Dynamics of evapotranspiration in semiarid grassland and shrubland ecosystems during the summer monsoon season, central New Mexico

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[1] To understand the coupled water and energy cycles in semiarid environments, we measured temporal fluctuations of evapotranspiration (ET) and identified key sources of the observed variability. Flux measurements are made using the Bowen ratio method, accompanied by measurements of soil moisture and radiation. We present data from semiarid grassland and shrubland sites, situated within 2 km of each other in New Mexico. The study includes three summer monsoon seasons. Midday available energy (Q_a) is higher at the grassland than at the shrubland by 20% or 70 W m^{-2} because of differences in net radiation (R_n) and soil heat flux (G). At both sites, midday evaporative fraction and daily ET are strongly correlated with surface soil moisture (θ_{0-5cm}) but poorly correlated with water content at greater depths or averaged throughout the entire root zone. The sensitivity of EF to θ_{0-5cm} is 30% lower at the grassland site. The differences in Q_a and EF cancel, yielding similar time series of ET at the two sites. Decreases in θ_{0-5cm}, ET, and EF following rainfall events are rapid: exponential time constants are less than 3 days. With the exception of the largest storms, infiltration following rainfall events only wets the top 10 cm of soil. Therefore the surface soil layer is the primary reservoir for water storage and source for ET during the monsoon season, suggesting that direct evaporation is a large component of ET. Given these results, predicting ET based on root zone−averaged soil moisture is inappropriate in the semiarid environments studied here.   *INDEX TERMS:* 1818 Hydrology: Evapotranspiration; 1833 Hydrology: Hydroclimatology; 1866 Hydrology: Soil moisture; 1878 Hydrology: Water/energy interactions; *KEYWORDS: Bouteloua eriopoda*, Bowen ratio, evapotranspiration, grassland, *Larrea tridentata*, shrubland

1. Introduction

[2] In arid and semiarid environments, ET is roughly equal in magnitude to precipitation on timescales longer than seasons [e.g., *Sala et al.*, 1992; *Phillips*, 1994; *Reynolds et al.*, 2000]. The variability of ET on shorter timescales is poorly constrained, but is critical for the coupled cycling of water, energy, and carbon in these environments. In particular, variability of ET affects (1) the amount of precipitation partitioned to runoff and recharge [*Phillips*, 1994; *Laio et al.*, 2001], (2) how land-atmosphere interactions influence weather and climate [*Eltahir*, 1998; *Pielke*, 2001], and (3) processes such as plant productivity, soil respiration, and biogeochemical cycling [*Noy-Meir*, 1973; *Lauenroth and Sala*, 1992; *Rodriguez-Iturbe*, 2000; *Porporato et al.*, 2003]. Considering the broad importance of ET, ecosystem-level observations of ET from semiarid environments are surprisingly limited (Table 1). The limited field studies that have been completed in semiarid environments demonstrate that ET varies greatly through time [*Gash et al.*, 1991; *Stannard et al.*, 1994; *Dugas et al.*, 1996; *Tuzet et al.*, 1997]. However, a more complete

understanding of the controls on ET in semiarid environments requires (1) continuous measurements of soil moisture, radiation, and other variables in conjunction with observations of ET and (2) ET and related variables measured across a range of soil and vegetation types.

[3] Meteorological and vegetation conditions control ET when soil moisture is not limiting [*Shuttleworth*, 1991]. However, because precipitation is much less than potential ET in arid and semiarid environments, ET is believed to be limited by soil moisture most of the time in dry land ecosystems [*Noy-Meir*, 1973; *Rodriguez-Iturbe*, 2000]. Both simple and complex hydrologic models typically include a dependence of ET on soil moisture [*Desborough et al.*, 1996; *Mahfouf et al.*, 1996]. Unfortunately, field observations required to test these relationships are often lacking, particularly in arid and semiarid environments where limitations from soil moisture are believed to be the most important.

[4] Recently, a simple ET-soil moisture relationship has been extensively used to investigate soil moisture dynamics [*D'Odorico et al.*, 2000; *Rodriguez-Iturbe*, 2000; *Fernandez-Illescas et al.*, 2001; *Laio et al.*, 2001] and the resulting impacts on plant productivity, species interactions, and nutrient cycling [e.g., *Ridolfi et al.*, 2000; *Porporato et al.*, 2001, 2003]. We summarize this ET-soil

Table 1. Studies of Evapotranspiration in Semiarid Environments

Reference	Location	Vegetation	Duration	Related Measurements[a]
Dugas et al. [1996]	Jornada, New Mexico	shrub/grass separate	April–May 1991, 1992; July–Nov. 1991, 1992	ET, soil evaporation
Gash et al. [1991]	Sahel, western Niger	shrub/grass mixed	first 6 weeks of dry season 1988	ET, soil moisture (weekly; gravimetric)
Kabat et al. [1997]	Sahel, western Niger	shrub/grass mixed	Aug.–Oct. 1992	ET
Malek and Bingham [1997]	Great Basin desert valley, Nevada	shrub	1993–1994 water year beginning Oct. 1	ET, soil moisture (weekly; TDR; not used in results)
Stannard et al. [1994]	Walnut Gulch, Arizona	shrub/grass separate	June 1990, July–Aug. 1990	ET (grass information limited)
Taylor [2000]	Sahel, western Niger	shrub/grass mixed	Aug.–Oct. 1992	ET, soil moisture (14 times in 40 days; neutron probe), soil evaporation
Tuzet et al. [1997]	Sahel, western Niger	shrub/grass mixed	Aug.–Sept. 1992	ET, shrub transpiration
Untland et al. [1996]	Sonoran desert, Arizona	shrub/succulents mixed	May 1993–May 1994	ET

[a]Sampling period and sampling method for soil moisture are in italics.

moisture relationship as follows: ET increases linearly with root zone water content (or effective saturation) between the wilting point and a moisture value at which ET equals potential ET. This relationship was fashioned according to data from lysimeter studies performed in relatively wet environments [*Black*, 1979; *Dunin and Greenwood*, 1986]. Modeling studies show that predicting ET based only on root zone averaged soil moisture may be an oversimplification, particularly if plants can compensate for a portion of their roots being in dry soil [e.g., *Guswa et al.*, 2002]. This relationship cannot be ruled out based on studies from wetter environments, where the only observations of soil moisture were from the surface soil [*Betts and Ball*, 1998; *Eltahir*, 1998]. This relationship has not been tested with ET and soil moisture data from semiarid environments (Table 1). The controls on ET from soil moisture should be greatest in these locations, and therefore the model should be considered a hypothesis that must be tested. More complex models calculate soil moisture stress independently in a series of distinct layers [e.g., *Mahfouf et al.*, 1996; *Feddes et al.*, 2001]. The assumption that restrictions on transpiration can be linearly summed across soil layers has not been sufficiently tested with field observations.

[5] A widespread vegetation change has been occurring over the past century in semiarid regions worldwide: herbaceous grasslands are being replaced by woody shrublands [*Burkhardt and Tisdale*, 1976; *Schultz and Floyd*, 1999; *Van Auken*, 2000]. Dramatic environmental changes are expected to accompany this shrub invasion, including changes in surface water, energy, and carbon cycling [*Schlesinger et al.*, 1990; *Abrahams et al.*, 1995; *Houghton et al.*, 1999; *Pacala et al.*, 2001; *Jackson et al.*, 2002; *Wilcox*, 2002]. As ET is the primary loss of water from soil in areas where shrub invasion has occurred, changes in ET may be integral to the observed and hypothesized ecosystem changes associated with shrub invasion [e.g., *Wilcox*, 2002; *Wilcox et al.*, 2003]. In addition, it is reasonable to expect that shrub invasion affects ET because the presence of woody species substantially changes ecosystem structure and function in four main ways (Table 2). First, leaf area index (LAI) is often lower in shrubland than grassland, which should yield lower transpiration rates [*Mahfouf et al.*, 1996]. However, there is more bare soil in shrubland, which is expected to yield more rapid evaporation [*Dugas et al.*, 1996]. Second, shrubs have deeper roots, permitting transpiration when surface soil moisture is limited [*Sala et al.*, 1992; *Pelaez et al.*, 1994; *Porporato et al.*, 2001]. Third, the phenology of C_4 grasses limits transpiration to the summer months whereas C_3 shrubs can transpire all year. Fourth, soil erosion in shrublands has modified soil texture [*Abrahams et al.*, 1995; *Kieft et al.*, 1998], yielding more spatially heterogeneous and deeper infiltration in this environment [*Bhark and Small*, 2003].

