Spatio-Temporal Variation of Reference Evapotranspiration and Its Climatic Drivers over the Tibetan Plateau during 1970–2018

Shanshan Hu 1, Rui Gao 1, Tao Zhang 2,*, Peng Bai 3 and Rui Zhang 2,*

1 Beijing Laboratory of Water Resources Security, College of Resource Environment and Tourism, Capital Normal University, Beijing 100048, China; hushanshan@cnu.edu.cn (S.H.); 2180902129@cnu.edu.cn (R.G.)
2 Land Satellite Remote Sensing Application Center (LASAC), Beijing 100048, China
3 Key Laboratory of Water Cycle & Related Land Surface Process, Institute of Geographic Sciences and Natural Resources Research, Chinese Academy of Sciences, Beijing 100101, China; baip@igsnrr.ac.cn
* Correspondence: zhangt@lreis.ac.cn (T.Z.); zhangrui@radi.ac.cn (R.Z.)

Abstract: Reference evapotranspiration ($ET_0$) is a key component of hydrologic cycle and it is important for water resources management. Analysis of $ET_0$ changes is particularly critical for understanding the impacts of climatic change on hydrology in ecologically fragile regions. In this study, using the Penman–Monteith method and the Mann–Kendall test, the variation characteristics of $ET_0$ on the Tibetan Plateau (TP) from 1970 to 2018 was analyzed, and the dominant climatic factors controlling the change of $ET_0$ was also explored. The result shows that in TP region: (1) there was an abrupt change in the trend of $ET_0$ around 1997, and the $ET_0$ declined at a rate of $-25.9 \text{ mm/decade}$ during 1970–1996 but increased by $31.1 \text{ mm/decade}$ during 1997–2018; (2) $ET_0$ is most sensitive to solar radiation, then relative humidity, wind speed and mean temperature; (3) the decrease of $ET_0$ before 1997 was mainly due to the decline of wind speed and the increase of relative humidity, while the increase of $ET_0$ after 1997 was mainly due to the decrease of relative humidity. The results of this study can provide data reference for the research of water balance on the TP.

Keywords: Tibetan Plateau; reference evapotranspiration; climate change; dominant factors

1. Introduction

As a link between water balance and surface energy balance [1], evapotranspiration is an important index reflecting climate variation and water circulation [2,3]. Reference evapotranspiration ($ET_0$) represents the maximum amount of evapotranspiration from a hypothetical reference surface under specific climatic conditions, given an adequate supply of water [4]. $ET_0$ plays a critical role in irrigation management [5] and is also an important parameter for agriculture water demand estimation and water balance analysis [6]. The Penman–Monteith (P-M) method is a fairly reliable method to estimate $ET_0$, which considers radiative and aerodynamic terms [7]. It focuses on the meteorological factors controlling evaporation process and is widely used in numerous research [8,9]. Understanding the spatial–temporal changes in $ET_0$ and the main causes can provide considerable understanding into the impact of climatic variation on water resources and hydrological cycles [10].

Extensive research has explored the spatial variation characteristics of $ET_0$ and the impact of climate factors on $ET_0$ in the late years [11]. It reported that pan-evaporation and $ET_0$ both have shown a steady downward trend in many regions of the world over the past several decades, including China, India, New Zealand and Italy [12–15], which is called ‘evaporation paradox’ [16]. The reasons for decreasing $ET_0$ have been widely discussed, global dimming and wind stilling were usually employed to explain the changes. For example, Han et al. [17] analyzed the entire trend in $ET_0$ in Jing-Jin-Ji region, finding that the decrease of wind speed and sunshine hours offset the influence of increased air temperature and led to a reduction in $ET_0$ before 1992. Xu et al. [18] found downward
trends in $ET_0$ in the Yellow River and the reduction of wind speed was the major driving force for decreasing trends. However, in recent years, increased $ET_0$ has been found in some regions, such as India and China [19]. Investigating long term change in $ET_0$ over India’s humid tropical Narmada River watershed, Pandey et al. [20] found an increasing trend in $ET_0$. Cong et al. [21] found that annual pan evaporation in China has increased since 1986. The changes in climatic variables such as air temperature, wind speed may explain the increase of $ET_0$. The increase of pan evaporation in China is mainly owing to the increase of vapor pressure deficit [21]. Liu et al. [22] investigated the $ET_0$ trends in the Northwest China from 1960 to 2010, finding the $ET_0$ began to increase in 1994, mainly caused by the increasing atmospheric temperature and wind speed. Since there are many factors affecting $ET_0$, the temporal and spatial differences between its changes and the changes of climate factors are complex and diverse. In addition, it is still uncertain how much $ET_0$ changes are caused by a certain factor. Therefore, it is necessary to further study the variation of $ET_0$ and its possible causes.

