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ORIGINAL ARTICLE

Effects of land use on spatial and temporal distribution of soil moisture within profiles

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Abstract
Soil moisture (SM) plays a key role in hydrological processes and the distribution and growth status of vegetation in arid and semi-arid regions. An understanding of SM dynamics can help to better explain runoff and soil erosion processes, enable vegetation restoration, and improve water resources management. This study investigated SM changes under different land cover types on hillslopes using fine-scale (every 10 cm and every hour) SM monitoring data at Dun Mountain in the semi-arid Loess Plateau. It was found that the SM of each soil layer generally followed the order of bare land > grassland > forestland. The mean annual SM of grassland and forestland was 71.8 and 65.4% of that of bare land, respectively. The SM of bare land generally displayed an increasing trend with depth. The SM of grassland and forestland generally increased first and then decreased with increasing depth. The mean SM of all three land cover types in different soil layers was largest in autumn. In grassland and forestland, there was a higher soil water replenishment (653.02 and 608.39 mm) and consumption (576.77 and 555.70 mm) than the corresponding values for bare land during the four seasons. The amount of soil water replenished in grassland and forestland in summer was 1.32 and 1.21 times that of bare land, respectively. The cumulative amounts of frozen soil water in bare land, grassland, and forestland were 495.98, 334.78, and 213.15 mm, respectively. The SM distribution among the different soil layers exhibited a strong temporal stability. The effect of meteorological factors on actual evapotranspiration displayed significant seasonal differences. In conclusion, vegetation cover reduced the SM at the slope scale, but the reduction was discontinuous at the annual scale. The results contribute to clarify the seasonal difference in actual evapotranspiration and provide new insights into soil moisture retention and freeze–thaw process in arid region.

Keywords Soil moisture process · Replenishment · Actual evapotranspiration · Freeze–thaw process · Meteorological factors

Introduction

Soil moisture plays a key role in ecological, hydrological, and biogeochemical processes at different scales, including plant growth, root water uptake, evapotranspiration, infiltration, runoff response, sediment, and nutrient transport (Heathman et al. 2009; Chaney et al. 2015; Hou et al. 2015; Xu et al. 2017; Xiao et al. 2019). A knowledge of SM dynamics is very important for understanding hydrological processes, enabling vegetation restoration, and improving water resources management (Huang et al. 2013; Ren et al. 2018). In arid and semi-arid areas, the SM is a major limiting factor affecting vegetation restoration and crop production (Chen et al. 2007; Yang et al. 2014). Inappropriate soil water use and management may lead to vegetation degradation and desertification. Land degradation forms a threat to the capacity of land to provide ecosystem services that are

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needed to reach the Sustainable Development Goals of the United Nations (SDGs). Understanding the role of soil water in a sustainably managed soil–water system is essential for the successful implementation and realization of the SDGs (Keesstra et al. 2016, 2018; Visser et al. 2019).

Stored water in soil is a dynamic property that is affected by several environmental factors, such as antecedent precipitation, precipitation, soil properties, topographic attributes, and land use type (Gómez-Plaza et al. 2000; Cantón et al. 2004; Brocca et al. 2007; Wang et al. 2013; Zhang et al. 2017a, b; Yu et al. 2018). The strong heterogeneity of environmental factors at different scales results in the spatial and temporal variation of the SM (Huang et al. 2016; Pradhan et al. 2019). Among these factors, soil properties and topography are considered relatively constant in the short term, while land use and climate are the dominant variables (Montenegro and Ragab, 2012; Wei et al. 2009; Shi et al. 2019a; Dang et al. 2020). Li et al. (2018) verified that soil texture and topography were the primary factors influencing the SM in a gully slope. Topography can indirectly affect SM by changing the redistribution of precipitation and surface runoff. However, topography mainly affects the SM of the shallow soil layers, with the SM of deep layers being mainly affected by vegetation. Some studies have also considered the interactive influences of vegetation restoration and land use changes on SM. Gómez-Plaza et al. (2001) found that vegetation played a key role in SM variability in vegetated areas, while soil texture and slope explained much of the soil water variability in non-vegetated areas. Introduced plants usually consume more soil water than native plants (Cao et al. 2011; Yang et al. 2012). On the other hand, studies across the globe indicate that 60–96% reductions in runoff and sediment compared with that of natural slopes can be achieved using diverse land preparation measures (Wang et al. 2015a). Proper land preparations and vegetation restoration can improve soil moisture and benefit land restoration (Feng et al. 2018).

The Loess Plateau is the largest and deepest loess deposit in the world. It covers an area of 640,000 km^2 and most of the plateau is located in a semi-arid zone (Fu et al. 2017). The disaster caused by the extensive Yangtze River flooding events of 1998 resulted in considerable attention being given to natural environmental protection in China (Cao et al. 2018). In response, the Chinese government launched the “Grain-for-Green” project on the Loess Plateau to restore cultivated hillsides (slope $\geq$ 25°) to forests or grassland to control soil erosion and land degradation (Chen et al. 2010; Wang et al. 2011; Shi et al. 2019b). The current area of forest plantation in China is approximately 0.69 million km^2, accounting for one-third of the world’s total forest plantation area (Chen et al. 2016). The vegetation cover has increased across more than 96% of the Loess Plateau, and the mean vegetation cover in the middle of the plateau increased by 20.06% from 2000 to 2015. However, the planting of forests usually depletes soil water due to strong evapotranspiration. Plants are forced to develop their root systems to consume deep soil water (Chen et al. 2008). The SM of forestlands is lower than that of native grassland and farmland, especially during dry periods (Yu et al. 2018). Yang et al. (2012) reported that afforestation reduces the available SM by more than 35% compared to traditional sloping cropland (Yang et al. 2012). More importantly, excessive reforestation has resulted in the drying of soil layers in some areas (Cao et al. 2011; Yang et al. 2012), decreased vegetation cover (Zhou et al. 2009), and the presence of “little old man trees” (Zhang and Song, 2003).