[6] Relatively few studies of ET have been completed in semiarid environments (Table 1); and only *Dugas et al.* [1996] focused on both grassland and shrubland. Surprisingly, the ET time series observed at adjacent grassland and shrubland sites at the Jornada Experimental Range of southern New Mexico were similar [*Dugas et al.*, 1996]. However, no explanation was provided for how the ET time series could be so similar, given the dramatic differences

Table 2. Primary Differences Between the Grassland and the Shrubland

Site Differences	Grass	Shrub
Percent cover	60%	30%
Plant type	C_4	C_3
Root depth	shallow	deep
Pattern of infiltration	uniform	variable
Canopy interception	more	less

between the ecosystems studied [*Dugas et al.*, 1996]. *Stannard et al.* [1994] measured ET in the Sonoran desert of southern Arizona in both grassland and shrubland. However, they focused only on the shrubland and did not describe similarities and differences in ET between the two different ecosystems. Although the measurements were not from grassland and shrubland, ET data from the HAPEX-Sahel site in western Niger showed that time series of ET from savanna and woodland are also very similar [*Kabat et al.*, 1997].

[7] Although ET measurements from semiarid environments have been described previously [e.g., *Dugas et al.*, 1996], the data collected were not sufficient to quantify how shrubland invasion changes ET and to identify the physical mechanisms responsible for any observed changes. In addition, these studies do not constrain the relationship between soil moisture and ET, which may vary with vegetation cover. First, the data sets used in these earlier efforts were typically short, and therefore did not provide information regarding the range of conditions that exist (Table 1). Second, the data required to identify the controls on temporal variations in ET were not collected. In particular, soil moisture at multiple depths and the components of the radiation budget were not measured continuously (Table 1). For example, *Gash et al.* [1991] conducted a 6 week ET study in a mixed grass-shrub environment, but only made gravimetric soil moisture measurements 6 times. Furthermore, the full suite of necessary measurements was not collected from different ecosystems, to identify the controls of vegetation on ET.

[8] Here we compare water and energy cycling in semi-arid grassland and shrubland at the Sevilleta National Wildlife Refuge (SNWR) of central New Mexico (Table 2 and Figures 1 and 2). The native grassland site is only 2 km from the shrubland site, where Creosotebush has recently invaded grassland. Because these sites are close together, differences in precipitation, radiation, and other hydroclimatological factors are minimized. This allows for a direct comparison of temporal variations in ET, and the factors that control these variations, between grassland and shrubland. Our study differs from previous studies in three key ways, and therefore offers new and important information concerning ET in semiarid environments. First, we provide a record long enough to quantify temporal fluctuations in ET. Second, we directly compare ET from semiarid grassland and shrubland ecosystems, allowing us to quantify how shrub invasion can influence hydrology, climate, and carbon cycling. Third, our ET measurements are accompanied by continuous measurements of soil moisture and radiation, allowing us to identify the key controls on ET. In particular, we test if the simple relationship between ET and soil moisture used in numerous previous studies is reasonable [e.g., *Laio et al.*, 2001].

2. Site Description

[9] Data were collected from the McKenzie Flats area of the Sevilleta National Wildlife Refuge (SNWR), central New Mexico. We collected measurements from a grassland and a shrubland which are separated by 2 km (Figure 1). The grassland is nearly monospecific, dominated by Black Grama (*Bouteloua eriopoda*) which covers about 60% of the ground surface (Figure 2). The shrubland site is also monospecific, dominated by Creosotebush (*Larrea tridentata*) which covers about 30% of the ground surface (Figure 2). At both measurement sites, the vegetation is uniform in all directions over a distance of several hundred meters. Livestock grazing has not been permitted at the Sevilleta NWR since the 1970s.

[10] Average annual precipitation in this area is 230 mm with more than half of the precipitation falling during the monsoon season (July–September). Daily rainfall rarely exceeds 15 mm. At both sites, the surface slope is less than 2 degrees. The top ~40 cm of soil is a sandy loam and the measured saturated hydraulic conductivity is high,
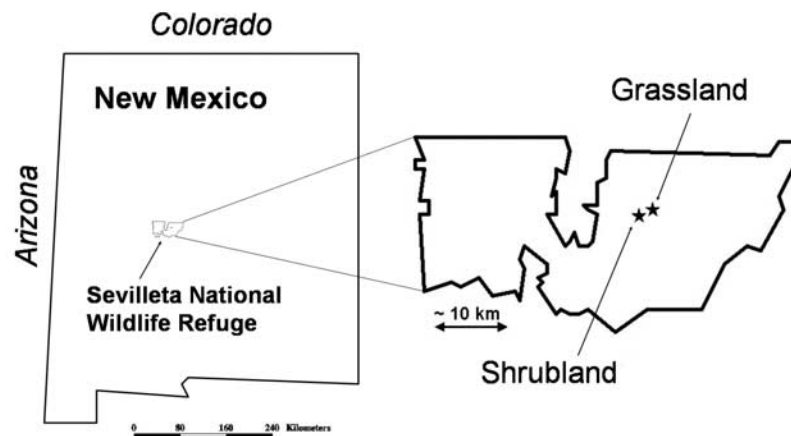

Figure 1. Location of the shrubland and grassland field sites in the Sevilleta National Wildlife Refuge, central New Mexico.
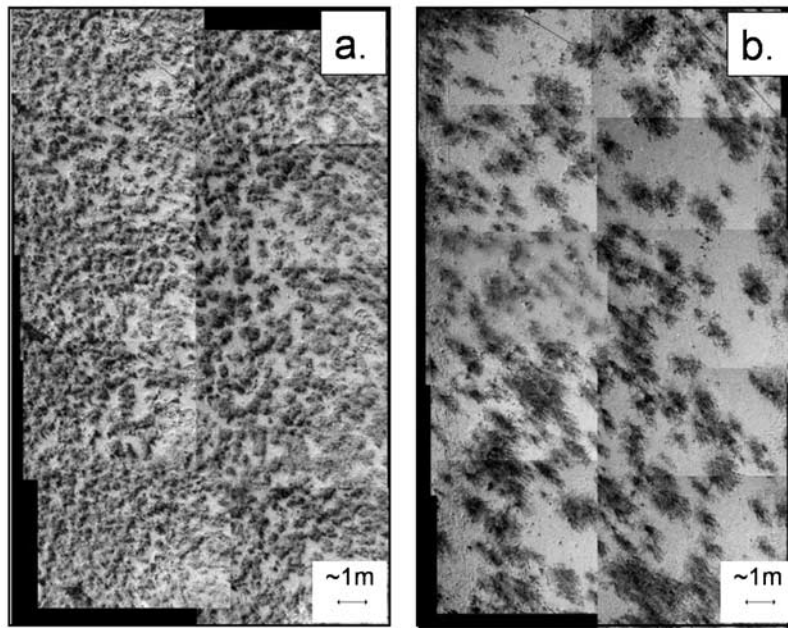
Figure 2. Overhead photo surveys (25 m × 15 m) of (a) grassland and (b) shrubland taken in January of 2002. The grassland has ~60% cover, and the shrubland has ~30% cover.

100–200 cm day^{-1} [*Bhark*, 2002]. As a result, there is very little runoff from either site, and nearly all rainfall infiltrates into the soil and is available for ET [*Bhark and Small*, 2003].

[11] Given that the grass-shrub ecotone is so narrow at the McKenzie flats research site, differences in the other factors that influence ET, e.g., rainfall, incident shortwave radiation, and soil type are minimized. Therefore we assume that vegetation is the primary source for differences in the water and energy balances between the two locations. As we demonstrate in Table 2, vegetation dictates the percent cover (Figure 2), root density profiles (Figure 3), and pattern of infiltration [*Bhark and Small*, 2003] at both the grass and the shrub sites.

3. Methods

[12] The instrumentation used in this study includes a Bowen Ratio system and typical micrometeorological devices at both the grassland and shrubland sites [*Shuttleworth*, 1993; *Moncrieff et al.*, 2000]. Precipitation, soil moisture, wind speed and direction, net radiation, and incident solar radiation were also measured. Here we present data from 1 June to 15 September for three years (2000–2002), intervals during which nearly complete data sets were collected at both sites. The only long data gap exists in the grassland record, starting on 5 August 2002, and we exclude data from both sites after this date. Measurements were made every 30 seconds and averaged to 30 minutes. The term "daily average" indicates that an average was made over an entire 24 hour day, beginning at midnight (local time). We report daily average values of ET and precipitation, as these are the basic components of the water budget. The term "midday average" indicates that an average was calculated from 10:00 am to 2:00 pm (local time). We report midday averages of evaporative fraction (EF; defined below) and components of the radiation

budget to show how soil moisture availability influences the partitioning of the surface energy budget.

