm} \cdot 10 \text{ yr}^{-2}$ and $0.91 \text{ mm} \cdot 10 \text{ yr}^{-2}$, respectively, without statistical significance.

The maximum and minimum values of seasonal ET_0 consistently occur in summer and winter, respectively, for the three regions. Whereas the seasonal ET_0 trends are different, ET_0 for the upper region has significant increasing trends in spring, autumn, and winter, with increasing rates of 2.41, 1.19, and $1.54 \text{ mm} \cdot 10 \text{ yr}^{-2}$, respectively. Seasonal ET_0 has no significant trend for the middle and lower regions.

The spatial patterns of seasonal and annual ET_0 in the HRB from 1961 to 2014 are plotted in Figure 6. There are clear spatial gradients for annual ET_0 from the upper region to the lower region. The maximum occurs in the lower region and is up to 1553 mm yr^{-1} near station L2, and the minimum is

found in the upper region and is as low as 757 mm yr^{-1} near station U2 in the upper region.

The spatial variation of seasonal ET_0 is smaller than that of annual ET_0 . The ET_0 changes in spring, summer, and autumn have similar spatial features. The ET_0 changes only in summer have a clear spatial pattern, ranging from 300 mm yr^{-1} to 700 mm yr^{-1} over the whole basin. Variations of ET_0 in the other three seasons have very small spatial gradients across the whole basin. The spatial difference in ET_0 in spring is between 232 mm yr^{-1} and 472 mm yr^{-1} , with a SD of 49 mm yr^{-1} , and the ET_0 variation in the autumn ranges from 145 mm yr^{-1} to 290 mm yr^{-1} , with a SD of 30 mm yr^{-1} . The spatial distribution of ET_0 in winter varies little, and its SD is only 5.8 mm yr^{-1} over the whole basin.

4.3. Trends in Meteorological Factors. According to the FAO P-M method described in (1), T_{max} , T_{min} , WS, RH, and R_s

FIGURE 5: Annual and seasonal ET_0 trends for the HRB during 1961 and 2014. NS means not significant at the level of $\alpha < 0.05$ by the MK test.

TABLE 2: Means of seasonal and annual ET_0 and their trends in the three subregions during 1961–2014.

Region		Annual	Spring	Summer	Autumn	Winter
Upper region	Mean (mm yr ⁻¹)	902	280	380	171	71.1
	Trend (mm·10 yr ⁻²)	6.61*	2.41*	1.33	1.19*	1.54*
Middle region	Mean (mm yr ⁻¹)	1051	330	446	196	78.6
	Trend (mm·10 yr ⁻²)	2.25	2.03	-0.33	-0.21	0.58
Lower region	Mean (mm yr ⁻¹)	1289	395	568	241	84.8
	Trend (mm·10 yr ⁻²)	0.91	2.02	-1.31	-0.26	0.31

Note: * means the significance level of 0.1.

FIGURE 6: Spatial patterns of the seasonal and annual mean ET_0 in the whole HRB from 1961 to 2014. Max and Min denote the maximum and minimum values, respectively, of ET_0 over the whole basin. SD indicates the standard deviation of the spatial variations of ET_0 .

are selected as the major meteorological factors having an important influence on ET_0 . T_{mean} (the average of T_{max} and T_{min}) is a comprehensive indicator for analyzing temperature variation.

Figure 7 shows monthly variations of meteorological factors in the upper, middle, and lower regions and the whole basin during 1961 and 2014. The variations of monthly T_{mean} and R_s are similar to those of monthly ET_0 (Figure 4), and their peak values occur in the middle of the year, with a

minimum at the ends of the year. The air temperature in the upper region is the smallest over the whole basin, which ranges from $-12.6^{\circ}\text{C month}^{-1}$ to $12.9^{\circ}\text{C month}^{-1}$. Although the average monthly T_{mean} in the middle and lower regions are both $8.1^{\circ}\text{C month}^{-1}$, the maximum value of T_{mean} in the lower region is larger than that in the middle region, and the minimum value in the lower region is smaller than that in the middle region. Moreover, the standard deviation of monthly T_{mean} in the middle region is the smallest.