Comparing the distinctive effects of land use on SM is critical to the successful restoration of vegetation in arid and semi-arid regions (Deng et al. 2016). Some studies have examined the spatial and temporal differences of SM under land use changes (Wang et al. 2013; Jia, et al. 2017; Yang et al. 2017; Yu et al. 2018). However, little work has been done to quantitatively analyze seasonal differences in SM for different land uses based on fine-scale SM monitoring data. The amount of soil water that is annually involved in freeze–thaw processes in different land uses is not known. Therefore, this study analyzed the seasonal differences of the SM within profiles and its response to meteorological factors for different land cover types in the semi-arid Loess Plateau. The objectives of this research were to: (1) assess the seasonal differences of SM within profiles and the corresponding major meteorological influencing factors, and (2) quantitatively analyze the soil water replenishment, actual evapotranspiration, and amount of soil water involved in freeze–thaw processes under different land uses.

Materials and methods

Study area

This study was conducted at the Ansai Soil and Water Conservation Station (109°19′23″, 36°51′30″ N) of the Chinese Academy of Science and Ministry of Water Resources, which is within a typical loess hill and gully region on the Loess Plateau (Fig. 1a, b). The altitude of the station ranges between 1068 and 1309 m a.s.l. The annual mean rainfall is approximately 540 mm, and is mainly concentrated in July–September. The average annual temperature is 8.8 °C. The average annual sunshine duration is 2416 h and the mean annual frost-free period is 143 ~ 174 d. The natural vegetation is mainly *Bothriochloa ischaemum* (L.) Keng., *Stipa bungeana* Trin., *Artemisia gmelinii*, and *Artemisia giraldii* Pamp. The plantation grassland mainly includes: *Robinia pseudoacacia* L., *Caragana Korshinskii* Kom., *Hippophae rhamnoides* Linn., *Astragalus adsurgens* Pall., and

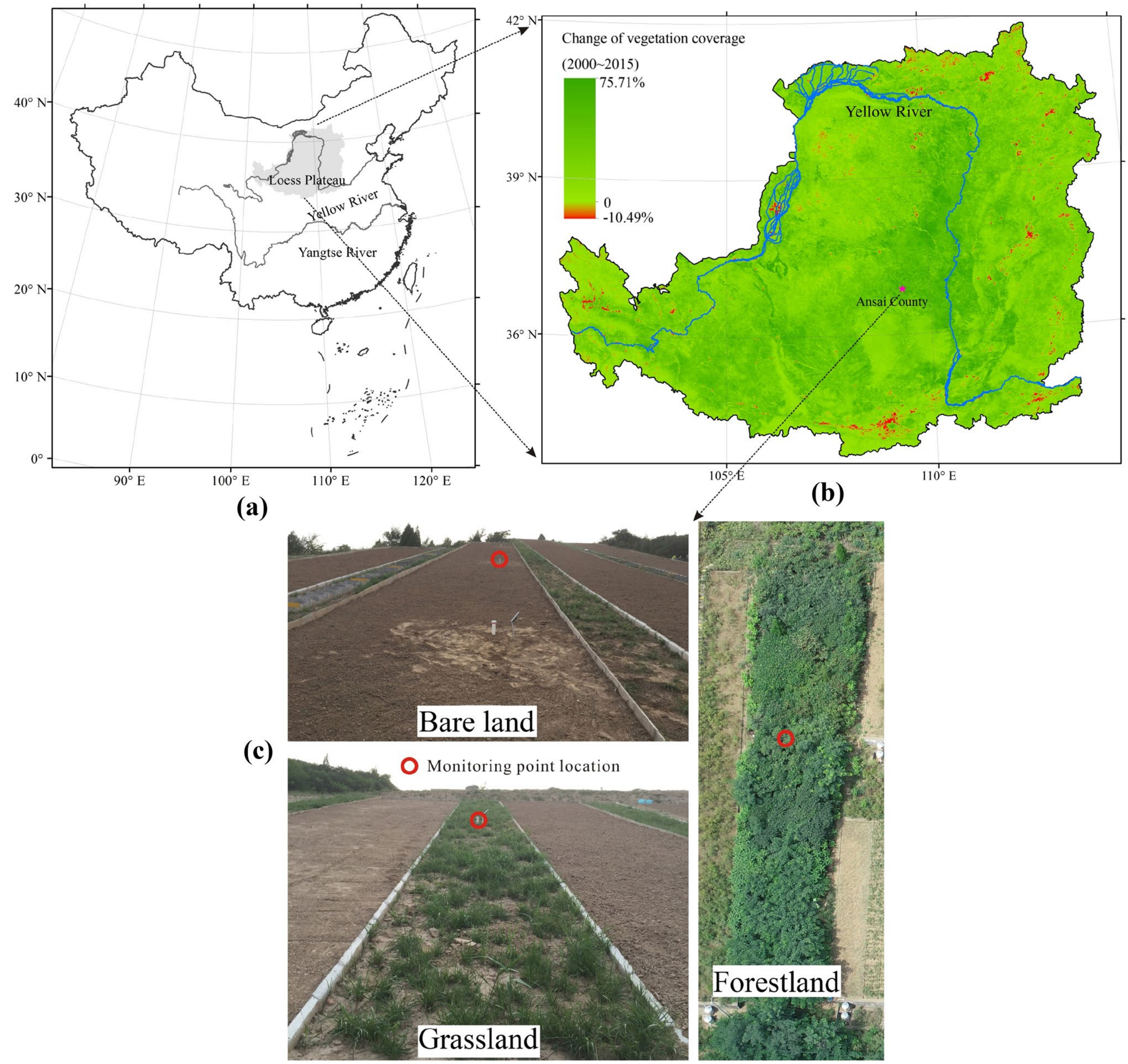

Fig. 1 Location of the study area (**a**, **b**) and the distribution of soil moisture-monitoring points (**c**)

Medicago sativa L.. The dominant crops are millet (*Setaria italica* L.), corn (*Zea mays* L.), and wheat (*Triticum aestivum* Linn.).

