

Energy balance and water use in a subtropical karst woodland on the Edwards Plateau, Texas

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SUMMARY

Woody encroachment into karst grasslands and savannas is presumed to reduce water availability and aquifer recharge, in part, because deep roots extract large quantities of water from perennial sources within the fractured bedrock underlying shallow soils. If true, energy balance partitioning and transpiration in woody ecosystems should be decoupled to an extent from rainfall, and sensitivity of the energy balance and evapotranspiration (ET) to rainfall and water deficits should be dampened. We evaluated responses of energy and water vapor fluxes to rainfall and water deficits in a live oak (*Quercus virginiana*)-Ashe juniper (*Juniperus ashei*) woodland on the karst Edwards Plateau, TX, USA, over a 2-year period using eddy covariance measurements of the turbulent fluxes. Total ET during the two years was 1416 mm, 92% of total rainfall. We observed large and rapid reductions in λE and increases in H during drying cycles, and high correlation between ET and soil water content in the upper 20 cm of the root zone. In most cases, ET declined at the same time as soil water content, indicating that the woodland relied heavily on water from recent rainfall events, rather than antecedent water. We found no evidence that deep roots were extracting significant amounts of water from a perennially stable supply of water. Excavations at the woodland site revealed a rock layer at 20 cm below the soil surface, with a dense root mat above the rock and penetration of relatively few roots into the rock through cracks and fissures. Thus, the most likely sources of water for trees were soil water and a limited supply of water stored in near-surface fractured rock layers.

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Introduction

Woody plant encroachment into arid and semi-arid grasslands and savannas is occurring worldwide (Archer et al., 2001; Van Auken, 2000), with potentially significant consequences for hydrology and biogeochemical cycling (Houghton et al., 1999; Baldocchi et al., 2004; Engel et al., 2005). The hydrologic impact of woody encroachment is particularly worrisome in karst landscapes because karst aquifers provide 25% of freshwater supplies for human consumption worldwide and 40% in the US (White et al., 1995). Karsts are landscapes formed from dissolution of soluble rocks, mainly limestone and dolomite. They generally have well-developed underground drainage systems, and strong interactions between surface and groundwater flow (Bonacci et al., 2009).

There is a wide-spread perception that woody encroachment, especially in karst landscapes, has significant negative effects on water yield (e.g. Tennesen, 2008). However, reality may be more complicated. A conceptual model by Huxman et al. (2005) laid

out criteria for impacts of woody encroachment on water yield. According to this framework, a reduction in streamflow is predicted in semi-arid uplands only if there is potential for rapid subsurface flow. Even though this condition may be characteristic of karst, it is by no means ubiquitous (Wilcox et al., 2005; Schwinning, 2008). For example, in the stepped landform of the Edwards Plateau largely unbroken rock layers close to the surface all but prevent downward fracture flow and deep root formation (Wilcox et al., 2007).

To date, perceptions of woodland ecohydrology on the Edwards Plateau have been shaped by studies conducted in the fault zone of the Balcones Escarpment. In this narrow zone along the eastern edge of the Edwards Plateau, roots of Ashe juniper (*Juniperus ashei* Buckholtz) and live oak (*Quercus virginiana* Miller) were observed in caves to depths of 9 m and 22 m, respectively, tapping into perched water tables (Jackson et al., 1999). Because these deep roots of juniper and oak have larger xylem conduit diameters and hydraulic conductances than shallow roots, they can contribute disproportionately more to water use if they encounter significant sources of water at depth (McElrone et al., 2004). Indeed, Jackson et al. (2000) found that water from depths greater than

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7 m was the source for 24% of growing season water used by a juniper whose roots had access to water in a cave. Pockman et al. (2008) reported that during prolonged drought, water uptake from deep roots of juniper accounted for as much as 60% of total daily transpiration. At night, hydraulic redistribution from deep to shallow soil maintained water flow through the roots. Using stable isotopes for tracing water sources, McCole and Stern (2007) concluded that juniper shifts its water use from predominately deep sources during the warm and dry summers to shallow sources during the cool and wet winters.

Recently, other studies conducted at sites lacking shallow caves in the stepped landform of the Edwards Plateau (Wilcox et al., 2007) have by contrast found no indication of water uptake from perched water tables (Schwinning, 2008; Eggemeyer and Schwinning, 2009). These studies concluded, based on stable isotope evidence, that water sources for juniper, oak and normally deep-rooted honey mesquite (*Prosopis glandulosa*) on the Plateau are largely shallow and unstable. Here we report on a study coming to the same conclusion based on measurements of energy balance and water vapor fluxes of a live oak-Ashe juniper woodland on the Plateau.

The surface energy balance is key to understanding how woody encroachment affects ecohydrology of Plateau ecosystems because water use is controlled by available energy and its partitioning between latent and sensible heat. If deep roots of woody species on the Plateau are exploiting a more stable supply of water than what is available to grasses, energy balance partitioning and transpiration in woody ecosystems should be decoupled to an extent from rainfall, and sensitivity of the energy balance and ET to rainfall and water deficits should be dampened.

Methodology

Site description

The Edwards Plateau, commonly referred to as the Texas Hill Country, is a 93,000 km² karst ecoregion that historically was vegetated by prairie grasses and live oaks (*Q. virginiana*) (Barnes et al., 2000). Chronic overgrazing by livestock, and suppression of wildfires has increased areal densities of woody species like Ashe juniper. Soils on the Plateau are generally shallow, and rainfall in excess of the soil's ability to retain is rapidly transported underground to aquifers, either through rapid subsurface flow through extensively fractured rock or through runoff into streams and rivers that connect with sinkholes and other recharge features. The Edwards-Trinity Aquifer is the only source of drinking water for over 2 million people residing primarily in the Austin-San Antonio corridor on the eastern edge of the Plateau. The recharge zone of the aquifer is characterized by highly faulted and fractured limestone, some of which outcrops at the surface. The aquifer is listed by the Karst Waters Institute as one of the ten most endangered karst systems in the world. The Plateau is also home to a number of threatened and endangered plant and animal species, many of them endemic, and at least 40 unique aquatic species live in the Edwards aquifer (Riskind and Diamond, 1986). Rainfall on the Plateau is highly variable, ranging from an annual average of 860 mm in the eastern portion of the Plateau to 380 mm in the west. Summers are usually warm and dry with sporadic rainfall, while winters are cool with frequent rainfall.

