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Is annual recharge coefficient a valid concept in arid and semi-arid regions?

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Abstract. Deep soil recharge (DSR) (at depth greater than 200 cm) is an important part of water circulation in arid and semi-arid regions. Quantitative monitoring of DSR is of great importance to assess water resources and to study water balance in arid and semi-arid regions. This study used a typical bare land on the eastern margin of Mu Us Sandy Land in the Ordos Basin of China as an example to illustrate a new lysimeter method of measuring DSR to examine if the annual recharge coefficient is valid or not in the study site, where the annual recharge efficient is the ratio of annual DSR over annual total precipitation. Positioning monitoring was done on precipitation and DSR measurements underneath mobile sand dunes from 2013 to 2015 in the study area. Results showed that use of an annual recharge coefficient for estimating DSR in bare sand land in arid and semi-arid regions is questionable and could lead to considerable errors. It appeared that DSR in those regions was influenced by precipitation pattern and was closely correlated with spontaneous strong precipitation events (with precipitation greater than 10 mm) other than the total precipitation. This study showed that as much as 42 % of precipitation in a single strong precipitation event can be transformed into DSR. During the observation period, the maximum annual DSR could make up 24.33 % of the annual precipitation. This study provided a reliable method of estimating DSR in sandy areas of arid and semi-arid regions, which is valuable for managing groundwater resources and ecological restoration in those regions. It also provided strong evidence that the annual recharge coefficient was invalid for calculating DSR in arid and semi-arid regions. This study shows that DSR is closely related to the strong precipitation events, rather than to the average annual precipitation, as well as the precipitation patterns.

1 Introduction

Recharge is an important source of groundwater budget and it is also a fundamental process that links the surface hydrological processes (e.g., precipitation), vadose zone process (e.g., infiltration and soil moisture dynamics), and the saturated zone process (e.g., groundwater flow) (Sanford, 2002; McWhorter and Sunada, 1977). How to accurately estimate recharge has remained a persistent challenge and an active research topic in the hydrological science community over many decades (Gee and Hillel, 1988; Scanlon, 2013; Sanford, 2002). It is generally accepted that recharge is correlated with precipitation in some fashions, and many studies adopt the concept of a recharge coefficient (Turkeltaub et al., 2015; Kalbus et al., 2006; Allocca et al., 2014), which is the ratio of the actual recharge to the precipitation, to estimate the recharge (Fiorillo et al., 2015; Allocca et al., 2014). The magnitude of such a recharge coefficient is controlled by a complex interplay of multiple factors such as moisture dynamics in the vadose zone (Schymanski et al., 2008), depth to water table, vegetation, etc., and the recharge coefficient is often regarded as a temporally invariant value at a given location (Fiorillo et al., 2015; Min et al., 2017; Vauclin et al., 1979). Specifically, it is assumed to be primarily controlled by the total precipitation and not too much by the temporal fluctuation of precipitation events (Hickel and Zhang, 2006; Acworth et al., 2016). In this study, we will challenge the concept of using a constant recharge coefficient to estimate the recharge in arid and semi-arid regions based on a multi-year field investigation.

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As water tables in many arid and semi-arid regions are relatively deep (greater than 2 m below ground surface) (Williams, 1999; Soylu et al., 2011), recharge in those regions is named deep soil recharge (DSR), which will be the concern of this study. DSR could ease the demand of sand-fixing vegetation on moisture during extremely dry seasons (Zhang et al., 2001; Shou et al., 2017), and it reduces water deficit, sustains life activities, and helps the vegetation live through extreme droughts (Zhang et al., 2004). In this sense, DSR is an important factor of the water cycle in arid and semi-arid regions (Adolph, 1947), and it could also provide much-needed references for the stability analysis of sand-fixing vegetation (Li et al., 2004, 2014). In the following, we will briefly review the existing methods of estimating DSR.

In general, there are three methods of measuring DSR in arid and semi-arid regions. The first method is an empirical approach which assigns a constant recharge coefficient associated with a certain precipitation event (Allison et al., 1994; Jiménez-Martínez et al., 2010). The empirical approach is simple to use but it lacks a rigorous theoretical base, and the recharge coefficient has to be calibrated through a groundwater flow model in the region, which is often not available.

The second method is a modeling approach involving numerical models such as HYDRUS (Šimůnek et al., 2012), SWAT (Arnold et al., 2012), UNSATH (Fayer, 2000), SWIM (Krysanova et al., 2005), and SWAP (van Dam, 2000) to calculate DSR. Modeling is an efficient way to test different hypothetical scenarios and it may be used to predict DSR in the future if the model is calibrated carefully. Detailed water balance models can be used for irrigated agriculture, but they usually cannot predict evapotranspiration accurately, especially when plants suffer seasonal water stress and plant cover is sparse (Gee and Hillel, 1988). When recharge is estimated as residual in water balance models, it can cause miscalculation as much as an order of magnitude (Scanlon, 2013; Voeckler et al., 2014). When using soil water flow models with measured or estimated soil hydraulic conductivities and tension gradients, similar miscalculation can also occur (Nyman et al., 2014; Gee and Hillel, 1988). In addition, the modeling usually involves upscaling of parameter values over a spatially and temporally discretized mesh from measurements which are made on specific moments and locations. Such an upscaling process is not always easy to execute and it could sometimes lead to serious errors. This is particularly true for arid and semi-arid regions where most precipitation may be episodic (occurring in short and unpredictable events) (Modarres and da Silva, 2007; Zhou et al., 2016) and may be confined to restricted portions of the area (Gee and Hillel, 1988).