FIGURE 7: Monthly variations of meteorological factors (T_{mean} , WS, RH, and R_s) in the upper (U), middle (M), and lower (L) regions and the whole basin (W). The 54-year mean (solid line) and standard deviation (error bar) are shown.

The variation of wind speed during a year is relatively small. The peak of monthly WS occurs in April. There are similar variation features of WS for the three subregions. The monthly WS in the lower region is the largest, with an average of $2.8 \text{ m s}^{-1} \text{ month}^{-1}$ during the year, whereas that in the lower region is the smallest, with an average of $1.8 \text{ m s}^{-1} \text{ month}^{-1}$. The higher error bar means that the monthly WS has significant fluctuations during the year.

The monthly RH from the lower to the upper region gradually increases and the fluctuations of RH are also substantial. The monthly RH in the upper region increases at first and then decreases, and its peak is during June and August. The monthly RH in the middle and lower regions decreases at first and then increases, and its bottom is in April.

The monthly R_s in different regions have the same variation features and standard deviations during the year. The high value of monthly R_s is found during June and August, and the low value occurs in winter. The standard

deviation during May and August is larger than that from November to February.

Figure 8 shows trends of annual T_{mean} , WS, RH, and R_s for the upper, middle, and lower regions and the whole basin during 1961 and 2014. Positive trends of annual T_{mean} during 1961 and 2014 are detected in the upper, middle, and lower regions and the whole basin, with significant rates of change of $0.32^\circ\text{C}\cdot 10 \text{ yr}^{-2}$, $0.33^\circ\text{C}\cdot 10 \text{ yr}^{-2}$, $0.38^\circ\text{C}\cdot 10 \text{ yr}^{-2}$, and $0.36^\circ\text{C}\cdot 10 \text{ yr}^{-2}$, respectively.

The mean annual WS in the lower region is $2.5 \text{ m s}^{-1} \text{ yr}^{-1}$ and is larger than that in other regions. The interannual oscillations of annual WS for the middle and lower regions and the whole basin are similar and have three phases: two relatively steady periods from 1961 to 1968 and 1969 to 1974, followed by a long-term statistically significant decline phase from 1974 to the 1990 s. However, the trend of annual WS in the upper region has only a statistically significant decline phase from 1961 to 2014. There are significant decreasing

FIGURE 8: Trends of annual T_{mean} , WS , RH , and R_s for the upper (U), middle (M), and lower (L) region and the whole basin (W). S indicates that the trend is statistically significant, and NS indicates that the trend is not significant at the 0.05 level.

trends for the upper, middle, and lower regions, with change rates of $-0.13 \text{ m s}^{-1} \cdot 10 \text{ yr}^{-2}$, $-0.17 \text{ m s}^{-1} \cdot 10 \text{ yr}^{-2}$, and $-0.20 \text{ m s}^{-1} \cdot 10 \text{ yr}^{-2}$, respectively.

The mean annual RH in the lower, middle, and upper regions is $41\% \text{ yr}^{-1}$, $50\% \text{ yr}^{-1}$, and $52\% \text{ yr}^{-1}$, respectively. During 1961 and 2014, decreasing trends in the upper and middle regions are not statistically significant, whereas the changes in annual RH in the lower region and whole basin have significant decreasing trends. Therefore, the changes in RH across the whole basin are mainly affected by the trend of RH in the lower region.

The change of annual R_s for the different regions has no significant decreasing trend during the 54-year period, whereas the interannual oscillations of R_s are clearer than those of RH .

