An Evaporation Correction Approach and Its Characteristics

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ABSTRACT

Evaporation is a principal factor in the hydrological cycle and energy exchange; however, estimations of evaporation include large uncertainties. In this study, a modified estimation of evaporation based on empirical linearly simplified Penman evaporation (PES) is proposed, soil moisture and precipitation are used to correct the land surface evaporation estimation, and the temporal and spatial characteristics of the corrected evaporation (CE) are investigated globally. The results show that CE is strong at low latitudes and weak at high latitudes. CE has obvious seasonal variation, ranging from 0.2 to 4.0 mm day⁻¹; CE is prominent in summer but feeble in winter. Compared to PES, CE is generally weaker in most regions, especially in arid regions, with differences of more than 9 mm day⁻¹. CE agrees well with evaporation derived from FLUXNET-Model Tree Ensemble (FLUXNET-MTE), MERRA, and GLDAS. In general, the root-mean-square error (RMSE) between annual CE and FLUXNET-MTE is less than 0.2 mm day⁻¹, and CE is about 5%–10% less than the evaporation of FLUXNET-MTE. In the arid regions, the maximum CE almost occurs in the month with the strongest precipitation; in the tropical regions, soil moisture enhances CE only when precipitation is less. In the context of global temperature rise, PES always shows an apparent increasing trend due to the water supply is not considered; however, CE decreases in western Asia, the western United States, the Amazon basin, and Central Africa, but weakly increases in the other study regions from 1984 to 2013. This study provides a method for estimating evaporation considering more restrictive factors on evaporation.

1. Introduction

Evaporation acts as an exchange function for soil moisture and air water vapor at the land–atmosphere interface (Monteith 1965; Yan and Chen 1990; Ali et al. 2008), and consumes more than half the solar energy absorbed by the land (Kiehl and Trenberth 1997). It is a critical component when measuring the underlying surface water balance and plays an important role in the global energy balance (Monteith 1965; Qian and Li 1996). There is a “complementarity” between the atmospheric and land moisture. In other words, evaporation is a balance produced between atmospheric vapor and soil moisture (SM). In practice, the problem of the accuracy of the evaporation estimation seriously interferes with our understanding of the hydrological cycle.

Evaporation is influenced by the water supply, dynamic and thermodynamic factors, and soil properties. In general, in arid regions, evaporation is limited by water, while in wet regions, evaporation is limited by atmospheric demand. Obviously, estimations of evaporation are affected by various factors in different regions. In addition, some studies suggest that evaporation cannot be measured directly from space at high resolution as a water variable; it must be physically derived as an energy variable such as the latent heat flux (Fisher et al. 2017). Accordingly, estimating evaporation is an open issue because it is regulated by multiple factors such as wind, temperature, specific humidity, soil, and vegetation.

Due to the difficulties involved in acquiring actual evaporation data, estimations of evaporation have received much attention. Traditional methods for single-point evaporation estimations include the Bowen ratio–energy balance method (Bowen 1926), the Penman method (Penman 1948), the aerodynamics method (Pelton 1960), the eddy covariance method (Swinbank 1955), the Penman–Monteith method (Monteith 1965),

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and many other approaches that have expanded on the Penman method (Wright 1982). The Penman method, which is widely used in the estimation of potential evaporation, combines energetic and atmospheric drivers (Fisher et al. 2011), but it does not consider explicit vegetation and the heat exchange with the ground. At coarse spatial–temporal scales, evaporation can be estimated using the water balance and hydrothermal method, which integrates the water and heat, and assuming that precipitation is the only source of water. However, while precipitation is the ultimate water supply source, it is not the only source available for evaporation; the soil can store antecedent precipitation (Li et al. 2016a) and provide another possible source for evaporation. SM can help improve evaporation estimations (Entekhabi et al. 2010; Purdy et al. 2018) because it represents the water supply conditions at the evaporation surface and has a continuous impact on evaporation. The results estimated by Li et al. (2016b) indicate that the evapotranspiration values range from 4 to 6 mm day^{−1} in the Northern Hemisphere (NH) in summer and even reach 8 mm day^{−1} in some regions. However, the average annual evaporation calculated via the water balance method is less than 3 mm day^{−1} in many regions of the NH. The large difference in the calculation results indicates that further studies on evaporation are necessary. Compared to other climatic regions at the same latitude, the average temperature in arid regions is higher and the SM content is lower. A distinct characteristic of arid regions is a precipitation deficiency. Insufficient precipitation further causes water budget imbalances, and the water budget has a significant impact on the hydrology of arid regions (Thornthwaite 1948). There is abundant precipitation in the tropical regions, which are important land water vapor sources globally, and precipitation recycling is strong there (Su et al. 2014).