Energy balance measurements were made in a live oak-Ashe juniper with a continuous interlocking canopy. The woodland is on the Freeman Ranch, a 1700 ha research area near San Marcos, Texas, USA (Fig. 1) operated by Texas State University-San Marcos. The entire ranch is on the Edwards Aquifer recharge zone. Both species of trees in the woodland are evergreen, but live oak re-

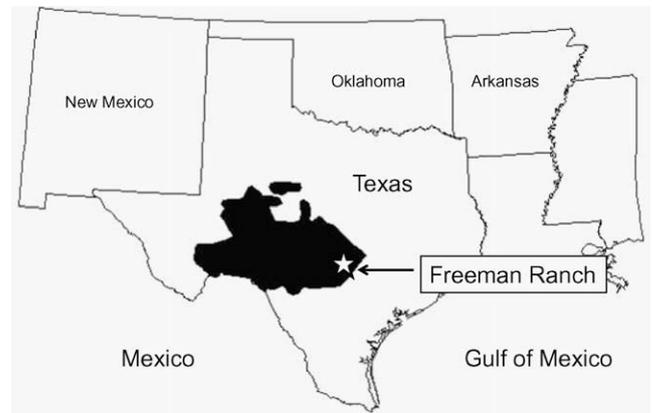


Fig. 1. Location of the Edwards Plateau (shaded) and the Freeman Ranch.

places its leaves more frequently than juniper. Most of the Ashe juniper is multi-stemmed, with an estimated areal density of 2850 stems ha⁻¹. At the site, mean diameter at breast height (*dbh*) and basal area for juniper are 0.08 m and 18.4 m² ha⁻¹, respectively. Mean *dbh* and basal area of live oak are 0.17 m and 23.5 m² ha⁻¹, respectively, and nearly all of the oak have single trunks. Areal density of above-ground biomass was estimated to be 5.6 kg m⁻² for juniper and 5.8 kg m⁻² for oak, based on the allometric equation of Jenkins et al. (2003). The soil in the woodland is Comfort stony clay, with a ~20-cm deep A horizon lying on fractured indurated limestone bedrock. Excavations showed that both oak and juniper formed dense root mats above the rock (Fig. 2), but some roots penetrated the rock through cracks and fissures.



Fig. 2. Photographs of the excavation at the woodland showing rooting patterns above the rock layer that was 20 cm below the surface.

Energy balance measurements

The surface energy balance is described by the equation

$$R_n = H + \lambda E + G + S \quad (1)$$

where R_n is net all-wave radiation, H is sensible heat flux, λE is latent heat flux, G is soil heat flux, and S is storage heat flux. Storage heat flux includes heat storage in above-ground biomass (S_v), and sensible (S_a) and latent (S_λ) heat storage in the canopy air space. Net radiation was measured at a height of 15 m above the soil using a model Q7.1 net radiometer (REBS, Seattle, WA). In addition, we had a Kipp and Zonen CRN1 net radiometer measuring short and longwave components of R_n from late May–mid-September in 2006.

Sensible and latent heat fluxes were determined by tower-based eddy covariance using the equations

$$H = \rho c_p \overline{wT'_a} \quad (2)$$

and

$$\lambda E = \overline{\lambda w' \rho'_v} \quad (3)$$

where ρ is density of air, c_p is specific heat of air, w is vertical wind speed, T_a is air temperature, λ is latent heat of vaporization, and ρ_v is vapor density. The prime in Eq. (2) denotes the fluctuation from a temporal average (here, 30-min) and the over bar a temporal average. Vertical wind speed and air temperature were measured acoustically using a CSAT-3 sonic anemometer (Campbell Scientific Inc., Logan, UT), while vapor density was measured using a LI-7500 open path infrared gas analyzer (LI-COR Inc., Lincoln, NE). The Schotanus et al. (1983) humidity correction was applied to sonic anemometer-derived H . The anemometer and open-path gas analyzer were placed at the same elevations as the net radiometer, and outputs sampled at 10 Hz. The gas analyzer was calibrated periodically using a span gas of known CO_2 concentration and a dewpoint generator (LI-610, LI-COR).

Soil heat flux (G) was determined by soil heat flux plates and calorimetry (Liebethal et al., 2005) using the equation

$$G = G_z + C_s \frac{\Delta T_s}{\Delta t} Z \quad (4)$$

where G_z is heat flux measured at depth z , C_s is soil heat capacity, T_s is average soil temperature above the heat flux plates, and t is time. Heat flux was measured at a depth of 5 cm at three locations at each site using heat flux transducers (HFT-3, REBS), and soil temperatures above the transducers were measured by thermocouple thermometry. Heat capacity was calculated from

$$C_s = \rho_b c_m + \rho_w \theta c_m \quad (5)$$

where ρ_b is soil bulk density, c_m is specific heat of the soil minerals, ρ_w is density of water, c_w is specific heat of water, and θ is volumetric water content. Water content was estimated by EC-10 capacitance sensors (Decagon, Pullman, WA).

Heat storage flux in the woodland biomass (S_v) was determined by calorimetry using the equation

$$S_v = m_v c_v \frac{\Delta T_v}{\Delta t} \quad (6)$$

where m_v is areal density of above-ground biomass, c_v is specific heat of vegetation, T_v is biomass temperature, and t is time. Biomass was estimated from measurements of diameter at breast height (*dbh*) using an allometric equation for a juniper-oak-mesquite woodland published by Jenkins et al. (2003). The *dbh* was obtained from stem (trunk) diameter measurements on all trees in 100 m² subplots along a 200 m transect in the woodland. Specific heat was estimated as 70% of the specific heat of water (Thom, 1975), and layer-weighted air temperature was used as a surrogate for

T_v , following Oliphant et al. (2004). Air temperatures were measured at heights of 0.5, 2, 5, 8 and 15 m above the soil using ventilated HMP35A temperature–humidity probes (Vaisala, Woburn, MA). Blanken et al. (1997) found that heat storage in leaves contributed <5% to total heat storage, and this was confirmed by Oliphant et al. (2004). We therefore neglected that contribution to S_v .

