Spatial variability and temporal stability of actual evapotranspiration on a hillslope of the Chinese Loess Plateau (postprint)

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Abstract

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Full Text

Spatial Variability and Temporal Stability of Actual Evapotranspiration on a Hillslope of the Chinese Loess Plateau

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Abstract: Actual evapotranspiration (ETa) is a key component of water balance. This study aimed to investigate the spatial variability and temporal stability of ETa along a hillslope and to analyze the key factors controlling its spatiotemporal variability. Potential evaporation, surface runoff, and the 0–480 cm soil water profile were measured along a 243 m transect on a Loess Plateau hillslope during normal (2015) and wet (2016) water years. ETa was calculated using the water balance equation. Results indicated that increased precipitation during the wet year did not alter the spatial pattern of ETa along the hillslope. Time stability analysis showed that a location with high temporal stability of ETa could be used to estimate the mean ETa of the hillslope. Time stability of ETa was positively correlated with elevation ($P<0.05$), indicating that, on a hillslope in a semi-arid area, elevation was the primary factor influencing the temporal stability of ETa.

Keywords: relative difference; water balance; surface runoff; soil texture; semi-arid areas

1 Introduction

Actual evapotranspiration (ETa) is an important component of the water and energy cycle, accounting for 95% of water balance in dry ecosystems (Wilcox et al., 2003). Moreover, water-limited environments cover almost half of Earth' s land surface and are expected to increase (Newman et al., 2006). Therefore, understanding spatiotemporal variations of ETa is crucial for efficient water resources management in water-limited regions (Lian and Huang, 2015, 2016).

ETa is perhaps the most difficult and complicated component of the hydrological cycle because it involves the combined effects of climate, terrain, vegetation, soil, and hydraulic properties (Thomas, 2000; Xu and Singh, 2005; McVicar et al., 2007; Vivoni et al., 2010; Zhao and Zhao, 2014; Li et al., 2017). Many studies have focused on spatiotemporal variability of ETa under different environmental conditions (Thomas, 2000; Mo et al., 2004; Gao et al., 2007; McVicar et al., 2007; Vivoni et al., 2010; Gao et al., 2019). For example, Thomas (2000) reported that evapotranspiration in arid and semi-arid areas of Northwest China decreased substantially across all seasons, primarily due to variations in wind, relative humidity, and maximum temperature. Nicholls et al. (2016) analyzed the effect of vegetation cover on ETa in the 52-hm^2 Sandhill Fen watershed of Canada and found that ETa increased as the peak leaf area index (LAI) increased during the growing season. Mo et al. (2004) simulated annual ETa over the Lushi basin in the middle reach of China' s Yellow River and suggested that the spatial pattern of annual ETa was similar to precipitation and LAI patterns. Vivoni et al. (2010) explored soil moisture and evapotranspiration distributions using a distributed hydrologic model in a semi-arid mountain basin and found hysteresis existed between soil moisture and evapotranspiration due to climate variability.

The above studies on spatiotemporal variability of ETa were mainly conducted at regional or watershed scales. The hillslope is the main unit of measurement in hydrological studies where ETa measurements are used to derive catchment-level ETa (Santiago et al., 2000; Ewers et al., 2002). However, systematic studies on spatiotemporal variability of ETa at the hillslope scale remain scarce. Carey and Woo (2001) investigated differences in ETa on four hillslopes with different vegetation types and stated that ETa was mainly limited by the phenology of deciduous forests and shrubs. Because there was only one measurement site for ETa on each hillslope, spatial variability of ETa within the hillslope was ignored. Mitchell et al. (2012) evaluated differences in ETa at different hillslope positions in a mixed-species eucalypt forest and discovered that ETa in the upper slope position was 40% lower than that in the middle and lower positions. However, because the ETa measurement sites were located on different hillslopes in the catchment, differences in ETa between sites were not necessarily due to slope position.

Many environmental factors might cause differences in ETa along a hillslope (Majidi et al., 2015), and measuring ETa at only one monitoring site makes it difficult to understand spatiotemporal variability at the hillslope scale. In this study, a transect line was established on a hillslope with a single vegetation type to obtain ETa at different positions and understand the spatial variability within the hillslope.

ETa always exhibits temporal variability at different hillslope positions. Time stability analysis can be performed to investigate spatial variability and temporal stability of ETa at the hillslope scale. This method has been successfully applied to examine spatiotemporal variability of soil moisture at hillslope and catchment scales, where the most time-stable location (MTSL) is identified to represent the mean soil moisture of the field (Vachaud et al., 1985; Starr, 2005; Hu et al., 2010; Gao et al., 2011; Gao and Shao, 2012; Jia and Shao, 2013; Hu and Si, 2014; Zhang et al., 2016; Liao et al., 2017). Recently, time stability analysis has been applied to analyze spatiotemporal variability of indices other than soil moisture, such as soil salinity and soil sulfur content (Douaik et al., 2006; Piotrowska-Długosz et al., 2017). However, no study has investigated the spatiotemporal variability of ETa using time stability analysis.