Soil sampling and analysis

Bare land, grassland, and forestland plots (Fig. 1c) in the upper part of the slope (Southeast) located at Dun Mountain near the Ansai station were selected for soil sampling. The recovery period of grassland and forestland is 3 years and more than 10 years, respectively. Soil samples were collected at 10-cm intervals down to a depth of 100 cm for each land use. Soil particle composition was measured by laser diffraction using a particle size analyzer (Mastersizer 2000: Malvern Instruments, Malvern, UK). Soil organic carbon (SOC) content was determined using an organic carbon analyzer (Multi N/C 3100: Analytik Jena AG, Jena, Germany). The soil total nitrogen (STN) content was determined using an automatic Kjeldahl apparatus (Kjeltec 8400, Foss Analytical, Hilleroed, Denmark). Details of the soil properties at different depths are provided in Table 1.

Three SM monitoring tubes (ET100-Pro, Insentek, Beijing, China) were installed in the middle of the three different land use plots to monitor changes in SM. The

Table 1 Characteristics of the soil properties at each soil depth in the three land uses

Property		L1	L2	L3	L4	L5	L6	L7	L8	L9	L10
Bare land	SOC, g/kg	4.76	3.35	2.41	2.30	3.32	2.04	2.39	2.70	4.08	2.00
	STN, g/kg	0.54	0.46	0.25	0.23	0.26	0.22	0.16	0.25	0.20	0.26
	Sand, %	42.82	41.17	37.94	36.78	35.21	38.26	35.13	33.01	35.70	33.61
	Clay, %	0.21	0.22	0.24	0.24	0.26	0.26	0.27	0.28	0.25	0.25
	Silt, %	56.97	58.61	61.81	62.98	64.53	61.48	64.60	66.71	64.05	66.14
Grassland	SOC, g/kg	2.52	4.49	3.63	3.86	2.70	2.52	2.51	2.36	2.15	3.04
	STN, g/kg	0.27	0.21	0.23	0.41	0.22	0.12	0.27	0.25	0.23	0.30
	Sand, %	40.46	39.29	38.83	39.15	38.58	39.67	39.13	40.57	39.35	40.66
	Clay, %	0.21	0.23	0.23	0.22	0.24	0.24	0.22	0.21	0.19	0.23
	Silt, %	59.33	60.48	60.94	60.64	61.18	60.09	60.65	59.21	60.46	59.12
Forestland	SOC, g/kg	5.15	5.29	5.16	4.52	3.48	3.45	3.75	2.07	1.91	1.67
	STN, g/kg	0.72	0.40	0.31	0.30	0.27	0.27	0.27	0.24	0.23	0.26
	Sand, %	39.64	34.56	32.00	28.58	29.59	34.35	29.31	31.04	33.60	31.39
	Clay, %	0.17	0.13	0.21	0.24	0.22	0.20	0.20	0.21	0.20	0.21
	Silt, %	60.19	65.31	67.79	71.19	70.19	65.45	70.49	68.76	66.20	68.40

SOC soil organic carbon; *STN* soil total nitrogen; *L1* 0 ~ 10 cm; *L2* 10 ~ 20 cm; *L3* 20 ~ 30 cm; *L4* 30 ~ 40 cm; *L5* 40 ~ 50 cm; *L6* 50 ~ 60 cm; *L7* 60 ~ 70 cm; *L8* 70 ~ 80 cm; *L9* 80 ~ 90 cm; *L10* 90 ~ 100 cm

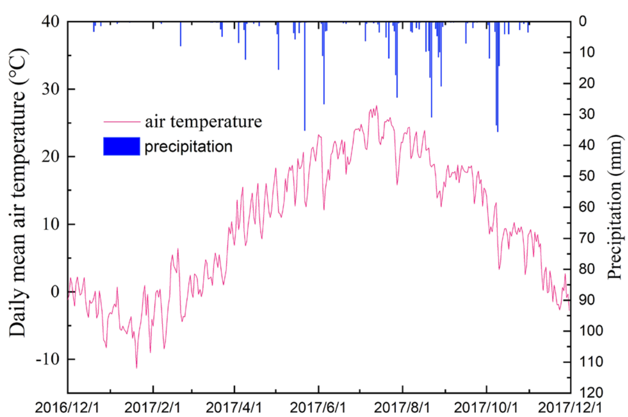

Fig. 2 The daily precipitation and air temperature at the Ansai station during the study period

distance between sensors and plot margins is about 2.0 m. Volumetric SM was obtained at 10-cm intervals down to a depth of 100 cm at each tube from 1 December, 2016 to 30 November, 2017 (four seasons). The ten soil layers from top to bottom were designated as L1, L2, L3, L4, L5, L6, L7, L8, L9, and L10, respectively. The SM and soil temperature of each layer were measured once an hour. Daily weather data (Fig. 2) were obtained from an automatic weather station (HOBO U30, Instrumart, Burlington, America) which is placed near the slope plots. The sampling interval of meteorological parameters (e.g., air temperature, precipitation, and wind speed) is 5 min.

The soil water consumption or replenishment amount per day (*SWD*) is calculated using the following equation:

$$SWD_i = SWS_{i+1} - SWS_i, \quad (1)$$

where *i* is the day of the study period; *SWS* is soil water storage (mm) at 8:00 am. Soil evapotranspiration can be negligible when precipitation occurs.

