Journal of Arid Environments xxx (xxxx) xxx–xxx

Contents lists available at ScienceDirect

Journal of Arid Environments

journal homepage: www.elsevier.com/locate/jaridenv

Simulated dynamics of soil water and pore vapor in a semiarid sandy ecosystem

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ARTICLE INFO

Keywords:
Dry soil layer
Evaporation zone
Heat transfer
HYDRUS
Soil evaporation

ABSTRACT

Understanding dynamics of soil water content (SWC) and pore air relative humidity (RH_{pa}), as influenced by wetting-drying cycles, is crucial for sustaining fragile ecosystems of desert lands across the world. However, to date, such an understanding is still incomplete. The objective of this study was to examine such dynamics at a typical desert site within the Horqin Sandy Land, located in Mongolian Plateau of north China. The results indicated that vaporization primarily occurred at a depth of around 10 cm below the ground surface. The diurnal variations of the SWC and RH_{pa} in the top 10 cm soils were much larger than those in the soils at a deeper depth. For a non-rainy day, the SWC and RH_{pa} were mainly determined by the relative magnitude of atmospheric temperature over soil temperature, whereas, for a rainy day, the SWC and RH_{pa} were primarily controlled by the rainfall pattern and amount. The retardation role of the top dry soil layer, which is about 10 cm thick and exists most time at the study site, can effectively prevent the beneath moist soils from being further dried up, and thus is beneficial for sustaining the desert ecosystem.

1. Introduction

In arid/semiarid regions, where the hydrologic cycle is dominated by vertical water movement (i.e., soil water evaporation) as influenced by heat transport (Dong et al., 2003; Wang, 2015), soil water content (SWC) and pore air humidity (RH_{pa}) are two important indicators of available water for sustaining their fragile sparse vegetation ecosystems (Goss and Madliger, 2007; Duan et al., 2011, 2015). Hereinafter, SWC is defined as the ratio of soil water volume to bulk soil volume, while RH_{pa} is defined as the ratio of the partial pressure of water vapor to the total pressure of air in soil pores (Farouki, 1981). When transported into soils, heat will cause increase of soil temperature, vaporizing soil water into pore vapor, in reverse, when soil temperature is cooled down, pore vapor can be condensed back into soil water (Wang, 2015). The heat emission out of soils will not only lower soil temperature but also cause loss of pore vapor to the ambient atmosphere (i.e., soil evaporation). For an area of interest, when such a vaporization-condensation dynamic process becomes insufficient to meet the water demand of sparse vegetation, this area will likely be subject to desertification. As a major reason for land deterioration in the arid regions of the world (Kassas, 1995; Zucca et al., 2011), including the Horqin Sandy Land (HSL) of Eurasian Grassland, desertification is closely related to the increasing soil evaporation as a result of inappropriate land management practices (e.g., removal of native grasses and plantation of deep-root trees) (Zhao et al., 2010; Smits et al., 2012; Wang, 2015; Li et al., 2016a). In order to develop practical measures for solving desertification-related eco-environmental problems (e.g., dust storm and loss of grassland production), it is needed to have a good understanding of soil water and pore vapor dynamics.

As noted by previous studies (e.g., Wang et al., 2006; Liu et al., 2013), such an understanding can be challenging because of the two-phase (i.e., liquid-vapor) fluid condition and the vapor flow resistance (VFR) effect of a thin (5–10 cm) top dry soil layer (DSL). The semiarid regions are usually dominated by bare sandy soils with such a DSL most of the time, within which soil moisture is dominantly in vapor phase with a very large capillary suction head (> 15,000 cm water height) (Goss and Madliger, 2007). In the past two decades, various laboratory and field studies (e.g., Daamen and Simmonds, 1996; Yamanaka and Yonetani, 1999; Saravanapavan and Salvucci, 2000; Wang et al., 2006; Goss and Madliger, 2007; Novak, 2010; Sakai et al., 2011; Liu et al., 2013) have been conducted to measure soil evaporation in water-limited environment. However, most of the studies were for short-term periods (i.e., from several days to a few months) and rarely measured vertical profiles of SWC and soil temperature, making it hard to determine dynamics of soil water and pore vapor as influenced by wetting-drying cycles. On the other hand, a few studies (e.g., Yamanaka

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https://doi.org/10.1016/j.jaridenv.2017.11.004
Received 12 February 2017; Received in revised form 12 October 2017; Accepted 10 November 2017
0140-1963/ Published by Elsevier Ltd.