4.4. Variations of the Sensitivity Coefficients. Mean daily sensitivity coefficients for major meteorological factors that exhibit large fluctuations during a year (Figure 11). Although annual T_{max} and T_{min} have the same trend, the variations of $S_{T_{\text{max}}}$ and $S_{T_{\text{min}}}$ are different. $S_{T_{\text{max}}}$ gradually increases from negative to positive at first and then decreases from positive to negative and achieves a larger and stable peak value during May and August (Figure 9(a)). The daily variation patterns of $S_{T_{\text{min}}}$ have a unimodal distribution, and the peak occurs on the 200th day of the year (Figure 9(b)). $S_{T_{\text{max}}}$ and $S_{T_{\text{min}}}$ are positive during summer, and the former is larger than the latter. $S_{T_{\text{max}}}$ and $S_{T_{\text{min}}}$ are negative during winter days, and the latter is smaller than the former. Thus, ET_0 is sensitive to T_{max} in summer but T_{min} in winter. The value of $S_{T_{\text{max}}}$ is greater in the lower region than in the other two regions. $S_{T_{\text{min}}}$ for the

FIGURE 9: Mean daily sensitivity coefficients for maximum temperature (a), minimum temperature (b), wind speed (c), relative humidity (d), and shortwave radiation (e) in the upper (U), middle (M), and lower (L) regions of the HRB. (f) Comparison of the mean daily sensitivity coefficients for major meteorological factors in the whole basin.

middle and lower regions is almost the same and is greater than that in the upper region.

The values of S_{WS} in the three regions are positive throughout the year. ET_0 is most sensitive to WS in the beginning and end of a year but is insensitive to WS in summer (Figure 9(c)). The variation patterns of S_{WS} for the

three regions are the same. The values of S_{WS} for the three regions have significant differences during a year. The value in the lower region is the largest; thus, ET_0 is more sensitive to WS in the lower region than in the other two regions.

Relatively strong negative sensitivity coefficients were obtained for RH (Figure 9(d)). ET_0 is less sensitive to RH in

FIGURE 10: Interannual variations of sensitivity of ET_0 in relation to $T_{\max}$, $T_{\min}$, WS, RH, and R_s .

the winter for the upper region compared with the two other regions. However, ET_0 is more negative sensitive to RH in the lower region during April and September.

The daily variation patterns of S_{R_s} agree with those of shortwave radiation (Figure 9(e)). ET_0 is insensitive to R_s in winter, and S_{R_s} increases and achieves its maximum value in summer. The variations of S_{R_s} for the three regions show similar patterns, whereas S_{R_s} in the lower region is significantly less than that in the upper and middle regions. The variation of daily S_{R_s} and S_{WS} appears to be an opposite pattern during a year. Similar findings were reported by Gong et al. [19]. $S_{T_{\max}}$ and S_{R_s} have a similar variation pattern, whereas $S_{T_{\min}}$ and S_{WS} appear to have opposite patterns. RH is the most sensitive factor and WS and $T_{\min}$ are the least sensitive factors in the whole basin throughout the year.

Figure 10 shows the interannual variations of annual sensitivity coefficients from 1961 to 2014. The variation of annual S_{WS} has a significant increasing trend, whereas the absolute values of $S_{T_{\min}}$ and S_{RH} show that they have statistically significant decreasing trends during 1961 and

2014. ET_0 becomes more sensitive to WS but less sensitive to $T_{\min}$ and RH. The annual $S_{T_{\max}}$ and S_{R_s} have increasing and decreasing trends, respectively, but their trends are not statistically significant during the period of 1961–2014. This shows that the relative changes of the meteorological factors $T_{\min}$ and R_s and the relative change of ET_0 maintain a stable ratio [41].

