Patterns of surface energy exchange and evapotranspiration in relation to water availability in an oasis-desert ecotone

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Abstract

A knowledge of the exchanges of energy and water over the terrestrial surface is the first step to understand the ecohydrological mechanisms, particularly in water-limited ecosystems in the dryland environments. However, patterns of energy exchange and evapotranspiration (ET) are not well understood in the oasis-desert ecotone, which plays an important role in protecting oasis against the threat of desertification in northwestern China's arid regions. Here the continuous measurements of surface energy fluxes were made using eddy covariance in conjunction with auxiliary measurements for two years (2014-2015) at a shrubland within an oasis-desert ecotone in the arid regions, northwestern China. Statistical analysis on 30-min time scale indicates that about 50% of daytime net radiation (R_n) over the shrubland is dissipated as H on average, which peaks in spring; one third R_n is consumed by soil heat flux (G). Only 9% of R_n was consumed for latent heat flux (λE), which peaks in summer (21% in 2014 and 16% in 2015), corresponding to the season with highest rainfall among all seasons. Daily mean ET is about 1 $mm \cdot d^{-1}$ during growing season of the shrub species. The rapid and transient increase in ET occurs following a rainfall event. A switch in surface soil moisture from 0.04 to 0.11 $m^3 \cdot m^{-3}$ causes an increase in R_n by about 11% and λE by 151% at the shrubland, respectively. Accumulated annual ET were 195 and 181 mm in 2014 and 2015, respectively, exceeding the corresponding P by about 87 and 77 mm, indicating that groundwater may be another important source of water for ET over the shrubland aside from P. These results provide valuable insight into the mechanisms of sustaining energy and water balance at the ecotone, and then produce some management guidelines for allocating water resources and protecting vegetation.

Keywords: energy exchange, evapotranspiration, water balance, phreatophyte shrubs, precipitation, groundwater.

1. INTRODUCTION

The exchange of energy and water vapor between the terrestrial surface and the atmosphere drives the Earth's climate from local to global scales (Dickinson, 1991; Betts & Ball, 1996; Sellers et al., 1997; Trenberth, Fasullo, & Kiehl, 2009; Krishnan, Meyers, Scott, Kennedy, & Heuer, 2012), and provides interfaces between its meteorological, hydrological and ecological components, involving all three of the carbon, water and energy cycles (Raupach, 1998; Katul, Oren, Manzoni, Higgins, & Parlange, 2012; Chen et al., 2013; Bonan, 2016). Therefore, a knowledge of surface energy exchange and evapotranspiration over land surface is fundamental to the understanding of the coupling existing between energy, ecosystem dynamics and the water cycle, in particular in arid and semiarid environments, where water is an important limiting resource not only for its scarcity but also for its intermittency and unpredictable presence (Warner, 2004; Rodríguez-Iturbe & Porporato, 2005; Williams & Albertson, 2005; D'Odorico and Porporato, 2006; Newman et al., 2006; Nicholson, 2011). Globally, drylands cover about 45% of Earth's land surface (Schimel, 2010) and therefore are a significant components of Earth's climate system because of their large surface albedo and high sensible heat flux (Kustas et al., 1991; Warner, 2004; Reynods et al., 2007; Nicholson, 2011).

Over the past decade, the eddy covariance (EC) technique has become the standard tool to study the exchange of energy, water vapor, and carbon dioxide between the Earth's surface and atmosphere (Baldocchi, 2001; Aubinet et al., 2012). EC has been widely used to study these fluxes in a number of ecosystems as part of FluxNet and its regional networks (e.g., AmeriFlux, AsiaFlux, CCPand FluxNet-Canada, Carbon-Europe, ChinaFLUX, OzFlux, Japan-Flux etc.), including the water-limited ecosystems in drylands (Bolle et al., 1993; Kustas & Goodrich, 1994; Goutourbe et al., 1997; Lloyd et al., 1997; Goodrich et al., 2000; Havstad, Kustas, Rango, Ritchie, & Schmugge, 2000; Scott et al., 2006; Kurc & Small, 2007; Renard, Nichols, Woolhiser, & Osborn, 2008; Evett et al., 2012; Krishnan et al., 2012; Issac et al., 2017).

In China, the drylands cover about 60% of the territory (Huang, Yu, Guan, Wang, & Guo, 2015), largely in the arid northwest, where a few oases of various dimensions are sustained by water originating from mountains in the upper-stream areas (Shen, Wang, Wu, & Han, 2001; Kang, Lu, & Xu, 2007). Drylands are home to more than 90–95% of the population in the areas, although representing only a very small portion (4–5%) of the total area of the region (Shen et al., 2001; Jia et al., 2004). There is an ecotone between these relative small-scale oases and large-scale desert, mainly consisting of shrub species with an indispensable ecological function of windbreak and shelterbelt protecting these oases from wind erosion and desertification-threaten (Zheng & Yin, 2010; Cao et al., 2011). It is noteworthy that the oasis-desert ecotone is substantially more fragile and more sensitive to climate change and anthropogenic disturbances than other ecotones due to water limitation (Huang et al., 2015; Liu et al., 2017; Bryan et al., 2018; Zen et al., 2020). Therefore, measurements of the energy and evapotranspiration at the land-atmosphere interface over the oasis-desert ecotone are crucial for determining the energy and water budgets, and for understanding the connections between the meteorological, hydrological, and ecological processes in the ecotone. However, there is a lack of field observations specifically designed to study the surface fluxes of the ecotone between oasis and desert, and the seasonal variations of energy exchange and evapotranspiration (ET) are not well understood in the ecotone.