The Tibetan Plateau (TP), as the ‘third pole’ of the earth, is very sensitive to global change due to its unique climate conditions and natural environment [23]. The TP is rich in lakes, glaciers, and wetlands and is the main source area of several major rivers in Asia. The management of affluent water resources on the plateau are not only a major issue related to China, but also the neighboring countries located in the lower reaches of the international rivers originating from the plateau. The management of abundant water $ET_0$ is the key constituent of the hydrologic cycle and has a great impact on regional water resources. Several studies have focused on the variation of $ET_0$ over the TP under the background of climatic change [3,24,25]. Zhang et al. [26] pointed out that the decrease of wind speed and sunshine duration were the dominant factors for the significant decline of $ET_0$ in the northern and southeastern TP from 1971 to 2004. Wang et al. [27] pointed out that $ET_0$ increased from 1997 to 2014 and the reduction of relative humidity played a crucial role in increasing of $ET_0$. This study will analyze and quantify the possible abrupt change characteristics of $ET_0$ and the impact of climatic factors on $ET_0$ during different stages over the TP.

This paper focused on the following objectives: (1) to analyze the temporal–spatial trends and transition characteristics of $ET_0$ from 1970 to 2018 on the TP region; (2) to investigate the temporal–spatial variation characteristics of climatic variables controlling the evaporation process; (3) to identify the main climate reasons for $ET_0$ change in different stages. The findings can improve the understanding of climate change impacts on $ET_0$ and provide useful information for water resources management and sustainable development.

2. Materials and Methods

2.1. Study Area

The TP is the largest geomorphologic unit on the Eurasian continent, ranging from 25°–40° N and 75°–105° E, covering most of the Tibet Autonomous Region and Qinghai Province in western China, with a total area of more than 2.5 million km² (Figure 1). The TP is the highest and most extensive plateau in the world, known as the ‘roof of the world’, with about 56% of the region over 4000 m above sea level [28]. Due to its high altitude and large volume, it forms a unique ‘plateau climate’. The annual average temperature of the TP is reduced from 20 °C in the southeast to below −6 °C in the northwest. The annual precipitation is also reduced from 2000 mm to below 50 mm. The main vegetation cover in this area is grassland and forest. The grassland is mainly divided into alpine meadow, alpine grassland and alpine desert [29].
2.2. Data Sources

Daily meteorological records from 73 meteorological stations in and around the TP during the period 1970–2018 were used in this study and the altitude of these stations ranges from 1231.2 to 4645.1 m. The meteorological data includes daily average air temperature, maximum air temperature, minimum air temperature, relative humidity, wind speed, air pressure, sunshine duration and the 20 cm pan evaporation. These datasets were provided by the China Meteorological Data Service Center (CMDC) (http://data.cma.cn/, accessed on 6 May 2019) and were processed with quality control. The 90 m DEM data set of the TP is provided by Geospatial Data Cloud site, Computer Network Information Center, Chinese Academy of Sciences (http://www.gscloud.cn, accessed on 12 August 2019).

2.3. Methods

2.3.1. Calculation of ET₀

Penman–Monteith (P-M) formula was used to estimate daily ET₀ in this article which was proposed by the FAO (Food and Agriculture Organization). The P-M method estimates ET₀ from hypothetical reference grass with an assumed height of 0.12 m, a fixed surface resistance of 70 m·s⁻¹, and an albedo of 0.23 [30]. The formula is:

\[
ET₀ = \frac{0.408\Delta(R_n - G) + \gamma \frac{900}{T_{\text{mean}} + 273} U_2(\varepsilon_s - \varepsilon_a)}{\Delta + \gamma(1 + 0.34 U_2)} \quad (1)
\]

where \(ET₀\) is the daily reference evapotranspiration (mm/d), \(\Delta\) is the slope of the vapor pressure curve and \(\gamma\) is the psychometric constant (kPa/°C), \(R_n\) is the net radiation and \(G\) is the soil heat flux density (MJ·m⁻²·day⁻¹), \(T_{\text{mean}}\) is the mean air temperature (°C), \(U_2\) is the wind speed at 2 m height (m/s), \(\varepsilon_s\) is the saturation vapor pressure and \(\varepsilon_a\) is the actual vapor pressure (kPa).