The third method includes a cluster of experimental techniques such as isotopic tracers (Klaus and McDonnell, 2013), water fluxes (Katz et al., 2016), and lysimeters (Scanlon, 2013). Among them, lysimeters are instruments that directly measure the hydrological cycle in infiltration, runoff, and evaporation. Generally, this instrument is located in an open observation field or as a controlled device, working either solely or in groups (Good et al., 2015). In a typical lysimeter, soil is filled into a column surrounded by impermeable lateral boundaries; thus, water can only enter or leave the column from upper or lower boundaries (Duncan et al., 2016; Fritzsche et al., 2016). A drainage system is usually placed at the bottom (Glenn et al., 2013). The depth of soil in the column depends on the experimental purpose. Experiments can be done with the same type of soil at different depths in a single column or in different columns but at the same depth. The soil surface can be cultivated with different crops or left as bare land. Observations can be recorded with weight or volume of water.

Application of above-mentioned methods for assessing DSR in arid and semi-arid regions has met a variety of challenges, primarily due to the fact that precipitation events often happen in the form of short pulses with highly variable intensity (Collins et al., 2014). The intermittent and unpredictable characteristics of precipitation events led to highly variable moisture and nutrient levels in the soils (Beatley, 1974; Huxman et al., 2004). It is unclear how the precipitation amount, time, and interval will affect the water moisture of arid and semi-arid regions, especially the change of deep soil water storage.

In this study, a new type of lysimeter is designed to accurately measure the amount of DSR in arid and semi-arid regions. With the help of a 3-year (2013–2015) field investigation with this new lysimeter, one can answer the following question: is the concept of an annual recharge coefficient valid or not for estimating DSR at a given location in arid and semi-arid regions? Before the introduction of this new type of lysimeter, it is necessary to briefly explain the challenges faced by the conventional lysimeter for studying DSR in arid and semi-arid regions.

2 Design of the new lysimeter for DSR measurement

2.1 Problems with the conventional lysimeter methods in arid and semi-arid regions

Lysimeters have been used to access the amount of water consumed by vegetation for more than 300 years (Howell et al., 1991). The type of lysimeter that is specifically designed to measure evapotranspiration (ET), called the precision-weighing lysimeter, has been developed within the past six decades. In order to satisfy different requirements and needs, there are various designs of weighing lysimeters, with surface areas ranging from 1.0 to over 29 m^2 (Howell et al., 1991). The stored media mass and the type of scale such as diameter and height are factors on which the accuracy of ET measurement depends, and many lysimeters have accuracy better than 0.05 mm (Howell et al., 1991). Figure 1a shows the schematic diagram of a conventional lysimeter installation in the field. It is basically a weight meter of soil with an open upper boundary at ground surface and a perforated bottom boundary and impermeable vertical side walls. The typical depth of lysimeters varies from 0.2 to 2 m but is rarely greater than 2.5 m (Howell et al., 1991). The horizontal cross-section area is usually in the range of 1 to 29 m^2. Precipitated water can freely infiltrate into the soil from the top, and downward flow of water at the bottom of the lysimeter is collected (through the perforation) as a function of time to calculate the recharge. Alternatively, the weight of combined water and soil inside the lysimeter can be accurately measured using a weight gauge to reflect any soil moisture change. Such information, combined with infiltration or evaporation at the surface, can yield the information of downward water flux at the depth of lysimeter.

The following issues deserve special attention when applying the conventional lysimeter for measuring recharge. Firstly, soil layers are inevitably disturbed when installing the instrument, so the result may not reflect the actual recharge in native (undisturbed) soils (Weihermüller et al., 2007). Secondly, the cost is too high to use multiple lysimeters to observe large-scale infiltration (Stessel and Murphy, 1992). Thirdly, when precipitation strength is relatively light and concentrated, a large lysimeter cannot sensitively and rapidly measure DSR (Goldhamer et al., 1999; Farahani et al., 2007). The conventional lysimeter often cannot answer the questions as to what soil layer different levels of precipitation can infiltrate and how much the infiltration amount is under different levels of precipitation (Gee and Hillel, 1988; Ogle and Reynolds, 2004).

The conventional lysimeter as shown in Fig. 1a may meet additional challenges when applied to arid and semi-arid regions. Firstly, the water table depths in arid and semi-arid regions may be much greater than the maximal depth of a conventional lysimeter (2.5 m). For instance, in Chagan Lake, southeast of Mu Us Sandy Land in the Ordos Basin of China, the water table depth was found to be greater than 4 m. In the Gobi Desert, the water table was reported to be at least 2.8 m deep (Ma et al., 2009). Therefore, the infiltration measured at the base of a conventional lysimeter may not represent the actual recharge that eventually enters the groundwater system. Secondly, the measurement accuracy of lysimeter often declines for soils with deep plant roots because the depth of lysimeter installation is limited and it may be less than the depth of those roots at site, which by itself can be important pathways for water migration. Consequently, the measured recharge of such disturbed soil by lysimeter may not represent the in situ recharge of the native (undisturbed) soil.

To resolve the above-mentioned issues faced by the conventional lysimeter, a new type of lysimeter is designed with specific considerations of the unique precipitation patterns and soil characteristics in arid and semi-arid regions. This new lysimeter is illustrated schematically in Fig. 1b.