This study explores a new global evaporation estimation approach that considers the water supplies of precipitation and SM based on a simplified Penman method. Subsequently, the corrected approach is examined by comparing the results of the new approach to those of the simplified Penman method and the traditional hydrothermal method which is calculated using only the precipitation. The contributions of influential evaporation factors in arid and humid regions as well as the trend of evaporation for the period of 1984–2013 are analyzed to improve our knowledge of the characteristics and the variations in evaporation. These results can benefit our deep understanding of hydrological cycle characteristics and their variability.

This paper is organized as follows. Section 2 describes the data, while section 3 proposes a corrected approach

for estimating evaporation. The characteristics of the evaporation with the corrected approach are given in section 4, and the evaluations of the evaporation estimation compared to FLUXNET–Model Tree Ensemble (FLUXNET-MTE), which is derived by empirical up-scaling of eddy covariance measurements from a global network of flux towers (FLUXNET), using a model tree ensemble (MTE) approach (Jung et al. 2011), and re-analysis data of Modern-Era Retrospective Analysis for Research and Applications (MERRA) and the Global Land Data Assimilation System (GLDAS) are shown. An analysis of the influence factors is presented in section 5. Section 6 shows the variations in the corrected evaporation (CE) over the period of 1984–2013. Discussions and conclusions are presented in section 7.

2. Data description and method

a. Data

Monthly global precipitation data from 1984 to 2013 with a horizontal resolution of $0.5^\circ \times 0.5^\circ$ are obtained from the Global Precipitation Climatology Centre (GPCC); these data are based on quality-controlled data from 67 200 stations worldwide with record duration of 10 years or longer. The surface net radiation flux, 2-m dewpoint temperature, 2-m temperature, 0–7-cm volumetric soil water, specific humidity, and U and V components of the wind data are obtained from the data product of the European Center for Medium-Range Weather Forecasts (ECMWF; ERA-Interim), which exhibits good performance for the atmosphere water budget and is commonly used in climate studies (e.g., Kauffeldt et al. 2015; Gao et al. 2014). The data cover the period of 1984–2013 with a resolution of $1.5^\circ \times 1.5^\circ$.

Three main prevalent independent evaporation datasets, including FLUXNET, MERRA-Land, and GLDAS, are used to evaluate CE. The FLUXNET dataset provides the evaporation through a global network of micrometeorological tower sites based on eddy covariance methods. The water flux is estimated by a machine-learning algorithm in FLUXNET-MTE (Jung et al. 2011). The monthly data cover the period 1984–2011 with a spatial resolution of $0.5^\circ \times 0.5^\circ$. The internal cross-validation results show that the correlation coefficient between evaporation product of MTE and FLUXNET sites data reaches $r = 0.91$, and the simulation results of Global Soil Wetness Project 2 (GSWP-2) have significant correlations with FLUXNET-MTE ($r = 0.91$) (Jung et al. 2010). The evaporation is converted by the FLUXNET-MTE latent heat by multiplying the inverse of the latent heat of vaporization. Due to the spatial distribution of the observation sites of

FLUXNET-MTE is sparse in the deserts of northern Africa and western Asia. The evaporation data from the monthly MERRA-Land (horizontal resolution of $0.667^\circ \times 0.5^\circ$ in the meridional and zonal directions) and GLDAS ($1^\circ \times 1^\circ$ grid) reanalysis datasets are also used to compare the performance of the new method. MERRA, assimilating recent satellite data from NASA and in situ observations with quality control and error corrections, can be regarded as a supplementary land surface reanalysis in the estimation of land surface hydrology (Rienecker et al. 2011). The GLDAS evaporation in the Community Land Model generates the optimal fields of the land surface data by integrating satellite and ground-based observational data products. Studies have reported that the MERRA evaporation can well reflect the temporal and spatial variation characteristics of global evaporation (Su and Feng 2015). The GLDAS evaporation is considered to be reliable because the model forcing data, including precipitation, temperature, and radiation, are observed and the models are physically based and subject to vigorous evaluations (Gao et al. 2014; Zhang et al. 2017).

