

Carbon, water, and energy fluxes in a semiarid cold desert grassland during and following multiyear drought

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[1] The net exchanges of carbon dioxide, water vapor, and energy were examined in a perennial Colorado Plateau grassland for 5 years. The study began within a multiyear drought and continued as the drought ended. The grassland is located near the northern boundary of the influence of the North American monsoon, a major climatic feature bringing summer rain. Following rain, evapotranspiration peaked above 8 mm d^{-1} but was usually much smaller ($2\text{--}4 \text{ mm d}^{-1}$). Net productivity of the grassland was low compared to other ecosystems, with peak hourly net CO_2 uptake in the spring of $4 \mu\text{mol m}^{-2} \text{ s}^{-1}$ and springtime carbon gain in the range of $42 \pm 11 \text{ g C m}^{-2}$ (based on fluxes) to $72 \pm 55 \text{ g C m}^{-2}$ (based on carbon stocks; annual carbon gain was not quantified). Drought decreased gross ecosystem productivity (GEP) and total ecosystem respiration, with a much larger GEP decrease. Monsoon rains led to respiratory pulses, lasting a few days at most, and only rarely resulted in net CO_2 gain, despite the fact that C_4 grasses dominated plant cover. Minor CO_2 uptake was observed in fall following rain. Spring CO_2 uptake was regulated by deep soil moisture, which depended on precipitation in the prior fall and winter. The lack of CO_2 uptake during the monsoon and the dependence of GEP on deep soil moisture are in contrast with arid grasslands of the warm deserts. Cold desert grasslands are most likely to be impacted by future changes in winter and not summer precipitation.

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1. Introduction

[2] During the early 2000s, much of the western United States experienced several consecutive years of drought that were severe enough in the Colorado River basin to reduce Lake Powell to its lowest level in 40 years [Pratson *et al.*, 2008]. The drought had major impacts on vegetation, leading to episodic death of warm desert shrubs [Bowers, 2005; Miriti *et al.*, 2007], widespread mortality of pinyon pine and juniper trees [Breshears *et al.*, 2005; Floyd *et al.*, 2009], severe insect outbreaks over much of the west [Raffa *et al.*, 2008], mortality of trees in montane and subalpine forests [Bigler *et al.*, 2007; Negron *et al.*, 2009], and decreased productivity or carbon loss in semiarid grasslands [Emmerich and Verdugo, 2008; Mielnick *et al.*, 2005; Scott *et al.*, 2009; Svecar *et al.*, 2008]. Widespread ecological responses across the western United States followed major

droughts of the 1950s and earlier [Allen and Breshears, 1998; Swetnam and Betancourt, 1998]. Tree ring reconstructions for the region reveal that long-term droughts, often with greater duration, have occurred many times over the last millennium [Cook *et al.*, 2007; Gray *et al.*, 2003]. Projections regarding future climate in the western United States suggest that it will become hotter and drier [Seager *et al.*, 2007], with potentially serious implications for semiarid vegetation communities [Archer and Predick, 2008] and the animals therein, as well as the ranching, farming, and hunting practices that depend on those ecosystems [Schwinning *et al.*, 2008].

[3] In general, little is known about the response of ecosystem carbon and water fluxes to sustained drought (lasting several years or more). Productivity of most grasslands is strongly linked to precipitation [Sala *et al.*, 1988], and this is particularly true for those in arid and semiarid climates [Noy-Meir, 1973]. Seasonal phenological stages of grasses and most arid land plants are strongly controlled by precipitation events [Beatley, 1974]. The response of grasslands to short-term droughts (such as seasonal drought, or a dry year) usually involves substantial reduction in evapotranspiration (ET) and CO_2 uptake and may involve sustained periods of CO_2 loss [Flanagan *et al.*, 2002; Meyers, 2001; Suyker and Verma, 2001]. During long-term drought there are likely to be substantial physiological changes [Kozlowski