In this paper, we present the surface energy partitioning and *ET* observations made over an ecotone between oasis and desert in the arid regions of northwest China during a two-year period over 2014-2015, using the eddy covariance technique in combination with standard meteorological and other auxiliary measurements. The main objectives of the paper are: (1) to investigate the diurnal and seasonal dynamics of energy fluxes and ET rates, (2) to examine the sensitivity of surface radiation and energy budget to changes in soil moisture from rainfall, and (3) to understand the hydrological mechanisms of sustaining surface water balance. The information and findings from this study will not only increase our collective knowledge about the mechanisms that control energy and water fluxes over the ecotone, but also improve our understanding of the ecohydrological processes occurring in this specific ecotone ecosystem under arid conditions. Consequently, insights can be offered for devising proper management strategies for sustainable utilization of land-water resources in the drylands and in other similar arid regions.

2. MATERIALS AND METHODS

2.1. Site description

The study site (39° 22′ 07″ N; 100° 08′ 48″ E; 1386 m a.s.l) is located in the ecotone between the Linze oasis and the Badain-Jaran Desert in the middle-part of the Hexi Corridor, northwest China. The detailed information about the site was presented in Ji, Zhao, Kang, Jin, & Xu (2016), and will be outlined briefly here. The study area has a typical continental, temperate arid climate. The meteorological data at the Linze Inland River Basin Research Station (LIRBRS), about 5 km apart from this study site, show that, on the average over 2005-2014, the study area has the daily mean temperature of 8.9 °C ranging from −10.4 °C in January to 24.0 °C in July, minimum temperature of −26.2 °C in January, and maximum of 38.6 °C in July. The average annual precipitation (P) is 123.5 mm, with nearly 80 % of the annual total falling between June and September. Soil at the site is loamy sand (sand 73.5%, silt 22.5%, and clay 1.4%) (USDA texture class) with a bulk density of 1.52 $g \cdot cm^{-3}$, very low organic materials and nutrients, a condition with water and nutrient deficits to plants.

The study site is representatively a land cover of shrubland with a relatively flat topography, which is mainly covered by phreatophyte shrubs such as *Haloxylon ammodendron*, *Nitraria tangutorum/sphaerocarpa*, and *Calligonum mongolicum*, interspersed with areas of completely bare crusted soil. The total vegetation cover at this study site was about 15%, and the peak leaf area index was estimated as 0.28 $m^2 \cdot m^{-2}$ during the growing season. According to our previous investigation on vegetation in the study area, the maximum depth the tap roots of *Haloxylon ammodendron*, *Nitraria tangutorum* and *Calligonum mongolicum* can reach to be 10.0

m, 3.6 m, and 4.2 m under the soil surface, respectively, with the average height of 1.65 m, 0.92 m, and 0.46 m during the measurement period. The groundwater table are generally at a depth of 4.4 m and fluctuates by less than ± 0.2 m below the ground surface during the entire two years.

2.2. Eddy covariance measurements and data processing

The EC instruments at the study site were mounted on a tower (at 8 m height above the ground surface) and consisted of a three-axis sonic anemometer (Model WindMaster Pro, Gill Solent Instruments, Lymington, Hampshire, United Kingdom) and an open-path infrared gas (CO_2/H_2O) analyzer (IRGA) (Model LI-7500A, LI-COR Inc., Lincoln, Nebraska, USA). The sonic anemometer was oriented into the predominant wind direction (from southeast). EC instruments were sampled at 10 Hz using a LI-7550 Analyzer Interface Unite controlled data logging system and were stored in a flash memory storage card. We use the EddyPro 6.2.1 software (LI-COR, Lincoln, Nebraska, USA) to calculating half-hourly mean fluxes from the raw EC data. The EddyPro software also applied the following corrections: low and high pass filtering (Moncrieff, Clement, Finnigan, & Meyers, 2004); compensation for air density fluctuations (Webb, Pearman, & Leuning, 1980), in which fluxes of sensible heat were corrected for the additional instrument-related sensible heat flux during the winter periods, due to instrument surface heating/cooling of the LI-7500 open-path gas analyzer (Burba, McDermitt, Grelle, Anderson, & Xu, 2008); sonic temperature correction for humidity (Van Dijk et al., 2004); time lag compensation; sonic anemometer tilt correction with double rotation (Wilczak, Oncley, & Stage, 2001); block averaging; angle-of-attack correction for wind components (Nakai & Shimoyama, 2012); and statistics tests (Vickers & Mahrt, 1997). Using the analytical footprint model of Kljun, Calanca, Rotach, & Schmid (2004) with typical daytime conditions, the peak contribution location to flux was typically less than 100 m with mean value of 25 m during the study period, and the along-wind extent of the 90% cross-wind integrated flux footprint was typically within the range of 1000 m distance from the tower with mean value of 285 m. Given the typical horizontal length scale of the shrubs and bare soil at this site (of the order of a few meters), the complete pattern was sampled. All the flux data were quality-flagged to rate stationarity and the development of the turbulent flow field (Foken et al., 2004). The quality criteria (QC) ranges from 1 (best) to 9 (poorest); the flux data were rejected if QC > 5.

We needed to gap-fill about 15% of the half-hourly turbulent fluxes data, which were either rejected by the quality control tests or missed due to unfavorable ambient conditions not satisfying the EC method. Several strategies were used to compensate for missing data. Small gaps (i.e., < 2 h) in the data were filled by linear interpolation. Longer gaps in turbulent fluxes data were filled by calculating the average monthly values for each half-hour of measurement, in a manner similar to the "Mean Diurnal Variation" method described by Falge et al., (2001). Additional data gaps in daytime H and λE were filled using the relationship between incoming solar radiation and H and λE using a moving window of 14 days and separated into morning or afternoon periods (Falge et al., 2001; Krishnan et al., 2012). Given the amount of gaps filled was small, the uncertainties in annual sensible and latent heat fluxes (or ET) associate with the gap-filling procedure were relatively low.