Reference evapotranspiration and pan evaporation both represent evaporative capacity of a basin under certain climatic conditions and their good relationship had been manifested by previous research [31]. The pan coefficient and correlation between them were calculated to verify the applicability of the P-M model (Figure 2). The correlation coefficient between pan evaporation and reference evapotranspiration was 0.71 with a pan coefficient of 1.75, which indicates that the P-M model is suitable for the analysis of reference evapotranspiration on the TP.
2.3.2. Cramer Test for Abrupt Change Point

The Cramer test is used to detect the mutation of $ET_0$ series. Cramer’s test has displayed fine performance in checking the stabilization of records by comparing the whole average value of the record and the means of certain parts of the record [32]. The Cramer’s test statistic $t_k$ is calculated as follows:

$$t_k = \sqrt{\frac{n(N-2)}{N-n(1+\tau_k^2)}} \times \tau_k \quad (2)$$

$$\tau_k = \frac{x_k - \bar{x}}{S}, \quad \bar{x}_k = \frac{1}{n} \sum_{i=k+1}^{k+n} x_i, \quad \bar{x} = \frac{1}{N} \sum_{i=1}^{N} x_i \quad (3)$$

where $\bar{x}$ and $S$ are the mean and standard deviation, for the entire period of $N$ years, and $\bar{x}_k$ is the mean of the sub period of the $n$ years to be compared with $\bar{x}$.

2.3.3. Mann–Kendall Test for Trend Analysis

The Mann–Kendall (MK) test is used to analyze $ET_0$ and climate factors trends. This method does not need to consider the distribution form of the detection sequence, and can exclude the influence of individual extreme values on the sequence test, which is suitable for testing the trend change of time series of meteorological, hydrological and other types of variables [32,33]. It is highly recommended for general use by the World Meteorological Organization [34]. Its principle is as follows: the original assumption is that $H_0$ is $n$ $(x_1, x_2, ..., x_n)$ independent time series samples with the same distribution; alternative hypothesis $H_1$ is a bilateral test, the statistical value $S$ is calculated as follows [17]:

$$S = \sum_{i=2}^{n} \sum_{j=1}^{i-1} \text{sign}(X_i - X_j) \quad (4)$$

$$\text{sign}(X_i - X_j) = \left\{ \begin{array}{ll}
1 & X_i - X_j > 0 \\
0 & X_i - X_j = 0 \\
-1 & X_i - X_j < 0 
\end{array} \right. \quad (5)$$
where: any $i \neq j$, $i \leq n$, $j \leq n$, when $S$ approximately obeys normal distribution, and the trend test statistical value $Z$ is obtained by means of $S$ standardization processing:

$$Z = \begin{cases} 
\frac{S-1}{\sqrt{\text{Var}(S)}} & S > 0 \\
0 & S = 0 \\
\frac{S+1}{\sqrt{\text{Var}(S)}} & S < 0 
\end{cases}$$ (6)

If $|Z| > Z_{1-\alpha/2}$, the null hypothesis is rejected with a given confidence level $\alpha$; namely, there is a significant trend in the time series data. When $\alpha$ is equal to 0.05, $Z_{1-\alpha}$ is equal to 1.96, respectively. The $Z$ value is used to assess the statistics trend; a positive $Z$ value expresses an upward trend, a negative $Z$ value expresses a downward trend.

The annual change rate of data is evaluated by the least square method. The principle is to look for the best function matching of data by minimizing the square sum of errors [35].

2.3.4. Spatial Interpolation Method

Spatial interpolation is very important for obtaining the spatial distribution characteristics of climate variables when meteorological stations are few and widely separated. In this paper, the gradient plus inverse distance squared (GIDS) method is used for interpolating climate variables and $ET_0$ [36]. It is an inverse distance weighting method that takes into account the changes of meteorological factors with latitude, longitude and altitude. Previous studies have shown that this method has good applicability in areas with large terrain undulations and uneven site distribution [37]. The interpolating climate variable is calculated as follows [36,37]:

$$Z = \left[ \sum_{i=1}^{N} Z_i + (X - X_i) \times C_x + (Y - Y_i) \times C_y + (E - E_i) \times C_e \right] / \left[ \sum_{i=1}^{N} \frac{1}{d_i^2} \right]$$ (7)

where $Z$ is the estimated climatic variable, $Z_i$ is the value at climate station $i$, $d_i$ is the distance from the site to climate station $i$ and $N$ is the number of climate stations used for the interpolation. $X$, $Y$ and $E$ are the longitude, latitude, and altitude of the point to be estimated, respectively, $X_i$, $Y_i$ and $E_i$ are the longitude, latitude, and altitude of climate station $i$, respectively, and $C_x$, $C_y$ and $C_e$ are regression coefficients for $X$, $Y$ and $E$, respectively.