Sensible heat storage in the canopy air-space (S_a) was calculated as described by Oliphant et al. (2004) using the equation

$$S_a = \rho c_p \sum_{i=1}^n \frac{\Delta T_a}{\Delta t} \Delta z_i \quad (7)$$

where T_a is the air temperature of layer i as measured by the HMP35A, and Δz_i is layer thickness. Similarly, latent heat storage (S_λ) was calculated as

$$S_\lambda = \lambda \sum_{i=1}^n \frac{\Delta \rho_v}{\Delta t} \Delta z_i \quad (8)$$

where ρ_v was determined from HMP35A measurements of relative humidity and air temperature.

Supporting measurements

A number of other meteorological, and soil measurements were made. Global irradiance (R_s) was measured with a pyranometer (LI-200, LI-COR) and rainfall with a tipping-bucket rain gauge (Texas Electronics, Inc., Dallas, TX, USA). Volumetric soil water content at depth intervals 0–0.10, and 0.10–0.20 m was measured continuously using capacitance sensors (EC-10, Decagon, Inc., Pullman, WA, USA) with three sensors installed at each depth interval. We encountered rock at depths greater than 0.2 m, so no water content sensors were installed beyond that depth.

Data processing and gap filling

All fluxes were calculated as 30-min averages. Eddy covariance calculations included spike removal, ‘natural wind’ coordinate rotation (Lee et al., 2004), and adjustments for variations in air density due to water vapor (Webb et al., 1980; Ham and Heilman, 2003). A friction velocity (u^*) filter was used to reject data obtained when turbulence was low (u^* less than a threshold value). We used a u^* threshold of 0.15 m s^{−1}, which was determined as the value above which further increases in u^* had little effect on flux calculations (Hastings et al., 2005). Gaps in meteorological data and turbulent fluxes were filled using the on-line tools of Reichstein (<http://www.bgc-jena.mpg.de/bgc-mdi/html/eddyproc/index.html>).

Energy balance closure

Energy balance closure is a requirement of the first law of thermodynamics, but it is seldom achieved due to systematic underestimation of the turbulent fluxes by eddy covariance, as discussed by Wilson et al. (2002). They examined energy balance closure across 22 sites in FLUXNET and reported that slopes of regressions of $\lambda E + H$ against $R_n - G - S$ for all sites and years ranged from 0.53 to 0.99 with a mean of 0.79. We achieved energy balance closure of 0.93 ($H + \lambda E = 0.93 (R_n - G - S) - 1.4$, $r^2 = 0.94$) over the two years, based on 30-min averages of the fluxes. We then forced energy balance closure using a Bowen ratio conservation approach discussed by Twine et al. (2000) and used by Oliphant et al. (2004), Scott et al. (2004), Barr et al. (2006), Steinwand et al. (2006), and Kosugi et al. (2007), among others. We multiplied both H and λE by the ratio of $(R_n - G - S)$ to $(H + \lambda E)$, thus preserving the Bowen ratio ($H/\lambda E$) measured by eddy covariance without favoring sensible or latent heat in the apportionment of the missing energy (Steinwand et al., 2006).

Results and discussion

Environmental conditions

Microclimatic conditions for 2005 and 2006 are shown in Fig. 3. The woodland received 6.0 GJ m^{-2} of solar radiation in 2005, and 6.1 GJ m^{-2} in 2006. Total rainfall in 2005 was 795 mm, while the total in 2006 was 744 mm. Both totals were below the annual mean of 858 mm. Total rainfall during the last 3 months of 2004 was 382 mm, 182 mm above the annual mean for those months, so that soil water storage was fully charged at the beginning of 2005. An additional 278 mm of rainfall occurred during the first 3 months of 2005, with rainfall occurring on 30 days during this period. In contrast, only 44 mm of rain fell during the last 3 months of 2005, resulting in low soil water content at the beginning of 2006. Also, total rainfall during the first three months of 2006 was only 129 mm, much lower than what occurred during the same time period in the previous year. There were several pronounced drying cycles during each year during which little or no rainfall oc-

curred, most notably between days 154 (3 June) and 187 (6 July) in 2005, and days 187 (6 July) and 247 (4 September) in 2006. Mean annual air temperatures and vapor pressure deficits (VPD) were 19.1°C and 0.72 kPa , respectively, in 2005, and 19.8°C and 0.85 kPa in 2006. Reference evaporation, calculated using the method of Allen et al. (1994), was slightly higher in 2006 (875 mm vs. 851 mm in 2005).

Net radiation

Seasonal fluctuations in R_n are shown in Fig. 4. The highest daily total recorded during the 2-year period was 20.9 MJ m^{-2} , while the lowest was 0.6 MJ m^{-2} . The annual total averaged 4.0 GJ m^{-2} , with <1% difference between years. During the four months in spring and summer of 2006 when components of R_n were measured by the CRN1 radiometer, midday albedo averaged 0.10, and the emitted longwave component of R_n averaged 464 W m^{-2} . Maximum and minimum outgoing longwave recorded during this period were 578 and 300 W m^{-2} , respectively. We did not observe any