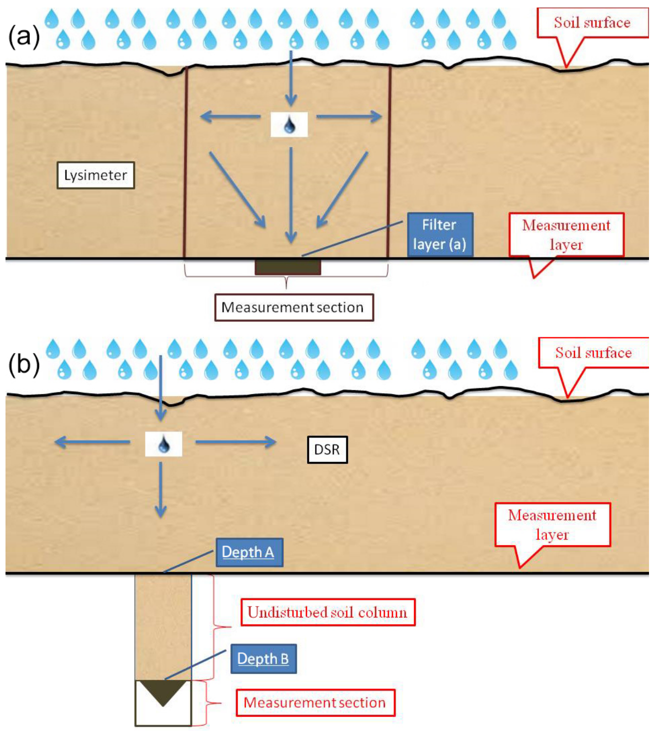

Figure 1. Schematic diagram of conventional lysimeter **(a)** and the new lysimeter **(b)**.

2.2 Design of a new lysimeter for measuring DSR in arid and semi-arid regions

This new lysimeter has a few innovations (see Fig. 1b) that can be outlined as follows. Instead of setting the upper boundary of the lysimeter at ground surface, the new design has its upper boundary at a designed depth (denoted as depth A in Fig. 1b) where infiltration will be measured. A cylindrical container with a diameter of 20 to 40 cm with impermeable walls is installed from depth A downward to a deeper depth B. The length of AB is determined according to the capillary rise of the in situ soil, which can be calculated using the average grain size of soils within AB. More specifically, the length of AB is greater than the capillary rise of soils within AB and it is usually greater than 0.6 m (Liu et al., 2014). At the soil surface, there is a device to measure the amount of the precipitation, and at the base of the instrument (depth B), a water collection device is used to measure the amount of water exit the base.

Before the measurement, one necessary preparation is to inject water from the top of the instrument at depth A using water pumps; the injection will stop until water starts to drip out from the base at depth B. One usually has to wait 10 days to allow the water profile in column AB to reach equilibrium. When water stops flowing out from depth B, the soil water in the column is regarded as reaching its equilibrium state, in which the soil moisture at depth B reaches the maximum field capacity. Under such an equilibrium status, the amount of infiltration entering the upper surface of the lysimeter will be discharged (with the same amount) from the base of the lysimeter after a certain delay time.

The proposed new method has a few innovative features that have not been considered in previous studies. Firstly, it can measure DSR at any given layer of a multi-layer soil system using a single apparatus installed in the field. Secondly, continuous real-time measurements can be recorded over any given time period; thus, a time series of DSR can be obtained, which will be very useful to understand the soil water dynamics at sandy areas of arid and semi-arid regions. Thirdly, the apparatus is portable and easy to install; thus, a large amount of data can be collected in various locations of a study area using multiple lysimeters, and spatial recharge distribution can also be obtained straightforwardly. This method is field tested in arid and semi-arid sandy regions of western China. It provides key references for the evaluation of water resources, water balance, and the stability assessment of sand-fixing vegetation in arid and semi-arid areas. It also provides data that are much needed for evaluating soil water contents and groundwater resources of those areas. An important feature of this new lysimeter is that it can provide reliable DSR data to examine the concept of the annual recharge coefficient when comparing with the precipitation data.

3 Field testing of the new apparatus

3.1 Description of the study area

Figure 2 shows the location of the study which is located in Ejin Horo Banner, on the eastern margin of Mu Us Sandy Land in the Ordos Basin of China (geographic location: 39°05′ N, 109°36′ E; altitude: 1070–1556 m a.s.l.). The groundwater tables between dunes are 5.3–6.8 m below ground surface. The area is located in a semi-arid continental monsoon climate zone. Precipitation concentrates from July to September, with relatively concentrated rainstorms. The average annual precipitation from 1960 to 2010 is 296.01 mm. The average annual temperature of this area is 6.5 °C, with about 151 days of frost-free season, 1809 mm total evaporation, an average of 2900 h of sunshine, and an average wind speed of 3.24 m s^{-1} (Wu and Ci, 2002; Karnieli et al., 2014). The study area is located in relatively gentle mobile dunes, and the soil type is Aeolian sandy soil.

In terms of geological structure, Mu Us Sandy Land is in the Ordos Basin, a large-scale syncline sedimentary basin with nearly north–south striking axis, and is of Mesozoic and Paleozoic ages. The basin covers an area of 640 km from north to south and 400 km from east to west. The axis of syncline is off west, and the east and west wings are asymmetric. The east wing is Monoclinic of westward tilt with a width of more than 300 km. The west wing is made of many fault-fold belts striking along the north–south direction and thrusting eastward with a width of less than 100 km. The southern boundary of the basin is Weibei plateau uplift. The southern part of this plateau uplift is descending in a ladder shape with blocks to Fenwei rift-subsidence basin. The northern boundary of this basin is Yimeng plateau uplift, with a lack of Lower Paleozoic, and its edge fault is connected to the Hetao fault basin. The basement of the Ordos Basin is of Precambrian crystalline metamorphic rocks.

Deposited in the basin are, in turn, Lower Paleozoic carbonate rocks, Upper Paleozoic–Mesozoic clastic rocks, and Cenozoic sedimentary rocks with a total depth of more than 6000 m. The discontinuous Cenozoic sediment is on top of the Mesozoic and Paleozoic layers, mainly of the Quaternary and partially of the Tertiary sediments. The Quaternary layer is mainly made of Aeolian sand and loess. Generally divided by a line along the Great Wall, the northwest land surface is mainly covered by wind-blown sand layers of varying thickness and a 40–120 m thick layer of alluvial lacustrine; the southeast land surface is mainly covered by loess with various thickness from tens of meters to more than 200 m. Below the loess layers, there is Tertiary Pliocene mud rock with thickness of a few meters to tens of meters.