The soil moisture-monitoring device (ET100-pro) can measure soil liquid water, but it does not measure soil solid water (ice). When the soil temperature becomes below zero, the soil liquid water becomes solid water, and the soil moisture measured by the device will rapidly decrease. When the soil temperature rises from below zero degrees, the soil solid water becomes liquid water again, and the soil moisture measured by the device will rise. In this way, the freezing and thawing process of soil water can be monitored and analyzed.

A one-way analysis of variance (ANOVA) was used to assess the extent of variation in the SM of each soil depth in different seasons (Mark and Workman 2018). A redundancy analysis (RDA) was used to explore the contribution of weather factors on the actual evapotranspiration of soil water in different seasons. The arrows of two variables pointing in the same direction indicate a positive correlation, and the angle between two arrows is inversely proportional to the degree of their correlation. The length of the arrow indicates the similarity of contributions (Shi et al. 2017). The non-parametric Spearman's rank correlation test was used to analyze the temporal stability of the SM distribution in different soil layers. The Spearman rank correlation coefficient can indicate the strength and direction of the same variable when observed at different times (Douaik, 2006).

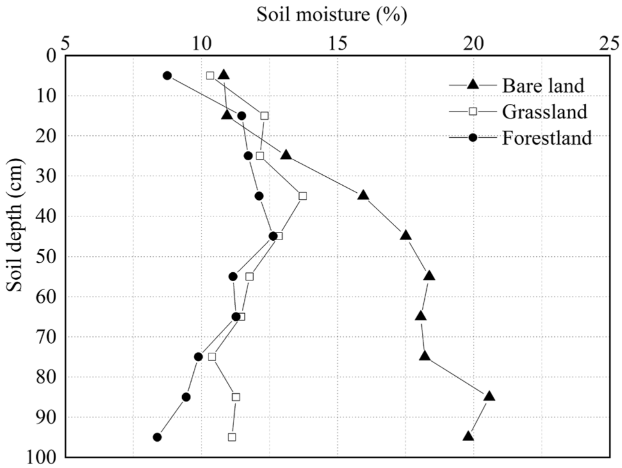

Fig. 3 Vertical changes of mean soil moisture under three land use types during the study period

Results

Vertical distribution and seasonal differences of soil moisture under different land uses

The vertical changes of SM under the three land use types are shown in Fig. 3. The SM at each soil depth generally followed the order of bare land > grassland > woodland. The maximum mean SM of bare land, grassland, and woodland was recorded at L9 (20.58%), L4 (13.72%), and L5 (12.64%), respectively. The mean SM for all three land uses was lowest at L1. The vertical distribution of mean SM for bare land was very different from that of grassland and forestland. The mean SM at different soil depths in bare land generally displayed an increasing trend. The SM of the ten soil layers for grassland and forestland generally increased first and then decreased. The mean SM of grassland and forestland was far lower than the field capacity (25%). The mean SM of bare land gradually approached field capacity with increasing depth, especially after L5. The coefficient of variation (CV) is an index of the magnitude of spatial variability (Nielsen and Bouma, 1985). The CV values of SM in the different layers of bare land were lower than those of forestland and grassland, indicating that the spatial variability of SM was lowest for bare land.

The seasonal differences in SM for each soil layer in the three land uses are shown in Table 2. There were significant seasonal differences in SM in the same soil layer for each land use ($p < 0.05$). The mean SM of the three land uses in the different soil layers was largest in autumn. The season with the lowest mean SM in bare land varied with soil depth. The mean SM in the grassland was lowest in winter for each soil layer. The mean SM of L1–L3 in forestland was lowest in winter, whereas the

Table 2 Seasonal differences of soil moisture in each soil layer under different land uses

Land use	Season	L1	L2	L3	L4	L5	L6	L7	L8	L9	L10
Bare land	Winter	**7.47 ± 2.55d**	**7.60 ± 3.01c**	**10.48 ± 3.81d**	**13.18 ± 4.12d**	16.07 ± 3.17c	18.03 ± 1.59b	17.98 ± 1.24b	18.25 ± 1.05b	20.76 ± 0.88b	20.00 ± 0.64b
	Spring	10.28 ± 3.46c	10.31 ± 2.44b	12.04 ± 1.82c	14.46 ± 1.84c	**15.35 ± 1.38d**	**16.57 ± 0.92c**	**16.40 ± 0.90d**	16.69 ± 0.73c	19.16 ± 0.59c	18.66 ± 0.41c
	Summer	12.15 ± 4.72b	12.52 ± 3.38a	14.35 ± 2.68b	17.11 ± 2.86b	17.96 ± 2.77b	17.96 ± 2.35b	16.93 ± 2.40c	**16.68 ± 2.25c**	**18.67 ± 2.20d**	**17.93 ± 1.56d**
	Autumn	**13.32 ± 3.37a**	**13.27 ± 2.46a**	**15.54 ± 1.96a**	**19.00 ± 2.11a**	**20.64 ± 1.91a**	**20.94 ± 1.58a**	**20.95 ± 1.49a**	**21.24 ± 1.38a**	**23.77 ± 1.18a**	**22.66 ± 1.02a**
Grassland	Winter	**6.00 ± 1.82c**	**7.67 ± 2.85c**	**7.42 ± 2.94c**	**8.88 ± 3.03d**	**8.35 ± 2.20d**	**7.98 ± 1.10d**	**8.27 ± 0.24d**	**7.17 ± 0.05c**	**7.95 ± 0.03c**	**8.14 ± 0.04b**
	Spring	10.61 ± 3.89b	12.24 ± 2.52b	11.25 ± 1.73b	12.01 ± 0.90c	10.56 ± 0.48c	9.15 ± 0.37c	8.84 ± 0.26c	7.57 ± 0.18c	8.24 ± 0.13bc	8.35 ± 0.11b
	Summer	10.15 ± 6.64b	11.80 ± 6.37b	11.87 ± 6.03b	13.38 ± 5.70b	12.26 ± 4.78b	10.40 ± 4.04b	9.44 ± 3.08b	8.05 ± 2.40b	8.43 ± 1.88b	8.27 ± 1.03b
	Autumn	**14.47 ± 4.19a**	**17.51 ± 3.38a**	**18.02 ± 3.59a**	**20.59 ± 3.57a**	**20.16 ± 2.99a**	**19.53 ± 2.51a**	**19.27 ± 1.76a**	**18.79 ± 1.62a**	**20.48 ± 1.61a**	**19.75 ± 1.32a**
Forestland	Winter	**5.82 ± 2.36c**	**8.46 ± 3.20c**	**9.19 ± 3.07c**	10.27 ± 3.10c	11.52 ± 2.41c	10.39 ± 1.15b	10.22 ± 0.40b	8.31 ± 0.31c	7.63 ± 0.26c	6.93 ± 0.10bc
	Spring	10.57 ± 3.68a	13.23 ± 3.48b	12.69 ± 2.66b	12.45 ± 2.25b	12.60 ± 2.01b	10.76 ± 1.47b	10.57 ± 1.05b	8.96 ± 0.65b	8.33 ± 0.34b	7.43 ± 0.18b
	Summer	7.34 ± 5.62b	9.46 ± 5.66c	9.37 ± 5.52c	**9.11 ± 4.66d**	**8.72 ± 3.62d**	**7.11 ± 2.63c**	**7.29 ± 1.83c**	**6.69 ± 0.97d**	**6.85 ± 0.52d**	**6.62 ± 0.43c**
	Autumn	**11.22 ± 4.30a**	**14.73 ± 4.93a**	**15.61 ± 5.06a**	**16.64 ± 5.03a**	**17.74 ± 4.74a**	**16.41 ± 3.84a**	**17.04 ± 3.26a**	**15.61 ± 2.91a**	**14.97 ± 3.22a**	**12.56 ± 3.59a**

