The severity of drought and precipitation prediction in the eastern fringe of the Tibetan Plateau

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Abstract

The trend, severity, and duration of drought in the eastern fringe of the Tibetan Plateau (EFTP) have been investigated using the Mann-Kendall (M-K) trend test, standardized precipitation index (SPI), and generalized extreme value (GEV), using data obtained from 438 rainfall stations and reanalysis datasets for the period 1961–2014. A recent drought trend is evident from a decrease in rainfall, with this mainly occurring on the eastern slope of the TP (< 3000-m elevation); this is attributed to downward air flows over the eastern slope of the Tibetan Plateau (TP) induced by TP heating. Recent droughts have also been more severe, with these again mostly occurring on the eastern slopes. The duration of drought illustrates that extreme droughts are becoming more frequent. The study also predicted summer precipitation, due to its crucial role in drought research in the EFTP. Results show that the preceding May–June–July (MJJ) averaged column-integrated meridional water vapor transport (MWVT) from the South China Sea (SCS), Philippine Sea, and tropical western Pacific is a vital predictor of summer precipitation in the EFTP. A partial least squares (PLS) regression prediction model is therefore constructed, using the leading PLS components of preceding MJJ-averaged column-integrated MWVT. Compared to the observed summer rainfall, the PLS prediction model performs an excellent reconstructed skill with a correlation of 0.81 (1961–2006) and exhibits a promising forecast skill with a correlation of 0.67 (2007–2014). Results suggest that southerly moisture transport in early summer would help prevent summer drought in the EFTP.

1 Introduction

Drought is one of the most important meteorological factors, affecting agriculture, water resources, and ecosystems (Dai et al. 2004). Severe droughts cause millions of casualties and large economic losses each year (Eisensee and Strömberg 2007; Hao et al. 2012). The eastern fringe of the Tibetan Plateau (EFTP) (23°–40° N, 99°–109° E) is located on the western boundary of the East Asia monsoon (EAM) area in China and includes provinces such as Sichuan, Guizhou, Yunnan, and Gansu (Gao et al. 2017; Fig. 1). Frequent drought events in the EFTP lead to landslides, rockfalls, and debris flows. For instance, a prolonged drought event resulted in the displacement of over 18 million people in 2006 (Zhang et al. 2015). During 2009–2010, an unprecedented drought in Sichuan led to 21 million residents suffering water shortages, with about 4 million crops devastated (Lu et al. 2011; Ma et al. 2017). A better understanding of drought in the EFTP would improve drought forecasting and water resource management and could help improve regional responses to drought.

Variables such as precipitation, evapotranspiration, temperature, and soil moisture can be used to measure

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Fig. 1 **a** Altitude of terrain in East Asia is color shaded (unit, m), with the red solid box showing the location of the EFTP. **b** U and V wind components at 850 hPa (in summer) during 1961–2014. **c** U and V wind components at 850 hPa (annual) during 1961–2014. Red arrows indicate wind in **a** and **b**

drought (Xin et al. 2006; Sheffield and Wood 2008; Briffa et al. 2010). Previous analyses of drought in the EFTP have used various variables. Li et al. (2011) attributed a dry spell in the southeastern Tibetan Plateau (TP) to topographic forcing, based on precipitation, evaporation, and humidity. Fang et al. (2010) proposed, based on temperature and precipitation, that the spatial distribution of drought in the southeastern TP is influenced by tropical oceans. Deng et al. (2017) used precipitation and evapotranspiration to attribute spatial variations of drought in the eastern TP to the North Atlantic Oscillation. It is worth noting that the eastern TP has recently tended towards drought. You et al. (2012) analyzed the drying trend of the eastern TP based on two precipitation datasets. Zhao et al. (2016) indicated that precipitation in the EFTP has generally decreased across rainfall stations, indicating a drought tendency. Based on precipitation and temperature, it would appear that southeastern TP droughts have been on the increase

since 2009 due to anthropogenic warming (Ma et al. 2017).