2.3 Environmental measurements

An automatic weather station was mounted on the same tower as the eddy flux tower in the shrubland. A 4-component radiometer (CNR4, Kipp & Zonen, Delft, The Netherlands) was located 6 m above the soil surface to measure the above-canopy radiation components and net radiation. Soil heat flux was measured with four soil heat flux plates (HFP01, Hukseflux Thermal Sensors, Delft, The Netherlands) installed 0.05 m below ground level: two beneath the canopy of shrubs, two in inter-canopy. The soil heat flux at the surface (G) was calculated by adding the soil heat flux measured by heat flux plates to the energy storage above the plates following Campbell & Norman (1998). Photosynthetically active radiation flux (PAR) was measured by a LI-190SB quantum sensor (LI-COR Inc., Lincoln, Nebraska, USA) at 6 m above the soil surface. Surface temperature was measured with a surface pyrometer (SI-111, Apogee Instruments Inc., Utah, US), installed at 6 m height above the ground. Measurements of air temperature and relative humidity were made using temperature/humidity probes (HMP155A, Vaisala, Vantaa, Finland) at 4 and 6 m above the ground. Wind speed and direction was measured by sonic anemometers (1405-PK-052, Gill Instruments Lrd., Lymington, UK) at the same height as measurements of air temperature and relative humidity. Barometer probe (CS100, Setra System Inc., Massachusetts, US) were applied to measure air pressure. Measurements of soil temperature were made with thermistor probes (109-L, Campbell Scientific Inc., Logan, Utah) at six depths (5, 10, 20, 40, 60 and 100 cm). Measurements of the rate of change of soil temperature above the heat flux plates (i.e., average soil temperature for 0–5 cm soil depth was determined by averaging the values at surface and 5 cm located) in combination with the specific volumetric heat capacity of heat of the soil allowed calculation of the soil heat flux at the surface by determining the changes in heat storage of the 0–5 cm soil layer. *P* was recorded in 15 min intervals by a tipping bucket rain gauge (TE525MM, Texas Electronics Inc., Dallas, TX) at 6 m above the ground. Because we were not equipped to measure *P* in the form of winter snowfall, *P* values from the LIRBRS weather station near the study site were used for the period extending from 1 November October to 28 February.

We monitored volumetric soil water content (θ) using commercial soil moisture probes (CS616, Campbell Scientific, Logan, UT, USA) placed at depths of 5, 10, 20, 40, 60, and 100 cm around the soil heat flux plates. Furthermore, the top soil moistures of four layers (0-5, 5-10, 10-20, and 20-40 cm) were measured around the soil heat flux plates by gravimetric method from 0 to 40 cm every day at 8:00 h. The gravimetric moisture content was converted into volumetric measurement using the bulk density of the soil. With the exception of the rain gauge, all probes and sensors on two automatic weather stations were sampled at intervals of 10 s and recorded as half-hourly averages using a data-logger (CR1000 XT, Campbell Scientific Inc., Logan, Utah, USA). Additionally, vegetation dynamics related to shrubs phonological regimes were measured on a weekly basis, including leaf area index (LAI) measurements using a plant canopy analyzer (LI-2200, LI-COR Inc., Lincoln, Nebraska, US), following the method proposed by Chen (1996), and shrub height measurements within four sampling plots at the study site.

3. RESULTS

3.1 Climate characteristics

The average monthly values of incoming solar radiation, air temperature, vapor pressure deficit (VPD), *P*, wind speed, and top soil moisture (0−40 cm) during the study period (2014-2015) are shown in Figure 1, along with the 11-year (2005-2015) monthly means observed at the weather station of the LIRBRS nearby the study site. The solar radiation, air temperature, vapor pressure deficit, and wind speed in 2014 and 2015 were similar and consistent with the long-term magnitudes and seasonal variations, indicating that both years were presentative of the typical annual weather at the study site. It should be noted that the monthly mean wind speed in both 2014 and 2015 exceeded universally the corresponding long-term averages, since the study site locates on a further fringe of the Linze oasis and has a higher wind speed than that at the weather station of the LIRBRS.

In contrast, the annual *P* in 2014 and 2015 were 108 and 104 mm, and were below the long-term average (i.e., 124 mm), even though the monthly *P* was both above or below the long-term means in some months. However, the *P* pattern during the two-year period closely mirrored the long-term trends (Figure 1d): most of the *P* fell during the months of June−September (c.a., 84% for 2014 and 73% for 2015) mainly due to the seasonal variations in the Southwest China monsoon. As expected, the top soil moisture (0−40 cm) at the study site was relatively low throughout the two-year period because of infrequent and small rainfall. The top soil moisture dynamics resulting from *P* history showed a similar seasonal trend, corresponding to the *P*. The top soil layer was slightly wetter in 2014 (i.e., 0.09 $m^3 \cdot m^{-3}$) than that in 2015 (i.e., 0.08 $m^3 \cdot m^{-3}$) due to the slightly more rainfalls in 2014 than those in 2015. The long-term values of soil moisture are not shown in Figure 1e because of the absence of the measurements at the LIRBRS.

3.2 Energy balance closure

The energy balance closure provides a useful test to check the plausibility of eddy covariance flux data sets obtained at different sites (Aubinet et al., 2000; Wilson et al., 2002; Oncley et al., 2007; Foken, 2008). In this approach, the sum of turbulent heat fluxes (i.e., sensible (H) and latent (λE) heat flux) measured from eddy covariance is compared with the available energy flux (the net radiation (R_n) minus the energy stored in the observed ecosystem, including soil (G), air (S) and biomass) measured from the meteorological sensors as the independent variable. In this study, the energy stored in vegetation component is assumed to be trivial due to little vegetation biomass. Accordingly, the term S in this sparse shrubland may only represent the change in latent heat associated with changes in the humidity of the air, and the change in sensible heat associated with temperature changes of the air between the soil surface and EC measurement height (calculated by the EddyPro 6.2.1 software).