*Means ± standard deviations in the same soil depth followed by the same letter are not significantly different for the same land use ($p > 0.05$)

SM of the other soil layers was lowest in summer. In general, the mean SM of the shallow soil layers was lowest in winter under each land use. The soil water storage at a depth of 0 ~ 1 m under each land use in winter and spring followed the order of bare land > forestland > grassland, whereas the soil water storage in summer and autumn followed the order of bare land > grassland > forestland.

Temporal stability of the soil moisture distribution among different soil layers

The temporal persistence of the spatial pattern of SM among different soil layers during the monitoring period was described using the Spearman rank correlation coefficient (Fig. 4). The spatial patterns of SM on a temporal scale for each soil layer were all highly significant ($p < 0.01$), which indicated that the SM pattern in the different soil layers had a strong temporal stability. The temporal stability of the spatial pattern of SM was strongest in three adjacent soil layers, with all correlation coefficients above 0.67. The temporal stability of the SM spatial pattern in non-adjacent soil layers was weaker than that in adjacent soil layers. The less adjacent is the two soil layers, the weaker the temporal stability of the spatial pattern of SM for the two soil layers. The temporal stability of the spatial pattern of SM was greatest for grassland and forestland in all soil layers. The Spearman rank correlation coefficients of the spatial pattern of SM on a temporal scale among different soil layers for bare land, grassland, and forestland ranged from 0.16 to 0.99, 0.42–0.97, and 0.38–0.95, respectively. The temporal stability of the spatial pattern of SM was stronger for grassland than for the other land use types.

Soil water consumption and replenishment under different land uses

The consumption and replenishment of soil water in different seasons for each land use are shown in Fig. 5. The amount of soil water replenished was much larger than the actual evapotranspiration (soil water consumption) in summer for each land use, which was due to the concentrated rainfall in summer. In other seasons, the amount of soil water replenished was generally lower than the actual evapotranspiration. The amount of soil water replenished for grassland and forestland in summer was 1.32 and 1.21 times that of bare land, respectively. The actual evapotranspiration for bare land in winter and spring was larger than that for grassland and forestland, while the actual evapotranspiration for bare land in summer and autumn was lower than that for grassland and forestland. The amounts of soil water replenished for bare land, grassland, and forestland during the study period were 526.21, 653.02, and 608.39 mm, respectively. The actual evapotranspiration for bare land, grassland, and forestland during the study period was 533.40, 576.77, and 555.70 mm, respectively. Therefore, the amount of soil water replenished and actual evapotranspiration for bare land during the study period were both lower than those of grassland and forestland. The difference between the amount of soil water replenished and actual evapotranspiration for bare land was small, and the amount of soil water replenished for grassland and forestland were 76.25 and 52.69 mm higher than the actual evapotranspiration, respectively.

Effects of meteorological factors on actual evapotranspiration

An RDA was conducted to quantify the effect of meteorological factors on actual evapotranspiration in different seasons