The daily $ET_0$ of 73 meteorological stations on the TP in 1970–2018 is calculated and transformed into $ET_0$ at annual scale. The annual average $ET_0$ on the TP is obtained by interpolating the annual $ET_0$ of 73 stations to the centroid of the area using the GIDS method.

2.3.5. Sensitivity Analysis and Contribution Rate Calculation

The sensitivity coefficient method is used to analyze the sensitivity of climate factors to $ET_0$ quantitatively [38]. The commonly used sensitivity analysis method is to assume that other variables are fixed, and to evaluate the impact of a variable change on the model output. The sensitivity coefficient of the $ET_0$ to a climate variable $x$ is defined by [39,40]:

$$S_x = \lim_{\Delta x \to 0} \left( \frac{\Delta ET_0/ET_0}{\Delta x/x} \right) = \frac{\partial ET_0}{\partial x} \cdot \frac{x}{ET_0}$$ (8)

A first-order Taylor series approximation was applied to calculate $S_x$ [41,42]:

$$S_x = \frac{\Delta ET_0}{\Delta x} \cdot \frac{x}{ET_0}$$ (9)

where $S_x$ is the sensitivity coefficient of the climate variable $x$, $\Delta x$ is the relative change of climate factor $x$, and $\Delta ET_0$ is the relative change in $ET_0$ induced by $\Delta x$. 

In this study, $ET_0$ was recalculated by making a $\pm 10\%$ change to each meteorological factor (solar radiation ($RS$), relative humidity ($RH$), air temperature ($T_{mean}$) and wind speed ($U_2$)) while keeping the other factors constant; sensitivity coefficient $S_x$ was calculated from Equation (9). A positive value of $S_x$ indicates that $ET_0$ increases as $x$ increases, and vice versa. The greater the absolute value of the sensitivity coefficient, the greater the effect of the variable on $ET_0$.

The sensitivity coefficient reflects the influence of meteorological variables on $ET_0$, but it is not equal to the $ET_0$ change caused by a certain variable. Here, the contribution rate is introduced to analyze the change of $ET_0$ caused by a single meteorological factor. It is the product of the sensitivity coefficient of a climate factor and its multi-year change rate [11]. A positive contribution means that the variety of a meteorological factor has a positive effect on the variety of $ET_0$ and vice versa. The larger the absolute value is, the greater the impact on $ET_0$ will be. The contribution rate was calculated using the following formulas [43,44]:

$$CR_x = S_x \cdot RC_x$$

$$RC_x = \frac{n \cdot a_x}{\bar{x}} \times 100\%$$

where $CR_x$ is the contribution rate climate variable $x$ to $ET_0$; $RC_x$ is the multi-year change rate of $x$; $n$ is the number of years; $a_x$ represents the trend rate of meteorological factor $x$ during the corresponding study period; $\bar{x}$ is the multi-year absolute average value of $x$ during the evaluation period. In this paper, we calculated the sensitivity of $ET_0$ to $RS$, $RH$, $T_{mean}$ and $U_2$, and calculated the contribution rate of these four climatic factors to $ET_0$. The evaluation period includes three periods: during the period from 1979 to 2018 and two periods before and after the mutation point.

$$C_\sim ET_0 = CR_{RS} + CR_{RH} + CR_{T_{mean}} + CR_{U2} + \varepsilon$$

where $CR_{RS}$, $CR_{RH}$, $CR_{T_{mean}}$, $CR_{U2}$ are the individual contributions to the long-term trend in $ET_0$ as a result of the change in $RS$, $RH$, $T_{mean}$ and $U_2$, respectively. $C_\sim ET_0$ is the sum of the contribution of variation in $RS$, $RH$, $T_{mean}$ and $U_2$ to the change in $ET_0$. $\varepsilon$ is the error between $C(ET_0)$ and $ET_0$ trend calculated by P-M model ($LR_\sim ET_0$).