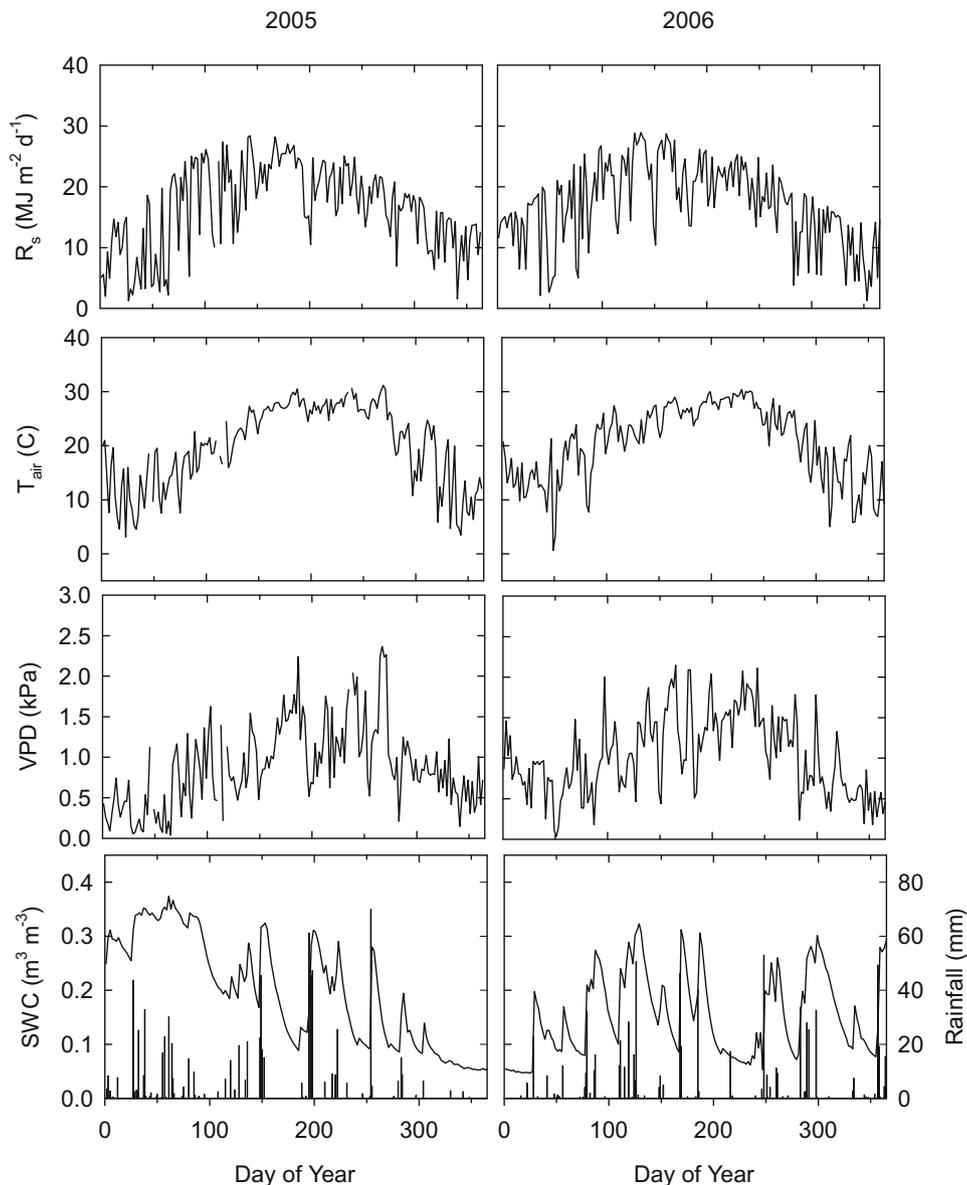


Fig. 3. Seasonal variation in global irradiance (R_s); average daily air temperature (T_{air}); vapor pressure deficit (VPD); soil water content in the upper 20 cm (SWC); and daily totals of rainfall (bars) in 2005 and 2006.

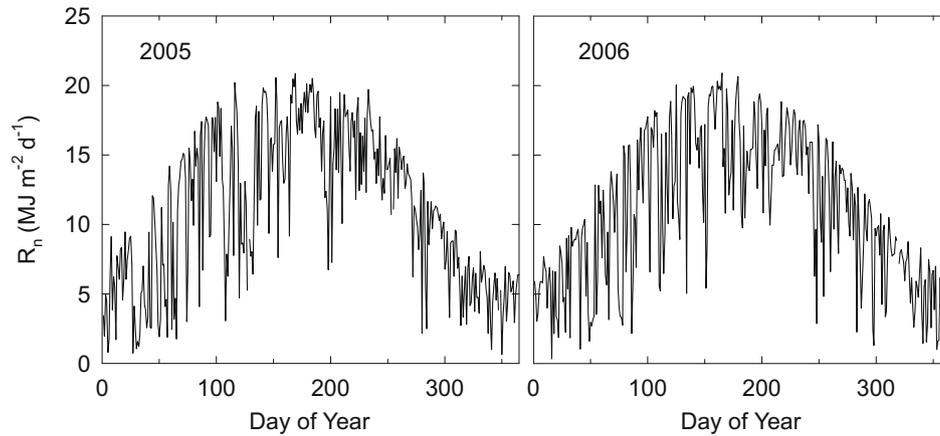


Fig. 4. Seasonal variations in net radiation (R_n) in 2005 and 2006.

consistent reductions in R_n during periods of water deficit that might have occurred due to changes in albedo or increases in canopy temperature.

Storage heat fluxes

On most days, diurnal fluctuations of G and S were of similar magnitude, with peak gains and losses in G lagging those in S by 4–6 h. However, on sunny days following cloudy days, there were large gains in S during the morning that considerably exceeded those in G . Seasonal changes in storage heat flux were dominated by G , as shown by the monthly averages of daily totals in Fig. 5. Averages of S were an order of magnitude lower than G because at night the canopy essentially lost all of the heat it gained during the daytime. Soil warming in spring and summer resulted in a heat gain and net downward transfer of G , and soil cooling in autumn and winter resulted in a heat loss and net upward transfer of G . In 2005, maximum heat storage in the soil occurred in June at the time of peak solar radiation and maximum soil surface temperatures. However, in 2006 maximum G occurred in August during a prolonged period without rain.

Cumulative heat loss in the woodland reached its peak in early spring, and cumulative heat gain in late summer (Fig. 6), 1/4 cycle out of phase with net radiation, as expected. The end of year deficits were similar to those found by Oliphant et al. (2004) for a deciduous forest in Indiana, USA, and are attributed to actual year-to-year variations, methodology, and unmeasured components.