In this study, we assume that time stability analysis can be utilized to investigate spatiotemporal variability of ETa at the hillslope scale. The objectives are: (1) to investigate spatial variability of water balance components along a hillslope; (2) to explore spatial variability and temporal stability of ETa along the hillslope; and (3) to determine the primary factors controlling the spatiotemporal variability of ETa on the hillslope.

2.1 Study Area

The study was conducted on a hillslope in Liudaogou Watershed (38°46 -38°51 N, 110°21 -110°23 E) on the Chinese Loess Plateau [Figure 1: see original paper]. The watershed is characterized by semi-arid continental climate conditions (Peel et al., 2007), with mean annual air temperature and precipitation of 8.4°C and 437 mm (1980-2015), respectively. Annual precipitation in 2015 (435 mm) and 2016 (612 mm) equals or exceeds the mean by 25%, thus 2015 and 2016 are termed normal and wet water years (Zhang et al., 2016). The watershed covers 6.89 km^2 and features deep gullies and large slopes. The dominant soil is Inceptisols with silt loam texture (Fu et al., 2012). The soil forms in loess parent material with substantial macropores and vertical fractures, leading to good drainage. Annual potential evapotranspiration is approximately 785 mm (Hu and Si, 2014). Elevation ranges from 1094 to 1274 m (Bo et al., 2011).

The studied hillslope is approximately 243 m long and 20 m wide, running from southeast to northwest with a mean slope gradient of about 19.8°. The hillslope is mainly vegetated with Korshinsk peashrub (*Caragana korshinskii* Kom). Relevant soil properties, terrain attributes, and vegetation features are provided in Table 1. Three plots were established at the upper, middle, and lower slope positions, each 2 m × 5 m with the longest side oriented downslope. Five, five, and six shrubs (*C. korshinskii*) with similar canopies were distributed in the upper, middle, and lower plots, respectively. Mean slope gradients of the upper, middle, and lower positions are 27.7°, 22.7°, and 12.1°, respectively .

2.2 Data Collection

A transect running down the middle of the hillslope was used for soil water content (SWC) measurements at 40 sampling locations established at approximately 5 m intervals [Figure 1b: see original paper]. Sampling locations were numbered 1 to 38 from top to bottom because measurements at locations 39 and 40 ceased in 2016. Locations 1-13, 14–26, and 27–38 represented the upper, middle, and lower slope positions, respectively. A 5-m long aluminum neutron probe access tube was installed at each location. Considering that the root system of Korshinsk peashrub is distributed in the 0–500 cm soil profile (Cheng et al., 2009), SWC was measured at 10-cm intervals between 10–100 cm depth and at 20-cm intervals between 120–480 cm depth. Measurements were made approximately every 12 days during observation periods (1 May–26 October 2015 and 15 April–29 October 2016). Volumetric SWC was measured by a neutron probe (CNC503DR hydro probe; Beijing Super Power Company, China). In the 0–10 and 10–20 cm soil layers, loss of fast thermal neutrons from the soil requires separate calibration for each layer (Bromley et al., 1997). Calibration was performed using standard methods by measuring counts at fixed depths in seven 480-cm long access tubes installed along the transect with a wide range of SWCs.

The standard count was obtained when the probe was immersed in water. The relationship between soil moisture counts and standard counts was then calibrated against volumetric SWC of 200 soil cores extracted from each fixed depth at the same horizontal distance from the access tube, measured using standard methods (Gardner et al., 2001).

During installation of neutron probe access tubes, disturbed soil samples were collected for soil texture determination at 20-cm intervals down to 200 cm depth at each location. Soil texture was relatively homogeneous throughout the profile, except at the surface. Thus, only particle size distribution of the 0–200 cm layer was measured using laser diffraction (Mastersizer 2000, Malvern Ltd., UK). At each location, three 40-cm deep pits were excavated 50 cm from the tube. Undisturbed samples were collected from pit walls at 30 cm depth using standard pre-weighed 100 ml Kopecky rings to determine bulk density (BD) by the gravimetric method (Baker, 1990) and saturated hydraulic conductivity (Ks) by the constant-head method (Libardi et al., 1980). Three replicates were collected per location. Statistics for BD and Ks are listed in Table 1.

LAI was measured with an LAI-2200 canopy analyzer (Li-Cor, Inc., USA). Plots of 2 m × 2 m were centered on each access tube. From each plot, three *C. korshinskii* plants were selected monthly during the monitoring period. Each plant was considered a replicate, and LAI values were averaged for each SWC monitoring location. Maximum LAI values were measured on 12 August 2015 and 18 August 2016. Mean maximum LAI values for the three slope positions are shown in Table 1. Elevation (E) and slope at each location were determined with an RTK-GPS receiver (Trimble 5700, Trimble Inc., USA) and a dinometer (PDC01, Everte Inc., China). Mean elevations and slopes of the three positions are shown in Table 1.

Three 2 m × 5 m plots were established for surface runoff measurement, centered at locations 6, 18, and 30, representing upper, middle, and lower slope positions [Figure 1b: see original paper]. Surface runoff was measured for each rainfall event that produced runoff in 2015 and 2016. Three D20 evaporation pans (diameter 20 cm, height 10 cm) with 5-cm mesh-wire screen covers were installed adjacent to runoff plots at each slope position. Pans were placed 70 cm above ground with mesh covers to prevent birds from drinking. Pan evaporation was measured every three days during observation periods in 2015 and 2016.