3. Results

3.1. Spatio-Temporal Variation Characteristics of $ET_0$

Figure 3a shows the tendency of annual $ET_0$ on the TP. The annual average $ET_0$ was 1037.17 mm, ranging from 983.53 to 1091.94 mm from 1979 to 2018. The annual $ET_0$ experienced a non-significant ($\alpha = 0.05$) decrease by 2.1 mm/decade (Table 1). According to the Cramer’s test, there are two abrupt change points in $ET_0$ series of TP, which are 1997 and 2000, respectively (Figure 3b). We found the decline rate of $ET_0$ was $-25.9$ mm/decade before 1997 and the change is significant, but the annual $ET_0$ increased significantly ($\alpha = 0.05$) at a rate of 31.1 mm/decade after 1997. Furthermore, the Cramer’s test was also performed on all 73 stations to verify the representativeness of 1997, and the results showed that 41 of them could detect the mutation point in 1997. Therefore, it can represent the main mutation characteristics of $ET_0$ in the region. Since the variation of $ET_0$ is the result of the comprehensive effects of climate variables, it is necessary to analyze each climate variable during the two periods separately. Using the inflection point of $ET_0$ as the basis for dividing the period, all the time series of the 73 stations were divided in these two periods for the subsequent analysis.
Spatial distribution of ET trends increased in the southeast TP from 1997 to 2018, while 21 stations showed a decreasing trend. During 1970–1996, there are 52 stations showing a downward trend, and these downward trends were remarkable (α = 0.05) at 34 meteorological stations; while 21 stations showed a decreasing trend. Only 24 stations in the eastern TP showed a decreasing trend.

Figure 4 shows spatial distribution of ET trends on the TP during 1997–2018 and 1970–1996. There were significant spatial differences in the trends of ET before and after 1997. During 1970–1996, there are 52 stations showing a downward trend, and these downward trends were remarkable (α = 0.05) at 34 meteorological stations; while 21 stations increased in the southeast TP from 1997 to 2018, there were annual ET increasing trends at 49 meteorological stations, of which 36 meteorological stations increased significantly (α = 0.05). Only 24 stations in the eastern TP showed a decreasing trend.
3.2. Spatio-Temporal Variation Characteristics of Climatic Factors

Figure 5 and Table 1 shows the trends of climate variables on the TP from 1970 to 2018. The $T_{\text{mean}}$ increased significantly ($\alpha = 0.05$) before and after 1997 at the rate of 0.125 and 0.373 °C/decade, respectively. The upward trend was steeper during the latter period than during the former period. $RS$ and $RH$ increased by 0.019 (MJ·m$^{-2}$·day$^{-1}$)/decade and 1.68%/decade from 1970 to 1996 and decreased significantly at the rate of $-0.163$ (MJ·m$^{-2}$·day$^{-1}$)/decade and $-2.816$%/decade after 1997, respectively. The variation of $U_2$ was opposite to the variation of $RS$ and $RH$. $U_2$ decreased dramatically by 0.294 m/s/decade during the former period but increased significantly by 0.158 m/s/decade during the latter period.

![Figure 5](image)

*Figure 5. Annual variation of climatic factors on the TP during 1970–2018. (a) Solar radiation ($RS$); (b) relative humidity ($RH$); (c) air temperature ($T_{\text{mean}}$); (d) wind speed ($U_2$).*

Figure 6 shows the spatial distribution of trends of four climate factors during 1970–1996 and 1997–2018. From 1970 to 1996, the increasing $RS$ distributed partially in the southeast and west of the TP, and the decreasing $RS$ majority distributed in the central area of the TP. From 1997 to 2018, the decreasing $RS$ was found in most stations and the increasing $RS$ distributed in some areas east and north-west of the TP. The increasing $RH$ was found in most stations and the decreasing $RH$ distributed in the north-western areas during 1970–1996, while the decreasing $RH$ was found in most stations and the increasing $RH$ distributed in some areas east of the TP during 1997–2018. $T_{\text{mean}}$ increased across the entire area before and after 1996. From 1970 to 1996, $U_2$ showed a downward trend in most stations, while only a few stations in the eastern part of the plateau showed an increasing trend. However, from 1997 to 2018, $U_2$ of most stations showed an upward trend, and only a few stations decreased.
Figure 6. Spatial distribution of climatic factors trends during 1970–1996 (a–d) and 1997–2018 (e–h); (a,e) RS ((MJ m\(^{-2}\) day\(^{-1}\))/decade); (b,f) RH (%/decade); (c,g) \(T_{\text{mean}}\) (°C/decade); (d,h) \(U_2\) (m/s/decade).

3.3. Spatial Distribution of the Sensitivity Coefficients

Figure 7 shows the spatial distribution of susceptibility coefficients on the TP. The sensitivity coefficient of RS (\(S_{\text{RS}}\)) ranged from 0.303 to 0.587 and decreased from southeast to northwest on the TP. The sensitivity coefficient of \(T_{\text{mean}}\) (\(S_{T_{\text{mean}}}\)) ranged from 0.025 to 0.172, which was similar to \(S_{\text{RS}}\). Different from \(S_{\text{RS}}\) and \(S_{T_{\text{mean}}}\), the sensitivity coefficient of \(U_2\) (\(S_{U_2}\)) increased progressively from southeast to northwest and varied from 0.080 to 0.259. The sensitivity coefficient of RH (\(S_{\text{RH}}\)) ranged from −0.444 to −0.019 and the
3.4. Dominant Climate Factors of ET₀ Change

In order to identify the dominant climate factors affecting ET₀, the contribution rate of each climatic factor was calculated using Equation (11). Comparing the ET₀ trends caused by the four climate factors (C_{ET₀}) and the ET₀ trends calculated by P-M model (LR_{ET₀}), the relative errors were 12.79%, 4.70% and 7.93% during 1970–1996, 1997–2018 and 1970–2018, respectively (Table 2). Therefore, equation (11) was a reasonable method to estimate the contribution rates of the four climatic factors to ET₀.