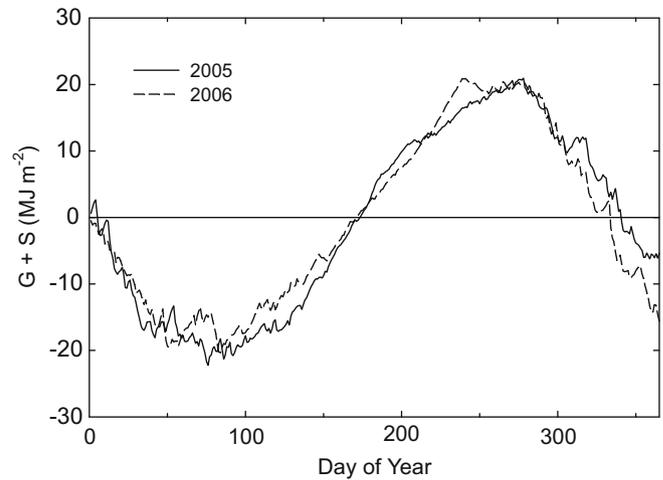


Fig. 6. Cumulative sum of soil heat flux (G) and storage heat flux in the canopy in 2005 and 2006.

Sensible and latent heat flux, and evaporation

Seasonal changes in λE and H , shown in Fig. 7(a–d), tracked changes in R_n , VPD and soil water content (Figs. 3 and 4). There were large reductions in λE when the soil dried, accompanied by increases in H . After forcing energy balance closure, peak values

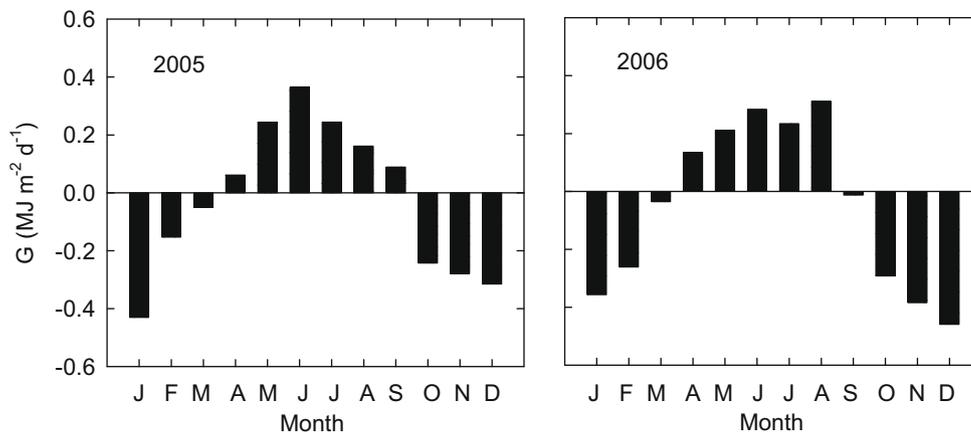


Fig. 5. Monthly averages of daily soil heat flux (G) in 2005 and 2006. Positive values denote a heat gain and negative values a loss.

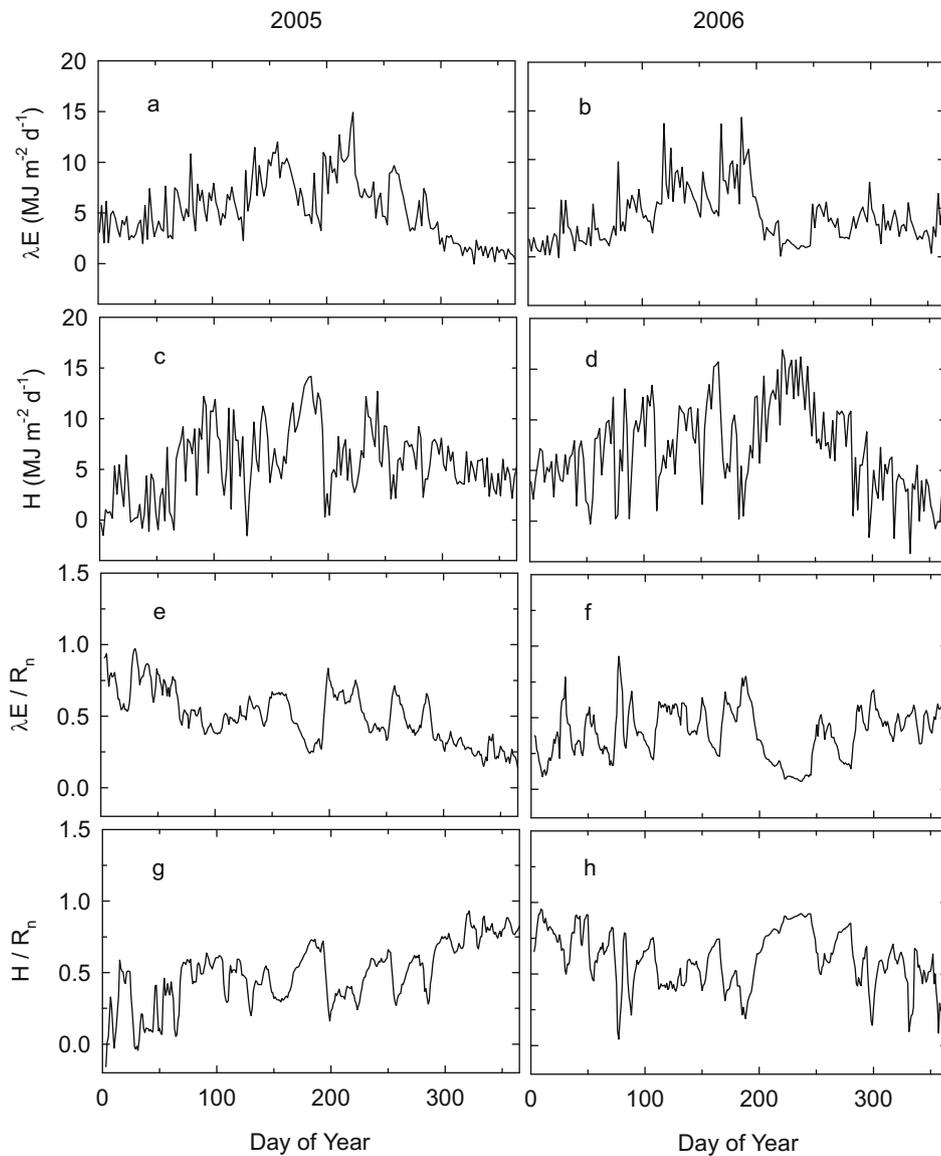


Fig. 7. Seasonal variations in latent (λE) and sensible (H) heat fluxes (a–d); and 5-d running averages of the ratio of daily totals of λE and H to net radiation (R_n) (e–h) in 2005 and 2006.