The EFTP lies in the EAM regime and is strongly influenced by precipitation throughout the year (Fig. 1b, c; Xu et al. 2008). Heavy rains often occur (Jiang et al. 2015). Precipitation in the EFTP is closely associated with local drought (Ueno and Sugimoto 2012); for example, Ren (2014) suggested that a severe rainstorm in the eastern TP during 2010 induced a severe subsequent drought. As a result of global warming, increased extreme precipitation is expected to lead to frequent drought events (IPCC 2013; Ge et al. 2017). A number of drought studies conducted in other parts of the world have also focused exclusively on precipitation (Byun and Wilhite 1999). For example, Huang and Xu (1999) analyzed the North China drought trend based on rainfall stations. Similarly, Zhai et al. (2010) reported that many river basins in north China have experienced frequent droughts since 1995, on the basis of data from rainfall stations. Paparrizos et al.

(2016) found that, based only on precipitation, drought in Greece appears to be becoming more severe. Precipitation could be easily applied in drought research (Hayes et al. 1999). However, there is limited drought research on the EFTP that uses precipitation as a meteorological variable. Though the above-mentioned studies indicated a drought tendency in the EFTP (You et al. 2012; Zhao et al. 2016; Ma et al. 2017), both the likely severity and duration of drought remain unclear. Precipitation prediction is further necessary to enable drought hazard prevention.

The goals of this study are therefore to investigate spatiotemporal variations in drought trends, and their severity and duration in the EFTP, on the basis of precipitation, and to establish a statistical model to predict summer precipitation in situ. The paper is structured as follows. Section 2 introduces data and methods. Section 3 describes the spatiotemporal pattern of the EFTP drought trend. Drought severity and duration are explored in Section 4. Section 5 focuses on prediction of EFTP summer precipitation. Finally, Section 6 presents a discussion and conclusions.

2 Data and method

2.1 Data

This study mainly analyzed observational datasets of 438 daily rainfall stations with long-term rainfall records from 2513 stations in China, with these spanning the period 1961–2014 (<http://www.nmic.cn/web/index.htm>). The 438 stations nearly cover the entire area considered in this study and are thus suitable for a representation of drought climate features in the EFTP. Reanalysis data, including monthly mean ($2.5^\circ \times 2.5^\circ$) wind field, was acquired from the National Centers for Environmental Prediction (NCEP). EFTP elevations > 3000 m are considered to constitute the eastern platform of the TP, while elevations < 3000 m refer to the eastern slope of the TP (the 3000-m boundary line is shown in Figs. 2, 3 and 4). In this paper, summer refers to the period June–July–August (JJA).

2.2 Method

2.2.1 The Mann-Kendall trend test

The Mann-Kendall (M-K) trend test is applied to analyze spatiotemporal positive or negative trends of variables such as rainfall (Mann 1945; Kendall 1975; Wang and Swail 2001). This method is used as a substitute for parametric linear regression analysis and to examine whether

Fig. 2 Locations of all stations (green dots) in the EFTP are shown in the red solid box (23° – 40° N, 99° – 109° E). North and south blue solid lines denote the Yellow and Yangtze Rivers, respectively. The gray-shaded colors indicate terrain height (unit, m)

non-zero values shown in the gradient of the estimated linear regression line are correct.

2.2.2 The standardized precipitation index method

The standardized precipitation index (SPI), based only on precipitation, is widely applied for study of drought events and drought classes (Hayes et al. 1999; Pai et al. 2011). The drought analysis conducted for the EFTP focuses only on the influence of rainfall and not on the influence of evapotranspiration. The water vapor necessary for precipitation weakens with northward progression from the South China Sea (SCS) and Bay of Bengal. The SPI can be applied to confirm the severity of drought in different climate contexts and is therefore adopted in this study.

The SPI estimates precipitation deficit for multiple timescales. For purposes of this study, a 24-month timescale during the period 1961–2014 was selected. The SPI has been shown to fit a gamma distribution (Thom 1958; Livada and Assimakopoulos 2007; Almedeij 2014).

The probability density function for the gamma distribution $g(x)$ is shown as follows:

$$g(x) = \frac{1}{\beta \Gamma(\alpha)} x^{\alpha-1} e^{-x/\beta} \quad (1)$$

The gamma function $\Gamma(\alpha)$ is defined as follows:

$$\Gamma(\alpha) = \int_0^\infty x^{\alpha-1} e^{-x} dx \quad (2)$$

where x is precipitation amount and α and β are shape and scale parameters, respectively.

Fig. 3 Inter-annual variations in annual mean precipitation (black curve and dots) and trend (solid blue and red lines) in the EFTP from 1961 to 2014

These are estimated as follows (Edwards and McKee 1997):

$$\hat{\alpha} = \frac{1}{4A} \left(1 + \sqrt{1 + \frac{4A}{3}} \right), \quad \hat{\beta} = \frac{\bar{x}}{\hat{\alpha}} \quad (3)$$

$$A = \ln(\bar{x}) - \frac{\sum \ln(x)}{n} \quad (4)$$

where n is number of precipitation amount.