A meteorological station adjacent to the upper slope position monitored daily solar radiation, air temperature, wind speed, relative humidity, precipitation, and other factors. Distributions of daily rainfall and mean air temperature from 2015–2016 are shown in Figure 2 [Figure 2: see original paper].

2.3.1 Variation in Potential Evapotranspiration (PET) Along the Hillslope

Daily PET was calculated using the Penman equation (Shuttleworth, 1993). PET calculated at the meteorological station was regarded as the PET at the upper slope position. Comparisons of PET values calculated by evaporation pan and Penman equation at the upper slope position are shown in Figure 3 [Figure 3: see original paper]. The mean ratios of measured D20 pan evaporation at the upper, middle, and lower slope positions were assumed equal to those of PET. Therefore, PET ratios for the upper, middle, and lower positions were 1.00:0.98:0.83 based on D20 pan measurements.

2.3.2 Actual Evapotranspiration (ETa)

Considering the 12-day interval of soil water measurements, 12-day ETa (mm/d) was estimated using the water balance method from a reduced equation (Eq. 1):

$$ET_{a,j} = \frac{(SWS_j - SWS_{j+1}) + P_{j+1} - R_{j+1} - G_{j+1}}{\Delta t}$$

where SWS_j is soil water storage (mm) in the 480-cm profile at measurement j; SWS_{j+1} is storage at measurement $j+1$; P_{j+1} is accumulated precipitation (mm) between measurements; R_{j+1} is accumulated runoff depth (mm); and G_{j+1} is deep percolation (mm), which was disregarded because annual rainfall infiltration rarely exceeds 100 cm depth in this area (Yang et al., 2011) and the groundwater table was generally below 20 m (Shangguan and Zheng, 2006). Precipitation was assumed homogeneous along the hillslope, while measured runoff depths in the three plots were assumed valid for locations 1–13, 14–26, and 27–38, respectively. Groundwater was not involved in the soil water cycle due to the high water table (Shangguan and Zheng, 2006), so its influence on water balance was disregarded.

2.3.3 Time Stability Analysis of ETa

The relative difference method of time stability analysis (Vachaud et al., 1985) was used. The relative difference of ETa, δ_{ij}, for location i at time j was defined by Equation 2:

$$\delta_{ij} = \frac{ET_{a,ij} - \overline{ET}_{a,j}}{\overline{ET}_{a,j}}$$

where i is the measurement location ($i = 1, 2, 3, \ldots, 38$) and j is the measurement count during 2015–2016 ($j = 1, 2, 3, \ldots, 24$). Due to limited datasets, hold-out cross-validation was used to evaluate time stability analysis for estimating mean ETa. A random number generator selected 14 datasets from 24 to form a training set for determining representative sites. The remaining 10 datasets constituted a validation set to evaluate prediction accuracy of representative sites for hillslope-mean ETa. $\overline{ET}_{a,j}$ is the mean ETa for all locations at measurement j (Equation 3):

$$\overline{ET}_{a,j} = \frac{1}{n}\sum_{i=1}^{n} ET_{a,ij}$$

The temporal mean relative difference (MRD), $\overline{\delta}_i$, and its standard deviation (SDRD), $\sigma(\delta_i)$, were calculated by Equations 4 and 5:

$$\text{MRD} = \bar{\delta}_i = \frac{1}{m}\sum_{j=1}^{m}\delta_{ij}$$

$$\text{SDRD} = \sigma(\delta_i) = \sqrt{\frac{1}{m-1}\sum_{j=1}^{m}(\delta_{ij} - \bar{\delta}_i)^2}$$

Generally, sampling locations with relatively low SDRD can be regarded as having high temporal stability (Grayson and Western, 1998). If MRD values are close to zero, ETa values approximate the area-mean ETa. Jacobs et al. (2004) suggested an index of time stability (ITS) considering MRD and SDRD to determine time-stable locations (Eq. 6):

$$\text{ITS} = \sigma(\delta_i)^2 + \bar{\delta}_i^2$$

The location with the lowest ITS value was regarded as the MTSL representative of the mean ETa on the hillslope.

2.3.4 Statistical Analysis

Significance levels of correlation were tested through two-tailed t-tests. Relationships between ETa at each sampling location and influencing factors were characterized by non-parametric Spearman' s test. Differences in soil properties among slope positions were tested through one-way analysis of variance. Tukey' s honest significance difference (HSD) method was used for post hoc multiple comparisons. K-S (Kolmogorov-Smirnov) test evaluated normal and non-normal distributions of SWC time series. All statistical analyses used SPSS 19.0 (IBM Corp., USA).

3.1 Variations in PET and Surface Runoff at Different Slope Positions

Figure 4 [Figure 4: see original paper] shows PET variations at different slope positions during normal (2015) and wet (2016) water years. Mean PET values at upper and middle positions were significantly higher ($P<0.05$) than at the lower position in both years. PET in the wet year was higher ($P=0.08$) than in the normal year for all positions.