of λE and H in 2005 were $14.9 \text{ MJ m}^{-2} \text{ d}^{-1}$ (6.1 mm d^{-1} of evaporation) and $15.4 \text{ MJ m}^{-2} \text{ d}^{-1}$, respectively, and 14.3 (5.8 mm d^{-1}) and $16.9 \text{ MJ m}^{-2} \text{ d}^{-1}$ in 2006. In 2005, there was a gradual decrease in λE as a percentage of available energy that was associated with declining water availability, with $\lambda E/R_n$ dropping from near 1 at the beginning of the year to 0.14 at the end (Fig. 7e). Sensible heat loss accounted for 82% of R_n at the end of 2005 (Fig. 7g). Similar trends were not observed in 2006, but like 2005, there were large reductions in $\lambda E/R_n$ and increases in H/R_n during drying cycles (Fig. 7f, and h). During the most severe dry period, λE dropped to 4% of R_n while H rose to 94%.

In 2005, turbulent fluxes were divided almost equally between λE and H (each $\sim 2 \text{ GJ m}^{-2} \text{ year}^{-1}$) (Fig. 8). For the first 9 months of that year, Bowen ratios ($\beta = H/\lambda E$) based on daily totals of H and λE were <1 on the majority of the days, and cumulative λE exceeded H . There were periods of water deficit during this time when β reached values as high as 4.8 (Fig. 9). During the last three months, β increased to values as high as 14.4 because rainfall and water availability were very low. In 2006, total H was much higher than λE ($2.5 \text{ GJ m}^{-2} \text{ year}^{-1}$ vs. $1.5 \text{ GJ m}^{-2} \text{ year}^{-1}$) because of lower water

availability, with β reaching 24 near the end of the extended dry period in late summer (Figs. 8 and 9). Oliphant et al. (2004) observed a much smaller increase in β in the Indiana forest during an unusually dry period. He attributed the small response to the effectiveness of deep roots in extracting water stored in the soil column.

Total ET in 2005, based on forced energy balance closure, was 806 mm, exceeding total rainfall by 11 mm. As mentioned previously, rainfall during the latter part of 2004 was unusually high, so it is likely that water availability in the root zone was very high at the start of 2005. In addition, 302 mm of rain fell during the first 4 months of 2005, and soil water content in the upper 20 cm did not drop below $0.2 \text{ m}^3 \text{ m}^{-3}$ until day 24 May 2005 (day 144). In contrast to 2005, total E in 2006 was 610 mm, 34 mm below total annual rainfall. Forcing energy balance closure increased estimates of E by 97 mm in 2005, and 119 mm in 2006.

We found that ET at our woodland site was highly responsive to rainfall, and during drying cycles, highly correlated with soil water content in the upper 20 cm (Figs. 10 and 11). ET in Figs. 10 and 11 was normalized by reference ET (ET_0) to minimize impact of

day-to-day changes in microclimate (irradiance, VPD, etc.). In most cases, ET started to decrease at the same time as soil water content, and the response was nearly linear (Fig. 11b,c,e,f). This suggests insufficient root water uptake from the bedrock to sustain transpi-

ration as soil water in the upper 20 cm was depleted. ET rates and slopes of the ET vs. water content relationships were lower in 2006 than in 2005, indicating there was less water available in the root zone at onset of drying cycles. There were some periods when ET

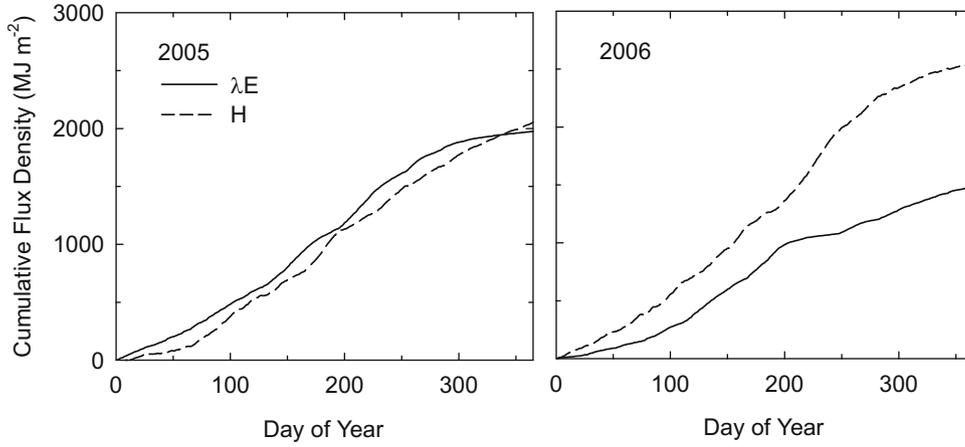


Fig. 8. Cumulative latent (λE) and sensible (H) heat fluxes in 2005 and 2006.