Table 2. The contribution rate of climate factors to ET₀ in different periods (%).

| Period      | CR_{RS} | CR_{RH} | CR_{Tmean} | CR_{U2} | C_{(ET₀)} | LR_{(ET₀)} | ε       |
|-------------|---------|---------|------------|---------|-----------|------------|---------|
| 1970–1996   | 0.10    | −2.43   | 0.80       | −4.41   | −5.94     | −6.70      | 12.79   |
| 1997–2018   | −0.96   | 3.33    | 1.58       | 2.34    | 6.29      | 6.60       | 4.70    |
| 1970–2018   | −0.67   | 1.04    | 3.05       | −4.29   | −0.87     | −0.94      | 7.93    |

Table 2 shows the contribution rate of the four climate factors to ET₀ in different periods. During 1970–1996, U₂ was the dominant factor influencing ET₀, followed by RH, T_{mean} and RS. The decrease of U₂ and increase of RH caused a −4.41% and −2.43% decrease to ET₀, respectively. However, the increases in T_{mean} and RS led to a 0.80% and 0.10% increase to ET₀, respectively. The comprehensive effect of climate factors finally led to a −6.7% decline to ET₀. During 1997–2018, the decrease of RH was the main reason causing the increase in ET₀ by 3.33%. In addition, the rise of U₂ and T_{mean} also aggravated the increase of ET₀ to a certain degree, resulting in the increase of ET₀ of 2.34% and 1.58%, respectively. The decrease of RS led to a variation in ET₀ of −0.96%. The effects of the climatic condition eventually caused the increase of ET₀ by 6.29%. In a word, the decrease of ET₀ during 1970 and 1996 was mainly due to the decrease of U₂, while the increase of ET₀ during 1997 and 2018 was mainly due to the decrease of RH.
Figure 8 gives the spatial distribution of contribution rates of four climatic factors to $ET_0$ during 1970–1996 and 1997–2018. From 1970 to 1996, the contribution rate of $RS$ ($CR_{RS}$) was positive in the southeast and western TP, and the negative $CR_{RS}$ was found in eastern and central parts of TP. From 1997 to 2018, $CR_{RS}$ was mainly positive and the negative contribution distributed in the northern and central sporadic areas. The contribution rate of $RH$ ($CR_{RH}$) was negative in most parts of TP and the positive $CR_{RH}$ was only distributed partly in northwest and east before 1997; while, most regions on the TP exhibited a positive $CR_{RH}$, except some areas of eastern TP after 1997. The distribution of contribution rate of $U_2$ ($CR_{U2}$) was similar to that of $RH$. The contribution rate of $T_{mean}$ ($CRT_{mean}$) was positive in most parts of TP during both periods.

Figure 8. Spatial distribution of contribution rates of four climatic factors to $ET_0$ during 1970–1996 (a–d) and 1997–2018 (e–h) (%); (a,e) $CR_{RS}$; (b,f) $CR_{RH}$; (c,g) $CRT_{mean}$; (d,h) $CR_{U2}$. 
4. Discussion

According to the analysis of this study, there are significant differences in climate factors between 1970–1996 and 1997–2018 (Figures 5 and 6; Table 1). The spatiotemporal dynamic of the four climate factors in this study is basically consistent with a lot of research on climatic change. Before 1997, the wind speed decreased significantly, but then increased significantly. The weakening of the Asian monsoon system may explicate the decrease of $U_2$ on the TP, while the strengthening of zonal wind may have a significant contribution to the increase of $U_2$ after 1997 [44]. An insignificantly decreasing trend of $RS$ was found during 1970–1996, but a significantly decreasing trend was found during 1997–2018. The solar dimming over the TP is mainly caused by the increase in water vapor and deep cloud cover [3]. $RH$ also decreased significantly after 1997. Most areas of TP belong to the arid region and the amount of water is limited. The rise of air temperature makes the saturated vapor pressure rise, which may be the reason for the decrease in $RH$ [45]. The climatic factors and $ET_0$ have complex mutual feed mechanism and give rise to an abrupt change in $ET_0$ in the meantime.