3.1. Measurements and Instrumentation

3.1.1. Radiation

[13] The components of the surface radiation budget were measured using identical radiation and energy balance systems (REBS) instruments at both the grass and shrub sites, with radiometers placed 3.5 m above the soil surface. Incident (SW_d) and reflected (SW_u) shortwave radiation were measured using double-sided pyranometers. Downward and upward total radiations were measured using total hemispheric radiometers (THRs). Downward (LW_d) and
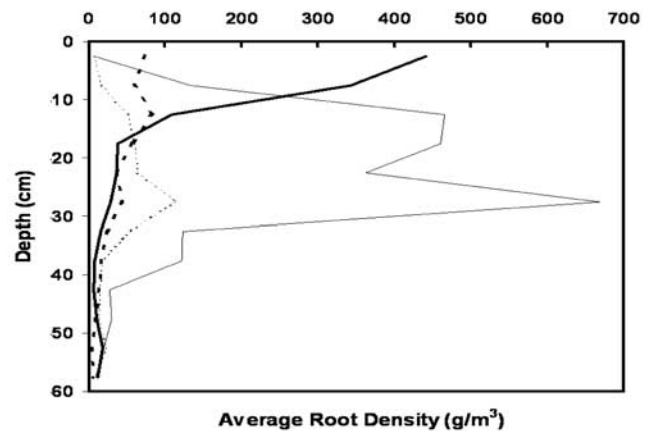

Figure 3. Root density (g m^{-3}) at the shrubland (thin lines) and the grassland (thick lines), measured beneath plant canopies (solid line) and bare soil (dashed line). Each line represents the average value from three separate profiles. Fine roots exist down to at least 1 m, but the density was below ~10 g m^{-3} in all samples.

upward (LW_u) longwave radiation were calculated by subtracting measured shortwave radiation from the corresponding THR measurements (i.e., $LW_d = THR_d - SW_d$). Total net radiation (R_n) was calculated using:

$$THR_d - THR_u = Rn \qquad (1)$$

All radiometers are equipped with built in ventilators so that corrections for wind speed variations were not necessary.

[14] Because of a miswired instrument at the grassland site, the downward and upward longwave radiation components were not recorded, although the net longwave measurement was not effected. We calculate LW_u at the grass site using net longwave from the grass site and downward longwave from the shrub site (i.e., $LW_{u,grass} = LW_{net,grass} - LW_{d,shrub}$). The accuracy of LW_u at the grassland depends on the assumption that downward longwave radiation does not vary significantly over a horizontal distance of $\sim$2 km, relative to variations in emitted longwave radiation.

3.1.2. Soil Heat Flux and Soil Moisture

[15] At both sites, ground surface variables (soil heat flux, soil moisture, and soil temperature) were measured 5 cm beneath the surface at (1) a plant canopy patch and (2) an upslope bare soil patch. Site-averaged values were calculated by weighting the canopy and bare surface measurements by the percent cover at each location. For example, the site-averaged volumetric water content (θ) is:

$$\theta = f\theta_c + (1 - f)\theta_b \qquad (2)$$

where f is the fraction of canopy cover and the subscripts c and b refer to canopy and bare. In desert ecosystems, spatial variability of soil heat flux is substantial. Therefore multiple canopy–bare soil pairs of soil heat flux measurements increase the accuracy of site-averaged soil heat flux values [Stannard et al., 1994; Kustas et al., 2000]. Results from these past studies suggest that using only a single canopy–bare soil pair yields errors of roughly 25%, equivalent to 25–40 W m^{-2} at midday at the sites studied here. These errors are discussed in more detail below.

[16] The soil heat flux was measured using a combined calorimetric-heat flux plant approach [Kimball et al., 1976]. Soil moisture was monitored at the same canopy and bare locations as soil heat flux and temperature. Volumetric water content was measured using Campbell Scientific water content reflectometers (WCR) at three soil depths: 2.5, 12.5, and 22.5 cm. Probes were inserted horizontally in the upslope direction from a shallow pit that was subsequently filled. The WCRs have two 30 cm rods spaced 3.2 cm apart. A probe with this geometry samples a semi-cyclindrical region around the rods, where 90% of the signal is derived from soil between 2.5 cm above and below the rods [Ferre et al., 1998]. Therefore a probe inserted at 2.5 cm effectively provides an estimate of volumetric water content in the top 5 cm of the soil, θ_{0-5}. Likewise, the deeper probes (12.5 and 22.5 cm) are considered to provide estimates of soil moisture at depths of 10–15 and 20–25 cm. We used the factory calibration for the WCRs, which was completed using low-salinity, sandy soils like those in our study area. The WCR measurements were not adjusted for soil temperature. The WCR values are consistent with synchronous, nearby gravimetric and TDR estimates of soil moisture.
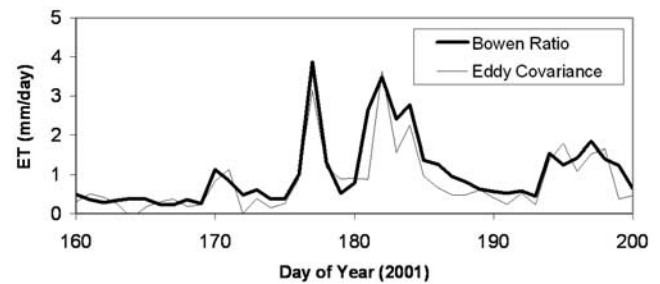

Figure 4. Comparison of daily ET measurements from colocated Bowen ratio and eddy covariance systems, for day of year (DOY) 160–200 in 2001. See Figure 6 for associated rainfall record.

[17] We use the average soil moisture as an estimate of root zone soil moisture, θ_{RZ}. Roughly one third of the shrub roots are at a depth greater than that represented by θ_{RZ} (Figure 3). As discussed below, the soil in the bottom of the root zone remains dry for most of the observation period, so θ_{RZ} would usually be lower if measurements had been taken from deeper in the soil profile.

3.1.3. Sensible and Latent Heat Fluxes

[18] Sensible and latent heat fluxes, H and λE, were calculated using the Bowen ratio method [Shuttleworth, 1993; Moncrieff et al., 2000]. We use REBS temperature and humidity probes to measure air temperature and vapor pressure at two heights above the surface, i.e., 0.5 m and 2 m in the grassland and 1 m and 2.5 m in the shrubland. Because the absolute magnitude of the temperature or vapor pressure difference is sometimes small compared to the bias in the instruments, we installed an automated exchange mechanism (AEM) at both the grassland and shrubland. The AEM exchanges the temperature and humidity probes every 15 minutes, thus eliminating instrument bias between the probes [Fritschen and Simpson, 1989]. At sunrise and sunset, the sensible and latent heat fluxes are opposite in sign and nearly equal, yielding a Bowen ratio (β) close to -1. Measurements from times when $-1.3 < \beta < -0.7$ are excluded because sensible and latent heat fluxes cannot be calculated using the Bowen ratio method during these intervals [Dugas et al., 1996; Unland et al., 1996]. For our study, this excludes about 10% of the measured data.

[19] Flux measurements made using the Bowen ratio method have been compared to those using the more direct eddy covariance method [Shuttleworth, 1993; Moncrieff et al., 2000] at both sites. The two methods yield similar values for daily and midday averages of both ET and evaporative fraction (Figure 4), confirming that the Bowen ratio data used here are reasonable. In general, the Bowen Ratio estimates of ET are higher by more than 10% as expected given eddy covariance closure problems [e.g., Unland et al., 1996; Thompson et al., 1999; Liu and Foken, 2001; Zhu et al., 2003]. The results of this comparison are similar to results from studies performed at other locations [Kabat et al., 1997].