At the 30 min time-scale, the global closure rate (derived from the slope of the linear fit of ($H + \lambda E$) vs. ($R_n -$ G − S)) is 0.94 for 2014 (Figure 2a) and 0.86 for 2015 (Figure 2b). To assess the potential for short-term variation in energy storage, energy balance closure was also tested using daily averages (summed over full 24-h periods) of $H + \lambda E$ and $R_n - G - S$ for each of the two years (Figures 2c and 2d). At this time scale, the slope of the regression statistics of $H + \lambda E$ and $R_n -$ G – S was also less than one. As expected, the relative agreement between turbulent heat flux and available energy is slightly better at the daily time step than that at the half-hourly time step, since the effect of G on energy balance closure was minimized or eliminated to a large extent for daily values (Figure 2d).

The ratio of total turbulent heat fluxes ($H + \lambda E$) to total available energy ($R_n - G - S$) computed on a 30-min basis was similar for 2014 (c.a., 0.73) and 2015 (0.74), as well as on a daily basis (c.a., 0.77 for 2014, and 0.78 for 2015), which is comparable with those reported for a sparse shrubland (Lloyd et al., 1997; Wilson et al. 2002; Heusinkveld, Jacobs, Holtslag, & Berkowicz, 2004; Kurc & Small, 2007), indicating that the typical energy partitioning among all energy fluxes at this site is acceptable in most cases.

3.3 Diurnal patterns of the surface energy budget

Figure 3 depicts the mean diurnal cycle of the components of the surface energy budget over the shrubland averaged over spring (March−May), summer (June−August), autumn (September−November), and winter (December−February) during the two-year study period. As shown in Figure 3, the mean seasonal surface energy fluxes show the quite similar diurnal patterns for two years, varying over the course of day in response to the diurnal cycle of solar radiation, and typically being strongest at noontime (c.a., around 12:00) and decreasing late in the afternoon when solar radiation diminishes. The nocturnal R_n and G losses over the shrubland are large (e.g., the mean lowest values in nighttime ranged from −55 to −76 W·m^{-2} for R_n and from −50 to −89 W·m^{-2} for G, respectively), like other desert landscapes (Warner, 2004). The peak values of R_n for a complete annual cycle occur in summer months, while seasonal maximum values of G are largest in spring and smallest in winter (Figure 3). The storage term was usually quite small, on the order of several Watts per square meters as a result of dry atmosphere. The peak values of H show a seasonal variation similar to G, since both components are strongly linked to surface temperature. With respect to λE, its peak values are largest in summer (91 in 2014 and 82 W·m^{-2} in 2015) (Figures 3b and 3f), and drop to its minimum in winter (c.a., 10 in 2014 and 13 W·m^{-2} in 2015) (Figures 3d and 3h), corresponding to lowest soil moisture, temperatures, solar radiation (Figure 1), and transpiration of the dominant shrub species (Ji et al., 2016).

The surface energy budget at midday (between 10 AM and 3 PM) and daytime (R_n > 0) are summarized in Table 1 by season for two years, as well as Bowen ratio ($H/\lambda E$), evaporative fraction ($\lambda E/(H+\lambda E)$), and the ratios of components to R_n. It is noticeable that H is always the dominant component of the surface energy balance throughout a year, followed by soil heat flux that is a sizable portion of the surface energy balance due to much sparse plant canopies in the shrubland, and latent heat flux is always small resulting from the lack of P in this arid climatic conditions

(Figures 1 and 3, Table 1). The dry environment of the shrubland is seen in a high Bowen ratio ($H/\lambda E$) and low evaporative fraction (defined as $\lambda E/(H+\lambda E)$), especially in spring and winter, both indicative of dry soil (Table 1). In summer, λE reaches up to its maximum with the largest flux ratio $\lambda E/R_n$ throughout a whole year among all seasons, resulting from the frequent and increasing rainfall as well as increasing shrub transpiration (Ji et al., 2016). For this period, the Bowen ratio is smallest, or the evaporative fraction is correspondingly largest, and the contributions of H and G to the surface energy balance slightly decline (Table 1). Annually, the surface energy balance is characterized by high H and weak λE, with the high Bowen ratios (i.e., averaged daytime value of 9.23 for 2014 and 6.89 for 2015) and low evaporative fraction (about 0.16 for two years) when compared with the irrigated cropland within the Linze oasis (e.g., Bowen ratio and evaporative fraction were about 0.24 and 0.80, respectively in summer) (Ji et al., 2011), and the contribution of H and G is approximately 50% and 33%, respectively.

3.4 Seasonal variation in surface energy fluxes

To better visualize the seasonal variations in the surface energy fluxes, the monthly averages of daily accumulated energy budget components throughout the year over the shrubland are illustrated in Figure 4a for 2014 and Figure 4b for 2015 (storage term S is ignored here, because of generally being much small in magnitude on a daily basis). As shown, the time series of surface energy fluxes over the shrubland are very similar, having a pronounced seasonal cycle, with a maximum in summer and a minimum in winter. R_n, λE, and G reach up to maximum in July (Figures 4a and 4b), while the peak values of H occurred in May for both years, corresponding to higher solar radiation and lower soil moisture (Figures 1a and 1f). The lowest values of R_n, G, and H occur in December. Whereas the lowest values of λE appear in January in 2014 and in March in 2015. The former is due to the low soil moisture and solar radiation in 2014 (Figures 1a and 1f), and the latter is due to low soil moisture in 2015 (Figure 1f), which greatly restricted latent heat flux.

The contributions of H, λE, and G to surface energy balance vary largely with season during the two-year study period, and are characterized by high *H* and weak λE and *G* on a basis of monthly average, with *H* being the dominant flux throughout a whole year (Figure 4). The largest values of $\lambda E/R_n$ on a monthly basis are about 0.31 in August for 2014 and 0.33 in September for 2015, with the Bowen radio of 1.53 and 1.67 and evaporative fraction of 0.40 and 0.37, respectively. In both months, the monthly accumulated *P* was maximum (Figure 1d). In general, statistical analysis indicates that approximately mean 54% and 55% of the R_n is transferred into *H* in 2014 and 2015, respectively, during the shrub species growing season when rainfalls are frequent and large; about 23% of the net radiation is transferred into λE in both years, when the Bowen ratio of about 3.0 and evaporative fraction of about 0.3, respectively. By comparison, during the off-growing season, more than 70% of R_n is dissipated as *H* (c.a., 75% in 2014 and 73% in 2015), corresponding to dry season (October−April), and only less than 10% of R_n was used for evapotranspiration (c.a., 7% and 9% for 2014 and 2015) due to both the dry soil and leaf senescence for shrub species. During this period, the Bowen ratio is high (c.a., 21 and 12 for 2014 and 2015) and evaporative fraction is low (c.a., 0.07 and 0.11 for 2014 and 2015).