Runoff comparisons among the three positions are presented in Figure 5 [Figure 5: see original paper]. Only 12 surface runoff events occurred in 2015–2016. Mean values exhibited a declining trend from upper (6.1 mm/d) and middle (3.9 mm/d) to lower (1.9 mm/d) positions. Significant difference was observed only between upper and lower positions ($P<0.05$).

3.2 Variations in SWC Along the Hillslope

Normal and non-normal distributions of SWC time series at each depth were observed via K-S test. Therefore, median and interquartile ranges characterized central tendency and dispersion during normal and wet years [Figure 6: see original paper]. Areas with highest SWCs during the normal year occurred mainly in the 240–480 cm layer at locations 13, 14, 15, 33, 34, 35, and 36 [Figure 6a: see original paper]. Other high-SWC areas were the 30–180 cm layer for all locations in 2016 due to approximately doubled precipitation compared with 2015 [Figure 6b: see original paper]. In 2015, lowest SWCs were within the 30–180 cm layer for most locations [Figure 6a: see original paper]. In 2016, the driest layer typically occurred at the surface (0–20 cm) [Figure 6b: see original paper].

Temporal variability in SWC was characterized by interquartile range [FIGURE:6c,d]. In 2015, highest temporal variability occurred in the 0–30 cm layer for all locations and in the 40–480 cm layer at locations 13 and 14 [Figure 6c: see original paper]. In 2016, highest temporal stability extended to 200 cm depth [Figure 6d: see original paper]. During transition from dry to wet periods, SWCs of the 200–480 cm layer remained relatively constant [FIGURE:6c,d].

3.3 Variations in ETa Along the Hillslope

Figure 7 [Figure 7: see original paper] presents ETa at each measurement location during the 2015 and 2016 growing seasons. Despite different absolute values, ETa exhibited similar trends during normal and wet years [Figure 7: see original paper]. ETa rapidly decreased between locations 1 and 11, reaching minima of 226 mm in 2015 and 522 mm in 2016 at location 11. ETa then increased to maxima of 291 mm at location 13 in 2015 and 687 mm at location 15 in 2016. After peaking, ETa gradually decreased by 41 mm (2015) and 177 mm (2016). Mean ETa in lower positions (locations 27-38) was lower than in middle positions (locations 13-26) in both years. Furthermore, hillslope ETa in 2016 was significantly higher ($P<0.01$) than in 2015, with averaged values of 593 mm and 270 mm, respectively, due to different precipitation amounts.

3.4 Time Stability of ETa

Relative difference analysis was performed to identify locations where ETa consistently equaled the mean ETa of the study area. Such locations could represent the area-averaged ETa value, reducing monitoring time, labor, and costs. Figure 8 [Figure 8: see original paper] presents ranked MRD in ETa, associated SDRD, and ITS for each sampling location. Descriptive statistics are displayed in Table 2 . Ranges (differences between minimum and maximum) of MRD, SDRD, and ITS were 64.04%, 91.96%, and 99.22%, respectively. Minimum values of MRD, SDRD, and ITS were −20.92%, 21.44%, and 22.10%, respectively. A significantly positive correlation ($P<0.01$) existed between MRD and SDRD values. The MTSL with the lowest ITS value was selected to estimate mean hillslope ETa. Location 13, with an ITS value of 22.10%, was identified for estimating hillslope-mean ETa. Validation confirmed that ETa at the MTSL (location 13) can predict mean hillslope ETa (R^2=0.931, P<0.01) [Figure 9: see original paper].

3.5 Factors Controlling Time Stability of ETa

Table 3 presents correlations between time stability indices (MRD, SDRD, ITS) and factors including clay content, Ks, BD, elevation, maximum LAI (LAIm), and SWS. MRD of ETa represents evapotranspiration magnitude. MRD values had the highest positive correlation with SWS (P<0.01); elevation (ELE) and profile mean clay content (PMCC) also correlated with MRD (P<0.05). SDRD and ITS represent ETa temporal stability. According to Table 3, only ELE significantly correlated with SDRD and ITS (P<0.05).

Among influencing factors, LAIm and BD were positively correlated (P<0.01). ELE had significant relationships with LAIm, BD, and surface clay content (SCC) (P<0.01). SCC was negatively correlated with BD (P<0.01). A significantly positive correlation existed between SWS and PMCC (P<0.05). However, a relatively weak correlation existed between MRD and LAIm (P>0.05).

4.1 Variations of Components in Water Balance Along the Hillslope

PET is influenced by climatic conditions such as solar radiation, air temperature, wind speed, and relative humidity, which can estimate ETa in water balance calculations. Since solar radiation is an important environmental control for PET (Table 4), the small PET at the lower slope position [Figure 4: see original paper] might be due to reduced solar radiation (Carey and Woo, 2011; Fyllas et al., 2017). Precipitation in 2016 was much greater than in 2015; however, PET did not differ significantly between years [Figure 4: see original paper]. This indicated that PET exhibited less inter-annual variability than precipitation, consistent with Oishi et al. (2010), who found that despite large inter-annual precipitation variation (934–1346 mm), annual evapotranspiration varied much less (610–668 mm) in a mature oak-hickory forest. This occurs because inter-annual evapotranspiration variability can be mitigated by vegetation responses that decrease soil water storage during drought periods (Jones, 2011).