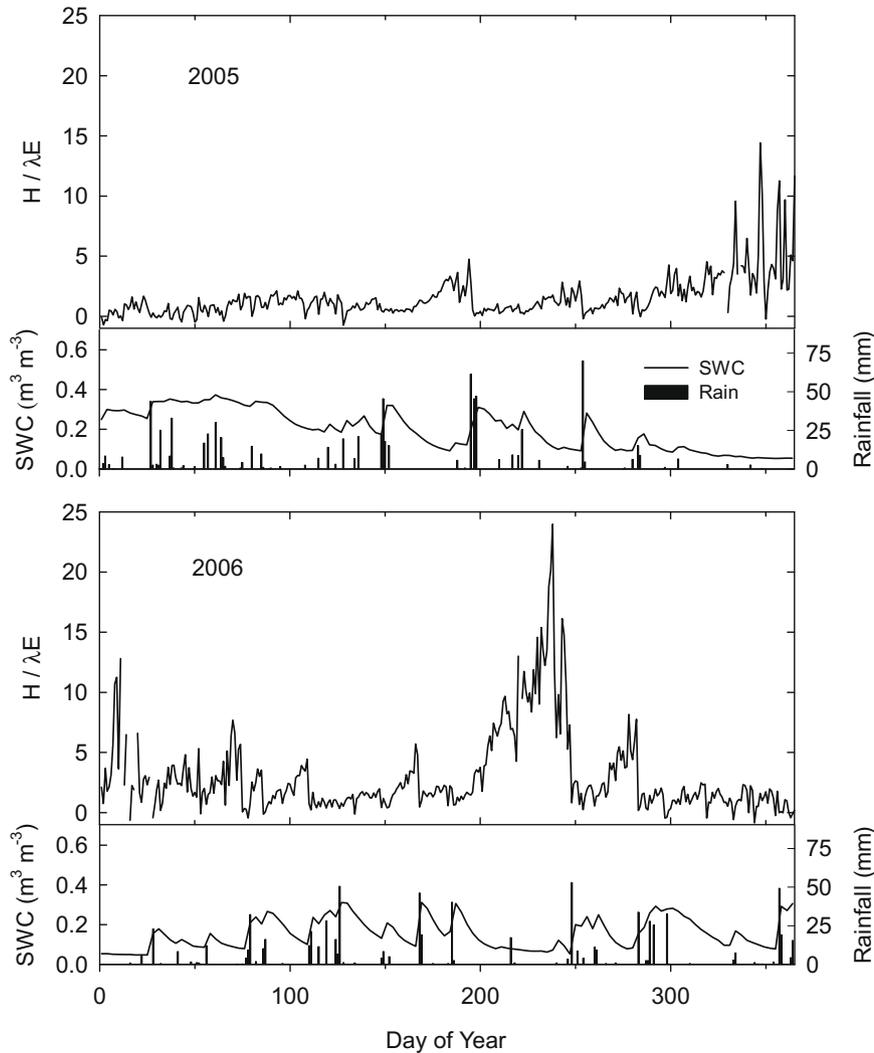


Fig. 9. Seasonal variations in the Bowen ratio calculated using daily totals of sensible (H) and latent (λE) heat fluxes. Also shown are soil water content (SWC) in the upper 20 cm and daily totals of rainfall.

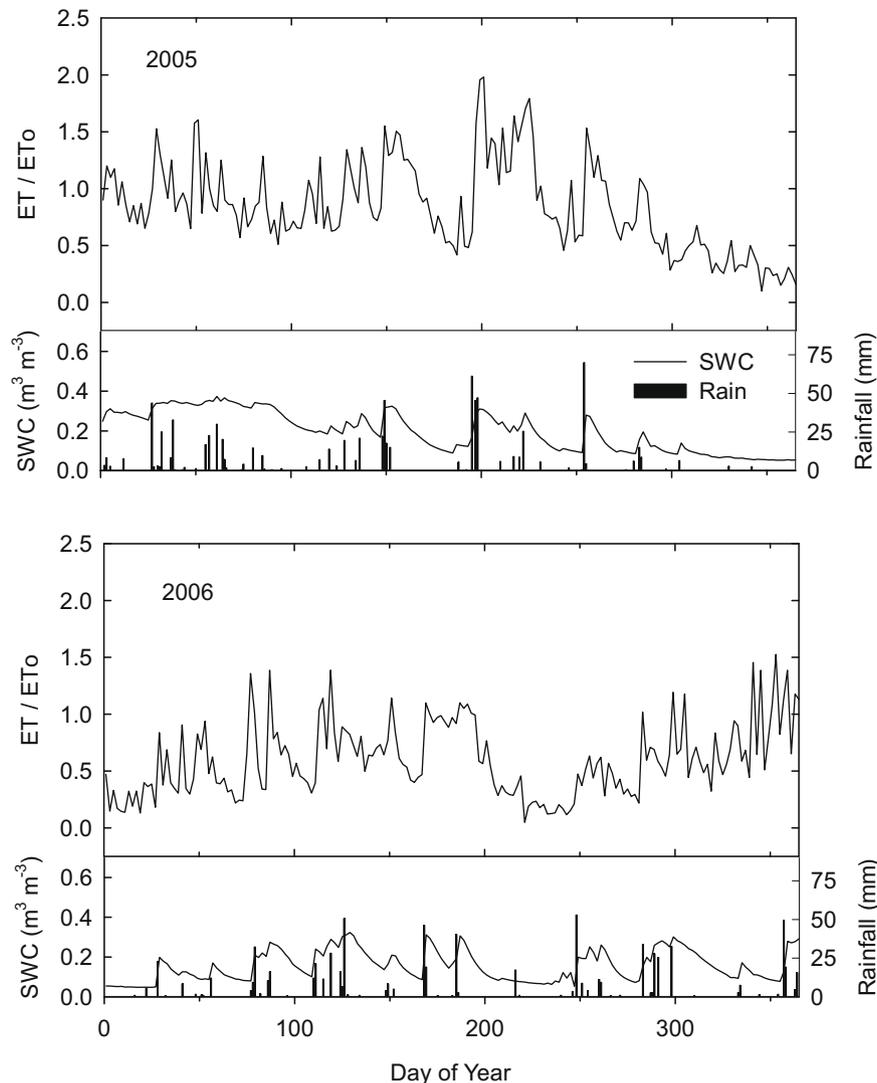


Fig. 10. Seasonal variations in evapotranspiration (ET), normalized by reference evapotranspiration (ETo). Also shown are soil water content (SWC) in the upper 20 cm and daily totals of rainfall.

response was delayed, and ET declined exponentially with soil water content (Fig. 11a and d). For example, ET during the most severe dry period (6 July–4 September 2006, days 187–247) did not begin to decline until 5 days after soil water content started to drop. This suggests a greater contribution from roots in the bedrock in meeting the evaporative demand than at other times. Apparently, rainfall that occurred prior to onset of this drying cycle was more effective in replenishing the root zone than similar rainfall events preceding other drying cycles. Cumulative ET during two of the most severe dry periods (days 154–187 in 2005, and days 187–247 in 2006) was 106 and 89 mm, respectively, which exceeded the water storage capacity of the 20 cm deep soil profile, estimated to be 70 mm. This indicates that roots did extract some water from the bedrock during drought. However, these water sources were ephemeral and depletable, just like the soil layer above.