The hydrostratigraphic units of the Ordos Basin are quite complex, consisting of multiple connected aquifers. Following the order from bottom to top, the multiple aquifers are primarily made from various rock types of a karst aquifer consisting of Precambrian and Ordovician limestone, a fractured aquifer consisting of Carboniferous and Jurassic clastic rocks, a porous-fractured aquifer of Cretaceous clastic rocks, and a porous aquifer consisting of unconsolidated Cenozoic and Quaternary sediments. Generally speaking, Mu Us Sandy Land has relatively rich groundwater resources. The shallow groundwater reservoir is estimated to hold about 120.3 billion metric tons of freshwater. Groundwater is mainly recharged by precipitation with an annual average recharge amount of 1.4 billion metric tons. Fine sands are the dominating sediments observed in the experimental site. In the upper 200 cm soil layer in the experiment area, the percentage of fine sand (0.5–0.1 mm) is 88.56, 77.88, 88.23, 88.89, 90.28, 83.90, and 84.21 % at depths of 0, 10, 30, 60, 90, 150, and 200 cm, respectively. The rest of the soils are primarily coarse sands. It is evident that the soil at the upper 200 cm is relatively homogeneous.

3.2 Statistical analysis of data

Research on the relationship between precipitation and DSR of bare sand land in arid and semi-arid regions is beneficial to understand the soil water dynamics of those regions. Because vegetation is absent, complexity related to transpiration process by plants is not a concern. Based on two time series of real-time data of precipitation and DSR, one can examine the relationship between DSR and precipitation. This study can serve as a basis for further study of DSR in semi-fixed and fixed sand lands with different fractional vegetation covers.

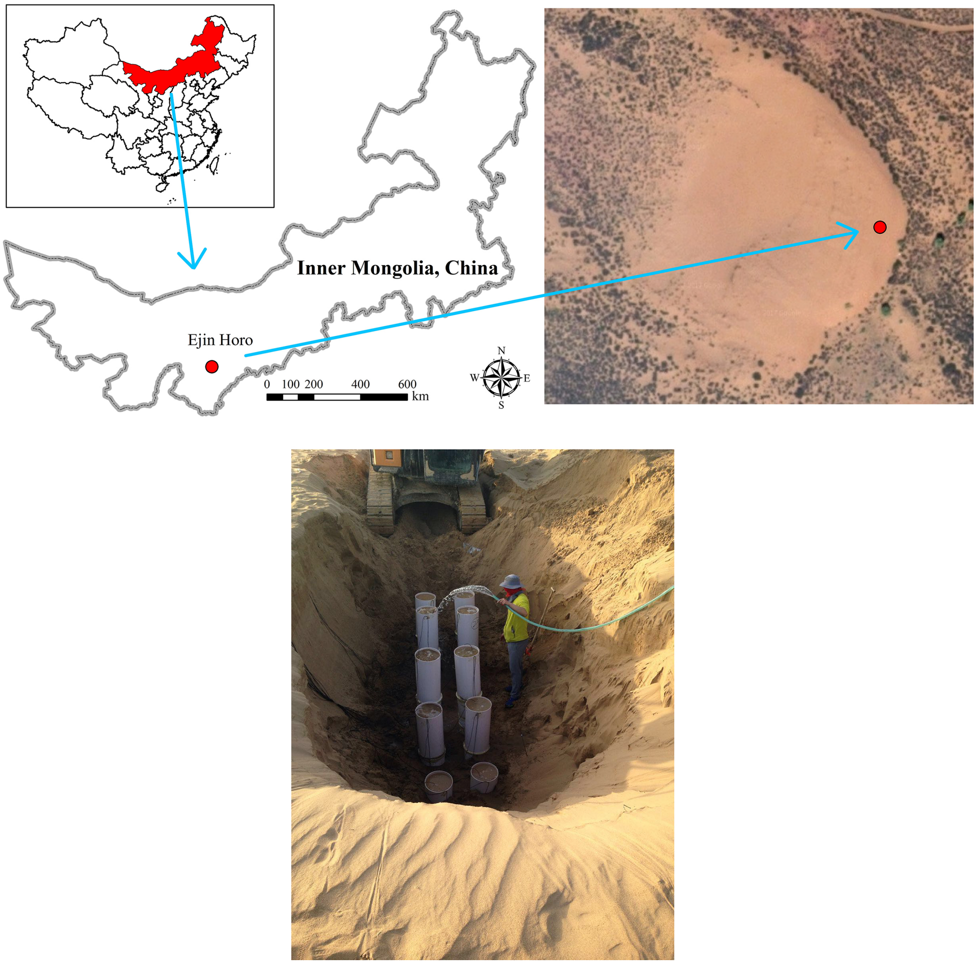

Figure 2. Geographic location of the experimental area.

On 1 September 2012, a mobile sand dune within the study site was set as the monitoring plot (geographic location: 39°05′ N, 109°36′ E; altitude: 1310 m) with the upper 300 cm soil profile excavated. The lysimeter as shown in Fig. 1b was installed following the procedure described in Sect. 2.2 and then backfilled using the excavated soil. Infiltration passing through the upper 200 cm depth is generally regarded as DSR in this study. It is worthwhile to point out that some other investigators may use a more or less different depth threshold for defining DSR. For instance, Zhang et al. (2008) used 140 cm instead of 200 cm depth as the threshold to define DSR. It was found that the water table depth was greater than 5 m in 2012–2015 at the study site, so its influence on DSR was negligible. A precipitation sensor (AV-3665R, AVALON, USA; precision: 0.2 mm) was placed above ground at the site. A data acquisitor (CR200X, Campbell, USA) was used to record DSR, of which DSR data were recorded every hour, and the precipitation data were recorded every half hour. In order to avoid the effect of freeze-and-thaw action, the experiment was conducted between 1 April 2013 and 30 November 2015. During such a 3-year period, no runoff occured at the studied area.