The cumulative distribution $G(x)$ during a time scale is as follows:

$$G(x) = \int_0^x g(x) dx = \frac{1}{\hat{\beta}^{\hat{\alpha}} \Gamma(\hat{\alpha})} \int_0^x \hat{\alpha}^{-1} e^{-x/\hat{\beta}} dx \quad (5)$$

Assuming $t = x/\beta$, thus,

$$G(x) = \frac{1}{\Gamma(\hat{\alpha})} \int_0^x t^{\hat{\alpha}-1} e^{-t} dt \quad (6)$$

The precipitation amount may be zero; however, the gamma function excludes this possibility. Cumulative distribution is therefore calculated as follows:

$$H(x) = q - (1-q)G(x) \quad (7)$$

where q is the probability of precipitation amount at zero.

According to Abramowitz et al. (1966), $H(x)$ could be transformed into the standard normal random distribution function:

$$SPI = -t + \frac{c_0 + c_1 t + c_2 t^2}{1 + d_1 t + d_2 t^2 + d_3 t^3}, \quad 0 < H(x) \leq 0.5 \quad (8)$$

$$SPI = t - \frac{c_0 + c_1 t + c_2 t^2}{1 + d_1 t + d_2 t^2 + d_3 t^3}, \quad 0.5 < H(x) \leq 1.0 \quad (9)$$

$$t = \sqrt{\ln \frac{1}{[H(x)]^2}}, \quad 0 < H(x) \leq 0.5 \quad (10)$$

$$t = \sqrt{\ln \frac{1}{[1.0 - H(x)]^2}}, \quad 0.5 < H(x) \leq 1.0 \quad (11)$$

where the coefficients in (8) and (9) are as follows: $c_0 = 2.515517$, $c_1 = 0.802853$, $c_2 = 0.010328$, $d_1 = 1.43788$, $d_2 = 0.189269$, and $d_3 = 0.001308$.

The above Eqs. (1)–(11) are used to estimate the SPI of precipitation amount.

Fig. 4 Multi-year averaged precipitation trends using M-K trend test based on 438 rainfall stations (1961–2014). Triangles denote an increasing rainfall trend and inverted triangles indicate a decreasing rainfall trend; in both cases, significance is at >90% confidence level. Color shading shows terrain height (unit, m). The black solid line represents the 3000-m contour

To measure drought severity in the EFTP, SPI drought classes (Table 1; <http://www.ncl.ucar.edu/Applications/spi.shtml>) were used.

2.2.3 The generalized extreme value distribution method

To confirm the distribution of drought duration, the generalized extreme value (GEV) distribution is used to calculate the probability density function (PDF) distribution. The GEV is a basic extreme value statistical method that has been widely applied to extreme temperature and precipitation (Rusticucci and Tencer 2008; Schubert et al. 2008; Yang et al. 2013; Rulfová et al. 2016).

The PDF of the GEV distribution method is computed as follows (Yang et al. 2013):

$$f(S, \sigma, \theta) = \frac{1}{\sigma} \begin{cases} (1 + \theta S)^{(-1/\theta)-1} \exp\left(-(1 + \theta S)^{-1/\theta}\right) & \theta \neq 0 \\ \exp(-S) \exp(-\exp(-S)) & \theta = 0 \end{cases} \quad (12)$$

where S is the standardized variable $S = (x - \mu)/\sigma$, $\mu \in \mathbb{R}$ is the location parameter (associated with record magnitude; units, mm), and $\sigma > 0$ is the scale parameter (associated with record variation; units, mm). θ is the shape parameter (related to the distribution tail). When $\theta > 0$, the formula is valid for $S > -1/\theta$, and when $\theta < 0$, the formula is valid for $S < -1/\theta$. When $\theta = 0$, density is always positive. Parameters are estimated according to maximum likelihood estimators.

2.2.4 Meridional water vapor transport

The meridional components of vertically integrated moisture transport can be represented as qv (Howarth 1983; Zhao et al. 2018), as follows:

$$qv(x, y, t) = \frac{1}{g} \int_{p_s}^{p_T} qv(x, y, p, t) dp \quad (13)$$

where x is the zonal field, y is the longitudinal field, p is the vertical level, and t represents time. g is accelerated speed because of gravity, q is specific humidity, and v is the meridional component of horizontal wind. p_s is pressure at sea surface level, and p_T is pressure at the top atmospheric level.