3.2. Available Energy, Evaporative Fraction, and Potential ET

[20] Throughout this study we refer to the terms available energy (Q_a) and evaporative fraction (EF). Q_a is the total
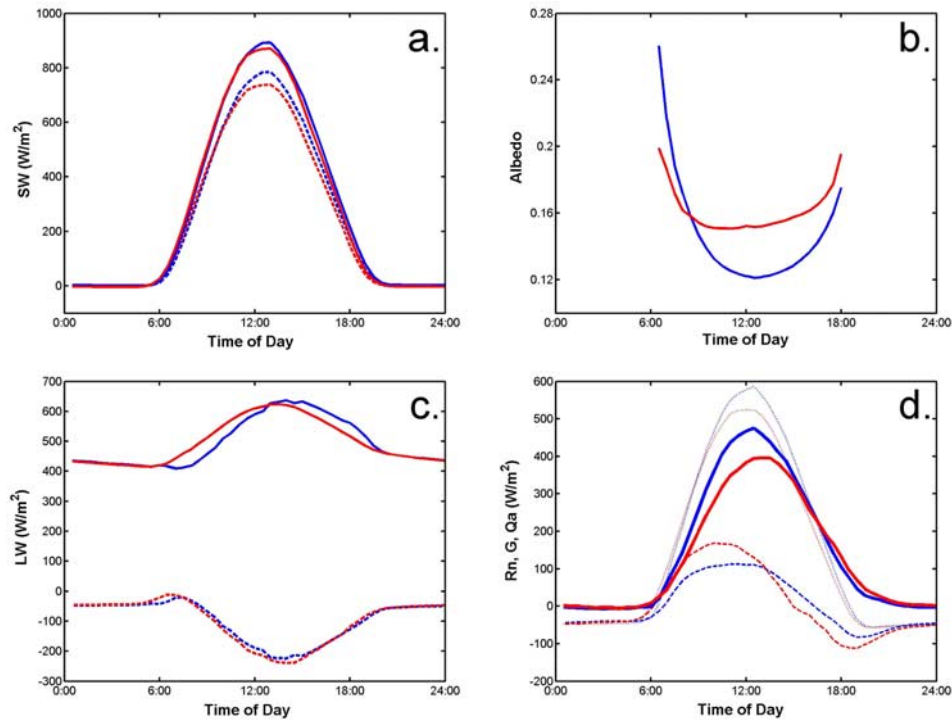
Figure 5. Average daily cycle of the surface radiation budget and soil heat flux. The blue lines represent the grassland, and the red lines represent the shrubland. (a) Net short wave radiation (dashed line) and incident shortwave radiation (solid line); (b) albedo; (c) net long wave radiation (dashed line) and long wave up (solid line); (d) net radiation (solid line), soil heat flux (dashed line), and available energy (dotted line).

energy available for the sum of latent and sensible heat, and is equal to:

$$Q_a = Rn - G \qquad (3)$$

EF is the ratio of the latent heat flux to Q_a. We calculate EF directly from temperature (dT) and vapor pressure gradient (de) information using the Bowen ratio, β:

$$EF = \frac{1}{1 + \beta}. \qquad (4)$$

where $\beta = \gamma dT/de$ and γ is the psychrometric constant [*Shuttleworth*, 1993].

[21] We calculated daily values of potential evapotranspiration (PET) from the 30-minute data sets using the Penman-Monteith equation as described by *Shuttleworth* [1993]. Different values of stomatal resistance and roughness length were used for grassland and shrubland, 165 and 179 s m^{-1} and 0.04 and 0.08 m, respectively. These values were set according to the Land Data Assimilation System database (http://ldas.gsfc.nasa.gov/LDAS8th/MAPPED. VEG/web.veg.table.html). The estimated PET was used to normalize the measured ET, yielding a value of ET/PET for each day in the record. ET/PET was then used to assess whether or not temporal variability in PET influenced the apparent relationship between ET and soil moisture.

3.3. Calculation of Timescales

[22] We calculate the time intervals over which soil moisture is depleted and ET and EF decrease following rainfall events following the method of *Hunt et al.* [2002].

Hunt et al. [2002] modeled the decrease in EF after rainfall as exponential through time, using EF data from New Zealand. We use the same method here, applied to our measured time series of soil moisture, EF, and ET. For example, the decrease of volumetric water content, θ, following rainfall events is modeled as:

$$\theta(t) = \left(\theta_1 - \theta_f\right)e^{(-t/\tau)} + \theta_f \qquad (5)$$

where t is time in days since the rainfall event, θ_1 is the water content observed on the first day following rainfall, θ_f is the final water content at the end of the drydown sequence, and τ is a best fit exponential time constant. We calculated τ using θ, EF, and ET data from intervals following large storms (>8 mm daily total) until the next measured rainfall (>2 mm daily total). We selected 8 mm as the cutoff so that only storms that yielded substantial EF and ET responses were included in the calculation. Higher and lower cutoff values yielded similar values of τ. The time series used were limited to 12 days, as longer dry intervals were uncommon. A similar approach was used by *Scott et al.* [1997] and *Lohmann and Wood* [2003] to evaluate the characteristic timescales of the evaporation response in land surface models.

4. Results

4.1. Net Radiation, Soil Heat Flux, and Available Energy

[23] When averaged over midday, Q_a is higher at the grassland than at the shrubland by 72 W m^{-2} (Figure 5 and Table 3), a difference of more than 20% of the total Q_a in

Table 3. Midday (10:00 A.M. to 2:00 P.M.) Values of Net Radiation (R_n), Soil Heat Flux (G), Available Energy (Q_a), Incident Short Wave Radiation (SW_d), Net Short Wave Radiation (SW_n), Emitted Long Wave Radiation (LW_u), and Net Long Wave Radiation (LW_n) at Both Grass and Shrub

	Grass	Shrub	Grass − Shrub
Rn, W m^{-2}	539	494	45
G, W m^{-2}	104	132	−28
Q_a, W m^{-2}	434	362	72
SW_d, W m^{-2}	823	814	9[a]
SW_n, W m^{-2}	721	690	30
LW_u, W m^{-2}	583	600	−17
LW_n, W m^{-2}	−182	−196	14
Albedo	0.12	0.15	−0.03

[a]The instrument bias is as large as the measured difference between the two sites.

either environment. Two factors account for this difference. First, average midday R_n is 45 W m^{-2} higher at the grassland site than at the shrubland site. This difference is nearly 10% of the average midday R_n at either site and is greater than the instrument error at midday [*Field et al.*, 1992; *Hodges and Smith*, 1997]. The difference in R_n is equal to the sum of the differences in net longwave (LW_n) and net shortwave (SW_n) radiation (Table 3). LW_n is less negative at the grassland site by ∼15 W m^{-2}. Given the wiring problem described in section 3, we cannot confirm that the measured difference

in LW_n is solely due to differences in longwave emitted, as calculated in Table 3. Differences in downward longwave may also contribute. Midday net shortwave radiation is 30 W m^{-2} higher at the grassland site. Within measurement error of the pyranometers, average midday SW_d is the same at the grassland and shrubland sites. Therefore we attribute the entire SW_n difference to the lower measured surface albedo at the grassland site, a difference of 3%.

[24] The second source of the Q_a difference is a result of contrasts in G. Differences in daily averaged G are negligible (<5 W m^{-2}), which is expected given that the day-to-day fluctuations of thermal energy storage in the soil are small. However, G averaged during midday is nearly 30% greater at the shrubland (∼30 W m^{-2}) (Table 3), with the largest differences occurring before noon (Figure 5). This difference is of similar magnitude to the maximum errors associated with using only a single canopy–bare soil pair of soil heat flux measurements at each site [*Stannard et al.*, 1994; *Kustas et al.*, 2000]. However, the differences reported here are reasonable given the more extensive bare soil in the shrubland. The combination of higher midday R_n and lower midday G at the grassland site results in higher Q_a throughout most of the day.

4.2. Midday Evaporative Fraction

[25] At both the shrubland and grassland, midday evaporative fraction (EF) varies considerably throughout the three monsoon seasons studied (Figure 6). The variability
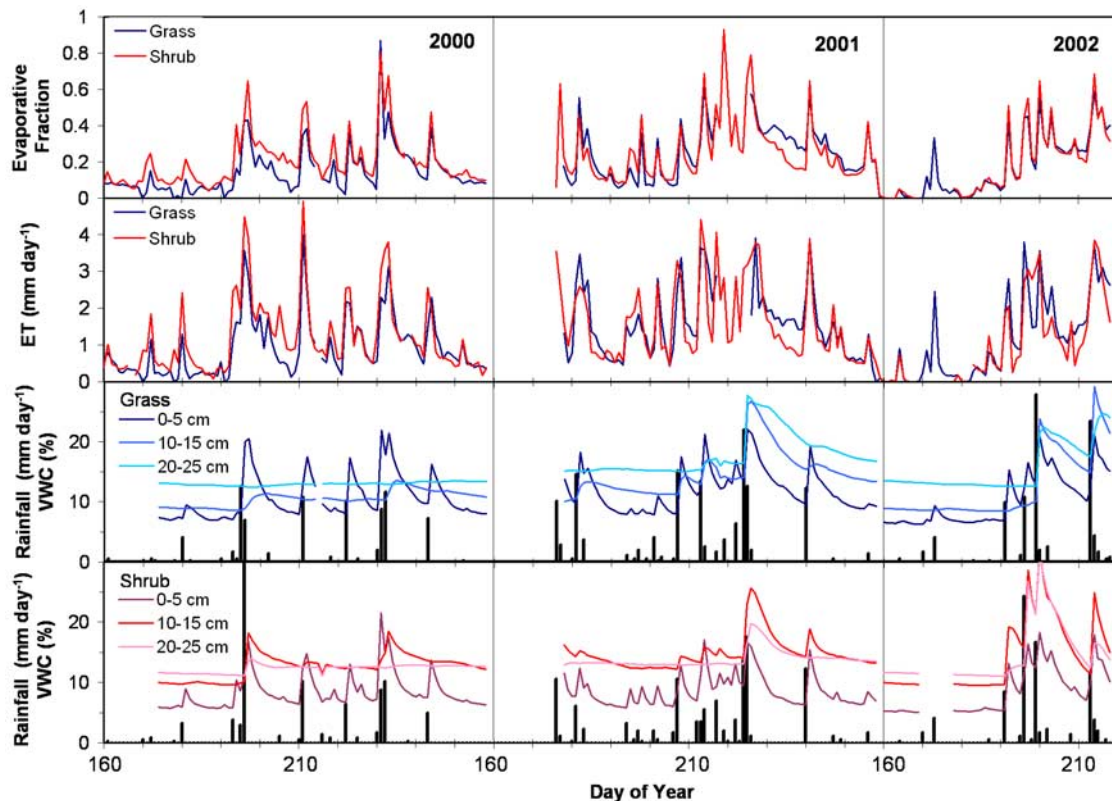