3.5 Seasonal variation in evapotranspiration

The time-series of daily *ET* from the shrubland in 2014 (Figure 5a) and 2015 (Figure 5b) are very similar, and correspond closely with seasonal variations in the available energy and surface moisture from rainfall, as well as leaf areas index (LAI), among which water availability is perhaps a more important factor than other factors in determining the *ET* over the shrubland in the water-limited environment. The rapid and transient increase in daily *ET* usually followed most *P* events at the shrubland, while *ET* decrease rapidly every day until the next *P* event occurs (Figures 5a and 5b). Maximum *ET* is observed immediately following *P* events (e.g., about 3.63 $mm \cdot d^{-1}$ on DOY 170 of 2014, and 2.92 $mm \cdot d^{-1}$ on DOY 149 of 2015 following an intense and large rainfall events of 20.6 mm between DOY 162 and 169 in 2014 and 8.4 mm between DOY 147 and 148 in 2015, respectively). Seasonally, the total *ET* between June and September contributed up to roughly 80% and 71% of the annual *ET* for

2014 and 2015, respectively, coincident with the summer monsoon season when the accumulated *P* contributed 85% in 2014 and 70% in 2015 to the annual *P*. The cumulative ET during the winter season were only about 5.39 mm in 2014 and 8.37 mm in 2015, respectively, corresponding to driest months with the absence of leaves and lowest R_n. Given these seasonal variations in *ET*, the magnitude and temporal dynamics of *ET* in relation to *P* events reflects the strong link between *ET* and *P* in the shrubland, similar to that in most arid regions (Warner, 2004; Nicholson, 2011).

The total annual ET over the shrubland in 2014 (c.a., 195 mm) was slightly more than that in 2015 (c.a., 181 mm), probably owning to more *P* (Figures 1d and 5b) and lower LAI during the most of growing season in 2015 (Figure 5c). Annually, it is worth of noting that the total *ET* exceeded the corresponding total *P* by 87 mm or 181% in 2014, and by 77 mm or 172% in 2015, respectively. This imbalance is likely due to the fact that these three dominant shrub species of the shrubland used largely deep soil water recharged by capillary rise, or may directly access shallow groundwater (Ji et al., 2016; Xu, Ji, Jin, & Zhang, 2016), especially for *Haloxylon ammodendron* (Zhou, Zhao, & Zhang, 2017). More details on water balance of the shrubland are discussed in Section 4.3.

4. DISCUSSION

4.1 The effects of soil moisture on surface radiation budget

Soil moisture can have a significant impact on surface radiation budget by altering the balance of longwave and shortwave radiation at the surface (Eltahir, 1998; Jones & Brunsell, 2008; Nicholson, 2011). The influence of soil moisture on surface radiation is likely greater in arid environments than in humid environments because the extreme contrast between wet and dry conditions (the occurrence of availability of surface soil moisture from rainfall in distinct pulses) (Malek, 2003; Small & Kurc, 2003; Warner, 2004; Nicholson, 2011).

Therefore, in order to investigate the effects of surface soil moisture (within the top 10 cm) in the shrubland on surface radiation budget quantitatively, two contrast sets data for soil moisture regime caused by five relatively larger rainfall events (between 7 and 15 mm) in summer (June−August) were selected: (a) dry soil moisture conditions (i.e., a clear-sky day prior to individual rainfall events); and (b) wet soil moisture conditions (i.e., a clear-sky day following individual rainfall events). By choosing this period with similar vegetation and environmental regimes but different soil moisture to minimize the influences of solar zenith angle, plant physiological behavior and other environmental factors on surface radiation budget. An equal number of observations (i.e., daytime (8:00 am to 18:00) data on 5 selected days included) that fall within each of two composites (that is, dry and wet soil moisture conditions) of similar soil moisture conditions are averaged to describe the typical surface radiation budget for those conditions.

Figure 6 illustrates the difference in daily average environmental variables (Figures 6a−6f) and surface radiation components (Figures 6g−6l) between dry and wet soil moisture conditions. It is interesting that the daily average downward shortwave radiation (S_d) increased by about 5% (Figure 6g) under the wet soil moisture conditions (average surface soil water content within the top 10 cm θ_s was 0.11 $m^3 \cdot m^{-3}$), comparing with the dry soil moisture conditions (θ_s was 0.04 $m^3 \cdot m^{-3}$), while the observed differences in upward shortwave radiation (S_u) showed a very small reduction (< 5 $W \cdot m^{-2}$). Consequently, surface albedo (0.147) under wet conditions decreased by 9% when compared with the dry conditions which leads to an increasing of 8% in net shortwave radiation absorbed by surface (526 to 560 $W \cdot m^{-2}$). The lower surface albedo is mainly attributed to the wet bare soil inter-canopy under wet conditions, which have a darker color and a lower reflectivity than under dry conditions.

For the surface longwave radiation components, the lower upward longwave radiation (L_u) is predictable when the surface is wet (Figure 6j) due to the reductions in surface temperature (T_s, Figure 6b) caused by evaporative cooling (i.e., ET is higher under the wet conditions, as discussed in the following section). Whereas it was observed that the downward longwave radiation (L_d) values also decreased by about 6% (371 to 348 $W{\cdot}m^{-2}$, Figure 6i) under wet conditions in comparison with dry conditions. Therefore, L_u loss (491 to 458 $W{\cdot}m^{-2}$) under wet conditions is partially diminished by the decreased L_d, resulting in slight difference in the net longwave radiation between wet (−110 $W{\cdot}m^{-2}$) and dry conditions (−120 $W{\cdot}m^{-2}$).