b. Penman estimation methods

The Penman evaporation (PE) proposed by Penman (1948) considers the effects of heat (radiation) and dynamics (wind speed) on evaporation, as shown in Eq. (1):

$$PE = \frac{\Delta}{\Delta + \gamma} \frac{(R_n)}{\lambda} + \frac{\gamma}{\Delta + \gamma} \frac{6.43(f_u)D}{\lambda}, \quad (1)$$

where PE is the potential open-water evaporation (mm day^{-1}); R_n is the net radiation at the surface (MJ m day^{-1}); Δ is the slope of the saturation vapor pressure curve ($\text{kPa } ^\circ\text{C}^{-1}$); γ is the psychrometric coefficient ($\text{kPa } ^\circ\text{C}^{-1}$); λ is the latent heat of vaporization (MJ kg^{-1}); f_u is the wind function; and D is the vapor pressure deficit (kPa).

Valiantzas (2006) simplified the Penman equation by computing the evaporation from readily available measured data; this simple algebraic formula is equivalent to the PE in accuracy. The simplified Penman evaporation (PES) is computed as

$$PES \approx 0.047R_s\sqrt{T+9.5} - 2.4\left(\frac{R_s}{R_A}\right)^2 + 0.09(T+20)\left(1 - \frac{RH}{100}\right), \quad (2)$$

where R_s is the solar radiation (MJ m day^{-1}), T is the air temperature ($^\circ\text{C}$) at 2 m, RH is the relative humidity, and R_A is the extraterrestrial radiation. In Eq. (2), the wind speed is set to be constant (2 m s^{-1}); therefore,

discrepancies in PES will inevitably increase in regions where the wind speed varies significantly. However, the results of Valiantzas (2006) show that the relative error in the estimation by Eq. (2) is small (only 4%) compared to Eq. (1), which does not require wind speed data and has a better calculation efficiency. The PE is a potential evaporation method assumes the underlying surface is sufficiently hydrated. In practice, the underlying surface is complexly limited by the water supply, and the actual evaporation is less than the PE. The actual evaporation would be equal to the PE only when the water supply is particularly sufficient.

3. An approach to estimating the evaporation

Equation (2) can be approximately regarded as the Penman evaporation due to the small discrepancy between the traditional Penman method and the PES. In this study, PES is treated as the evaporation capacity. Figure 1a shows the precipitation distribution. According to the general definition, there are six main arid regions with less than 200 mm of annual precipitation throughout the world (the red boxes in Fig. 1a). And three typical humid regions where the annual precipitation is more than 1500 mm are also shown in Fig. 1a (the orange boxes). Figure 1b presents the distribution of PES, illustrating that PES varies with latitude and decreases toward high latitudes coinciding with decreases in the temperature. The high latitudes and the Tibetan Plateau are weak evaporation regions. North Africa, western Asia, the western United States, Australia, and southern Africa are regions with high PES, more than 8.0 mm day^{-1} , where there is less rainfall. The difference between the annual PES and precipitation is shown in Fig. 1c, which indicates that PES is greater than the precipitation in most regions, especially in China–Mongolia, central Asia, western Asia, North Africa, the western United States, western Namibia, and northern Australia, which represent arid and semiarid regions. These positive differences imply that the evaporation capacity and the vapor condition are not in balance and these regions retain their arid characteristics without sufficient precipitation, which is different from the distribution of arid regions. This is because the calculation of PES assumes that the underlying surface is a water surface and that water vapor can be provided continuously. The actual evaporation should be less than PES due to the limited underlying water supply.

In the hydrothermal balance method, precipitation reflects the conditions of the land surface water supply. Fu (1981a) proposed an estimation of the land surface evaporation based on the evaporation capacity that

FIG. 1. (a) Distribution of the total annual precipitation P (mm), (b) the open-water underlying surface simplified Penman evaporation (PES; mm day^{-1}), and (c) the difference between the PES and precipitation calculated for the period of 1984–2013 ($\text{PES} - P$; mm day^{-1}).

considers the effect of the moisture supply and the degree of wetness in dry and humid climate regions:

$$ER \approx \begin{cases} P \left[\frac{PES}{P} - \frac{1}{m} \left(\frac{PES}{P} \right)^m \right], & \frac{PES}{P} < 1, \quad m = 2 \\ PES \left[\frac{P}{PES} - \frac{1}{m} \left(\frac{P}{PES} \right)^m \right], & \frac{PES}{P} > 1, \quad m = 3 \end{cases}, \quad (3)$$

where ER is the land surface evaporation, P is precipitation, and m is a characteristic parameter of the underlying surface. Under humid climate conditions ($PES < P$), $m = 2$, and under dry climate conditions ($PES > P$), $m = 3$.