Figure 6. Daily time series of evaporative fraction (midday), evapotranspiration (daily total), volumetric water content (0−5 cm, 10−15 cm, and 20−25 cm), and precipitation (bars) at the grassland and shrubland sites. For ET and EF, grass is represented by a blue line, and shrub is represented by a red line. In third and fourth panels, grassland (blue) and shrubland (red) plots are separate for water content and precipitation.
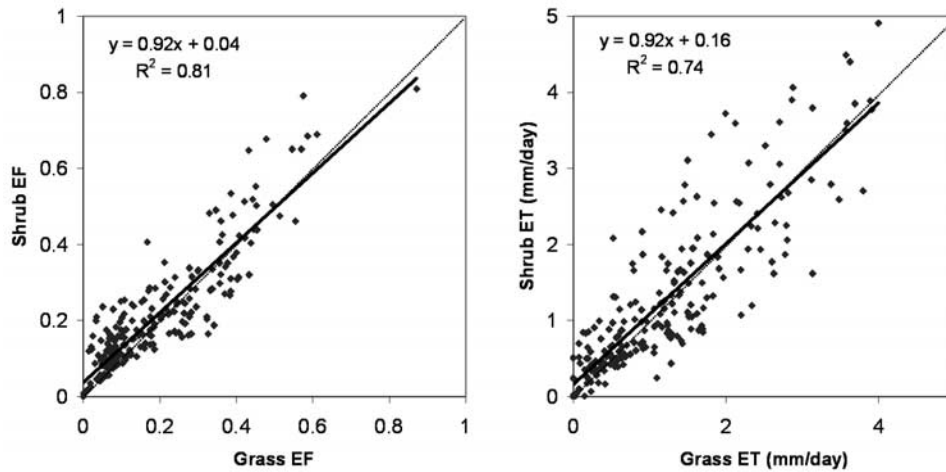
Figure 7. Scatterplot of (a) midday evaporative fraction and (b) daily total evapotranspiration at the grass and shrub sites. Straight line is best fit line via linear regression: $d\mathrm{EF}_{shrub}/d\mathrm{EF}_{grass} = 0.92$ and $d\mathrm{ET}_{shrub}/d\mathrm{ET}_{grass} = 0.92$.

is very similar between the two sites. Peak EF values are approximately 0.8 at both sites, observed following the largest rainfall events (e.g., day 230 in year 2000) with local maxima following smaller rainfall events. In both the grassland and the shrubland, EF decreases to about 0.1 within a week as the soil dries out, unless there is another rainfall event. Before the onset of monsoon rainfall in July (day 180), even lower EF values ($\sim$0.05) are observed.

[26] We quantify the similarity between the grassland and shrubland EF time series using linear regression (Figure 7, left). The plot of EF at the shrubland (EF_{shrub}) versus EF at the grassland (EF_{grass}) has only little scatter ($r^2 = 0.81$). Some of the scatter that does exist is due to different rainfall inputs at the two sites. Although the rainfall records are very similar (Figure 6), there are differences in daily rainfall totals for some of the events. For example, on day 195 of 2000, 15 mm more rainfall was measured in the shrubland than in the grassland, which resulted in a higher EF over the subsequent days at the shrubland. The slope of the regression (0.92) primarily reflects the fact that EF is greater at the

shrubland than at the grassland on days when EF is low at both sites.

[27] Midday EF is strongly, linearly related to volumetric water content in the top 5 cm of the soil (θ_{0-5}) at both sites (Figure 8). Regressions of θ_{0-5} versus midday EF yield an r^2 value of 0.83 at the shrubland site and 0.80 at the grassland site (Table 4). The r^2 values calculated using only clear-sky days are very similar. This demonstrates that θ_{0-5} explains 80% or more of the day-to-day variations in EF at both sites, on both clear and cloudy days. The slope of the regression line ($M_{EF} = d\mathrm{EF}/d\theta$) for the shrubland ($M_{EF-shrub} = 4.7$) is 25% greater than the slope at the grass site ($M_{EF-grass} = 3.6$). This difference in slope is significant at the 99% confidence interval. Therefore, although EF varies linearly with θ_{0-5} at both sites, EF is more sensitive to variations in θ_{0-5} at the shrubland than at the grassland.

[28] A significant positive, linear relationship also exists between EF and soil moisture in deeper soil layers at both sites (Table 4). However, this relationship is not nearly as clear as with surface soil moisture and EF. For example, the
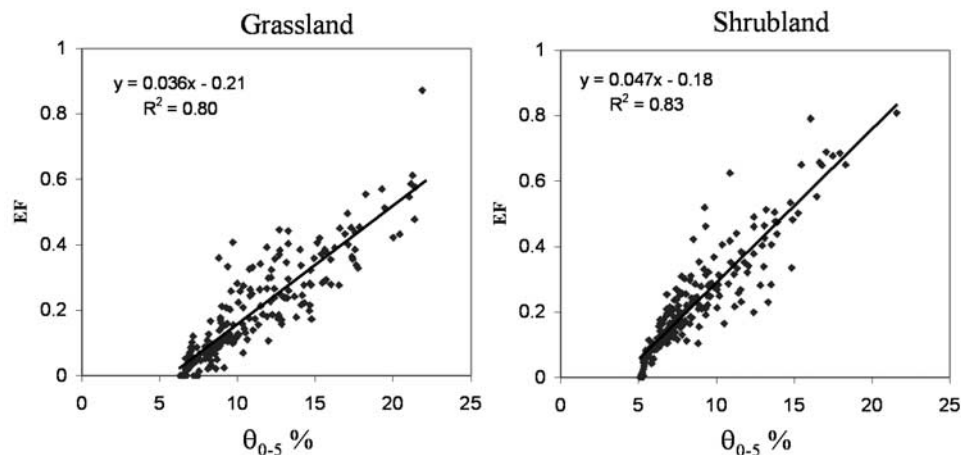

Figure 8. Midday evaporative fraction (EF) versus surface soil (0−5 cm) volumetric water content for (a) grass and (b) shrub locations.

Table 4. Slopes for Scatterplots of EF and ET Versus Volumetric Water Content ($M_{EF} = dEF/d\theta$, $M_{ET} = dET/d\theta$) at Various Depths for Both Shrub and Grass Locations[a]

	Grass EF		Shrub EF		Grass ET			Shrub ET		
	Slope	r^2	Slope	r^2	Slope	r^2	r^2 (No Rain)	Slope	r^2	r^2 (No Rain)
Top	3.63 ± 0.24	0.80	4.70 ± 0.29	0.83	20.4 ± 2.1	0.63	0.74	24.8 ± 3.2	0.53	0.64
Middle	2.50 ± 0.37	0.45	2.70 ± 0.44	0.41	12.6 ± 2.7	0.28	0.39	13.8 ± 3.3	0.25	0.41
Bottom	2.48 ± 0.48	0.32	2.37 ± 0.69	0.18	11.9 ± 3.3	0.18	0.13	11.9 ± 4.8	0.10	0.20
All	3.52 ± 0.38	0.59	2.75 ± 0.46	0.40	18.2 ± 3.0	0.39	0.50	14.1 ± 3.4	0.24	0.36

[a]The 99% confidence intervals for these slopes are also displayed; r^2 values from the linear fit are shown for all days and for days without rainfall in the case of ET.

r^2 values between soil moisture at a depth of 22.5 cm and EF are only 0.32 and 0.18, at the grass and shrub sites respectively. The slope of the regression line ($M_{EF} = dEF/d\theta$) also decreases with depth at both sites. The relatively limited predictive value of deeper soil moisture for EF is expected given that wetting fronts typically do not propagate below 10 cm in these environments (Figure 6) [e.g., *Bhark and Small*, 2003]. When root zone soil moisture is used in the regression, the r^2 and slope values are intermediate between those derived using the surface and deeper soil layers.