As a result, the wet soil moisture conditions would tent to enhance R_n over the shrubland by approximately 11% (Figure 6l) through increase in net shortwave radiation (34 $W{\cdot}m^{-2}$) and in net longwave radiation (10 $W{\cdot}m^{-2}$), which contributed about 85% and 15% to the difference in surface net radiation between wet and dry soil moisture, respectively. These results demonstrate that soil moisture regime does have a significant on surface radiation budget by lowering of surface albedo and cooling of surface temperature (Eltahir, 1998; Malek, 2003; Warner, 2004; Jones & Brunsell, 2008; Nicholson, 2011). It is interesting to note that, however, S_d and L_d also play a crucial role on R_n with a positive (increase of 31 $W{\cdot}m^{-2}$) and a negative (decrease of 23 $W{\cdot}m^{-2}$) contribution to R_n for the shift from dry to wet conditions, respectively. This negative contribution of L_d offsets dramatically the positive contribution from reduction of L_u to R_n under wet conditions. This behavior differs from Small & Kurc's (2003) observations conducted under similar conditions, in which the increase in R_n under wet conditions is responsible for and even equal to the decrease in L_u, and that S_d and L_d do not or slightly vary with soil moisture. This difference may be most likely related to the specific atmospheric opacity resulting from aerosol load (Warner, 2004; Nicholson, 2011) and atmospheric emitting temperature (Miller, Slingo, Barnard, & Kassianov, 2009), and even to the complex interaction between land surface and atmosphere involving atmospheric boundary layer mixing (Betts & Ball, 1996; Eltahir, 1998; Maxwell, Chow, & Kollet, 2007; Jones and Brunsell, 2008; Miller et al., 2009; Bonan, 2016), which is out of the scope of this study.

4.2 Response of surface energy partitioning to soil moisture

In the dryland environments, P is much less than potential ET. The actual ET is believed to be limited by soil moisture availability most of time because P is not only scarce but also intermittent and unpredictable (Kurc & Small, 2004; Rodríguez-Iturbe

& Porporato, 2005). Even small differences in soil moisture differences induced by rainfall events can have a very large impact on the surface energy balance (Eltahir, 1998; Nicholson, 2011) because the energy and moisture budgets at the land surface are principally linked by evaporation, which is an expenditure of both energy and water mass (Brubaker & Entekhabi, 1996; Sellers et al., 1997; Shuttleworth, 2012). Hence, we used the data acquired in the same periods as the previous section (i.e., section 4.1) to identify the impacts of soil moisture on surface energy balance. Such analysis is the necessary for a quantitative understanding of the response of surface energy partitioning to soil moisture, because the seasonal and intra-seasonal variations in ET sensitively reflect the episodic pulses of rainfall in the shrubland to a large extent (Figure 5).

The average surface energy components in daytime over the shrubland under the dry and wet conditions are presented in Figure 7. The daily average λE increased significantly by roughly a factor of 2.5 under wet conditions, in comparison with dry conditions (Figure 7a). This is closely linked to increasing availability of water for surface evaporation from the wet canopy (i.e., rainfall interception) and soil surface. In contrast to λE, interestingly, H shows a small decrease (about 9%) under wet conditions when compared with dry conditions. This behavior can be explained by the small changes (1.7 °C) in temperature gradient between surface and air (Figures 6b and 6c) for the two contrasting conditions, and greater values of wind speed under dry conditions than wet conditions (Figure 6f), which can partly balance the reduction of H at the surface if the differences in atmospheric properties (such as air density and specific heat capacity of air) are neglected, since the surface H is proportional to both T_s-T_a and aerodynamic conductivity that increases with wind speed (Campbell & Norman, 1998). The resultant Bowen ratio dropped dramatically from about 3.8 for wet conditions to 1.4 for dry conditions (Figure 7e), and evaporative fraction increases significantly in an order of over 100 % (0.21−0.43).

The availability of surface moisture also has an important influence on G by altering the thermal properties, by which the temperature gradient between the surface and subsurface is also controlled (Nicholson, 2011). Our results showed that *G* for the wet surface was 20% lower than for the dry surface (Figure 7c). This behavior can be understood when linking the thermal properties of the soil and the temperature gradients within the top soil layer. *G* is proportional to both the thermal conductivity of soil and the surface-subsurface temperature difference (Campbell and Norman, 1998). Wet soils have a higher thermal conductivity and heat capacity than dry soils (Betts & Ball, 1998; Warner, 2004; Nicholson, 2011), enhancing heat transfer to the subsurface, then creating a small temperature gradient just beneath the surface, and produces a lower *G*. Our observations showed that the difference between T_s and temperature at 10 cm in soil is 2.4 °C smaller for the wet surface (0.4 °C) than that for the dry surface (2.8 °C) (not shown). So, we speculate that the reduction of temperature gradient between surface and subsurface might make a larger contribution to the observed decrease in *G* than the increasing thermal conductivity under wet conditions for this contrasting comparison with respect to soil moisture.

Based on the analyses above, the increase in λE under wet conditions is not solely due to decrease in the corresponding *H*. The increase available energy at surface may also contribute the most energy for increased *ET* from surface. In this study, we observed that an increase in R_n and a reduction in *G*, on average, resulted in a higher available energy (c.a., 355 W·m^{-2}) under wet conditions than under dry conditions (288 W·m^{-2}, Figure 7d). The increase of available energy is comparable for the increase in λE (c.a., 74 W·m^{-2}) under wet conditions, in comparison with dry conditions, given the fact that *H* and *G* showed small reductions (Figures 7b and 7c). These results differed slightly from previous studies conducted under similar conditions (Kustas et al., 1991; Stannard et al., 1994; Kabat, Dolman, & Elbers, 1997; Small & Kurc, 2003), in which observed λE from the surface increases at the expense of the corresponding *H*. In summary, the observations from this and previous studies, confirm that the availability of surface moisture does have a decisive influence on surface energy partitioning, and this influence may also highly depend on the specific nature of surface (in terms of species composition of vegetation, phenological pattern, vegetation structure or leaf area index, thermal and hydraulic properties of soil), and even on local climatic and hydrological regimes.