The statistics of precipitation and DSR are shown in Table 1, which reveals that there is an obvious difference of precipitation at the experimental plot from 2013 to 2015. The annual precipitation is 83 mm in 2013, 205.6 mm in 2014, and 186.4 mm in 2015. This is to say that the annual precipitation in 2014 and 2015 was 2.48 and 2.25 times that in 2013, respectively. Such a dramatic fluctuation and uneven distribution of annual precipitation is typical of arid and semi-arid regions. The corresponding annual DSR is 20.2 mm in 2013, 20.6 mm in 2014, and 9.2 mm in 2015. This is to say that the annual DSR values in 2014 and 2015 are 1.02 and 0.46 of that in 2013. The annual DSR/precipitation ratios (or the so-called annual recharge coefficients) for 2013, 2014, and 2015 are 24.33, 10, and 4.94 %, respectively.

It appears that there is no clear correlation between the annual DSR and the annual precipitation according to the data of 2013–2015. In another words, use of the annual recharge coefficient for the study site becomes questionable as such

Table 1. The annual precipitation–DSR relationship from 2013 to 2015.

Year	Precipitation (mm)	DSR (mm)	DSR/precipitation ×100 %
2013	83	20.2	24.33 %
2014	205.6	20.6	10 %
2015	186.4	9.2	4.94 %

a coefficient implies that there is a close correlation between the annual DSR and the annual precipitation, which is not supported by the data of 2013–2015 here. Therefore, we will scrutinize the precipitation pattern and intensity more closely to decipher the connection of precipitation and DSR in the following.

3.2.1 The relationship between precipitation pattern and DSR

Research on bare sandy soil water dynamic processes usually focuses on temporal and vertical differences (Ritsema and Dekker, 1994; Postma et al., 1991). In terms of temporal soil moisture variation over an annual cycle, the process could be divided into soil moisture replenishment, depletion, and relatively stable periods. In terms of vertical soil moisture variation, soil water content usually first increases with depth and then decreases based on an interplay of mutual infiltration and evaporation processes. In general, soil could be divided as a surface dry sand layer, a layer with drastic moisture change, and a layer with relatively stable moisture content. Specifically, the soil deeper than 160 cm in arid and semi-arid regions would have a relatively stable moisture content. This is because of two reasons. Firstly, soil water will not be uptaken to the surface by capillary force at such depths; secondly, ground water table in arid and semi-arid regions is usually much lower than 160 cm.

In our study site, 2013 is an especially dry year with only 83 mm precipitation compared to 296.01 mm of average annual rainfall calculated over a period from 1960 to 2010. The precipitation and DSR patterns of 2013 are shown in Fig. 3. The measurement accuracy of the lysimeter is 0.2 mm. During the observation period from 1 April to 30 November, there are 25 total recorded precipitation events, mostly concentrated in the period from May to August. There is a one-time strongest precipitation event with a 24 h precipitation amount reaching 32 mm on 3 August. The DSR correlated to this event can be identified from 21 September to 30 November and reaches 17.2 mm. The delay time from the precipitation event to the start of DSR is approximately 48 days. The DSR/precipitation ratio for this particular event is as high as 53.75 %. Such a DSR/precipitation ratio appears to be the highest in 2013. It is notable that although the strongest precipitation event on 3 August contributes the greatest to DSR observed from 21 September to 30 November, a few precipitation events with amount of 6.6 mm prior to this strongest precipitation event also contribute a minor part for DSR from 27 July to 1 August. It is also notable that the DSR/precipitation ratio for the strongest precipitation event is substantially higher than the average annual recharge coefficient of 24.33 % in 2013. This leads to the conclusion that DSR is closely related to the strong precipitation events rather than to the average annual precipitation.

In 2014, the annual precipitation is 205.6 mm, and DSR is 20.6 mm, leading to a 10 % annual average recharge coefficient, which is less than half of that in 2013. As shown in Fig. 4, the frequency of precipitation events in 2014 is obviously higher than that of 2013. From 1 April to 30 November, there are total 68 instances of precipitation events, compared to 41 instances in 2013. Furthermore, the precipitation distribution in 2014 is more uniform than that in 2013. Specifically, precipitation events are concentrated in the period from June to August, with the highest 24 h accumulative precipitation of 15 mm on 30 July. As shown in Fig. 4, recorded DSR data cover a period from 1 April to 30 November, and the maximum DSR occurs on 1 October. Because the experimental plot is located in a transition zone between arid and semi-arid regions, summer evaporation is strong, leading to relatively less DSR during the summer season. During the period from 1 September to 30 November, atmospheric temperature drops and sunshine duration becomes shorter, which results in less surface evaporation and greater DSR during this period. Compared to 2013, there are more summer precipitation events in 2014. That is one reason why precipitation in 2014 (205.6 mm) is greater than 2013 (83 mm) but the overall DSR in 2014 is less than that in 2013.

The strongest single-day precipitation in 2014 is 15 mm (occurring on 30 July), which is less than half of the strongest single-day precipitation event of 32 mm that occurred in 2013 (3 August); annual DSR/precipitation ratio is 24.33 % in 2013 but drops to 10 % in 2014. This once again supports the conclusion that the strong precipitation events rather than the average annual precipitation are mostly responsible for the average annual DSR. This is the other reason why precipitation in 2014 (205.6 mm) is greater than that in 2013 (83 mm) but the overall DSR in 2014 is less than that in 2013.