Equation (3) shows that, when the evaporation capacity (PES) is much larger than the precipitation, the maximum ER is approximately equal to the precipitation. According to the water balance, local simultaneous precipitation is not the only water source for evaporation; antecedent precipitation stored in the soil is also an important factor that affects evaporation in a continuous and slow process. Therefore, SM should be considered in Eq. (3). Kelliher et al. (1995) and Su et al. (2014) proposed to use SM to improve evaporation estimations. The relationship between SM and potential evaporation is thought to be linear for actual evaporation (Allen et al. 1998; Walter et al. 2000). However, Fu (1981b) suggested that evaporation from SM follows a complicated process. The estimation of the actual evaporation assumes that all the SM can evaporate into the atmosphere (Li 2017); however, the important factor of SM retention is ignored. According to Jung et al. (2010), evaporation cannot continue with limited moisture in the soil; furthermore, the air vapor pressure needs to be less than the surface vapor pressure of the soil for water vapor to be transported to the atmosphere via diffusion or convection. Therefore, the water in the soil cannot be completely evaporated into the atmosphere.

Studies suggest that the mean SM memory can reach approximately 30 days in the NH (Dirmeyer et al. 2009; Li et al. 2016a) and that the global time scale of the SM evaporation can reach 42 days (Wang-Erlandsson et al. 2014). Therefore, land evaporation is defined as the sum of ER and the monthly changes in SM, $\Delta SM_i = SM_{i+1} - SM_i$, where ΔSM_i indicates the changes in SM in the i th month. When the value of SM is negative, the moisture in the soil is assumed to evaporate into the atmosphere. Considering the changes in soil moisture cannot be completely used for evaporation, some of them could also transfer to runoff or infiltration. To avoid overcompensating, a coefficient Ψ , which quantifies the sensitivity of soil moisture to precipitation (Wei et al. 2008),

is adopted by using monthly average soil moisture v_{sm} and precipitation v_p :

$$\psi = \frac{v_{sm}}{v_p}, \quad (4)$$

where

$$v_x = \frac{1}{n} \sum_{i=1}^n \left[\frac{1}{x_i} \sqrt{(x_i - \bar{x}_i)^2} \right], \quad (5)$$

the x denotes soil moisture (sm) or precipitation (p). The term x_i is the monthly average of the variable x in i month, and n is the number of months of 30 years; $\bar{x}_i$ is the average for n months.

The corrected evaporation (CE) is shown in Eq. (6):

$$CE = \begin{cases} ER + \psi |\Delta SM|, & \Delta SM < 0 \\ ER, & \Delta SM \geq 0 \end{cases}. \quad (6)$$

4. Characteristics of the evaporation using the corrected approach

The annual and seasonal evaporation corrected by the precipitation and the changes in SM are presented in Fig. 2. On an annual scale, CE is larger than 3.0 mm day^{-1} at 15°S – 15°N but less than 1.4 mm day^{-1} at high latitudes, that is, 45° – 90°N (Fig. 2a); the CE has its minimum in arid and cold regions, less than 0.4 mm day^{-1} , which might be attributed to less precipitation and lower temperatures. In terms of seasonal variations, from March to May (Fig. 2b), the CE is much stronger, approximately 2.2 mm day^{-1} in eastern America, western Eurasia, and at low latitudes over the NH, and it is more than 3.2 mm day^{-1} over the regions of 0 – 15°S . From June to August (Fig. 2c), CE in most parts of the NH reaches its maximum. It is larger than 2.6 mm day^{-1} , except in arid regions. A conspicuous feature of CE is that it is small in North Africa, western Asia, and northwestern China, which are prominent arid regions. In most of the land areas at south of 15°S , CE is less than 0.6 mm day^{-1} . From September to November (Fig. 2d), CE decreases in the NH but increases in the Southern Hemisphere (SH). The largest CE is shown at 20°S – 10°N and on the eastern coasts of China, where precipitation is abundant, and the temperature is higher than in the high-latitude zone. In most of the NH, the CE from December to February of next year is less, which is nearly zero (Fig. 2e). In most regions of SH, CE is greater than 3.2 mm day^{-1} except in arid regions. CE is stronger in summer and weaker in winter. Compared to PES, CE obviously differs on annual and seasonal scales (Figs. 3a–e). In particular, large differences occur in arid

FIG. 2. Corrected evaporation (CE; mm day^{-1}) distribution, calculated during the period from 1984 to 2013 via the precipitation and monthly changes of soil moisture on the basis of the PES: (a) annual, (b) spring, (c) summer, (d) autumn, and (e) winter.