$ET_0$ decreased significantly before 1997 while increased significantly after 1997 on the TP in this study (Figures 3 and 4; Table 1). The change of $ET_0$ is influenced by climate factors, and the weight of each climate factor is different. The dominant factor causing the decrease of $ET_0$ on the TP was $U_2$ before 1997 (Figure 8; Table 2), and this is similar to many studies on the TP. For example, Chen et al. [46] found that wind speed is the dominant meteorological factor affecting the $ET_0$ of the TP. Zhang et al. [26] considered that wind speed caused changes in the $ET_0$ of the TP, especially in the northern part of the TP. Yin et al. [47] have shown that the decrease of wind speed can explain the variation of the potential annual evapotranspiration in the temperate zone of the north of China and the TP. In this study, the dominant factor that caused the increase of $ET_0$ on the TP after 1997 is $RH$, which is consistent with Wang’s [27] study that relative humidity may be an important factor affecting $ET_0$. However, this is not consistent with the view of Xie Hong [48] that relative humidity is considered to have little effect. This is due to the difference in the research periods and methods.

Some uncertainties remain for the estimation of $ET_0$ in this study. Some of the parameters in the P-M equation are empirical and local. Therefore, regional optimization of parameters is an important means to improve the accuracy of $ET_0$ calculation. Additionally, the number of meteorological stations used in this study is limited and unevenly distributed. The TP has large area and complex natural environment. It is difficult to fully describe the spatiotemporal dynamic of the climatic factors and $ET_0$. More data from different sources and interpolation methods should be introduced in the future.

5. Conclusions

In this study, we analyzed the temporal–spatial trends, transition characteristics of $ET_0$ and identified the main reasons for the change of $ET_0$ in different stages over the TP based on data from 73 weather stations during 1970–2018. The main conclusions are:

1. There is an abrupt point in the TP's annual $ET_0$ series around 1997. The $ET_0$ decreased remarkably at a rate of $-25.9$ mm/decade before 1997, while increasing significantly at a speed of $31.1$ mm/decade after 1997;
2. Second item: before 1997, $T_{mean}$ and $RH$ increased significantly while $U_2$ decreased significantly and $RS$ decreased insignificantly. After 1997, $T_{mean}$ and $U_2$ increased insignificantly, $RH$ and $RS$ decreased significantly;
3. The sensitivity analysis of each climate factor to $ET_0$ indicated the $ET_0$ on the TP is most sensitive to $RS$, followed by $RH$, $U_2$ and $T_{mean}$;
4. Third item: from 1970 to 1996, $U_2$ was the most important meteorological factor contributing to the decline of $ET_0$, followed by $RH$, $T_{mean}$, and $RS$. During 1997–2018, the decreasing of $RH$ was the dominant factor causing an increase in $ET_0$. The increase of $U_2$ and $T_{mean}$ also intensified the increase of $ET_0$ to a certain extent.
Author Contributions: Conceptualization, S.H. and T.Z.; data curation, R.G.; formal analysis, R.G.; funding acquisition, S.H.; methodology, P.B.; resources, T.Z.; visualization, R.Z.; writing—original draft, S.H. and R.G.; writing—review and editing, S.H., T.Z., P.B. and R.Z. All authors have read and agreed to the published version of the manuscript.

Funding: This research was funded by National Natural Science Foundation of China, grant number 41501027, the Beijing Outstanding Young Scientist Program, grant number BJJWZYJH01201910028032.

Institutional Review Board Statement: Not applicable.

Informed Consent Statement: Not applicable.

Data Availability Statement: There are no relevant statements for this research.

Acknowledgments: The authors would like to express their sincere thanks to National Natural Science Foundation of China (41501027) and Beijing Outstanding Young Scientist Program (BJJWZYJH01201910028032) for their valuable support. The authors also would like to thank the peer reviewers for their valuable suggestions and comments on improving this paper.

Conflicts of Interest: The authors declare no conflict of interest.