Table 1 SPI classes (<http://www.ncl.ucar.edu/Applications/spi.shtml>)

SPI	Drought category
$SPI \leq -1.6$	Extreme dry
$-1.6 < SPI \leq -1.3$	Severely dry
$-1.3 < SPI \leq -0.5$	Dry
$-0.5 < SPI \leq 0.5$	Nearly normal
$SPI > 0.5$	Wet

2.2.5 Partial least squares regression

Due to its wide variable coverage and ability to overcome limitations of high collinearity and small sample size, partial least squares (PLS) regression has been widely applied in statistical prediction (Haenlein and Kaplan 2004; Smoliak et al. 2010; Yu et al. 2017). Precipitation in the EFTP is closely related to southerly moisture transport, mainly from the SCS and Philippine Sea (Xu et al. 2002). PLS regression is thus utilized to find the PLS components of preceding column-integrated meridional water vapor transport (MWVT) as predictors and to predict summer precipitation in the EFTP. In other words, the PLS regression method aims to build predictors Z that combine factors X linearly as PLS components; PLS components are then used as predictors to obtain the predictand Y (Wu and Yu 2015).

The calculations of the PLS components Z and the predictand Y are defined as follows:

$$Z_k = a_{k1}X_1 + a_{k2}X_2 + \dots + a_{kj}X_{kj} \quad (14)$$

$$Y = b_1Z_1 + b_2Z_2 + \dots + b_kZ_k \quad (15)$$

The final formula of X and the predictand Y is computed as follows:

$$Y = w_1X_1 + w_2X_2 + \dots + w_kX_k \quad (16)$$

where X_{ij} indicates the 3-month averaged column-integrated MWVT variation prior to summer precipitation in the EFTP (i denotes space times; j denotes grid points). Z refers to leading PLS modes (k refers to mode number). Y_i represents predictand time series of summer rainfall in the EFTP.

2.2.6 Calculation of apparent heat source Q_1

The apparent heat source (AHS) (unit, $K s^{-1}$) is defined as follows (Yanai et al. 1973; Li et al. 2014):

$$Q_1 = C_p \left(\frac{\partial T}{\partial t} + \vec{V} \cdot \nabla T + \omega \left(\frac{P}{P_0} \right)^k \frac{\partial \theta}{\partial P} \right) \quad (17)$$

where T is temperature, and $\vec{V}$ and ω represent the horizontal and vertical components of wind, respectively. P_0 is pressure at sea surface level, C_p denotes specific heat at constant pressure, θ indicates potential temperature, and $k = 0.286$.

3 Spatial and temporal drought trends in the EFTP

This section presents the precipitation trend in the EFTP based on analysis of data from 438 rainfall stations (Fig. 2), using the M-K trend test. No significant trends were observed for the

period 1961–1990 (Fig. 3, blue line), and the curve of annual mean precipitation exhibited inconsistent fluctuations. During the period 1991–2014, rainfall exhibited a decreasing trend, as reported in previous studies (Duan et al. 2012; You et al. 2012; Zhao et al. 2016). The lowest value of annual mean precipitation (64 mm) occurred in 2011.

To illustrate spatial variations in precipitation, precipitation trends at each station are calculated using the M-K trend test (Fig. 4). The rainfall trends of 111 of 438 stations are significant at the 90% confidence interval. Of these stations, 88 show negative precipitation trends, accounting for 79.3% of total significant stations. Only 20.7% (23 stations) of total significant stations display positive rainfall trends. Notably, all stations with significant increased precipitation trends are located on the eastern platform of the TP. However, most stations with significant negative rainfall trends are located on the eastern slope of the TP. Results are similar to those obtained in other studies that have shown rainfall decreases on the TP slope over the last three decades (Yang et al. 2011, 2014).

Many researchers have concluded that the TP is a profound heat source during boreal summer (Flohn 1957; Yanai et al. 1992; Wu and Zhang 1998), but features of the AHS over the TP throughout the year remain unknown. We analyzed the spatial distribution of multi-year averaged AHS (unit, K s^{-1}) > 500 hPa over the TP during the period 1961–2014 (Fig. 5). A surprising maximum center of multi-year averaged AHS occurs over the eastern platform of the TP, implying that this is a unique heat source throughout the year (purple box in Fig. 5). Previous studies have also indicated that, as a heat source,

the TP attracts moisture from the tropics. Southerly moisture triggered by TP heating would greatly influence convective rainfall over the TP (Xu et al. 2014).