As shown in Fig. 5, the total annual precipitation of 2015 is 186.4 mm, and DSR is 9.2 mm, leading to a 4.94 % annual average recharge coefficient, which is significantly smaller than that of 2013 (24.33 %) and 2014 (10 %). There are total 66 observable precipitation events in 2015. Such precipitation events are mostly concentrated from 4 April to 6 July, with a total precipitation of 155 mm during this period, which represents 83.15 % of the total precipitation in 2015. The measured DSR from 4 April to 6 July is only 7 mm, representing 77.78 % of the total DSR in 2015. Throughout 2015, the two strongest precipitation events happen on 4 April and 5 June; both 24 h precipitation events reach 17.2 mm. We observe a single-day DSR peak of 0.8 mm, 36 days after 4 April, one of the two greatest single-day DSR values ob-

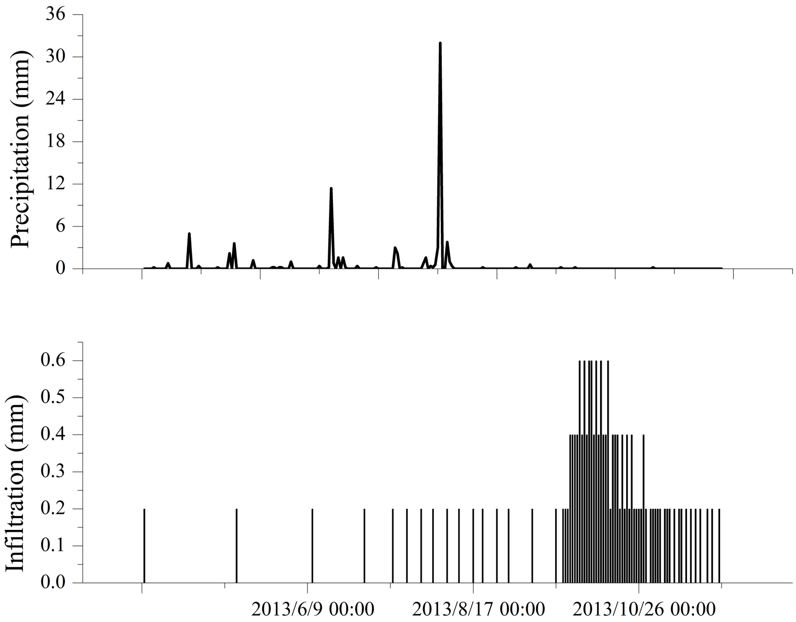

Figure 3. Precipitation and DSR patterns in 2013.

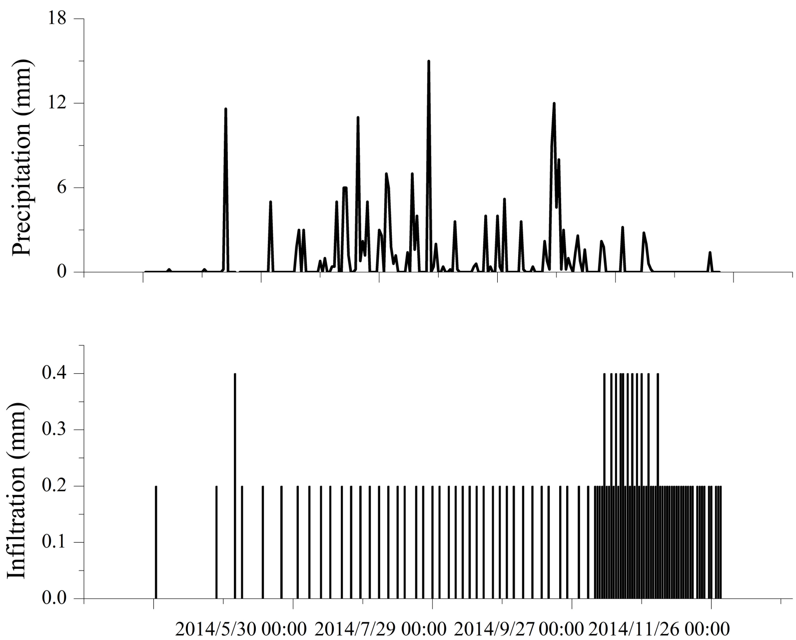

Figure 4. Precipitation and DSR patterns in 2014.

served in 2015, but no peak value of DSR response to the strong precipitation on 5 June. As explained before, summer stronger evaporation leads to relatively less DSR during the summer season compared with other seasons. The third greatest precipitation is 16.8 mm on 5 October, which leads to a peak value of 0.8 mm of DSR on 21 October, with a 16-day delay time. If comparing the precipitation events that occurred on 4 April (17.2 mm) and 5 October (16.8 mm), one can see that these two precipitation events are similar in strength (17.2 mm for 4 April and 16.8 mm for 5 October) but different in the DSR delay time (36 days for 4 April and 16 days for 5 October). When comparing two precipitation events which are similar in strength but different in the DSR delay time, temperature is the most likely factor responsible for such delay, so this leads to a conclusion that temperature influences the DSR rate. To investigate how the soil tem-