regions. In other words, the CE over land is much less than the potential evaporation, especially in arid regions with difference more than 8.0 mm day^{-1} . Annual mean results of the difference between CE and ER indicate that ER coincides well with CE in most high-latitude regions (Fig. 3f); however, over freezing ground regions in NH, CE is obviously higher than ER from March to August, while in equatorial areas CE is stronger than ER all seasons (Figs. 3g–j). From March to May, snow melting and ice thawing in seasonal freeze–thaw regions, such as western Europe, leads to more liquid water stored in soil, which can even reach saturation; meanwhile, due to less precipitation, the rapid increase of near-surface temperature and wind speed strengthen the evaporation capacity and lead to the rapid decrease of SM, and then CE increases accordingly (Fig. 3g). In the SH, SM has an obvious promoting effect on evaporation at low latitude. From June to August, as the air temperature increases over the NH, the large value regions of difference move northward (Fig. 3h). From

September to February of next year (Figs. 3i,j), CE is the same as ER in the north of 15°N . In the SH, $0^\circ\text{--}20^\circ\text{S}$, the SM enhances the evaporation obviously. These results not only illustrate the SM change in freeze–thaw areas, where the SM change is prominent (Yang et al. 2016) and would promote evaporation, but also indicate that the impacts of SM at different latitudes on evaporation exhibit distinct seasonal differences.

In general, if evaporation is greater than precipitation, the region should be identified as a source of vapor and can be defined as hydrologically arid. The annual precipitation is lower than evaporation in some regions where the climate is becoming drier. Figure 4 shows the spatial distributions of humid and arid regions, that is, the difference between the CE and precipitation. The annual CE is about $0.2\text{--}0.4 \text{ mm day}^{-1}$ greater than precipitation in the western United States, North Africa, western Asia, China–Mongolia, southern Africa, and South Australia around the globe, where are remarkable arid regions. The distributions of arid regions defined by

FIG. 3. (left) Difference between CE and PES (CE - PES; mm day⁻¹) and (right) difference between CE and ER (CE - ER; mm day⁻¹) for (a),(f) annual, (b),(g) spring, (c),(h) summer, (d),(i) autumn, and (e),(j) winter.

FIG. 4. Annual difference between CE and precipitation (mm day^{-1}), i.e., CE minus precipitation.

the differences between CE and precipitation basically agree with previous definitions of arid regions (Hulme 1996; Kocurek 1998; Thomas 1997). This agreement implies that the corrected approach reasonably represents the land evaporation and can distinguish between dry and wet regions.

5. Preliminary verification of the corrected evaporation estimation

To evaluate the annual performance of the CE, the root-mean-square errors (RMSEs) between CE and the FLUXNET-MTE, MERRA, and GLDAS evaporation are calculated:

$$\text{RMSE} = \sqrt{\frac{\sum_{j=1}^m (\text{CE} - E)^2}{m}}, \quad (7)$$

where E is the evaporation in FLUXNET-MTE, MERRA, or GLDAS and m is the number of measurements (number of years). The RMSE represents the deviation between CE and the FLUXNET-MTE, MERRA, or GLDAS evaporation values.

a. Comparison between CE and evaporation in the FLUXNET-MTE, MERRA, and GLDAS

Figure 5 shows the RMSE and bias ratio between evaporation of FLUXNET-MTE, MERRA, and GLDAS and CE, PES, respectively. The RMSE between FLUXNET-MTE and CE is about 0.2 mm day^{-1} in most areas, and is larger at lower latitude at 5°S – 5°N , more than 0.4 mm day^{-1} (Fig. 5a). In most mid- to high latitudes, the RMSE between CE and MERRA is less than 0.3 mm day^{-1} , indicating that CE is consistent with evaporation of MERRA (Fig. 5b). In lower latitudes (5°S – 5°N), where the annual CE is more than 3.4 mm day^{-1} , the RMSE is approximately 1.0 mm day^{-1} . The RMSE

between CE and GLDAS is less than 0.3 mm day^{-1} in most areas but is more than 0.7 mm day^{-1} at lower latitudes (Fig. 5c), indicating that there is a slight bias in CE and evaporation of GLDAS at low latitudes. The annual CE variation coincides well with FLUXNET-MTE globally and closes to the evaporation of GLDAS and MERRA in most regions, the evaporation of MERRA is greater than GLDAS. Correspondingly, the RMSE between PES and FLUXNET-MTE is around 2.0 – 3.0 mm day^{-1} over the nonarid regions. In southern Africa, Australia, western Asia, and the western United States, RMSE is more than 6.0 mm day^{-1} (Fig. 5d). It notes that the RMSEs between PES and evaporation of MERRA and GLDAS are greater than 6.0 mm day^{-1} in North Africa, and are about 3.0 – 4.0 mm day^{-1} in the middle and low latitude (Figs. 5e,f). Compared with Figs. 5a–c, the RMSEs showed in Figs. 5d–f are one order of magnitude as large as those RMSEs calculated by CE. It means that CE is much better coincide with the evaporation products.