To further confirm convective precipitation variations over the TP and surrounding areas affected by the TP heat source, Fig. 6 shows vertical sections of the correlation vectors of multi-year averaged AHS over the eastern platform of the TP (Fig. 5, purple box) to meridional mean circulations along 97.5° – 105° E and zonal mean circulations along 26° – 39° N, respectively, during 1961–2014. In meridional mean circulations at the vertical sections, air ascent motions induced by TP heating are strong over the eastern platform of the TP, favoring convection with strong upward air flows. However, an air flow downdraft could be observed over the southeastern slope of the TP, with this being unfavorable for convective motion (Fig. 6a). In zonal mean circulations at the vertical sections, upward air motions induced by TP heating are active over the eastern platform of the TP, also favorable for convection over this region. Conversely, obvious downward air flows occur over the eastern slope of the TP, weakening convective motion (Fig. 6b). The above results indicate that the air ascent motions over the eastern platform of the TP induced by TP heating would lead to an increasing precipitation trend. However, the decreasing rainfall trend over the eastern slope of the TP may be attributed to obvious local downward air flows induced by TP heating.

Based on the above, drought trends in major parts of the EFTP emerge clearly and distinctly through the spatiotemporal characteristics of precipitation variations. However, the severity degree of drought in situ remains uncertain.

Fig. 5 Spatial distribution of multi-year averaged AHS (unit, K s^{-1}) > 500 hPa over the TP and adjacent regions during 1961–2014. The black curve refers to the 3000-m boundary of the TP. The purple box denotes the maximum center of multi-year averaged AHS over the eastern platform of the TP

Fig. 6 **a** Section showing meridional correlations between multi-year averaged AHS over the eastern platform of the TP (Fig. 5, purple box) and V and W wind components along 97.5°–105° E (1961–2014). **b** Section presenting zonal correlations between multi-year averaged AHS

over the eastern platform of the TP (Fig. 5, purple box) and U and W wind components along 26°–39° N during 1961–2014. Colors indicate significance at >90% confidence level. The plateau section is colored black

4 Severity degree of drought in the EFTP based on SPI analysis

The SPI is used worldwide for the study of drought (Stagge et al. 2013; Wang et al. 2015; Zarch et al. 2015). In this study, the SPI has been employed to determine the degree of severity of monthly drought at 438 rainfall stations over the period 1980–2014 (Table 1; Fig. 7). Overall, 226 months with dry conditions (representing 56% of total), 96 months with wet conditions (23%), and 87 months with nearly normal conditions (21%) are recorded over this period. Of the 226 months with drought conditions, extreme drought, severe drought, and drought occur for 32 (14.1%), 59 (26.1%), and 135 (59.7%) months, respectively (Table 2). Drought has evidently been more severe recently, and the drought intensity peak occurred during an

Fig. 7 Trend of monthly drought degrees in the EFTP based on SPI classes (1980–2014). The gray-shaded area indicates near normal conditions. SPI values >0.5 indicate wet conditions, while SPI values <−0.5 indicate drought conditions. Different dashed red lines represent different SPI levels of −0.5, −1.3, and −1.6, respectively. The red circle indicates most severe drought, with minimum SPI values

exceptionally dry year in 2011, with an SPI value of nearly −3.0 (Fig. 7, blue circle). This result is identical with that shown in Fig. 3 and is similar to findings of previous research (Lu et al. 2014; Ma et al. 2017).

Furthermore, the spatial pattern of drought degree in the EFTP has been investigated using SPI classes, according to the number of drought months at 438 rainfall stations during the 1980–2014 period (Fig. 8). When SPI values are between −0.5 and −1.3, major parts of the research area experience droughts. The maximum number center of drought months is observed in the southeast TP (26°–29° N, 104°–107° E), located on the eastern slope of the TP (Fig. 8a). When SPI values are between −1.3 and −1.6, the most severe droughts occur on the eastern slope of the TP (30°–35° N, 104°–108° E). The maximum number centers of severe drought months are situated between the upper reaches of the Yellow River and the Yangtze River on the eastern slope of the TP (Fig. 8b, red circles). When SPI values are <−1.6, the main areas subject to extreme droughts are concentrated <3000 m, and all maximum number centers of extreme drought months are also located between the upper reaches of the Yellow River and the Yangtze River on the eastern slope of the TP (Fig. 8c,