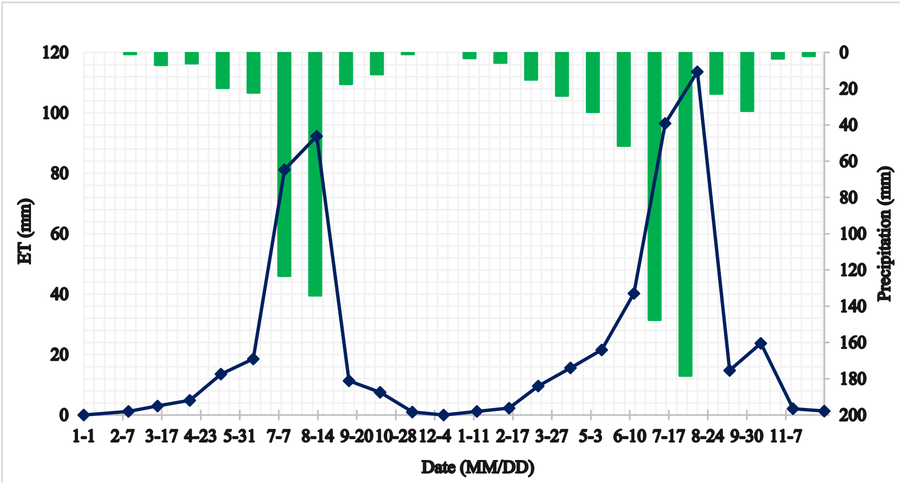

Fig. 1. Monthly precipitation and actual evapotranspiration (AET) at the study site. Each bar/dot represents a month of the plotting date.

Fig. 2. Pictures of the: (a) plane view (with the sensor-based weather station tower); and (b) soil vertical profile (with three-needle heat pulse or TNHP sensors), of the study site.

Table 1
Precipitation and rainy days in each month of the two study years.

Year	Precipitation (mm) (Number of Rainy Days)												
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Annual
2013	0 (0)	1.1 (2)	7.1 (3)	6.2 (3)	19.7 (5)	22.3 (5)	123.4 (9)	134.2 (11)	17.6 (3)	12.2 (4)	1.1 (2)	0 (0)	344.9 (47)
2014	3.3 (4)	5.8 (5)	15.1 (5)	24.0 (6)	33.7 (6)	51.6 (3)	147.7 (4)	178.5 (7)	22.9 (5)	32.4 (4)	3.5 (2)	2.1 (1)	520.6 (52)

et al., 1998; Sakai et al., 2011) used mathematical models to reproduce the vaporization-condensation-movement dynamics observed in laboratory soil columns, without determining the dynamics of soil water and pore vapor within different soil layers. Wang (2015), the correspondence author of this paper, conducted an overview of existing studies attempting to measure and/or model soil water evaporation in arid/semiarid environment, concluding that to fill the knowledge gap of soil water and pore vapor dynamics, field measurements need to be extrapolated using mathematical models.

The liquid-heat-vapor processes in a soil profile can be described by the Philip and de Vries (1957) theory (hereinafter designated as the PdV for description purposes), which in turn can be modeled using a set of partial differential equations. In reality, these processes interactively determine the soil evaporation phase at any time of interest. Philip (1958) classifies the desiccation of a soil profile into three phases (i.e., Phase I to III). Phase I occurs when the soil is sufficiently moist and thus has a soil water evaporation (E) indistinguishable from that with saturated surface (E_s), whereas, Phase III occurs when the soil surface layers are very dry and E is sensitive to, and may be negatively correlated with, heat flux into soil. Phase II occurs when the soil has an intermediate moisture content and thus E is independent of E_s and depends on soil moisture distribution only. During Phase I, liquid water is supplied from the lower layers and vaporized at the soil surface at rate E_s, and then the vapor is transferred into the ambient atmosphere. On the other hand, during Phase II and III, vaporization takes place not at the surface but within the soil mass (e.g., the evaporation zones or EZs), and the vapor diffuses upward through a top DSL to the surface and then into the atmosphere. Such classification of evaporation phases is verified by field observations presented by Goss and Madliger (2007).

Based on a few previous studies (e.g., Yamanaka and Yonetani, 1999; Goss and Madliger, 2007), during Phase I and II, E tends to be maximal when total heat flux into the soil is greatest (e.g., at noon), whereas, during Phase III, E tends to be maximal when total heat flux out of the soil is greatest (e.g., at midnight), and vice versa. This is because for Phase I and II, liquid water is vaporized at/near soil surface, which is positively related to the solar radiation (i.e., heat) input into