Figure 5g illustrates that, in most regions of the globe, the bias between CE and FLUXNET-MTE is about -10% . The difference between CE and FLUXNET-MTE is small. However, the bias ratio between PES and FLUXNET-MTE is more than 100% , even more than 400% in the arid regions (Fig. 5h). The evaporation calculated by the corrected approach is evidently improved, compared to the PES. It is suggested that the estimation of CE can show the characteristics of land evaporation distribution and variation reasonably.

b. The monthly variation in CE and meteorological factors

To further examine how reasonable the CE is, factors affecting and changing the characteristics of evaporation in arid and humid regions are investigated. The definition of arid regions proposed by Hulme (1996) and Kocurek (1998) applies to regions with the annual precipitation of less than 200 mm . Most arid regions result from the descending movement of the meridional Hadley circulation in subtropical areas. Some arid regions can be attributed to areas where it is difficult to achieve vapor convergence due to their long distance from an ocean or their large topographic relief (Voice and Hunt 1984; Qian et al. 2017). On the contrary, the convection is strong and the moisture in the atmosphere is sufficient in tropical regions, where the annual rainfall is more than 1500 mm .

To examine the dominant factors in arid and humid regions, Fig. 6 shows the monthly variation of SM, precipitation, the vapor pressure deficit (VPD) indicating the air saturation condition, the difference between the ground and air temperatures $T_s - T$, and the 2-m air

FIG. 5. The RMSE (mm day^{-1}) and bias ratio between evaporation of independent datasets and CE and PES, respectively. (a)–(c) The RMSE between CE and evaporation of FLUXNET-MTE, MERRA, GLDAS, (d)–(f) the RMSE between PES and evaporation of FLUXNET-MTE, MERRA, GLDAS, and (g), (h) the bias ratio (CE and PES minus FLUXNET-MTE divided by FLUXNET-MTE, respectively).

temperature T . In the China–Mongolia arid region (Fig. 6a), precipitation is more than 25 mm from June to August. In spring, the temperature starts to rise, and the ice in the soil starts to thaw. The SM peaks in March. Evaporation causes SM to decrease slightly from April to September. Figure 6a shows that the VPD, which dominates evaporation, is always greater than 0 hPa; in particular, it is stronger than 10 hPa from June to August, which implies that the evaporation capacity is strong during this period. The greater the VPD, the greater the evaporation will be while water resources are sufficient.

In the North Africa arid region (Fig. 6b), precipitation is low and SM is small. Precipitation is primarily concentrated in the period from July to September. SM changes little over the entire year but decreases slightly in September. Further, Fig. 6b shows that temperature and VPD are positive all year in this region. Warming and drought environmental conditions maintain strong evaporation capacity; however, due to insufficient water resources, the actual evaporation is small. In western Asia (Fig. 6c), where the climate is similar to a Mediterranean climate, precipitation is primarily concentrated in winter and spring, and the minimums in precipitation occur in June. SM decreases during

March–July, and at its minimum in July. Note that VPD, temperature, and the difference between the ground and air temperatures are positive over the entire year, implying that the evaporation capacity is strong; however, SM and precipitation are insufficient, causing the amount of actual evaporation to be small. Compared to the other three arid regions in the NH, there is more precipitation in the western United States, especially from July to September, more than 50 mm (Fig. 6d). SM starts to decrease from March to June, when the surface temperature is higher than 0°C and the VPD is strong, which illustrates that the evaporation capacity is strong; then, the SM maintains a low value. Due to the high precipitation from July to September, the SM correspondingly begins to increase, and then SM decreases as precipitation decreases again. In the arid regions of the SH (Figs. 6e,f), the temperature is always higher than 0°C , and VPD is stronger than 0 hPa in all months. The precipitation in Australia from November to March in next year is more than 10 mm; SM decreases from March to April and in August, and the CE is much stronger. In southern Africa, precipitation is concentrated in the period from January to March; SM decreases from April to June.