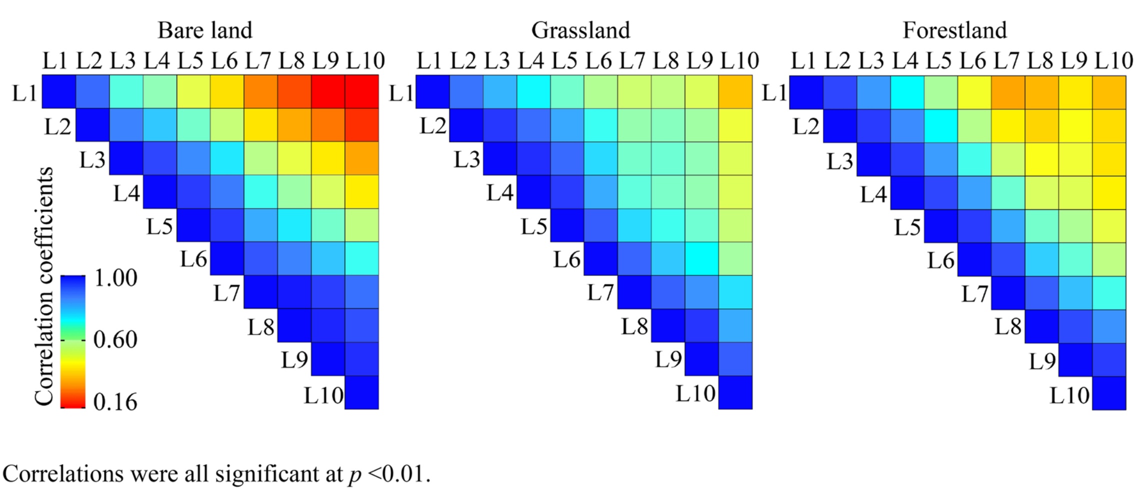

Fig. 4 Spearman correlation coefficients for the soil moisture distribution on a temporal scale among different soil layers

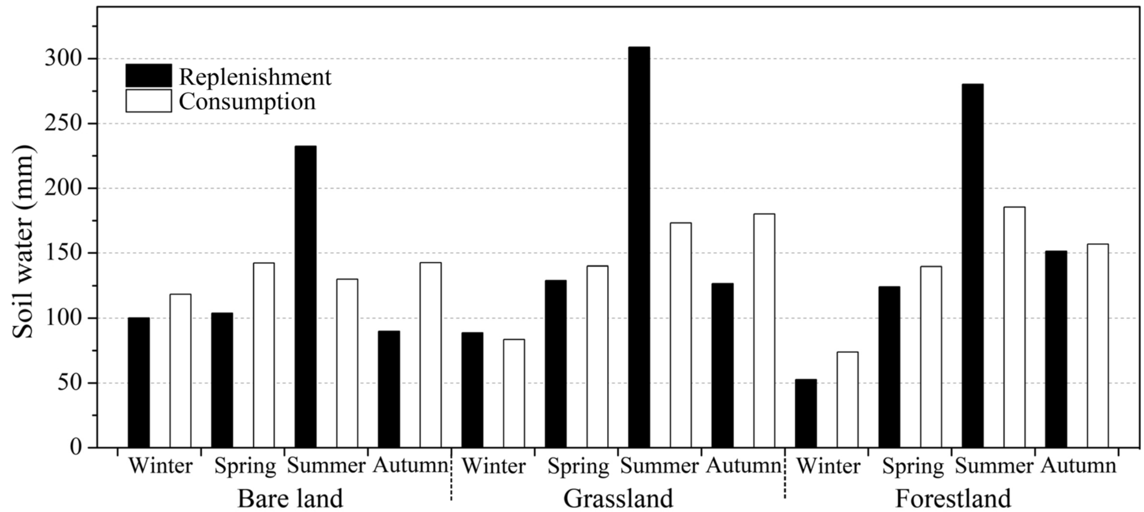

Fig. 5 Soil water consumption and replenishment in different seasons for different land uses

for each land use (Table 3). The RDA results showed that the effect of meteorological factors on actual evapotranspiration had significant seasonal differences. The air temperature (AT) and total solar radiation (TSR) were the main meteorological factors affecting actual evapotranspiration. Vapor pressure (VP) and average wind speed (AWS) had relatively weak effects on actual evapotranspiration for the different land uses. For bare land, atmospheric pressure (AP) and relative humidity (RH) explained 16 and 10% of the variation in actual evaporation in summer and autumn, respectively. Meteorological factors accounted for more than 50% of the variation in the actual evapotranspiration of grassland and forestland in spring and autumn, while the contribution in summer and winter was relatively low. The total contribution of meteorological factors to the actual evaporation of bare land in spring and winter was 65 and 42%, respectively. The impact of meteorological factors on actual evapotranspiration in spring and autumn was greater for forestland and grassland than for bare land, while the impact of meteorological factors on actual evapotranspiration was weaker for all land uses in summer and winter. Therefore, vegetation restoration would influence the impact of meteorological factors on actual evapotranspiration.

The amount of soil water involved in freeze–thaw processes under different land uses

The daily freeze–thaw process of soil water was analyzed with regard to the changes of soil temperature and SM under different land uses (Fig. 6). The freeze–thaw process was

Table 3 The impact of meteorological factors on actual evapotranspiration in different seasons

Land use		AT	RH	VP	AP	AWS	TSR	Total explanation
Bare land	Spring	23%	6%	3%	2%	1%	**30%**	65%
	Summer	8%	2%	1%	**16%**	1%	3%	31%
	Autumn	7%	**10%**	8%	0%	2%	1%	28%
	Winter	**16%**	1%	8%	2%	10%	5%	42%
Grassland	Spring	**32%**	5%	7%	6%	1%	29%	80%
	Summer	3%	2%	7%	6%	1%	**9%**	28%
	Autumn	18%	2%	11%	4%	2%	**19%**	56%
	Winter	**10%**	0%	4%	1%	8%	3%	26%
Forestland	Spring	**33%**	6%	6%	6%	2%	17%	70%
	Summer	2%	2%	1%	**5%**	3%	3%	16%
	Autumn	**30%**	2%	20%	12%	1%	28%	93%
	Winter	**14%**	1%	6%	3%	3%	4%	31%

AT Air temperature; *RH* Relative humidity; *VP* Vapor pressure; *AP* Atmospheric pressure; *AWS* Average wind speed; *TSR* Total solar radiation