Surface runoff is affected by rainfall intensity, duration, vegetation presence, soil, and microtopography (Dunne et al., 1991; El-Hassanin et al., 1993). Surface runoff decreased gradually from upper to lower slope positions [Figure 5: see original paper]. Since rainfall was consistent across the hillslope, runoff variations could be mainly attributed to differences in slope gradient (Table 1), supported by El-Hassanin et al. (1993), who reported that runoff increased with slope gradient. Median runoff values for the three positions were as small as 3.5 mm/d, with only 12 runoff events during two years. These results indicated that in semi-arid areas, runoff amount and frequency were so limited that they had little influence on water balance calculations.

Larger water storage was concentrated in the deep soil layer (240–480 cm) of the lower slope position [FIGURE:6a,b]. A similar observation was reported by Fu et al. (2013), who found soil water downhill accumulation in deep layers (300–400 cm) for fallow and cropland due to elevation. Due to rapidly increased precipitation, the spatial pattern of SWC in the vertical profile changed significantly from 2015 to 2016 [Figure 6: see original paper]. For instance, the 30-180 cm layer was the driest in 2015 because it was the main root distribution zone of *C. korshinskii* (Cheng et al., 2009), becoming the driest layer due to root water uptake. This layer then became the wettest area because of doubled precipitation in 2016 [Figure 6b: see original paper]. This agreed with Chen et al. (2008), who observed that stored water in the 200–300 cm layer of grassland greatly increased due to doubled precipitation on the Chinese Loess Plateau. The highest temporal variability of SWC in 2015 and 2016 occurred in the 0-30 cm and 0–200 cm layers, respectively [FIGURE:6c,d]. The former resulted from combined effects of precipitation, solar radiation, soil evaporation, and topography on the surface layer (Wythers et al., 1999; Pan and Wang, 2009; Gao et al., 2015). The latter likely resulted from precipitation recharge to root-zone SWC, causing transition from driest to wettest conditions in the root zone.

ETa varied with terrain, climate, soil, and plant characteristics, exhibiting a slightly declining trend [Figure 7: see original paper]. Based on the simplified water balance (not considering rainfall, deep percolation, and groundwater recharge (Shangguan and Zheng, 2006; Yang et al., 2011)), ETa variations were mainly ascribed to SWS and surface runoff variations. For SWS, no significant difference was found in temporal variability (displayed by interquartile range) of SWC among sampling sites [FIGURE:6c,d], indicating similar variation degrees of water storage for all sites. Therefore, the slightly declining ETa trend mainly resulted from surface runoff variations [Figure 5: see original paper]. Additionally, the significantly low PET ($P<0.05$) at the lower slope position [Figure 4: see original paper] supported the declining ETa trend. ETa exhibited similar trends during normal and wet years [Figure 7: see original paper], indicating that the horizontal spatial pattern of ETa was hardly altered by precipitation in semi-arid areas. Based on the simplified water balance, ETa was mainly affected by SWS and surface runoff depth variations. For SWS, the horizontal spatial pattern changed little because increased SWS in semi-arid areas typically resulted in vertical water flow controlled by soil properties (Grayson et al., 1997; Green and Erskine, 2011). For runoff, horizontal overland flow could influence soil water distribution, but runoff amount and frequency were quite limited [Figure 5: see original paper].

Among locations 1–17, location 11 had almost the lowest ETa in both 2015 and 2016. Location 11 also had the lowest SWC on the entire hillslope [Figure 6: see original paper]. The low ETa at location 11 might be because low SWC limited plant growth in semi-arid areas (Linacre, 2004), resulting in lower transpiration water consumption. ETa around location 14 was higher than most of the hillslope [Figure 7: see original paper], which could be ascribed to high water storage in this area [FIGURE:6a,b]. Increased water storage in arid and semi-arid areas can greatly promote plant growth, thus leading to high ETa (Garcia and Tague, 2014).

4.2 Time Stability of ETa and Its Influencing Factors

Time stability has been substantially applied to SWC in previous studies (Vachaud et al., 1985; Jacobs et al., 2004; Gao and Shao, 2012; Zhang et al., 2016). In addition to SWC, time stability as an effective method to analyze spatiotemporal variability was first applied to soil sulfur content and soil salinity (Douaik et al., 2006; Piotrowska-Długosz et al., 2017). This study performed time stability analysis on ETa for the first time. The ranges of SDRD and ITS in ETa were 145.38 and 151.42 (Table 2), significantly higher than those in SWC (Gao and Shao, 2012; Zhang et al., 2016). This indicated that ETa had relatively higher temporal variability than SWC because ETa components (soil evaporation and transpiration) and their controls are numerous, greatly increasing ETa variability (Zeng and Cai, 2016). The minimum SDRD and ITS in ETa were not significantly higher than those in SWC (Table 2), indicating that the location with the lowest ITS could accurately estimate hillslope-mean ETa. MRD and SDRD values at each location had a significantly positive relationship ($P<0.01$), suggesting that high ETa (presented as high MRD) caused high temporal variability (presented as high SDRD). A similar observation was reported by Vivoni et al. (2010), who investigated spatiotemporal variability of soil moisture and evapotranspiration in a mountainous basin. This also indicated that locations with high temporal stability in ETa (low SDRD) tended to be in areas with relatively low ETa (low MRD). Figure 8 shows that the MTSL (location 13) could estimate hillslope-mean ETa ($R^2=0.931$, $P<0.01$). In future monitoring, ETa at the MTSL can be continuously monitored automatically to represent the hillslope average, saving considerable labor, time, and expenditure.