4.3. Evapotranspiration

[29] Overall, daily ET behaves in a similar fashion to midday EF (Figure 6). Maximum ET (4 mm d^{-1}) is observed immediately follow rainfall events. Then, ET decreases to relatively low values (0.5 mm d^{-1}) within a week, unless there is additional rainfall. As was the case with EF, the time series of ET from the grassland and shrubland are very similar (Figure 6). The regression between ET at the two sites has somewhat more scatter ($r^2 = 0.74$) than the equivalent regression for midday EF ($r^2 = 0.81$) (Figure 7). Still, the temporal fluctuations are very similar at the two sites. The rainfall that accumulates during each storm may be higher at either of the sites, which is likely the primary source of the scatter in Figure 7 (right), as was the case for EF. The additional scatter in the ET plot could be related to a number of factors, for example, (1) dissimilar variations in Q_a through time at the two sites (2) or errors related to measuring temperature and humidity gradients throughout the day for ET [*Perez et al.*, 1999], versus only at midday for EF.

[30] Although the rainfall total on any day may differ between the sites, the total rainfall during the three monsoon seasons examined is very similar at the grassland and shrubland sites, 313 and 294 mm respectively. During the same interval, we measured a total of 273 mm of ET at the grassland site. We estimate the actual ET total to be 295 mm, after the 10 day data gap in 2001 is filled (Figure 6). We estimate ET during this interval to be about 20 mm (or 2 mm d^{-1}), based on the following two proxies: (1) the soil moisture measured at the grassland site and the relationship between soil moisture and ET (19 mm) (see Figure 9); and (2) the ET measured at the shrubland site during the same interval (22 mm). At the shrubland site, we measured 306 mm of ET during the 3 season interval, and estimate the total to be 310 mm once the data gap around day 170 in 2002 is filled. Given reasonable measurement errors for ET and precipitation, these water balance calculations

show that (1) ET is equal to rainfall at both sites; and (2) the total fluxes (P or ET) are equal at the two sites. The former is expected because there is no runoff or recharge from either site and changes in soil moisture storage calculated from the soil moisture profiles are small (<20 mm). The latter is expected given the close proximity of the two sites.

[31] Daily ET increases with surface soil moisture (θ_{0-5}) at both sites (Figure 9). However, the relationship is not nearly as clean as between midday EF and θ_{0-5}. When all days in the record are included, regressions of ET versus θ_{0-5} yield r^2 values of 0.63 and 0.53 at the grassland and shrubland, respectively (Table 4). Many of the outliers in the relationship are from rainy days, when daily ET is high and soil moisture is relatively low compared to other days in the record. These outliers are an artifact of how the comparison was completed. Rainstorms typically occur during the late afternoon at the field sites. Daily ET is high on rainy days because the soil and plant surfaces are wet and ET is very rapid following the storm. However, daily averaged soil moisture is low because the soil was dry for most of the day. When rainy days are excluded, the linear relationship between ET and θ_{0-5} is more obvious, with r^2 values of 0.74 and 0.64 at the grassland and shrubland, respectively. As is the case for EF, the slope of the regression line ($M_{ET} = dET/d\theta$) is greater for the shrubland ($M_{ET-shrub} = 0.25$) than for the grassland ($M_{ET-grass} = 0.21$) (Table 4), with or without rainy days.

[32] Daily ET does not clearly increase with higher root zone soil moisture (θ_{rz}), as it does with surface soil moisture (Figure 9 and Table 4). In the grassland, daily ET tends to be higher on days when θ_{rz} is relatively high, particularly if rainy days are excluded. However, there is considerable scatter in the relationship ($r^2 <= 0.5$). The relationship is even less clear in the shrubland ($r^2 <= 0.36$) where a linear relationship does not fit the data well. At both sites, the observations fall within a broad zone: ET may be high or low when θ_{rz} exceeds ~10%. As was the case for EF, the correlation between ET and soil moisture at 12.5 or 22.5 cm is limited at both sites (Table 4).

[33] In Figure 10, we show how measured ET normalized by the PET calculated for each day (ET/PET) varies with soil moisture. The relationship between ET/PET and soil moisture (Figure 10) is similar to that between ET and soil moisture (Figure 9). ET/PET increases with surface soil moisture (θ_{0-5}) at both sites. In contrast, there is no consistent relationship between ET/PET and root zone soil moisture (θ_{rz}), particularly in the shrubland. As expected, *x* is typically greater on rainy days: ET is relatively high following storms while PET tends to be low because it is
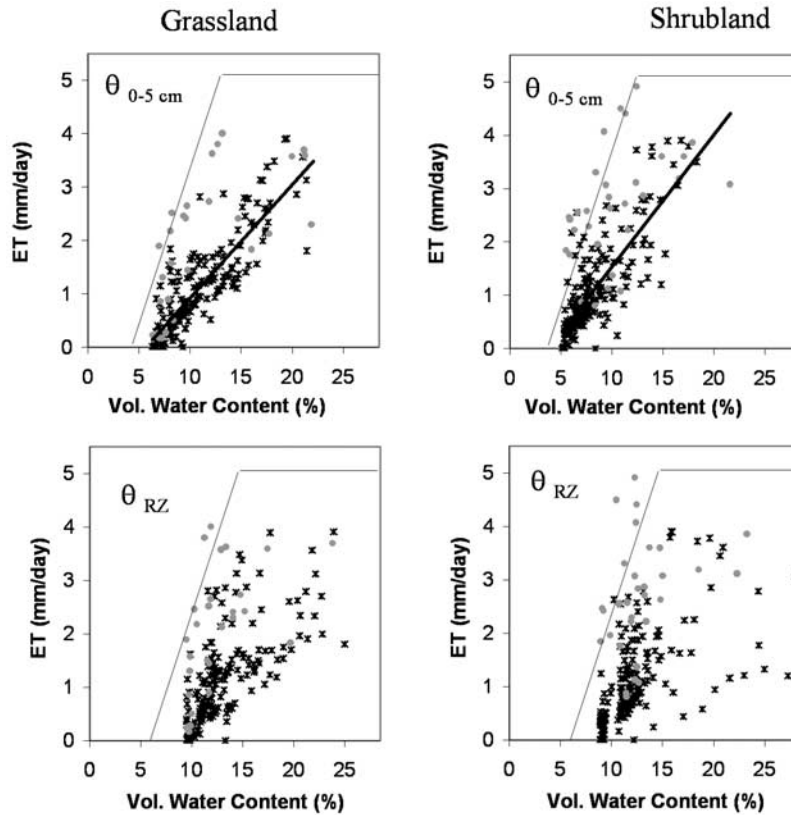
Figure 9. Daily ET versus volumetric water content for (left) grass and (right) shrub locations. (top) Comparison of ET to $0-5$ cm soil moisture. (bottom) Comparison to root zone averaged soil moisture. Days with rainfall (>2 mm) are represented by the shaded circles, and other days are represented by asterisks. The regression (solid line) is calculated using data from nonrainy days and is only shown for the $0-5$ cm plots (top) where a linear fit is reasonable. The thin shaded lines are envelopes fit by eye, fashioned after the ET-θ relationship used by *Laio et al.* [2001].

cloudy. ET/PET is <0.5 on most other days. Because ET and PET do not positively covary at our sites, it makes sense that we found similar results using absolute or normalized values of ET (Figures 9 and 10).

4.4. Timing of Decreases in Soil Moisture, EF, and ET

[34] Surface soil moisture, EF, and ET all decrease rapidly following rainfall events at both the grassland and shrubland sites (Figure 11): the best fit exponential time constants (τ) are less than three days in all cases. The exponential model fits the surface soil moisture time series well and yields values of $\tau_{\theta} = 2.8$ days and $\tau_{\theta} = 2.5$ days for grass and shrub, respectively. The timescales for EF and ET decreases are slightly shorter than for θ_{0-5}. We found that τ_{EF} is 2.0 days for the grassland and 1.8 days for the shrubland, and that τ_{ET} is 2.1 days for the grassland and 1.9 days for the shrubland. The exponential curves do not fit the ET and EF data as closely as they fit the water content data. The decreases in EF and ET between days 1 and 2 are more rapid than the exponential model. Following day 4, the observed decreases in EF and ET are slower than that predicted with the exponential model, particularly in the shrubland. A two-component model may fit the EF and ET data more closely, with the initially rapid decrease representing direct evaporation from bare soil and plant surfaces [e.g., *Scott et al.*, 1997; *Lohmann and Wood*, 2003].