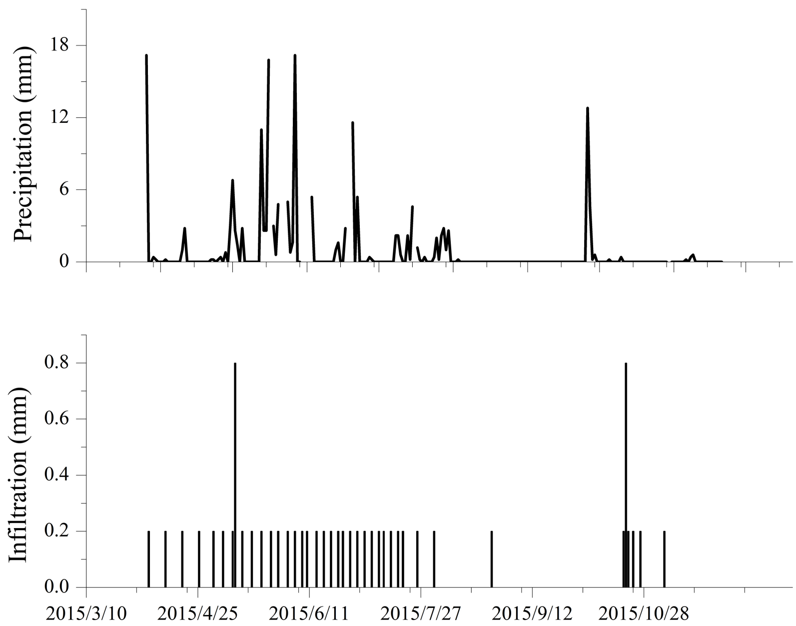

Figure 5. Precipitation and DSR patterns in 2015.

perature affects the DSR rate, further field experiments are needed in the future study.

3.2.2 Relationship between precipitation intensity and DSR

Based on observational data and analysis in Sect. 3.2.1, one can see that precipitation intensity, to some extent, influences DSR. For the sake of illustration, the precipitation intensity for bare sand land is roughly classified into light, moderate, and strong events with precipitation amounts less than 6 mm, between 6 and 10 mm, and greater than 10 mm, respectively. In general, light precipitation rarely can reach the soil zone deeper than 40 cm because of evaporation; thus, it makes almost no contribution to DSR (Zhang et al., 2016). Such a classification may be revised under different vegetation-covering conditions in different regions (Kosmas et al., 2000).

According to this classification, statistics of moderate to strong precipitation events and their percentage shares in the annual precipitation from 2013 to 2015 are shown in Table 2. In 2013, there are only two precipitation events with intensity greater than 6 mm. The total amount of these two precipitation events is 43.4 mm, which represents 52.29 % of the annual precipitation in 2013. In 2014, there are 11 precipitation events with intensity greater than 6 mm, much more frequent than in 2013 (2 events) and moderately more frequent than that of 2015 (8 events). The total moderate-to-strong precipitation in 2014 is 98.6 mm, representing 47.96 % of the annual precipitation in 2014. In 2015, there are eight precipitation events with intensity greater than 6 mm, accounting for 53.54 % of the annual precipitation in 2015.

Among these 3 years, 2015 has the largest percentage of moderate-to-strong precipitation over the annual precipitation. However, in this same year, one has seen the smallest ratio of annual DSR/precipitation ratio or annual recharge coefficient (see Table 1). This implies that the annual DSR does not seem to be positively correlated to the annual total precipitation. This finding has a few profound consequences. It basically states that assigning a constant annual recharge coefficient for a particular soil regardless of precipitation patterns is not a good practice, because annual DSR is not always proportional to the total annual precipitation. Instead, it appears to be more closely related to individual precipitation events stronger than 10 mm.

Table 3 lists the number of strong precipitation (with amount greater than 10 mm) and also the strongest precipitation amount for 2013, 2014, and 2015. In 2013, there are only two strong precipitation events, but the maximum single-day precipitation amount reaches 32 mm (3 August). The accumulative strong precipitation of 2013 is 43.4 mm, which is 52.28 % of the annual precipitation in 2013. In 2014, there are four strong precipitation events, and the maximum single-day precipitation amount is 15 mm. The accumulative strong precipitation of 2014 is 49.6 mm, which is 24.12 % of the annual precipitation in 2014. In 2015, there are six strong precipitation events, and the maximum single-day precipitation amount is 17.2 mm. The accumulative strong precipitation of 2015 is 86.6 mm, which represents 46.46 % of the annual precipitation in 2015. The annual DSR vs. annual precipitation ratios are 24.33, 10, and 4.94 % for 2013, 2014, and 2015, respectively.

Table 2. Percentage of valid precipitation in total precipitation amount.

Year	Number of precipitation > 6 mm (24 h cumulative)	Amount of precipitation > 6 mm (mm)	Valid precipitation/ annual precipitation (%)
2013	2	43.4	52.29
2014	11	98.6	47.96
2015	8	99.8	53.54

Table 3. Interannual statistics of strong precipitation and its percentage in total annual precipitation amount.

Year	Number of strong precipitation	Maximum precipitation event (mm)	Annual DSR (mm)	Annual DSR/annual precipitation (%)
2013	2	32	20.2	24.33
2014	4	15	20.6	10
2015	6	17.2	9.2	4.94

As shown in Table 3, the strongest single-day precipitation (32 mm in 2013) appears to affect DSR the most in 2013. For 2014 and 2015, as the strongest precipitation events in these 2 years are significantly smaller than those in 2013. Such a positive correlation is particularly strong for 2013 which has the largest maximum precipitation event of 32 mm, showing that the strong single-day precipitation affects DSR. This positive correlation is weaker for 2014 and 2015, which have moderate and somewhat similar maximum precipitation events (15 and 17.2 mm, respectively). As shown in Figs. 4 and 5, precipitation patterns in 2014 and 2015 are quite different despite the fact that the maximum precipitation events are similar to each other. The precipitation in 2014 is somewhat uniformly distributed from April to November, while the precipitation in 2015 is mostly concentrated from May to June. This observation suggests that DSRs for these 2 years are related to the precipitation pattern as well as the precipitation strength. However, precisely quantifying such a correlation between DSR and the precipitation pattern and precipitation strength requires further investigations.