References

1. Lu, X.; Zang, C.; Burenina, T. Study on the variation in evapotranspiration in different period of the Genhe River Basin in China. Phys. Chem. Earth 2020, 120, 102902. [CrossRef]
2. Maeda, E.E.; Wiberg, D.A.; Pellikka, P.K.E. Estimating reference evapotranspiration using remote sensing and empirical models in a region with limited ground data availability in Kenya. Appl. Geogr. 2011, 31, 251–258. [CrossRef]
3. Kun, Y.; Baoqiong, D.; Jun, Q.; Wenjun, T.; Ning, L.; Changgui, L. Can aerosol loading explain the solar dimming over the Tibetan Plateau? Geophys. Res. Lett. 2012, 39. [CrossRef]
4. Boisier, J.P.; de Noblet-Ducoudré, N.; Ciais, P. Historical land-use-induced evapotranspiration changes estimated from present-day observations and reconstructed land-cover maps. Hydrol. Earth Syst. Sci. 2014, 18, 3571–3590. [CrossRef]
5. Wu, H.; Wang, X.; Wang, Y.; Xu, Y.; Han, X. Temporal variations in reference evapotranspiration in Hubei Province, China, from 1960 to 2014. Theor. Appl. Climatol. 2019, 135, 433–448. [CrossRef]
6. Liu, C.; Zhang, D. Temporal and Spatial Change Analysis of the Sensitivity of Potential Evapotranspiration to Meteorological Influencing Factors in China. Acta Geogr. Sin. 2011, 66, 579–588.
7. Rachid, H.; Tarik, B.; Hayat, L.; Adil, S. Comparative assessment of different reference evapotranspiration models towards a fit calibration for arid and semi-arid areas. J. Arid Environ. 2021, 184, 104318.
8. Zhao, J.; Xia, H.; Yue, Q.; Wang, Z. Spatiotemporal variation in reference evapotranspiration and its contributing climatic factors in China under future scenarios. Int. J. Climatol. 2020, 40, 3813–3831. [CrossRef]
9. Fan, J.; Wu, L.; Zhang, F.; Xiang, Y.; Zheng, J. Climate change effects on reference crop evapotranspiration across different climatic zones of China during 1956–2015. J. Hydrol. 2016, 542, 923–937. [CrossRef]
10. Li, Y.; Wei, X.; Su, H. Quantitative analysis of causes for temporal and spatial variation characteristics and evolution of potential evapotranspiration in Gansu Province during recent 30 years. J. Water Resour. Water Eng. 2015, 26, 219–225.
11. Wang, Z.; Ye, A.; Wang, L.; Liu, K.; Cheng, L. Spatial and temporal characteristics of reference evapotranspiration and its climatic driving factors over China from 1979–2015. Agric. Water Manag. 2019, 213, 1096–1108. [CrossRef]
12. Liu, X.; Zheng, H.; Zhang, M.; Liu, C. Identification of dominant climate factor for pan evaporation trend in the Tibetan Plateau. J. Geogr. Sci. 2011, 21, 594–608. [CrossRef]
13. Deepak, J.; Yogob, D.; Ercan, K.; Vijay, P.S.; Ahmad, F.F. Trends in reference evapotranspiration in the humid region of northeast India. Hydrol. Process. 2012, 26, 421–435.
14. Michael, L.R.; Graham, D.F. Changes in New Zealand pan evaporation since the 1970s. Int. J. Climatol. 2005, 25, 2031–2039.
15. Moonen, A.C.; Ercoli, L.; Mariotti, M.; Masoni, A. Climate change in Italy indicated by agrometeorological indices over 122 years. Agric. For. Meteorol. 2002, 111, 13–27. [CrossRef]
16. Brutsaert, W.; Parlange, M.B. Hydrologic cycle explains the evaporation paradox. Nat. Int. Wkly. J. Sci. 1998, 396, 30. [CrossRef]
17. Han, J.; Wang, J.; Zhao, Y.; Wang, Q.; Zhang, B.; Li, H.; Zhai, J. Spatio-temporal variation of potential evapotranspiration and climatic drivers in the Jing-Jin-Ji region, North China. Agric. For. Meteorol. 2018, 256, 75–83. [CrossRef]
18. Xu, S.; Yu, Z.; Yang, C.; Ji, X.; Zhang, K. Trends in evapotranspiration and their responses to climate change and vegetation greening over the upper reaches of the Yellow River Basin. Agric. For. Meteorol. 2018, 263, 118–129. [CrossRef]
19. Zhang, Q.; Wang, W.; Wang, S.; Zhang, L. Increasing Trend of Pan Evaporation over the Semiarid Loess Plateau under a Warming Climate. J. Appl. Meteorol. Clim. 2016, 55, 2007–2020. [CrossRef]
20. Brij, K.P.; Deepak, K. Identification of trend in long term precipitation and reference evapotranspiration over Narmada river basin (India). Glob. Planet. Chang. 2018, 161, 172–182.
21. Cong, Z.T.; Yang, D.W.; Ni, G.H. Does evaporation paradox exist in China? Hydrol. Earth Syst. Sci. Discuss. 2008, 5, 357–366.