ETa involves soil evaporation and plant transpiration through stomata, influenced by hydrological processes such as precipitation, runoff, and groundwater. Therefore, ETa temporal stability was directly or indirectly affected by terrain, soil, climate, and vegetation characteristics. MRD represents ETa magnitude. A significantly positive relationship ($P<0.01$) existed between SWS and MRD (Table 3). Regarding this, Manabe (1969) proposed two SWC thresholds: SWCCRIT (critical water content above which ETa equals PET) and SWCWILT (wilting coefficient below which ETa is zero), suggesting ETa increases with SWC between these thresholds. Shuttleworth (1993) found SWCCRIT typically equals 50%-80% of SWCFC (field capacity). In this study, most SWCs were within these thresholds. The positive ETa-SWS relationship ($P<0.05$) supported Manabe (1969) and was consistent with other studies (Vivoni et al., 2010; Garcia and Tague, 2014).

Furthermore, a significantly positive relationship existed among MRD, SWS, and PMCC (Table 3). Thus, soil texture influenced SWS distribution and ETa spatial variability because it controlled vertical water distribution (Grayson et al., 1997). Lateral overland flow was infrequent [Figure 5: see original paper], so horizontal soil water redistribution seldom occurred. Only ELE significantly correlated with SDRD and ITS (Table 3) because ELE could affect ETa temporal stability by indirectly influencing soil and vegetation on a hillslope. This was supported by Table 3, which showed ELE was negatively correlated with BD and LAI but positively correlated with SCC ($P<0.01$). A weak correlation existed between LAIm and ETa (Table 3) because considerable bare soil areas between shrubs during the growing season caused high evaporation. Soil evaporation was not associated with plant productivity compared with transpiration (Kool et al., 2014), weakening the positive ETa-LAIm correlation.

5 Conclusions

In this study, soil water storage, surface runoff, and precipitation along a hillslope were determined during normal (2015) and wet (2016) water years. ETa was calculated from a simple water balance equation. Spatial variability and temporal stability of ETa were investigated, and key factors influencing spatiotemporal variability were analyzed. Conclusions are: (1) The lower slope position showed lower potential evapotranspiration, probably due to reduced solar radiation compared with upper and middle positions; (2) The horizontal spatial pattern of ETa was hardly altered by precipitation in semi-arid areas, although doubled precipitation increased ETa; (3) Time stability analysis indicated that locations with high temporal stability tended to be in areas with relatively low ETa, and the most time-stable location can accurately estimate hillslope-mean ETa; and (4) Temporal stability of ETa was mainly ascribed to elevation on the Chinese Loess Plateau hillslope.

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References

Baker J M. 1990. Measurement of soil water content. Remote Sensing Reviews, 5(1): 263–279.

Bo X, Wang Q H, Fan J, et al. 2011. Application of the SCS-CN model to runoff estimation in a small watershed with high spatial heterogeneity. Pedosphere, 21(6): 738-749.

Bromley J, Brouwer J, Barker A, et al. 1997. The role of surface water redistribution in an area of patterned vegetation in a semi-arid environment, south-west Niger. Journal of Hydrology, 198(1-4): 1-29.

Carey S K, Woo M. 2001. Spatial variability of hillslope water balance, wolf creek basin, subarctic yukon. Hydrological Process, 15(16): 3113-3132.

Chen H S, Shao M G, Li Y Y. 2008. The characteristics of soil water cycle and water balance on steep grassland under natural and simulated rainfall conditions in the Loess Plateau of China. Journal of Hydrology, 360(1-4): 242-251.

Cheng X R, Huang M B, Shao M G, et al. 2009. A comparison of fine root distribution and water consumption of mature *Caragana korshinkii* Kom grown in two soils in a semiarid region, China. Plant and Soil, 315(1-2): 149-161.

Douaik A, van Meirvenne M, Toth T. 2006. Temporal stability of spatial patterns of soil salinity determined from laboratory and field electrolytic conductivity. Arid Land Research and Management, 20(1): 1-13.

Dunne T, Zhang W, Aubry B F. 1991. Effects of rainfall, vegetation, and microtopography on infiltration and runoff. Water Resources Research, 27(9): 2271-2285.

El-Hassanin A S, Labib T M, Gaber E I. 1993. Effect of vegetation cover and land slope on runoff and soil losses from the watersheds of Burundi. Agriculture Ecosystems & Environment, 43(3-4): 301-308.