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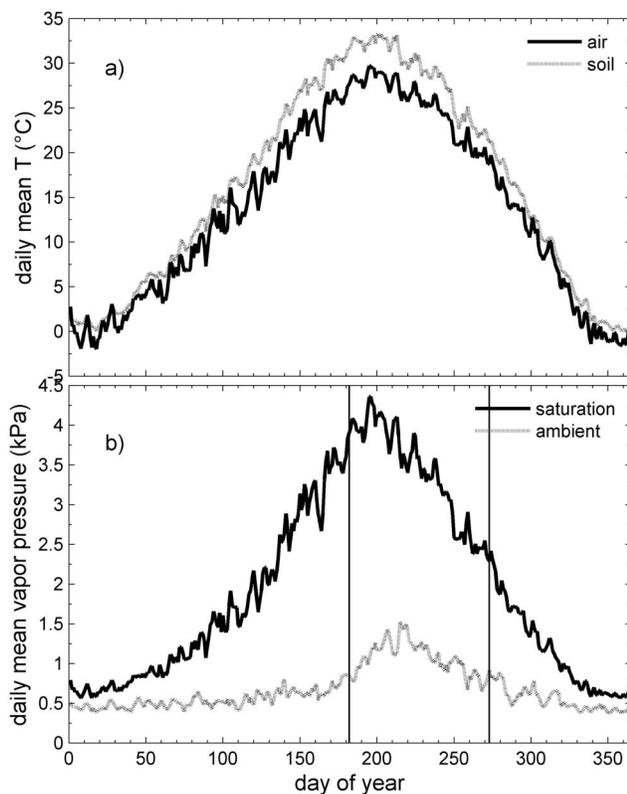


Figure 1. Daily mean (a) temperature of the air (1.85 m height) and soil (10 cm depth) and (b) atmospheric saturation and ambient vapor pressure during 1999–2008. Vertical lines show the July, August, September monsoon period.

and Pallardy, 2002], including severe water stress, carbon starvation, and xylem embolism [McDowell et al., 2008]; depletion of carbon and nutrient storage reserves [Reynolds et al., 2004]; and senescence or death of foliage and roots. Decreased plant productivity over many years will have cascading effects on animals, likely causing famine, range or habitat shifts, and disease or mortality [e.g., Brown and Ernest, 2002; Ernest et al., 2000]. Long-term drought may also lead to physical changes such as decreased groundwater recharge [Small, 2005], decreased stream flow [Wilcox et al., 2008], and wind or water erosion of soil with concomitant biogeochemical changes [Okin et al., 2006]. Some of these changes may be irreversible, possibly leading to fundamental changes in ecosystem structure or function such as desertification [Schlesinger et al., 1990].

[4] The North American monsoon (NAM) is a dominant seasonal climatic feature of the southwestern United States and northern Mexico [Adams and Comrie, 1997]. The warm Chihuahuan and Sonoran deserts of North America experience relatively high rainfall during July through September (45%–60% of total annual precipitation [Reynolds et al., 2004]). In contrast, the cold desert of the Great Basin in the western United States is largely uninfluenced by the NAM and does not receive significant summer rain [Higgins et al., 1997]. The cold desert of the Colorado Plateau region, just to the north of the warm deserts, experiences variable NAM influence, with high summer rain in Arizona and New

Mexico, and summer rain generally decreasing toward the north and west (Colorado and Utah).

[5] Located in the center of the Colorado Plateau in southeast Utah, Corral Pocket (the perennial grassland of the present study) receives, on average, about one third of the 216 mm total annual moisture in summer. Daily mean air and soil temperature peak at 30°C–35°C (Figure 1a), with surface soils hotter when in the sun. Monsoon rains arrive in the summer when soils are hottest, and although there is a minor increase in ambient humidity, evaporative demand remains high (Figure 1b). In general, C_4 grasses respond favorably to warm wet summers [Teeri and Stowe, 1976], exhibiting enhanced photosynthesis, transpiration, and growth following rain [Nobel, 1980; Schwinning et al., 2002], and sometimes flowering more than once within favorable seasons [Everett et al., 1980]. Both C_3 and C_4 photosynthetic types can be active in southern Utah from March to October, with C_3 grasses active a bit earlier in spring and C_4 grasses somewhat more active following late summer rain [Belnap and Phillips, 2001]. Corral Pocket has a substantial proportion of C_4 grasses, and with the warm summer and monsoon moisture, the grassland might be expected to be physiologically active during summer, as in the warm deserts [Cable, 1975] or tallgrass prairies of North America [Suyker et al., 2003].