In summary, one may conclude that annual DSR in arid and semi-arid regions mainly relies on strong precipitation events, but the determination of the threshold for strong precipitation events that directly contribute to DSR is still unclear and requires further investigation.

Under the condition of continuous precipitation, it may be difficult to discretize precipitation into individual events. The following example illustrates a procedure to deal with this situation. As shown in Fig. 6, there is a 13-day continuous precipitation process in 2013 from 27 July to 8 August, and the accumulative precipitation is 43.8 mm. The start of a continuous DSR distribution corresponding to this 13-day continuous precipitation event is observed 3 days after the end of this precipitation process, and the peak value of DSR occurs 46 days after the end of this precipitation process. The DSR distribution gradually recedes to zero around 78 days after the end of the precipitation process. The accumulative DSR amount over a 75-day period is 18.4 mm. The ratio of the 75-day cumulative DSR over the 13-day precipitation event is 42 %.

4 Discussion

This improved lysimeter is on the real-time dynamic monitoring of DSR, and it provides reliable evidence for an accurate evaluation of precipitation-related recharging capability of bare sand lands in arid and semi-arid regions. However, there are a number of issues that deserve further attention and require additional investigations in the future. The moisture evaporation, the soil absorption of moisture, and the water infiltration of post-evaporative redistribution are all very complex processes, especially in arid and semi-arid regions. It is sometimes difficult to clearly distinguish the amount of evaporation and DSR with conventional methods as outlined in the introduction. This study selects precipitation and infiltration data during the period from 1 April to 30 November, so the influence of freeze–thaw process during winter is avoided, and the experimental design and data analysis are simplified. For these reasons, the next steps should be a full-term monitoring, systematic study on DSR, as well as a study on the soil temperature and daily temperature influences on DSR.

Although this experiment does not address the issue of soil temperature effect on DSR in great detail, the relationship between DSR and soil temperature is evident. In general, a higher temperature means a stronger evaporation demand and thus an often smaller DSR.

Through the analysis of this study, one can see that the use of an annual recharge coefficient for the study area is not

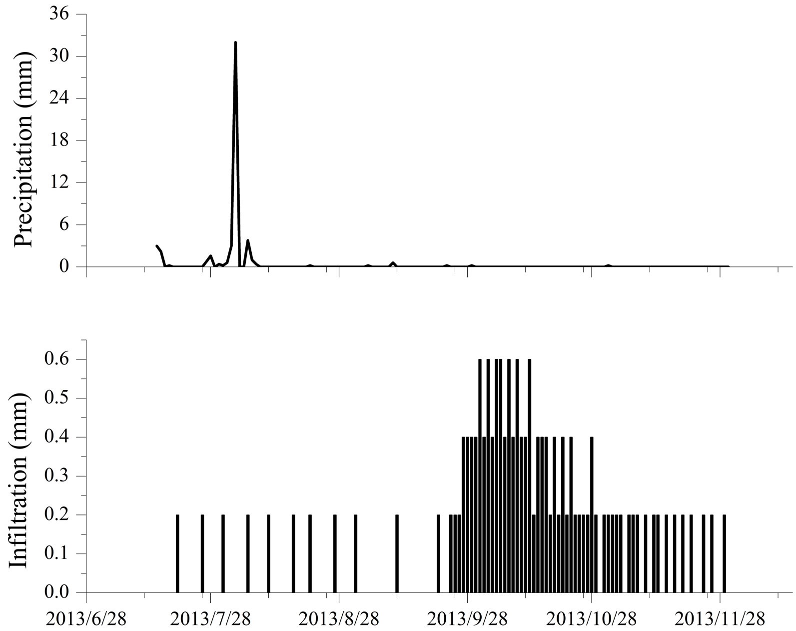

Figure 6. The single-day intensive precipitation contribution to DSR in 2013.

supported by the data collected from the new lysimeter, as the annual recharge is not positively correlated with the annual total precipitation. Instead, we find that the recharge is somewhat positively correlated with a few strong precipitation events (greater than 10 mm), and is closely correlated with the strongest precipitation event (considerably greater than 10 mm), as well as the precipitation patterns. It is probably reasonable to assign different weighting factors for different precipitation strengths to calculate DSR. However, the threshold to define a strong precipitation event that makes direct contribution to DSR is not precisely quantified, and this is a subject that should be investigated in more details in the future. The determination of weighting factors for different precipitation strengths is also a subject that requires further investigation.

This investigation is based on detailed analysis of precipitation and DSR data at the study site without involving modeling effort which certainly will be explored in the future as well. This study represents our first attempt of questioning the application of a recharge coefficient concept in arid and semi-arid regions.

5 Conclusions

This study uses a newly designed lysimeter to study three consecutive years (2013–2015) of DSR underneath bare sand land on the eastern margin of Mu Us Sandy Land in the Ordos Basin of China. The objective is to identify the characteristics of the DSR distribution and the factors affecting the DSR distribution. Specifically, we like to examine if the commonly used recharge coefficient concept can be applied for arid and semi-arid regions such as the eastern margin of Mu Us Sandy Land of China. The following conclusions can be drawn from this study:

- The annual recharge coefficient concept is generally inapplicable for estimating DSR in the study site.
- Precipitation pattern, including precipitation intensity and precipitation season, significantly influences DSR.
- The temperature influences the DSR/precipitation ratio, which is lower in summer than in other seasons, given the similar precipitation intensity.
- DSR is not correlated with the annual precipitation. Instead, it is correlated with the strong precipitation (greater than 10 mm) events at the site. However, quantitative determination of the thresholds for such strong precipitation events that make direct contributions to DSR is not entirely understood. Further investigation is needed on this subject.

Data availability. The data used in this study can be accessed by contacting the corresponding author directly.

Competing interests. The authors declare that they have no conflict of interest.

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