[35] The average time series of θ at the deeper soil depths (12.5 and 22.5 cm) and averaged throughout the root zone (θ_{RZ}) are also plotted in Figure 11 (top). The decreases in θ_{RZ} clearly reflect the rapid decline in surface soil moisture following rainfall, given the much smaller declines in θ at 12.5 and 22.5 cm. We did not use the same approach to quantify the rate of soil moisture drying as for the surface soil for several reasons. First, most storms did not wet the soil to depths greater than 10 cm (Figure 6), so the average time series included many samples when soil moisture was not actually decreasing. In addition, including only the cases when the deeper soil was wetted yielded only several drydowns, given the constraints on storm size described in section 3. Second, in cases when deeper soil was wetted, there is often a gradual increase followed by a slow decrease (e.g., DOY 195 in 2000 at the grassland), so the exponential model was not appropriate. As expected, the deeper soil clearly dries out more slowly than the surface soil.

5. Discussion and Conclusions

5.1. Similarities in ET Between Grassland and Shrubland

[36] *Dugas et al.* [1996] showed that ET in grassland and shrubland were similar, although they did not quantify the similarities as in section 4. Consistent with the work of
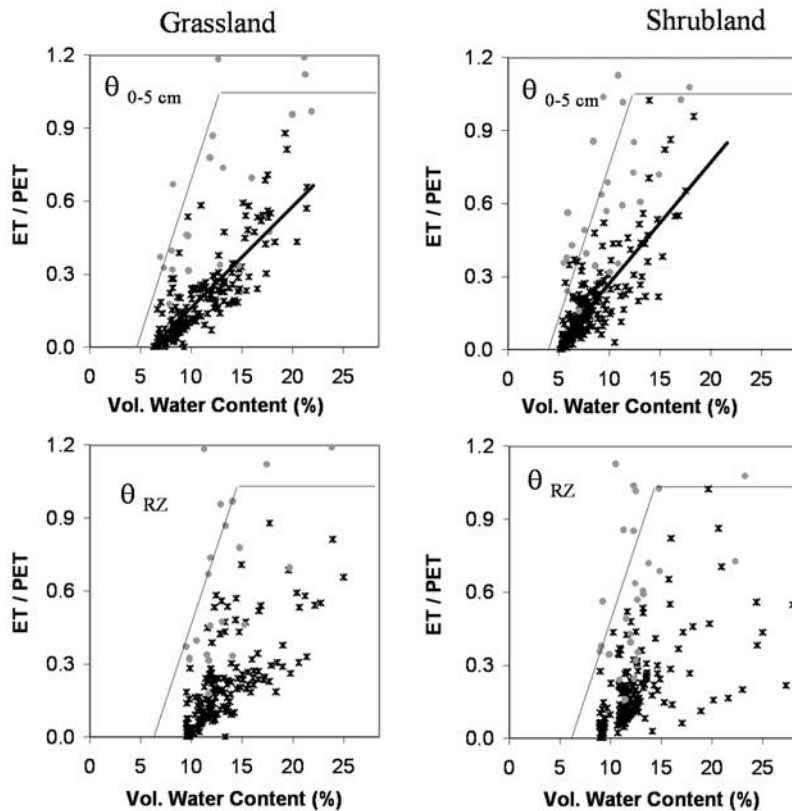
Figure 10. Same as Figure 9, except the ratio of ET to PET is plotted instead of the actual magnitude of ET.

Dugas et al. [1996], the ET time series are very similar at our grassland and shrubland sites. First, total ET is equal to precipitation throughout the monitoring interval, within measurement error. Second, the ET values vary from 0.5 to 4 mm d^{-1}. Third, ET declines rapidly from relatively high values after rainstorms to relatively low values within several days or a week. At both sites, the exponential time constants for decreases of θ_{0-5}, EF, and ET are all less than three days. This is rapid compared to that observed at other locations. For example, τ_{EF} is 6 days for a semiarid tussock grassland and more than 10 days for rye grass pasture [*Hunt et al.*, 2002]. Although equivalent time constants have not been calculated in most studies, we expect τ_{EF} from most ecosystems to be longer than that observed at the Sevilleta.

[37] Although *Dugas et al.* [1996] showed that grassland and shrubland ET were similar, they provided no explanation as to how ET from these two different ecosystems (Table 2 and Figures 2 and 3) could be so alike. Our data show that the similarities in ET are the product of two compensating differences. First, the observed midday Q_a is higher in the grassland by 70 W m^{-2}, equivalent to 20% of the total Q_a. This difference is of similar magnitude to the changes in Q_a that accompany land surface changes, like deforestation [*Gash and Nobre*, 1997], or extremes in soil moisture state in semiarid environments [*Meyers*, 2001; *Small and Kurc*, 2003].

[38] The primary source of the observed Q_a contrast is the difference in bare soil coverage between grassland and shrubland, 30% and 60% respectively (Table 2). More bare soil in shrubland enhances midday G because a lack of canopy shading yields higher skin temperature and stronger

temperature gradients near the soil surface [*Breshears et al.*, 1997; *Tuzet et al.*, 1997; *Breshears et al.*, 1998]. Our data shows that midday G is higher in the shrubland by 30 W m^{-2}, although this difference is based on only a single canopy–bare soil pair of soil heat flux measurements at each site. The higher skin temperature also leads to more longwave radiation being emitted in shrubland, by about 15 W m^{-2}. Finally, the more extensive bare soil in shrubland yields an albedo that is higher by 3%, resulting in SW_n that is lower by 30 W m^{-2}. A similar difference in albedo between grassland and shrubland has been observed at the Jornada Range [e.g., *Dugas et al.*, 1996; *Schlesinger and Pilmanis*, 1998; *Barnsley et al.*, 2000]. Similar contrasts in albedo between woody and herbaceous environments have been observed in the Sahel [*Nicholson et al.*, 1998].

[39] Alone, higher Q_a in the grassland would yield more ET at this site. However, in the grassland, EF is often lower and the sensitivity of EF to soil moisture is lower (Figures 6 and 8). The greater sensitivity of EF to surface soil moisture in the shrubland may be the product of rapid direct evaporation from bare soil in this environment (discussed below). The differences in Q_a and EF tend to cancel, yielding similar time series of ET. *Kabat et al.* [1997] found a similar result in a comparison of savanna and woodland in the Sahel. The woodland had higher Q_a but a lower EF; therefore a smaller portion of Q_a went to latent heating and ET in the savanna.

5.2. Shallow Soil Moisture and ET

[40] The observations of soil moisture, EF, and ET described above indicate that the surface soil layer (0–
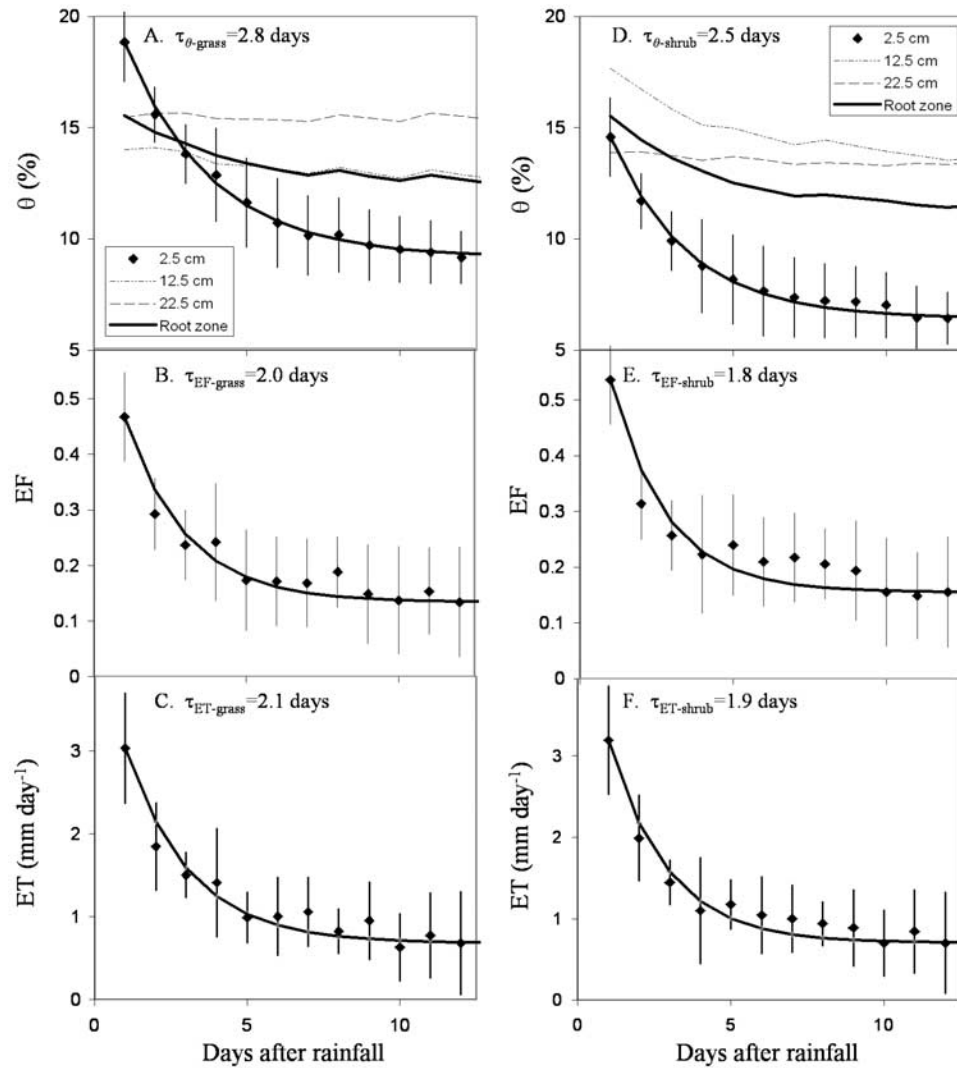
Figure 11. Volumetric water content in the top 5 cm of soil, midday evaporative fraction, and daily evapotranspiration content plotted against days since last rainfall at the (a, b, and c) grass and (d, e, and f) shrub locations. Points show average values as a function of days since last rainfall over 8 mm. Error bars show one standard deviation. Solid line shows exponential fit to data, and exponential time constants are included for each graph. For water content, average values for 12.5 cm and 22.5 cm soil depths and the average for the entire root zone are also included.