[6] Rain events during the NAM occur primarily via convective thunderstorms and are highly variable in time and space. The size of a rain pulse strongly influences the organisms that can respond (see review by Huxman et al. [2004b]). Although some plants will respond to rain events as small as 5 mm [Sala and Lauenroth, 1982], in general larger pulses are required for plants to become physiologically active following a dry period [Huxman et al., 2004b; Schwinning et al., 2002, 2003], particularly to initiate growth. Some plants will increase photosynthesis and nutrient uptake following a pulse [BassiriRad et al., 1999; Huxman et al., 2004a; Schwinning et al., 2002], while some plants ignore summer rain entirely [Ehleringer et al., 1991; Phillips and Ehleringer, 1995; Williams and Ehleringer, 2000]. Microbial communities can respond fairly rapidly to even very small moisture pulses, with important influences on carbon and nutrient cycling [Austin et al., 2004; Fierer and Schimel, 2002; Schaeffer and Evans, 2005]. If wetted during the daytime, biological soil crusts will gain carbon and fix nitrogen, but small events or nighttime rain can have limited N fixation and respiratory carbon loss, depending on conditions prior to the rain event [Belnap, 2002]. Microbial biomass and community composition can change within hours to days of wetting [Fierer et al., 2003; Saetre and Stark, 2005]. Soil invertebrate communities are also likely responsive to moisture pulses [Whitford et al., 1995]. Hence, the ecosystem-level responses of desert grassland to moisture pulses are quite complex.

[7] The principal aim of this study was to examine the energy, water vapor, and CO_2 fluxes of the Corral Pocket grassland during the multi-year drought and following recovery from drought. The cold desert location and the relatively smaller local influence of the NAM contrast this grassland with those studied in the Sonoran and Chihuahuan deserts. A secondary objective was to examine the ecosystem-level gas exchange responses of the grassland to monsoon moisture. During severe drought, we hypothesized

Table 1. Percent Live Cover for Dominant Plant Species at Corral Pocket in Spring 2003^a

Species	Percentage of Total Plant Cover (%)	Photosynthetic Type
<i>Hilaria jamesii</i> (galleta grass)	49.9	C ₄ bunchgrass
<i>Stipa hymenoides</i> (Indian ricegrass)	19.7	C ₃ bunchgrass
<i>Coleogyne ramosissima</i> (blackbrush)	14.0	C ₃ shrub
<i>Gutierrezia sarothrae</i> (broom snakeweed)	7.5	C ₃ subshrub
<i>Aristida purpurea</i> (purple three-awn)	5.6	C ₄ bunchgrass
<i>Sporobolus contractus</i> (spike dropseed)	2.9	C ₄ bunchgrass
Other	0.5	N/A

^aTotal plant cover equaled 14% of the ground surface; the rest was bare ground. Nomenclature follows *Welsh et al.* [2003].

that all grasses would be dormant, with limited net ecosystem CO₂ uptake, or more likely, respiratory carbon loss. We expected that available energy would be dominated by sensible heat when moisture was limiting. Once favorable moisture conditions returned, we hypothesized that both C₃ and C₄ physiological types would be active in the spring and during the summer monsoon. The soil biological community was expected to respond to moisture pulses at any time of the year.

2. Methods

2.1. Study Location

[8] This study was conducted at Corral Pocket, a grassland in southeast Utah (38.09°N, 109.39°W, 1520 m), during 2001–2007 (no observations were made during 2004 and 2005). The grassland is composed of native perennial C₃ and C₄ bunchgrasses (Table 1), with low plant cover (9%–35% perennial plants and 7%–15% annual plants) and most of the interspaces covered with cyanobacterially dominated biological soil crusts [*Belnap et al.*, 2009]. As is typical of the region, cattle are grazed on the site for about 6 weeks during the late fall or winter, and aboveground vegetation regrows each spring. Crusts are impacted by grazing and maintain fairly low biomass (level of development class 1 [*Belnap et al.*, 2008]). The site is part of the Climate Impact Meteorological (CLIM-MET) network (<http://esp.cr.usgs.gov/info/sw/clim-met/index.html>). Mean annual temperature is 12.0°C, and mean annual precipitation is 216 mm (1965–2008, Western Regional Climate Center, <http://www.wrcc.dri.edu/>). Soils are fine sandy loams, with 63% sand (consisting of 52% fine and very fine sand fractions), 27% silt, and 9% clay. Soil field capacity was 22.6% by mass. Further site details can be found elsewhere [*Belnap et al.*, 2009; *Goldstein et al.*, 2005].