Figure 11 describes the spatial patterns of the sensitivity coefficients of ET_0 to the major meteorological factors across the whole HRB. The mean annual values of sensitivity for $T_{\max}$, $T_{\min}$, WS, RH, and R_s are 0.28, -0.04 , 0.27, -0.38 , and 0.29 at the basin scale, respectively. RH is the most sensitive factor, and $T_{\min}$ is less sensitive to ET_0 over the whole basin. It seems that $S_{T_{\max}}$ and $S_{T_{\min}}$ have similar spatial patterns, whereas spatial distributions of the absolute values of $S_{T_{\max}}$ and $S_{T_{\min}}$ are opposite due to the negative sign of $S_{T_{\min}}$. Overall, there are three different spatial distributions for the five meteorological factors. (1) $S_{T_{\max}}$ and S_{WS} have a similar spatial pattern, increasing from the south to the north of the basin with significant spatial gradients. (2) The spatial patterns of

FIGURE 11: Spatial distribution of the mean annual sensitivity coefficients for ET_0 to the major meteorological factors (T_{max} , T_{min} , WS, RH, and R_s) during 1961–2014.

$S_{T_{min}}$ and S_{R_s} are similar, and the sensitivity for the two factors decreases from the upper region to the lower region. (3) The spatial variation of S_{RH} has no significant gradient from the lower region to the upper region.

4.5. Contribution of the Trends of the Meteorological Factors to That of ET_0 . The sensitivity coefficient describes the response of ET_0 to changes in meteorological factors but is not able to reflect change magnitude in ET_0 caused by meteorological factors. Namely, ET_0 change is strongly sensitive to a meteorological factor, but the meteorological factor must not cause a significant change in ET_0 . This is because, other than the sensitivity coefficients, changes in ET_0 are influenced by changes in meteorological factors as well. Consequently, (15)

is used to diagnose the contribution of meteorological factors to ET_0 changes.

As shown in Figure 12, the relative changes of monthly, seasonal, and annual ET_0 calculated using (15) well fit those of the actual ET_0 from observed data. This result illustrates that sensitivity coefficients and changes in meteorological factors could be used to analyze the contribution of one or more meteorological factors to ET_0 changes in the HRB.

Figure 13 shows the contributions of meteorological factor changes to relative changes in annual and seasonal ET_0 for the 9 stations in the HRB during 1961–2014. WS is the largest contributor to ET_0 change among meteorological factors in the middle and lower regions. The decreasing trends of WS cause ET_0 decreases, with relative changes in ET_0 of -3%

FIGURE 12: Relationship of $C(ET_0)$ to $TR(ET_0)$ at different time scales for all stations in the HRB. $C(ET_0)$ is the sum of the relative change of ET_0 contributed by changes in meteorological factors using (15). $TR(ET_0)$ is the long-term relative change of ET_0 from the observed data.

to -18% , corresponding to changes of -30 to -250 mm. However, T_{min} and R_s trends from 1961 to 2014 have little influence on the changes in ET_0 for the middle-lower regions. For the upper region, the trends of T_{max} and T_{min} for all stations significantly increase ET_0 , and relative changes of ET_0 are between 2.3% and 3.2% , corresponding to changes of 19 to 26 mm. The positive effects of WS and RH on ET_0 change are similar to air temperature, which cannot be ignored for station U3.

The contribution of the seasonal change of meteorological factors to ET_0 change is similar to that at an annual scale. WS is still the dominant contributor to ET_0 change for the middle-lower regions at all seasons. The relative changes of ET_0 caused by WS change are greater than 5% for most stations in the middle and lower regions, whereas the relative changes of ET_0 caused by other factors are less than 5% . The trends of seasonal T_{max} and T_{min} still result in an increase in ET_0 for the upper region. However, there are differences for the contribution levels of each meteorological factor in different seasons and regions. For example, the trends of R_s for stations U1, U2, and M1 have more significant contributions to the changes of ET_0 only in summer, whereas the R_s trends for all stations have little effect on the changes of ET_0 in other seasons. Moreover, the T_{min} trends in lower regions do not contribute to changes in ET_0 in autumn, whereas the contribution of T_{min} to ET_0 change is strong in the other three seasons. RH and WS for station U3 have similar effects on ET_0 change, for which the effect is stronger in summer than that in other regions.