5 cm) is often the primary source of water for ET in semiarid grassland and shrubland, during the summer monsoon season. First, both EF and ET are strongly correlated with surface soil moisture, θ_{0-5} (Figures 8 and 9). In comparison, both EF and ET are only weakly correlated with soil moisture at greater depths (Table 4). The correlation between EF or ET and root zone soil moisture (θ_{RZ}) is stronger than for the individual deeper layers, but this likely reflects the overwhelming influence of θ_{0-5} fluctuations on variability of θ_{RZ}. Second, the exponential time constants for decreases of EF and ET following rainstorms are about 2 days, slightly shorter than the timescale over which θ_{0-5} tends to decline. The similarity of these timescales suggests drying of the surface soil is largely responsible for the rapid temporal fluctuations in ET. In contrast, the soil at greater depths is not wetted following most storms, and it dries much more slowly in the events when it is wetted.

[41] Our observations show that the monsoon season water cycles of the grassland and shrubland ecosystems examined here can often be characterized as follows. First, rainfall events are small: daily storm totals are less than 10 mm on 80% of rainy days during the monsoon season, according to the 70-year rainfall record from Socorro, NM. Second, the infiltration following storms only wets the top ~10 cm of the soil. Such shallow wetting fronts are expected given that storms typically occur when the soil is dry and storm totals are so small [*Sala et al.*, 1992; *Bhark and Small*, 2003]. Third, ET returns nearly all rainfall to the atmosphere on a timescale of several days. Fourth, the soil typically remains in a dry state for several days or longer before the next storm. During the monsoon season, the average interval between storms with rainfall >2 mm is 13 days, much longer than the timescale on which the soil dries out and ET decreases to relatively low values. When

this model applies, the surface soil layer (0–10 cm) is effectively the only reservoir in which water is temporarily stored following rainfall events.

[42] The time series in Figure 6 show that important exceptions to this shallow soil moisture model exist. Soil moisture increases below a depth of 10 cm during the largest storms, or when several storms occur over a short period. The storms on around DOY 225 in 2001 and DOY 200 in 2002 delivered more than 30 mm of rainfall. In response, soil moisture increased at all three monitored depths. Excluding the surface soil layer, the subsequent drying occurred over a period of weeks. The initial decrease in ET was still rapid following both events. However, daily ET did not fall below 1 mm d^{-1} for several weeks, in contrast to what was observed following smaller storms that did not wet the deeper soil layers. The importance of relatively large storms, and the resulting effects on soil moisture and ET, varies from year to year. The 2001 and 2002 monsoon seasons both had large storms. In contrast, soil wetting was primarily restricted to the surface soil throughout the entire monsoon seasons in 2000 or 2003 (not shown), as particularly large storms did not occur in these years, with the exception of DOY 195 in 2000 at the shrubland. The shallow soil moisture model is not appropriate during the cold season, as both potential and actual ET are much lower during the winter. This allows wetting fronts to propagate deeper into the soil, particularly in wet years [Scott et al., 2000].

5.3. Partitioning of ET to Evaporation Versus Transpiration

[43] The dominance of surface soil moisture as the source for ET suggests that the contributions from evaporation and transpiration are different in grassland and shrubland during the monsoon season, even though the times series of ET are very similar. At both locations, ET and EF are very strongly correlated with surface soil moisture (θ_{0-5}) but show little dependence on soil moisture at greater depths (Table 4). In addition, the soil is not wetted to depths greater than 10 cm in most storms. In the shrubland, very few roots are found between the surface and a soil depth of 10 cm (Figure 3). Therefore most of the ET at the shrubland must be the result of direct evaporation from the soil (E). The portion of ET from direct evaporation in the shrubland is also enhanced by the extensive bare soil in this environment. In contrast, root density in the grassland is greatest in the top 5 cm of soil (Figure 3). Because ET is strongly related to θ_{0-5}, the possibility of substantial transpiration (T) by the grasses cannot be excluded. In addition, the more extensive canopy coverage in the grassland shades the soil surface, lowering soil temperature and slowing evaporation from the soil. In both ecosystems, the portion of ET from transpiration is likely higher following large storms when deeper soil is wetted.

[44] This model of partitioning of ET into evaporation and transpiration is consistent with the Dugas et al. [1996] study at the Jornada Range. Using minilysimeters, they determined that the ratio of evaporation to ET was higher in shrubland than in grassland. Additionally, sap flow measurements from the HAPEX-Sahel site in semiarid Africa show that shrub transpiration contributes a minimal amount to ET, even during the rainy season [Tuzet et al., 1997]. Additional work is required to quantify the contri-

bution of evaporation and transpiration to total ET from semiarid grassland and shrubland.

[45] Our results demonstrate that the commonly used model for predicting daily ET from root zone soil moisture [Black, 1979; Dunin and Greenwood, 1986; Rodriguez-Iturbe, 2000; Laio et al., 2001] is not appropriate for semiarid environments similar to those studied here. Daily ET does not increase with higher root zone soil moisture (Figures 9 and 10). Instead, nearly the entire range of ET was measured (0–4 mm d^{-1}) within several percent of the lowest θ_{rz} values observed (~10%). The same result holds when ET is normalized by PET (Figure 10). The limited dependence of ET on θ_{rz} is reasonable. Wetting fronts rarely propagate below 10 cm during the monsoon season, so a significant portion of plant roots exist at soil depths that are not wetted following most monsoon storms. The surface soil layer is the primary source of ET and direct evaporation from bare soil is likely a large component of the total flux [e.g., Dugas et al., 1996]. Predicting ET from θ_{rz} is only reasonable in environments where most or all of ET is from transpiration, and where plants do not compensate for a portion of their roots being in dry soil [Guswa et al., 2002].

5.4. Implications of the Observed Dynamics of θ, EF, and ET

[46] The short intervals when both the soil is wet and ET is high may also influence land-atmosphere interactions [Charney, 1975; Zheng and Eltahir, 1998; Small, 2001]. The soil moisture-induced fluctuations of latent heating observed in the Sevilleta are large. High latent heat values are accompanied by enhanced Q_a, due to the influence of soil moisture on surface temperature and longwave radiation [Small and Kurc, 2003]. Elevated latent heating and Q_a, due to relatively high soil moisture, should have the greatest impact when atmospheric conditions promote convective rainfall. However, because the duration of soil moisture and ET anomalies is short, the influence of soil moisture on boundary layer development and rainfall may be negligible. In wetter environments, the duration of soil moisture anomalies is expected to be longer, perhaps leading to a more substantial soil moisture-rainfall feedback [Findell and Eltahir, 1997].

[47] Quantifying the nature of land-atmosphere interactions has been hindered by the lack of data to constrain how soil moisture varies spatially and temporally at large scales [e.g., Entekhabi et al., 1999]. Point measurements of soil moisture exist [e.g., Robock et al., 2000], but scaling and interpolating these measurements to regional and contintental domains is problematic. This data gap may be filled via remote sensing of soil moisture from satellite platforms, such as the Advance Microwave Scanning Radiometer [Njoku et al., 2003]. Perhaps the most significant problem associated with this type of data is that the instruments are only sensitive to soil moisture within 1 to several cm of the soil surface. Our results clearly show that water content fluctuations in the surface soil layer provide a fundamental constraint on soil moisture storage and ET in semiarid grassland and shrubland environments, at least during the monsoon season.

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