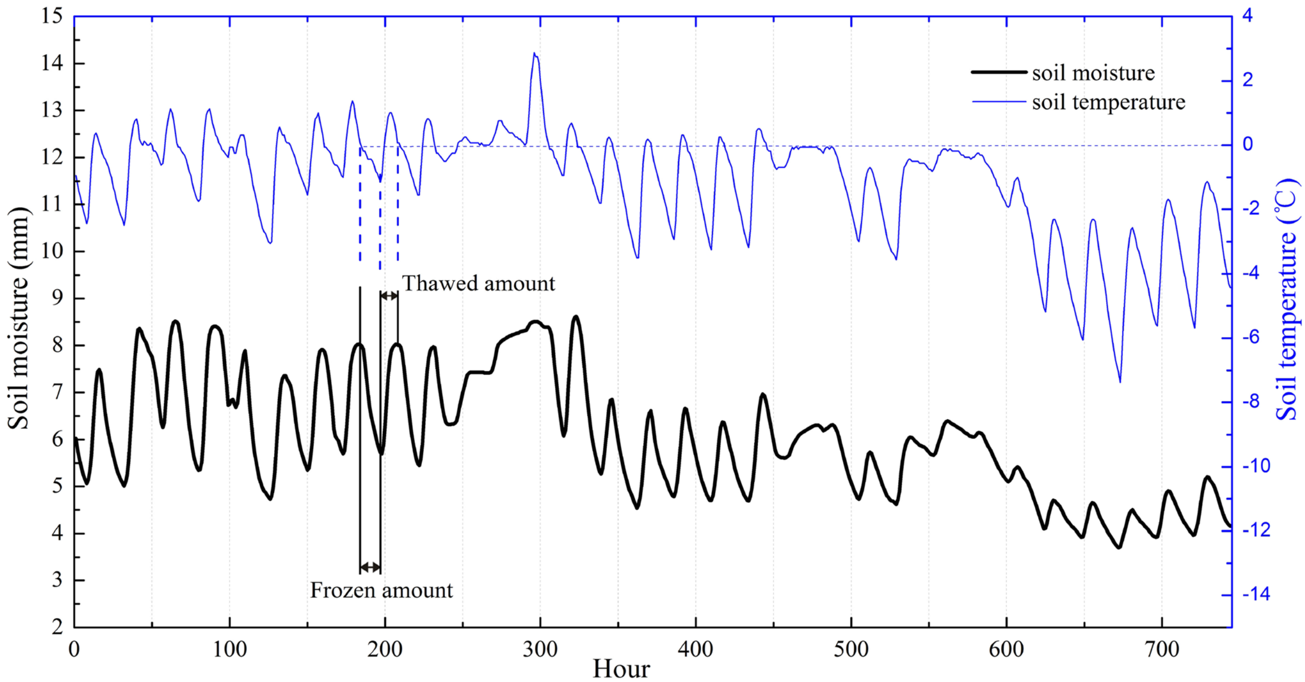

Fig. 6 The soil water freeze–thaw process under grassland in the L1 layer in December 2016

consistent with the changes in soil temperature. The cumulative amount of frozen soil water over 1 year under the different land uses is shown in Table 4. There was little difference between the cumulative amounts of frozen and thawed soil water under each land use during the study period. The cumulative amounts of frozen soil water for bare land, grassland, and forestland were 495.98, 334.78, and 213.15 mm, respectively. The freeze–thaw depth of grassland and forestland was 10 cm deeper than that of bare land. The number of freeze–thaw days in L1 followed the order of bare land > forestland > grassland, accounting for about 30% of the whole year. The number of freeze–thaw days for different land uses in other soil layers did not follow the same order. The cumulative frozen amount and number of freeze–thaw days decreased with increasing soil depth. The freeze–thaw process mainly occurred in the surface layer (i.e., 0–20 cm), which contained more than 80% of the cumulative frozen soil water.

Discussion

Temporal and spatial differences of soil moisture under different land uses

Fine-scale evaluation of the temporal and spatial changes of soil moisture in different land uses is essential for vegetation restoration and watershed management (Huang and Shao, 2019). The soil water storage at 0–100 cm depth for grassland and forestland was 72 and 65% of that for bare land, respectively. The soil water storage at 50–100 cm depth for grassland and forestland was 59 and 53% of that for bare land, respectively. This indicated that vegetation restoration does reduce SM. This agreed with previous studies (Cao et al. 2011; Yang et al. 2014; Jian et al. 2015), which reported that vegetation restoration can decrease the SM in both the shallow and deep soil layers. Lower soil

Table 4 The amount of frozen soil water over 1 year under different land uses

Land use		L1 (mm)	L2 (mm)	L3 (mm)	L4 (mm)	L5 (mm)	L6 (mm)
Bare land	CFA	359.89	68.41	43.99	19.63	4.06	/
	FTD	116	72	59	23	3	/
Grassland	CFA	202.93	80.57	21.16	13.03	8.58	8.51
	FTD	104	85	54	33	21	19
Forestland	CFA	133.13	38.36	17.61	10.48	9.16	4.41
	FTD	109	67	44	23	16	15

CFA Cumulative frozen amount of soil water during the study period; *FTD* Frozen and thawed days water-holding capacity, higher plant density, less rainfall and more concentrated precipitation distribution are the main driving factors for the formation of dry soil layer (Jia et al. 2015).

The mean vegetation cover in the Loess Plateau increased by 14.28% from 2000 to 2015 (Xu et al. 2018). Vegetation cover in the study area has increased significantly (Fig. 1b), which may have led to a decrease in the SM over a large area. Continuous reduction of soil water can induce vegetation degradation. Jia et al. (2017) and Yang et al. (2012) showed that the SM of deep soil depth can decrease by more than 35% after vegetation restoration. However, this study found that the amount of soil water replenished in grassland and forestland was greater than the amount consumed in the four seasons of the study period. Therefore, changes in the SM are closely related to the vegetation restoration stage and annual recharge. A reduction of SM does not occur every year.