Table 2 The numbers of drought months, based on SPI classes

SPI	Number of months
Extreme dry ($\text{SPI} \leq -1.6$)	32
Severely dry ($-1.6 < \text{SPI} \leq -1.3$)	59
Dry ($-1.3 < \text{SPI} \leq -0.5$)	135
Nearly normal ($-0.5 < \text{SPI} \leq 0.5$)	87
Wet ($\text{SPI} > 0.5$)	96

Fig. 8 Spatial patterns relating to drought months (unit, number) in the EFTP (1980–2014) at different SPI levels: **a** $-1.3 < \text{SPI} \leq -0.5$, **b** $-1.6 < \text{SPI} \leq -1.3$, and **c** $\text{SPI} \leq -1.6$. Black dots represent rainfall stations. The north and south blue curved lines represent the Yellow and Yangtze

Rivers, respectively. The gray line shows the 3000-m contour. The red circles in **b** and **c** indicate the maximum number centers of drought months

red circles). The number centers of drought months in different drought level have clear patterns, with elevations < 3000 m tending to suffer severe and extreme drought

events, implying that such events easily occur on the eastern slope of the TP. It is worth noting that most of the maximum number centers of severe and extreme drought

months occupy the region between the upper reaches of the Yellow River and the Yangtze River on the eastern slope.

To analyze changes in drought duration in the EFTP in recent decades, variations in the number of drought months with SPI values < -0.5 are analyzed at each station through the GEV distribution method. The 10-year running window is used to compare extreme drought every 15 years. Total drought months for all stations in the study area have increased when drought months were > 140 each month (Fig. 9, purple arrow). Results thus suggest that extreme droughts are becoming more frequent.

5 Summer precipitation prediction in the EFTP based on a statistical PLS prediction model

Precipitation always occurs during summer in the EFTP, leading to local drought (Jiang et al. 2015). Precipitation prediction is clearly crucial for drought research in this region, and for this purpose, we select stations on the eastern slope of the TP, at which precipitation declines significantly (at the 90% confidence interval) in summer (Fig. 4 inverted triangle). Dryness and wetness conditions in the EFTP are closely related to the southerly moisture transport characteristics of the TP-monsoon. Moisture from the Bay of Bengal is transported eastward towards the SCS and Philippine Sea from spring to summer, eventually affecting the TP (Xu et al. 2002). Zhu et al. (2011) also demonstrated that southerly water vapor from the SCS and the western Pacific bypasses the TP, resulting in the summer TP drought.

The atmosphere is a dissipative nonlinear system with a predictability limit of about 2 weeks (Lorenz 1963; Li and Chou 1997). The preceding May–June–July (MJJ) averaged column-integrated MWVT is therefore chosen

Fig. 9 GEV distribution of mean drought months at different stations (1980–2014) for PDF. The green, blue, and red lines represent GEV results per 15 years

to predict summer precipitation. The correlation between preceding MJJ-averaged column-integrated MWVT and summer precipitation is illustrated in Fig. 10. Summer precipitation is obtained from stations on the eastern slope of the TP which have experienced significant rainfall reduction (at the 90% confidence interval). A significant positive correlation appears in the SCS, Philippine Sea, and tropical western Pacific in the area of 0° – 25° N and 110° – 150° E. This significant research area of MJJ-averaged column-integrated MWVT is applied in a PLS regression prediction model.

PLS regression is employed to identify the dominant PLS components of MJJ-averaged column-integrated MWVT variations (hereafter referred to as PLS modes) preceding summer rainfall as the predictors. The PLS modes can best explain the covariance between MJJ-averaged column-integrated MWVT variations and summer rainfall variations simultaneously; conversely, the empirical orthogonal function (EOF) mode can only explain MJJ-averaged column-integrated MWVT.