In low-latitude humid regions, such as the Amazon basin, Central Africa, and southeast China (Figs. 6g–i),

FIG. 6. Monthly variation in the factors affecting evaporation, SM (mm; white bar), P (mm; black bar), VPD (hPa; circle line), T ($^{\circ}\text{C}$; square line), and $T_s - T$ ($^{\circ}\text{C}$; triangle line) over (a) China–Mongolia, (b) North Africa, (c) western Asia, (d) the western United States, (e) Australia, (f) southern Africa, (g) the Amazon basin, (h) Central Africa, and (i) southeast China.

the annual rainfall exceeds 1500 mm and the soil moisture is sufficient. From June to September in the Amazon basin, precipitation is small and SM decreases. In Central Africa, the precipitation is less from both June to August and from December to February, the SM decreases in the same period, which implies more evaporation occurring. Similarly, there is more precipitation in southeast China in summer. The temperature is always higher than 20°C throughout the year in the Amazon basin and Central Africa, and it is more than 25°C in summer in southeast China, the high temperature appears to benefit to evaporate. However, in these humid regions, VPD is always nearly 4 hPa, which is smaller than arid regions, and surface temperatures are smaller than air temperature, apparently, the evaporation capacity would be restrained.

Figure 7 shows the monthly evaporation of PES, ER, and CE. In China–Mongolia (Fig. 7a), PES is much stronger than CE. The largest difference between CE and ER appears in April and June. The SM increases and then reaches its maximum in March, afterward, because of the increasing evaporation capacity, SM decreases rapidly, and CE becomes stronger, especially in May (Fig. 6a). The precipitation in North Africa (Fig. 7b) is less than that in other arid regions, and evaporation is less because of the limited water supply

there. CE is maximal in the month with the most precipitation, August. From February to June in western Asia, the transportation of SM into the atmosphere enhances the evaporation. In April, the SM contribution to CE is at its maximum for the year. This result is due to the large amount of precipitation in March (Fig. 6c). In the western United States (Fig. 7d), CE is strong from August to October, more than 1.5 mm day^{-1} . In Australia and southern Africa, the SM decreases in autumn and winter, in this period, the loss of SM is conducive to the evaporation (Figs. 7e,f). Figure 7 also indicates that there is a large difference between the CE and the PES in the arid regions. In the humid regions (Figs. 7g–i), the difference between CE and PES, which is much smaller than arid regions, is about 1 mm day^{-1} . In these humid regions, the monthly variation of CE is consistent with the variation of precipitation; it means that CE in humid regions is also sensitive to the water supply. It is worth noting that the increasing and decreasing trend of CE is contrary to PES in the tropical humid regions such as the Amazon basin and Central Africa. This phenomenon may be similar to the Budyko hypothesis, namely, if the energy condition is constant, the potential evaporation will decrease as the precipitation increases (Su and Feng 2015).

Figures 6 and 7 show that SM promotes CE one month later than the precipitation increase in arid regions;

FIG. 7. Monthly variation of the three different evaporation models over (a) China–Mongolia, (b) North Africa, (c) western Asia, (d) the western United States, (e) Australia, (f) southern Africa, (g) the Amazon basin, (h) Central Africa, and (i) southeast China (ER, mm day^{-1} , white bar; CE, mm day^{-1} , black bar; PES, mm day^{-1} , solid circle line).

somewhat differently, SM enhances the CE when precipitation in the humid region is less. There is less SM in arid regions, and only more precipitation would increase SM; however, due to the strong potential evaporation, then SM will be quickly evaporated. In the tropics, abundant precipitation provides sufficient moisture for evaporation; when the precipitation decreases, soil moisture can transfer to the atmosphere gradually.

6. Changes in CE and PES over the last 30 years

Global temperature has increased over the last 30 years (Guan et al. 2017). To investigate the changes in evaporation with temperature increase, Fig. 8 shows that, in the last 30 years from 1984 to 2013, there are different changes in CE and PES. CE has a weakly increasing trend in China–Mongolia (Fig. 8a), North Africa (Fig. 8b), Australia (Fig. 8e), southern Africa (Fig. 8f), and southeast China (Fig. 8i) for 4.02, 4.02, 13.1, 2.19, and 25.19 mm decade^{-1} , respectively. However, the increasing trends of CE are much smaller than those of PES. The CE decreases in western Asia (Fig. 8c), the western United States (Fig. 8d), the Amazon basin (Fig. 8g) and Central Africa (Fig. 8h), and the annual variability rates of CE are -5.48 , -21.17 , -14.97 , and $-2.19 \text{ mm decade}^{-1}$, however, PES has an increasing trend. CE in arid regions ranges from 0.1 to 1.2 mm day^{-1} and in humid regions CE ranges from

2.2 to 3.5 mm day^{-1} . PES significantly increases in these study regions; the maximum trend is $45.63 \text{ mm decade}^{-1}$ in western Asia, and the minimum trend is $1.10 \text{ mm decade}^{-1}$ in the Amazon basin.