The largest mean SM of the three land uses in autumn was due to the 118.6 mm of rainfall received from 18 to 31 August 2017 (Fig. 2). The rainfall during this period caused the high soil moisture state in autumn. This was consistent with other published reports (Liu and Shao 2014; Wang et al. 2015b; Xu et al. 2017). This phenomenon is common in the Loess Plateau because rainfall is mainly concentrated from July to September. There were not only seasonal differences in SM, but also significant differences in the SM of different soil layers. The levels of significant difference and range in the SM of the different soil layers for grassland and forestland were generally smaller than that for bare land (Table 5).

Table 5 Differences in soil moisture among different soil layers for each land use

Soil layer	Bare land	Grassland	Forestland
L1	10.82 ± 4.22 g	10.32 ± 5.38f	8.74 ± 4.71e
L2	10.94 ± 3.59 g	12.31 ± 5.36bc	11.48 ± 5.12bc
L3	13.11 ± 3.32f	12.15 ± 5.43bcd	11.72 ± 5.00bc
L4	15.94 ± 3.64e	13.72 ± 5.66a	12.12 ± 4.86ab
L5	17.51 ± 3.15d	12.84 ± 5.38b	12.64 ± 4.68a
L6	18.37 ± 2.32c	11.77 ± 5.18cde	11.16 ± 4.18c
L7	18.06 ± 2.39c	11.45 ± 4.87de	11.27 ± 4.06c
L8	18.21 ± 2.37c	10.39 ± 5.07f	9.89 ± 3.75d
L9	20.58 ± 2.37a	11.27 ± 5.46e	9.44 ± 3.62d
L10	19.80 ± 2.07b	11.12 ± 5.05ef	8.38 ± 3.02e

*Means ± standard deviations of the different soil layers followed by the same letter are not significantly different in the same land use ($p > 0.05$)

The main factors affecting soil moisture changes

The amounts of soil water replenished under bare land, grassland, and forestland from 1 December 2016 to 30 November 2017 were 526.21, 653.02, and 608.39 mm, respectively, which were higher than the precipitation (490.60 mm) received during the period. This was because dew is an important source of soil water in semi-arid regions. The average daily amount of dew was 0.75 mm in Jujube forest from July to October (Gao 2014). Glenn et al. (1996) showed that the amount of dew received from 27 September to 6 November was 33% of the daily transpiration. Zhang et al. (2012) found that in the semi-arid region of the Loess Plateau in central Gansu, the non-rainfall land surface water from the atmosphere accounted for 15% of the total land surface water source.

The temporal variation in SM was also driven by meteorological properties such as seasonal changes in temperature and evapotranspiration (Jia and Shao 2014; Zhang et al. 2017a, b). The temporal variability of the spatial pattern of SM for grassland and forestland on a temporal scale was higher than that of bare land. This was likely because AT and TSR were generally the main meteorological factors affecting actual evapotranspiration. In grassland and forestland, the effect of AT and TSR on the variation in soil temperature were weaker than for other land uses. There were differences in the main factors influencing the change of SM in different seasons. The main factors influencing SM in bare land in spring were meteorological, while the main influencing factors in other seasons were internal soil factors. The main factors influencing SM in grassland and forestland in spring and autumn were meteorological, while the main influencing factors in summer and winter were soil internal factors.

Land use is the main influencing factor of SM. The soil internal factors affecting the spatial and temporal changes of SM were mainly roots, soil texture, and SOC. Gao and Shao (2012) found that the soil clay content was the main factor affecting the spatial distribution of soil water on a semi-arid hillslope. Li et al. (2018) also reported that soil clay content and topography were the most important factors influencing SM in gully areas. The current study area is similar to the above study areas, but the topography of the different land uses is the same. Other studies have reported that soil particles, root density, and SOC are the primary factors influencing the temporal characteristics of SM for hillslope-scale vegetation (Crave and Gascuel-Odoux 1997; Famiglietti et al. 2008; Jacobs et al. 2004; Cheng et al. 2017; Xu et al. 2017).

Impact of land use on the freeze–thaw process

The freeze–thaw process can effectively alter soil structure (Bullock et al. 2001; Sun et al. 2018), which has an important impact on soil erosion processes (Sun et al. 2018; Wang et al. 2014, 2017; Wu et al. 2018) and nutrient loss processes (Cheng et al. 2018; Xiao et al. 2019). Moreover, the freeze–thaw cycle ultimately affects runoff generation, flow concentration and runoff yield by changing soil properties (Wu et al. 2020). However, previous studies have focused on the effects of temperature and SM on the freeze–thaw process, and have given little attention to the differences in freeze–thaw depth, amounts of water involved in the freeze–thaw process, and the number of freeze–thaw cycles. Xiao et al. (2019) found that SOC was more sensitive to freeze–thaw cycles in forests than in natural-succession grassland, but ignored the difference in freeze–thaw cycles caused by SM and soil temperature for different land use types. Wang et al. (2017) reported that the freeze–thaw cycle increased the amount of soil erosion compared to a control slope, but the study was based only on a single freeze–thaw cycle. It was found that for different land uses, there were significant differences in freeze–thaw depth, the cumulative amount of frozen soil water, and freeze–thaw days. Grassland and forestland reduced the cumulative amount of frozen soil water compared to bare land. Special consideration should be given to the freeze–thaw process at a soil depth of 0–20 cm, where the cumulative amount of frozen soil water in different land uses exceeded 80% of the total cumulative frozen amount.

Conclusions

Soil moisture was not only significantly lower under grassland and forestland, but the spatial variability was also higher at a depth of 1 m than that of bare land. Following the consumption and replenishment of soil moisture, grassland and plant density should be taken into consideration when implementing vegetation restoration in arid regions to achieve sustainable vegetation development. Moreover, lowering ground temperature and increasing dew recharge can effectively improve soil water storage. In addition, the freeze–thaw process mainly occurred at soil depths of 0–20 cm. The soil freeze–thaw experiment should pay attention to the soil depth. The number of freeze–thaw cycles and the setting of soil temperature also need to be optimized.

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