To determine the contribution of the predictors to summer rainfall in the EFTP, the 46-year training period (1961–2006) is used to build the PLS model for prediction. An 8-year hindcast period (2007–2014) is selected for testing the forecast ability of the PLS prediction model. The PLS formula for prediction is given as follows:

$$Rainfall = MWVT \times \beta + Y_{residual} \quad (18)$$

where $Y_{fit}(nt) = MWVT \times \beta$, β is the coefficient of the PLS model, and nt represents years. For instance, the summer rainfall forecast for 2007 is based on the preceding MJJ-averaged column-integrated MWVT (1961–2007) and β is obtained from the PLS modes employing the preceding MJJ-averaged column-integrated

Fig. 10 Correlation between MJJ-averaged column-integrated MWVT (units, $\text{g m}^{-1} \text{s}^{-1}$) and summer precipitation (units, mm) in the EFTP (1961–2004). Colored areas are significant at $> 95\%$ confidence level. Summer precipitation is obtained from stations on the eastern slope of the TP where precipitation has declined significantly (at the 90% confidence level)

Fig. 11 Summer precipitation (unit, mm) forecast for the EFTP based on the PLS prediction model. The black curve (1961–2014) is observed summer precipitation. The red curve (1961–2006) is summer precipitation, reconstructed with PLS modes of preceding MJJ-averaged column-integrated MWVT variations. The blue curve (2007–2014) is PLS model forecast results

MWVT and summer rainfall (1961–2006). Results $Y_{fit}(nt)$ are the fitted time series for the period 1961–2006. Finally, the summer rainfall forecast for 2007 is obtained from the last Y_{fit} and residence.

The PLS prediction model has shown a remarkable skill in reconstructing summer rainfall (Fig. 11, red curve) for the training period, and the coefficient between observed and reconstructed summer rainfall time series is 0.81, significant at the 99% confidence level. For the 2007–2014 forecast period, the PLS prediction model performs reliably (Fig. 11, blue curve). The coefficient between observed and forecasted summer rainfall is 0.67, significant at the 95% confidence level.

Our results suggest that southerly moisture transport from the SCS, Philippine Sea, and tropical western Pacific in early summer could be a precursor of summer rainfall in the EFTP. Improved predictive abilities are expected to contribute to summer drought prevention in the EFTP.

6 Discussion and conclusions

This study aims to elucidate the spatiotemporal distribution of drought in the EFTP, focusing on an analysis of drought severity and duration and on improving the ability to predict droughts using the rainfall variable.

First, spatiotemporal drought distribution in the EFTP is studied. An apparent decreasing rainfall trend is observed in 1991–2014, indicating recent drought. Areas with a decreasing rainfall trend are mainly concentrated on the eastern slope of the TP, with drought here attributed to local downward air flows induced by the TP heat source. Second, drought severity and duration are analyzed based on SPI classes and the GEV method, respectively. Results indicate that drought has become more severe, with severe and extreme droughts tending

to occur on the eastern slope of the TP. Extreme droughts are also becoming more frequent.

To help in preventing drought hazards, the early summer column-integrated MWVT from the SCS, Philippine Sea, and tropical western Pacific is used to predict summer precipitation in the EFTP. A statistical PLS prediction model is built to predict summer rainfall, using the PLS components of preceding MJJ-averaged column-integrated MWVT. Due to its excellent forecasting ability, wide variable coverage, and multiple correlation solving capabilities, the PLS prediction model is widely used in statistical prediction (Ye et al. 2017; Yu et al. 2017); for example, Wu and Yu (2015) analyzed the key role of the mega-El Niño Southern Oscillation (ENSO) in EAM seasonal prediction utilizing the PLS model. In this research, a PLS model is used for rainfall prediction in the EFTP. This shows excellent performance in reconstructing summer rainfall for 1961–2006. Promising forecast abilities are also shown for the period 2007–2014. Correlation coefficients of the observation with the reconstructed and forecast summer rainfall are 0.81 and 0.67, respectively. Results obtained suggest that southerly moisture transport in early summer would enable summer drought prevention in the EFTP.

Previous studies have observed a decreasing rainfall trend in the EFTP, noting that rainfall reduction often occurs on the TP slope (Yang et al. 2011, 2014; Zhao et al. 2016). Our results confirm these conclusions and we attribute rainfall reduction on the eastern slope of the TP to local downward air flows induced by the TP heat source.

This study indicates a trend towards more severe drought and provides a model for reliable summer precipitation prediction in the EFTP on an inter-annual timescale, which would provide a powerful supplement for drought research in situ. However, seasonal and intra-annual drought characteristics still require further investigation. In addition, only station observations and reanalysis datasets are employed in this

research. In future, use of more accurate satellite data and current climate models is warranted.

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