4.3 Water balance and its implications

ET and *P* are by far the dominant components of the land-surface water balance in dryland regions (Huxman et al., 2005; Rodríguez-Iturbe & Porporato, 2005; Nicholson, 2011; Qiu, Li, & Yan, 2015). A knowledge of the land-surface water balance in the shrubland is thus necessary for a better understanding of the energy and water exchange between the atmosphere and the surface in this ecotone. Among the rest of components in the land-surface water balance, surface runoff is practically absent over the shrubland, and leakage losses generally are also minuscule for deep soils (the wetting front were confined to depths shallower than 15 cm) owing to the small rainfall events with low intensity and less rainfall amounts in the shrubland (Figures 5a and 5b); the lateral subsurface flow in the unsaturated zone can be considered to be negligible due to the relative flat surface; there is also no any lateral surface water flow into or out the shrubland. It is thus speculated that the *P* must be balanced by the *ET* for this shrubland on the annual basis, given the change in soil water storage in the first 2 m of the soil profile in the shrubland being much smaller over the entire year (not shown here).

However, the total *ET* was 195 mm and 181 mm in 2014 and 2015, respectively, exceeding the corresponding *P* by 87 mm and 77 mm (Figure 8) and resulting in a negative annual hydrological excess ($P-ET$), suggesting that the shrubland is a net consumer of water resources. This behavior is not surprising since the main source for the imbalances in the shrubland may be attributed to the dominant phreatophyte shrub species with deep roots that can utilize shallow groundwater (i.e., stores in unconfined aquifers) for transpiration, given the fact that the groundwater depth was shallow (i.e., about 4.4 m) and fluctuated little throughout the year. According to the previous study conducted in the shrubland (Ji et al., 2016; Zhou et al., 2017), the dominant phreatophyte shrub species at the site are more or less dependent on the availability of groundwater. Note that the groundwater sources occurring at 4.4 m depths may be consistently recharged by the lateral groundwater flows away from the Linze oasis (located close to the shrubland about 3 km in distance) towards the oasis-desert ecotone areas (Chen & Qu, 1992), as the groundwater sources in oasis areas are recharged principally through infiltration from the Heihe river that cuts across the Lineze oasis nearby the shrubland site, and through the deep drainage from the irrigation water for agricultural purpose (Ji, Kang, Zhao, Zhang, & Jin, 2009; Li et al., 2018). This behavior of negative annual hydrological excess was also reported in other dryland regions with deep-roots vegetation cover (Eamus, Froend, Hose, Loomes, & Murray, 2006; Steinwand, Harrington, & Or, 2006; Scott, Huxman, Cable, & Emmerich, 2006; Qiu et al., 2015).

In addition, the land-surface water balance resulting from the observations in this study indicated that P is not the only one source for water supply in the shrubland. Groundwater may be another important source of water for ET over the shrubland, comparable with P. Groundwater may also play a crucial role in regulating the energy exchange between land surface and atmosphere over the shrubland, as ET links the water balance to an important component of the energy balance (λE). From the knowledge of the annual energy and water balances discussed above, assuming that P was completely depleted by ET and thus there was no P penetrated to recharge the groundwater, it was estimated that the water supply for ET from groundwater was responsible for near half of λE (about 45% in 2014, and 43% in 2015) on the annual basis, consuming approximately 17% of R_n over the shrubland in both years. Thus, these characteristics of the land-surface water balance have foremost ecohydrological implications for protecting vegetation in the oasis-desert ecotone, as well as for water resources management in oasis. For example, a reasonable allocation of water resources, such as avoiding heavy groundwater pumping for extensive agricultural irrigation use within the oasis, can sustain groundwater availability for these shrub species over the oasis-desert ecotone to ensure the maintenance of the ecotone ecosystem health (that is, declines in groundwater depth may have crucial negative impact on these shrub species), because these dominant shrub species are typically dependent on groundwater availability.

5. CONCLUSIONS

We have described diurnal and seasonal variations in surface energy balance at a shrubland site within the ecotone during a two-year study period, using measurements of a flux tower and an automatic weather station, in combination with auxiliary field observations. The remarkable seasonal variations in the surface energy exchange and evapotranspiration is closely related to the combined effect of solar radiation, precipitation, and plant phonologic regime (or seasonal variation in vegetation dynamics), given that R_n is in phase with P and LAI under the local climatic conditions. Our results demonstrated that the availability of surface moisture from precipitation plays a major role in modulating the rates of ET over the shrubland, as well as on surface radiation budget or energy partitioning, although λE is consistently far lower than H throughout the year. Note that an increased available energy (as a result of increased R_n and decreased G) may contribute the most energy for increased λE from surface to the rapid and transient increase following rainfall events at the shrubland, whilst the corresponding H present a very small reduction. These behaviors are slightly different from the observed variations under similar conditions in literature, concerning vegetation characteristics and climatic regimes.

Interestingly, in the present study, the observed land-surface water balance for the shrubland showed that the total annual ET exceed greatly the corresponding P for the two-year study period, suggesting that these dominant phreatophyte shrubs with deep roots can utilize shallow groundwater for transpiration to some extent. So, it is concluded that groundwater may play a crucial role in regulating the water exchange between land surface and atmosphere over the shrubland, as well as in modulating surface energy partitioning through the coupling between the water cycles and energy exchanges. From the vegetation point of view, groundwater availability perhaps plays a more important role in sustaining vegetation settlement and growth than does rainfall in the water-limited ecosystem, given the scarce precipitation with small amounts and specific hydrological conditions. The analysis and findings from this study can provide a better understanding of the energy and water balances in the shrubland and associated underlying mechanisms in the arid environments, which helps offer guidelines for allocating water resources, protecting vegetation, and maintaining well-being of the oasis-desert ecotone.