Ewers B E, Mackay D S, Gower S T, et al. 2002. Tree species effects on stand transpiration in northern Wisconsin. Water Resources Research, 38(7): 1103, doi: 10.1029/2011WR010636.

Fu W, Huang M B, Gallichand J, et al. 2012. Optimization of plant coverage in relation to water balance in the Loess Plateau of China. Geoderma, 173-174: 134-144.

Fu X L, Shao M G, Wei X R, et al. 2013. Effects of monovegetation restoration types on soil water distribution and balance on a hillslope in northern loess plateau of China. Journal of Hydrologic Engineering, 18(4): 413-421.

Fyllas N M, Bentley L P, Shenkin A, et al. 2017. Solar radiation and functional traits explain the decline of forest primary productivity along a tropical elevation gradient. Ecology Letters, 20(6): 730-740.

Gao G, Chen D L, Xu C Y, et al. 2007. Trend of estimated actual evapotranspiration over China during 1960-2002. Journal of Geophysical Research-Atmospheres, 112: D11120, doi: 10.1029/2006JD008010.

Gao L, Shao M. 2012. Temporal stability of shallow soil water content for three adjacent transects on a hillslope. Agricultural Water Management, 110: 41-54.

Gao L, Lv Y J, Wang D D, et al. 2015. Can shallow-layer measurements at a single location be used to predict deep soil water storage at the slope scale? Journal of Hydrology, 531: 534-542.

Gao X D, Wu P T, Zhao X N, et al. 2011. Estimating spatial mean soil water contents of sloping jujube orchards using temporal stability. Agricultural Water Management, 102(1): 66-73.

Gao Y F, Zhao C Y, Muhammad W A, et al. 2019. Actual evapotranspiration of subalpine meadows in the Qilian Mountains, Northwest China. Journal of Arid Land, 11(3): 371-384.

Garcia E, Tague C. 2014. Climate regime and soil storage capacity interact to effect evapotranspiration in western United States mountain catchments. Hydrology And Earth System Sciences, 11(2): 2277-2319.

Gardner C M K, Robinson D, Blyth K, et al. 2001. Soil water content. In: Smith K A, Mullins C E. *Soil and Environmental Analysis: Physical Methods* (2nd ed.). New York: Marcel Dekker, 1-64.

Grayson R B, Western A W, Chiew F H, et al. 1997. Preferred states in spatial soil moisture patterns: Local and nonlocal controls. Water Resources Research, 33(12): 2897-2908.

Grayson R B, Western A W. 1998. Towards areal estimation of soil water content from point measurements: time and space stability of mean response. Journal of Hydrology, 207(1-2): 68-82.

Green T R, Erskine R H. 2011. Measurement and inference of profile soil-water dynamics at different hillslope positions in a semiarid agricultural watershed. Water Resources Research, 47(12): W00H15, doi: 10.1029/2010WR010074.

Hu W, Shao, M A, Han F, et al. 2010. Watershed scale temporal stability of soil water content. Geoderma, 158(3-4): 181-198.

Hu W, Si B C. 2014. Can soil water measurements at a certain depth be used to estimate mean soil water content of a soil profile at a point or at a hillslope scale? Journal of Hydrology, 516: 67-75.

Jacobs J M, Mohanty B P, Hsu E C, et al. 2004. SMEX02: Field scale variability, time stability and similarity of soil moisture. Remote Sensing of Environment, 92(4): 436-446.

Jia Y H, Shao M A. 2013. Temporal stability of soil water storage under four types of revegetation on the northern Loess Plateau of China. Agricultural Water Management, 117: 33-42.

Jones J. 2011. Hydrologic responses to climate change: Considering geographic context and alternative hypotheses. Hydrological Process, 25(12): 1996-2000.

Kool D, Agam N, Lazarovitch N, et al. 2014. A review of approaches for evapotranspiration partitioning. Agricultural and Forest Meteorology, 184: 56–70.

Li Q Y, Sun Y W, Yuan W L, et al. 2017. Streamflow responses to climate change and LUCC in a semi-arid watershed of Chinese Loess Plateau. Journal of Arid Land, 9(4): 609–621.

Lian J, Huang M. 2015. Evapotranspiration estimation for an oasis area in the Heihe River Basin using Landsat-8 images and the METRIC model. Water Resources Management, 29(14): 51–57.

Lian J, Huang M. 2016. Comparison of three remote sensing based models to estimate evapotranspiration in an oasis-desert region. Agricultural Water Management, 165: 153–162.

Liao K, Zhou Z, Lai X, et al. 2017. Evaluation of different approaches for identifying optimal sites to predict mean hillslope soil moisture content. Journal of Hydrology, 547: 10–20.

Libardi P L, Reichardt K, Nielsen D R, et al. 1980. Simple field methods for estimating soil hydraulic conductivity. Soil Science Society of America Journal, 44(1): 3–7.

Linacre E. 2004. Evaporation trends. Theoretical and Applied Climatology, 79: 11–21.

Majidi M, Alizadeh A, Vazifedoust M, et al. 2015. Analysis of the effect of missing weather data on estimating daily reference evapotranspiration under different climatic conditions. Water Resources Management, 29(7): 2107–2124.