The energy balance and ET responses of the woodland to rainfall and water deficits showed little evidence of a decoupling of transpiration from recent rainfall, as would be expected if deep roots were providing significant amounts of water from stable sources in the bedrock. Instead, our observations support the conclusions of Schwinnig (2008), that water sources for oak and juniper on the Plateau are generally shallow and unstable. While deep roots

can extract water from perennial sources such as perched water tables and cave streams in the narrow geographical range where these features exist (Jackson et al., 2000), the implication by Tennesen (2008) that this is the norm on the Edwards Plateau, and is responsible for large-scale depletion of surface and groundwater, is quite likely a gross overstatement.

Summary and conclusions

We examined energy balance partitioning and water use in a live oak-Ashe juniper woodland on the karst Edwards Plateau over a 2-year period (2005 and 2006), with a focus on response to rainfall and water deficits. We found large differences in total ET between years, and in partitioning of available energy between H and λE , due to differences in water availability. On an annual basis, turbulent fluxes in 2005 were divided almost equally between λE and H , whereas in 2006, λE was only 60% of H because overall water availability was lower than in the previous year. Total ET over the two years, based on forced energy balance closure, was 1416 mm, 92% of total rainfall. We observed large and rapid reductions in ET (λE) and increases in H during drying cycles, and high correlation between ET and near surface soil water content,

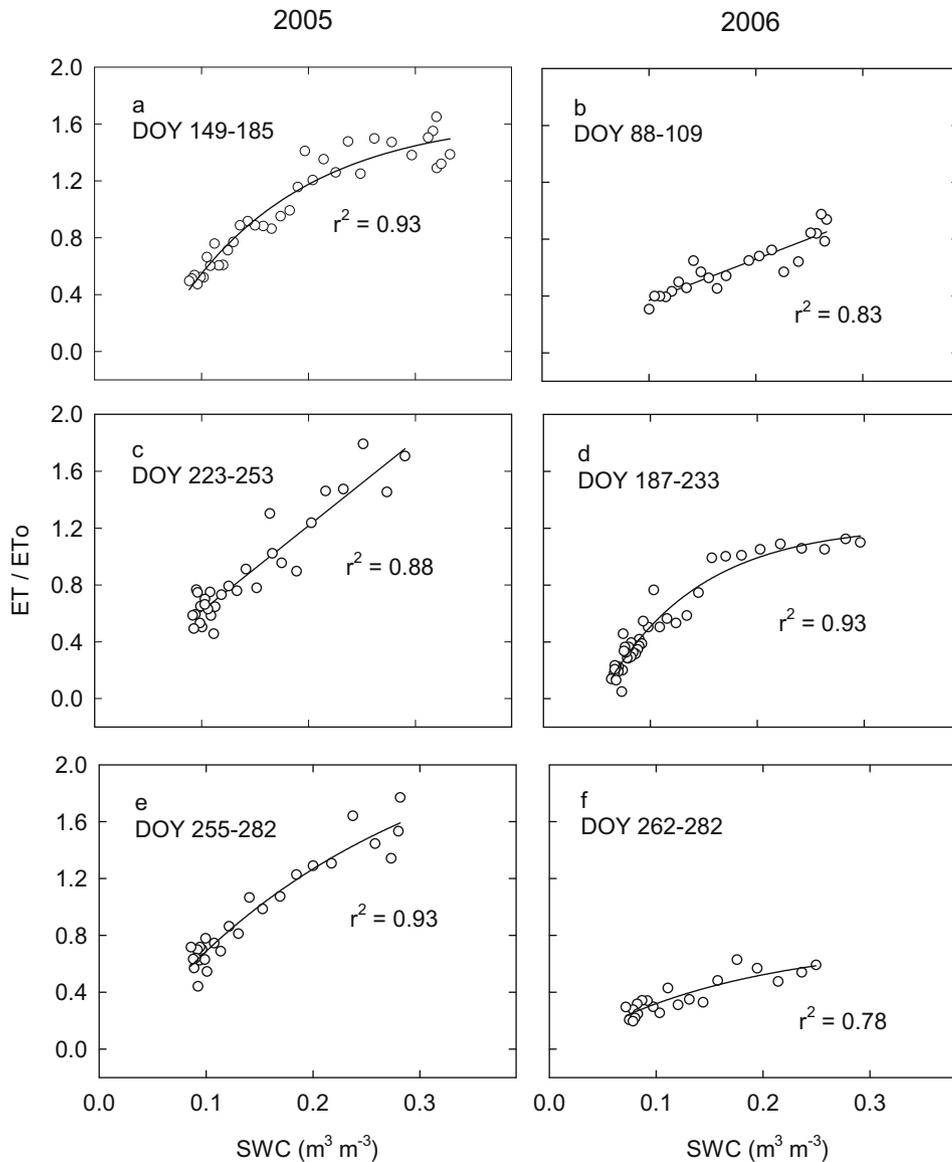


Fig. 11. Response of evapotranspiration (ET), normalized by reference evapotranspiration (ET_o), to soil water content (SWC) in the upper 20 cm of the root zone during drying cycles in 2005 and 2006, identified by the first and last day of year (DOY) of the drying period.

indicating that the woodland relied heavily on water from recent rainfall events, rather than antecedent water.

The woodland we studied was on the Edwards Aquifer recharge zone where the limestone bedrock is highly fractured and infiltration rates are high. It has long been assumed that woody species on the Plateau withdraw substantial amounts of water stored deep within the fractured bedrock which otherwise would find its way into the aquifer. This assumption has been used, in part, to justify removal of juniper and other species to increase water availability (Jones and Gregory, 2008). Our results cast doubt on that assertion, indicating instead that the water storage capacity of the root zone is limited, with little evidence that deep roots extract significant amounts of water from a more stable supply of water than what is available to shallow roots. Because storage capacity of the fractured bedrock is low, deep roots likely experience a hydraulic environment similar to that in shallow soil, fast recharge followed by rapid depletion. As a result, transpiration and partitioning of available energy between latent and sensible heat fluxes are closely coupled to rainfall.

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