5. Discussion

This paper carefully and thoroughly analyzed the trends and spatial variations of the annual and seasonal ET_0 for

different regions over the HRB. The spatial patterns of annual and seasonal ET_0 during the last 54 years in this study are consistent with the previous studies [34, 35]. However, the overall increasing trend ($2.01 \text{ mm} \cdot 10 \text{ yr}^{-2}$) of annual ET_0 for the whole basin in this paper is different from the significant decreasing trend reported by previous studies [29, 30]. After serious comparison and analysis, the causes of the differences come from inconsistent study areas and from differences in the data time series, treatment of missing data, and analysis methods. (i) Because the lower region of the HRB is the desert area and is difficult to fix the basin divides, four different basin areas have been defined by the Yellow River Conservancy Commission during different periods. The basin area defined most recently in 2005 is larger than the basin areas defined in 1985, 1995, and 2000 and can better describe the hydrological characteristics, especially for the lower region of the basin. This study adopted the latest basin area data defined in 2005, and previous studies adopted the earlier basin area data defined in 1995. (ii) Different data time series may result in different trends of annual ET_0 . The trends of annual and seasonal ET_0 calculated by the data series of 1959–1999 or 1961–2000 were earlier and shorter than the data series of 1961–2014 in this study. This latest data series, covering more than 50 years of climate stage and data quality during this latest period, is more reliable and is without missing data. (iii) Because meteorological stations are scarce in the inland arid basin in China, the stations around the basin must be considered to increase the precision of calculation of regional ET_0 . Clearly, the results obtained using only the 10 stations in the previous studies are less reliable than those using 16 stations related to the basin.

Equation (15) was used to assess the contribution of meteorological factors to ET_0 trends. Figure 12 shows that correlation of the estimated and the actual relative changes of ET_0 are very good, whereas the correlation coefficients decrease with increasing time scales from monthly scale to annual scale. This illustrated that the accuracy of (15) decreases with increasing time scale. The error sources of (15) are that (i) the five major meteorological factors cannot completely cover all impact factors of the FAO P-M equation; (ii) the selected factors interact with each other and are not totally independent; and (iii) the annual averaging variations of the daily sensitivity coefficient could produce different offsets to ET_0 changes contributed by different meteorological factors.

6. Summary

In arid regions, investigating the causes of reference evapotranspiration (ET_0) change is important for understanding hydroclimatic change and the response of ecoenvironment. The Heihe River Basin (HRB), the second largest inland river basin in China, is divided into the upper, middle, and lower subregions to diagnose the causes of ET_0 changes in different dryness environment.

First, the ET_0 changes for the HRB were obtained by FAO P-M method and meteorological data series from 16 stations during 1961–2014. The seasonal and annual ET_0 have no significant increasing trends for the whole basin, whereas

FIGURE 13: Contributions of meteorological factor changes to relative changes in ET_0 at annual and seasonal time scales for the stations in the HRB. An upward bar means that the factor trend causes a positive change in ET_0 , and a downward bar means that the factor trend causes a negative change in ET_0 .

there is a clear increasing spatial gradient from the upper region to the lower region.

Second, the dimensionless sensitivity analysis showed that relative humidity is most sensitive to ET_0 change and negative, followed by maximum temperature and shortwave radiation but with positive sensitivity. The sensitivity of minimum temperature is weakest and negative.

Finally, to quantify the influence magnitude of the major meteorological factors on ET_0 changes, an approach to integrating the sensitivity and changes of meteorological factors is proposed. Contribution analysis showed that wind speed is the dominant factor to cause the decrease of ET_0 for the middle and lower regions. And the maximum and minimum temperatures are the main contributors to the increasing trends of ET_0 for the upper region. Therefore,

the ET_0 changes are mainly affected by aerodynamic factors rather than radiative factors as dryness increase.

Conflict of Interests

The authors declare that there is no conflict of interests regarding the publication of this paper.

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