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DATA AVAILABILITY

The data that support the findings of this study are available on request from the corresponding author.

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TABLES

Table1. Comparisons of the various terms of the energy budget and flux ratios at midday time (10:00 AM−3:00 PM) and daytime (R_n > 0 W·m^{-2}) over the shrubland for different seasons. R_n: net radiation; G: ground heat flux; S: storage term; H: sensible heat flux; and λE: latent heat flux. The ratio of G to R_n is ignored in consideration of G being quite small in magnitude and only a fraction of less than 3 % of R_n when compared to other fluxes terms.

	Ratio	
	Bowen	Evaporative

		Surface energy flux ($W \cdot m^{-2}$)					ratio	fraction			
		R_n	G	S	H	λE	$H/\lambda E$	$\lambda E/(H+\lambda E)$	H/R_n	$\lambda E/R_n$	G/R_n
Midday time (10:00 AM – 3:00 PM)											
2014	Spring	422	164	2	261	18	14.50	0.06	0.62	0.04	0.39
	Summer	500	163	2	218	85	2.56	0.28	0.44	0.17	0.33
	Autumn	378	133	3	176	31	5.68	0.16	0.50	0.09	0.38
	Winter	277	110	4	130	8	16.25	0.06	0.47	0.03	0.40
2015	Spring	469	166	3	248	23	10.78	0.08	0.53	0.05	0.35
	Summer	530	173	1	247	74	3.34	0.23	0.46	0.14	0.33
	Autumn	386	136	2	170	39	4.36	0.19	0.44	0.10	0.35
	Winter	277	102	5	113	10	11.30	0.08	0.40	0.03	0.37
Daytime ($R_n > 0$ $W \cdot m^{-2}$)											
2014	Spring	297	92	3	174	13	13.38	0.07	0.58	0.04	0.31
	Summer	330	83	2	142	67	2.12	0.32	0.43	0.21	0.26
	Autumn	263	86	3	129	25	5.16	0.19	0.49	0.09	0.33
	Winter	213	75	4	95	6	15.83	0.06	0.45	0.03	0.35
2015	Spring	315	94	3	172	17	10.12	0.09	0.54	0.05	0.30
	Summer	366	96	2	168	60	2.80	0.26	0.46	0.16	0.26
	Autumn	276	80	3	119	31	3.84	0.21	0.43	0.11	0.29
	Winter	214	71	5	86	8	10.75	0.09	0.40	0.03	0.34

FIGURE LEGENDS

Figure 1. Monthly average incoming solar radiation (a), air temperature (b), vapor pressure deficit (c), total precipitation (d), wind speed (e), and top soil moisture (0−40 cm) (f) at the study site in both 2014 and 2015 as compared with the long-term 11-year (2005−2015) monthly average values.

Figure 2. Energy balance scatter plot showing degree of energy balance closure based on 30-min data (a, b) and daily values (c, d) of latent plus sensible heat flux ($H + \lambda E$) versus net radiation minus ground and storage heat flux ($R_n - G - S$) for both 2014 and 2015. Also shown are the linear regression results including the equation of best fit, coefficient of determination (R^2), and the number of samples (n) included in the regression.

Figure 3. Diurnal cycle of net radiation (R_n), soil heat flux (G), storage heat flux (S), sensible heat flux (H), and latent heat flux (λE) for both 2014 and 2015. Symbols

represent the half-hourly values averaged over spring (March−May), summer (June−August), autumn (September−November), and winter (December−February), respectively.

Figure 4. Monthly averages of daily accumulated net radiation (R_n), latent heat flux (λE), sensible heat flux (H), and soil heat flux (G) over the shrubland for a complete annual cycle of 2014 (a) and 2015 (b).

Figure 5. Daily evapotranspiration (ET) and precipitation (P) over the shrubland for a complete annual cycle of 2014 (a) and 2015 (b), and leaf area index (LAI) at the shrubland during the two-year period (c). The LAI values represent the measurements for about 10-day periods, defined as green leaf area per unit ground area.

Figure 6. Difference in the environmental conditions (a-f) and in the surface radiation budget (g-l) under dry (i.e., orange columns) and wet (i.e., cyan columns) soil moisture conditions at the surface of the shrubland in summer. Each column represents the mean daytime (8:00 − 18:00) values, and bars are 1 standard error. θ_s denotes surface soil volumetric moisture of 0-20 cm layer; T_s and T_a denote surface and air temperature, respectively; RH denotes relative humidity; VPD denotes vapor pressure deficit; u_c denotes wind speed at canopy height; S_d and S_u denotes downward and upward shortwave radiation, respectively; L_d and L_u represent downward and upward longwave radiation, respectively; Albedo is surface albedo (k); and R_n denote surface net radiation. Differences in θ_s, T_s, RH, VPD, S_d, L_u, Albedo, and R_n between dry and wet surfaces are statistically significant at the 95% confidence interval, at which differences in T_a, u_c, S_u, and L_d are not significant.

Figure 7. Comparison of surface energy components under dry (i.e., orange columns, θ_s = 0.04 $m^3 \cdot m^{-3}$) and wet (i.e., cyan columns, θ_s = 0.11 $m^3 \cdot m^{-3}$) soil moisture conditions at the surface of the shrubland in summer. Each column represents the mean daytime (8:00 − 18:00) values, and bars are 1 standard error. λE, H, G, and Q_a represent latent heat flux, sensible heat flux, soil heat flux, and the available energy

($R_n - G$), respectively; Bowen ratio is $H/\lambda E$. Differences in λE, Q_a, and Bowen ratio between dry and wet surfaces are statistically significant at the 95% confidence interval, at which differences in H and G are not significant.

Figure 8. Comparison of cumulative evapotranspiration (*ET*) and cumulative precipitation (*P*) in the shrubland in 2014 (a) and 2015 (b).