Previous studies have shown that the annual precipitation has increased in North Africa and China–Mongolia but has decreased in western Asia and the Amazon basin (Li et al. 2016b; Gao et al. 2018), which implies that the CE trend might be more related to precipitation. PES primarily depends on the radiation, temperature, and VPD, which assume that the underlying surface is pure water; CE, however, is also affected by the realistic water supply in the underlying land surface. PES is an indicator of how much solar energy a region receives, rather than the change in the water content. The changes in CE are relatively complex processes. The difference between PES and CE illustrates that the evaporation ability is not the determining factor that affects the land evaporation and that the wetness degree of the underlying surface needs to be considered. Further studies on other factors such as wind and the vapor pressure deficit will be discussed in future studies.

7. Discussion and conclusions

This study proposed a new modified approach for an evaporation estimation using SM changes and

FIG. 8. Annual evolution of CE (mm day⁻¹; solid circle line) and PES (mm day⁻¹; dashed line with triangles) over (a) China–Mongolia, (b) North Africa, (c) western Asia, (d) the western United States, (e) Australia, (f) southern Africa, (g) the Amazon basin, (h) Central Africa, and (i) southeast China from 1984 to 2013.

precipitation, which is different from the traditional method of calculating the evaporation based on a water balance theory assuming that precipitation is the only source of evaporation, SM, and runoff (Fisher et al. 2011). This new approach further considers the antecedent precipitation stored in the soil effect on evaporation. Based on this, the characteristics of the evaporation calculated by modified approach are analyzed, particularly in arid regions and humid tropical regions. This new approach effectively corrects the evaporation estimation throughout the global land area. The main results are showed below.

The evaporation estimation corrected by soil moisture changes and precipitation is reasonable in most regions globally. A serial of comparisons among CE, PES, and three evaporation datasets show that CE can better reflect the evaporation distribution, and it is well consistent with FLUXNET-MTE globally. The evaporation estimated by the corrected method considers more restrictive factors on the evaporation, making CE suitable for use in estimating the evaporation in seasonal freeze–thaw regions, deserts, and humid regions.

CE illustrates the changes in the evaporation characteristics on annual and seasonal scales well, that is, CE is smaller in winter but larger in summer and is smaller at high latitudes but larger at low latitudes, with a range of 0.2–4.0 mm day⁻¹. CE can be used to distinguish between arid regions and humid regions. CE is stronger than precipitation in arid regions and, conversely, much weaker than precipitation in humid regions. Obviously, CE is approximately 10 times

smaller than PES in arid regions while is close to PES in the humid tropical regions. In arid regions, precipitation and SM limit CE; while the evaporation ability is strong in these regions, CE is at a minimum, and it can reflect the evaporation close to the actual conditions. In the humid tropical regions, low VDP might restrain more evaporate from soil in the period of abundant precipitation; oppositely, soil moisture could transfer to the atmosphere in less precipitation period, that is, soil and air moisture are complementary.

CE reveals a different trend in the evaporation temporal characteristics from PES. With increasing global temperature, the vapor content capacity of the air also increases; this further strengthens the evaporation ability. PES shows a prominent increasing trend in the last 30 years ideally without considering water supply. However, CE shows a slightly weakening trend in western Asia and the western United States in the NH and in the Amazon basin and Central Africa in the tropics. The trends of CE consistent with studies and observations that have reported that precipitation in China–Mongolia and North Africa have increased over the last 30 years, causing these two regions to experience a period with a warming and humid climate (Shi et al. 2007). In addition, CE has decreased with the decline in precipitation in western Asia, the western United States, and the Amazon basin (Li et al. 2016b; Gao et al. 2018). CE is sensitive to the water supply globally.

The approach proposed in this study only considers monthly changes in the SM and precipitation; how to describe submonthly changes in the SM and precipitation needs to be further explored.

In addition, other paths that water can take to evaporation, such as surface runoff and groundwater flow, remain unexplored.

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