Manabe S. 1969. Climate and the ocean circulation: I. The atmospheric circulation and the hydrology of the earth' s surface. Monthly Weather Review, 97(11): 739–774.

McVicar T R, van Niel T G, Li L T, et al. 2007. Spatially distributing monthly reference evapotranspiration and pan evaporation considering topographic influences. Journal of Hydrology, 338(3–4): 196–220.

Mitchell P J, Benyon R G, Lane P N. 2012. Responses of evapotranspiration at different topographic positions and catchment water balance following a pronounced drought in a mixed species eucalypt forest, Australia. Journal of Hydrology, 440–441: 1–15.

Mo X G, Liu S X, Lin Z H, et al. 2004. Simulating temporal and spatial variation of evapotranspiration over the Lushi basin. Journal of Hydrology, 285(1–4): 125–142.

Newman B D, Wilcox B P, Archer S R, et al. 2006. Ecohydrology of water-limited environments: A scientific vision. Water Resources Research, 42(6): W06302, doi: 10.1029/2005WR004141.

Nicholls E M, Carey S K, Humphreys E R, et al. 2016. Multi-year water balance assessment of a newly constructed wetland, Fort McMurray, Alberta. Hydrological Processes, 30(16): 2739–2753.

Oishi A C R, Oren K A, Novick S, et al. 2010. Interannual invariability of forest evapotranspiration and its consequence to water flow downstream. Ecosystems, 13: 421–436.

Pan Y X, Wang X P. 2009. Factors controlling the spatial variability of surface soil moisture within revegetated-stabilized desert ecosystems of the Tengger Desert, Northern China. Hydrological Process, 23(11): 1591–1601.

Peel M C, Finlayson B L, McMahon T A. 2007. Updated world map of the Köppen-Geiger climate classification. Hydrology and Earth System Sciences Discussions, 4(2): 439–473.

Piotrowska-Długosz A, Siwik-Ziomek A, Długosz J, et al. 2017. Spatio-temporal variability of soil sulfur content and arylsulfatase activity at a conventionally managed arable field. Geoderma, 295: 107–118.

Santiago L S, Goldstein G, Meinzer F C, et al. 2000. Transpiration and forest structure in relation to soil waterlogging in a Hawaiian montane cloud forest. Tree Physiology, 20(10): 673–681.

Shangguan Z P, Zheng S X. 2006. Ecological properties of soil water and effects on forest vegetation in the Loess Plateau. International Journal of Sustainable Development & World Ecology, 13(4): 307–314.

Shuttleworth W J. 1993. Evaporation. In: Maidment D R. *Handbook of Hydrology*. Sydney: McGraw-Hill.

Starr G. 2005. Assessing temporal stability and spatial variability of soil water patterns with implications for precision water management. Agricultural Water Management, 72(3): 223–243.

Thomas A. 2000. Spatial and temporal characteristics of potential evapotranspiration trends over China. International Journal of Climatology, 20(4): 381–396.

Vachaud G, Passerat de Silans A, Balabanis P, et al. 1985. Temporal stability of spatially measured soil water probability density function. Soil Science Society of America Journal, 49(4): 822–828.

Vivoni E R, Rodriguez J C, Watts C J. 2010. On the spatiotemporal variability of soil moisture and evapotranspiration in a mountainous basin within the North American monsoon region. Water Resources Research, 46(2): W02509, doi: 10.1029/2009WR008240.

Wilcox B P, Breshears D D, Seyfried M S. 2003. Water balance on rangelands. In: Stewart B A, Howell T A. *Encyclopedia of Water Science*. New York: Marcel Dekker, Inc., 791–794.

Wythers K R, Lauenroth W K, Paruelo J M. 1999. Bare-soil evaporation under semiarid field conditions. Soil Science Society of America Journal, 63(5): 1341-1349.

Xu C Y, Singh V P. 2005. Evaluation of three complementary relationship evapotranspiration models by water balance approach to estimate actual regional evapotranspiration in different climatic regions. Journal of Hydrology, 308(1-4): 105-121.

Yang L, Wei W, Mo B, et al. 2011. Soil water deficit under different artificial vegetation restoration in the semi-arid hilly region of the Loess Plateau. Acta Ecologica Sinica, 31: 3060-3068. (in Chinese)

Zeng R J, Cai X M. 2016. Climatic and terrestrial storage control on evapotranspiration temporal variability: Analysis of river basins around the world. Geophysical Research Letters, 43(1): 185-195.

Zhang D B, Yao P W, Na Z, et al. 2016. Soil water balance and water use efficiency of dryland wheat in different precipitation years in response to green manure approach. Scientific Reports, 6: 26856, doi: 10.1038/srep26856.

Zhang Y K, Xiao Q L, Huang M B. 2016. Temporal stability analysis identifies soil water relations under different land use types in an oasis agroforestry ecosystem. Geoderma, 271: 150-160.

Zhao L W, Zhao W Z. 2014. Evapotranspiration of an oasis-desert transition zone in the middle stream of Heihe River, Northwest China. Journal of Arid Land, 6(5): 529-539.

Note: Figure translations are in progress. See original paper for figures.