2.2. Meteorological Measurements

[9] Weather was monitored hourly, including net radiation (REBS Q*7, Radiation and Energy Balance Systems, Seattle, Washington), global radiation (LI-200SX, calibrated annually, LI-COR Biosciences, Lincoln, Nebraska), air temperature and relative humidity at flux measurement height (see below, un aspirated HMP45C, Vaisala, Boulder, Colorado), and soil heat flux at 5 cm depth (REBS HFT3, Radiation and Energy Balance Systems, Seattle, Washington). Additional weather data were obtained from the CLIM-MET database. Beginning in July 2002, soil moisture was measured in a single nest of horizontally aligned water content reflectometry (WCR) probes at 5, 10, 20, 40, and

60 cm depths (CS615, Campbell Scientific, Inc., Logan, Utah). Soil moisture was also measured using a single 10 cm deep WCR probe at a separate location during 2001–2006. These data were scaled to match the data from the nest at a 10 cm depth so a comparison could be made across all years. Soil temperature (T) was measured with thermocouples at 5, 10, 20, 40, and 60 cm depths. Soil heat storage above the heat flux plate was calculated following *Mayocchi and Bristow* [1995].

2.3. Flux Measurements

[10] Net fluxes of sensible heat, ET, and CO₂ were measured using eddy covariance. Flux measurement height was 1.25 m above the ground during most of 2001 and 1.85 m afterward (energy balance closure did not differ, data not shown). Wind velocity and air T were measured at 10 Hz with a three-dimensional sonic anemometer (CSAT3, Campbell Scientific, Logan Utah). The CO₂ and H₂O were measured at 10 Hz with two separate systems: first, with an open-path infrared gas analyzer (IRGA, LI-7500, LI-COR Biosciences, Lincoln, Nebraska) from April 2001 through the end of 2003, and second, with a closed-path IRGA (LI-7000, LI-COR Biosciences, Lincoln, Nebraska) from April 2002 through the end of 2007. Flux measurements using both analyzers overlapped for ~1.5 years.

[11] Fast time series data were linearly detrended, then despiked following *Højstrup* [1993]. Virtual heat fluxes were converted to sensible heat fluxes [*Schotanus et al.*, 1983]. Spectral corrections were applied after *Massman* [2000, 2001], including detrending, block averaging, sensor separation, tubing attenuation, volume averaging, and instrument frequency response. Tubing and electronics delays were accounted for. Fluxes were determined hourly, referenced to a site-specific long-term coordinate system, following *Finnigan et al.* [2003]. Flux analysis time periods from 30 min to 4 h were investigated, and the hourly time step and long-term coordinate system provided the best energy balance closure. Flux data were omitted during nonstationary periods by rejecting hours where the mean of 10 min covariances differed from the 60 min covariance by more than 30% [*Foken and Wichura*, 1996]. A friction velocity threshold of 0.1 m s⁻¹ was used to exclude data during stable periods (other thresholds were also examined, see section 2.4).

[12] The CO₂ and ET fluxes were corrected for air density fluctuations [*Webb et al.*, 1980]. The sensible heat correction for the open-path fluxes was much larger than the actual net flux (sometimes an order of magnitude larger), leading to uncertainty if the CO₂ density was incorrect [*Serrano-*

Table 2. Comparison of Fluxes Measured Using the Open and Closed Path Infrared Gas Analyzers^a

	Conditions	Slope	Intercept ($\times 10^{-3}$)	r^2	n
2002 CO ₂	unstable	0.98	-5.7	0.70	1996
	stable	1.00	1.1	0.58	1384
2003 CO ₂	unstable	0.84	3.9	0.77	2705
	stable	0.92	0.4	0.76	1813
2002 H ₂ O	unstable	1.05	1.1	0.98	2118
	stable	1.12	1.1	0.89	1645
2003 H ₂ O	unstable	1.08	34	0.98	2784
	stable	1.11	14	0.66	2174

^aShown are results of linear regressions between open path (y axis) and closed path (x axis) hourly fluxes. Regressions were performed for unstable and stable conditions for each year separately (n = number of hours). Units for the fluxes are $\mu\text{mol m}^{-2} \text{s}^{-1}$ (CO₂) or mm d^{-1} (H₂O).

Ortiz *et al.*, 2008]. The open-path analyzer signal drifted because of dust buildup and rain between infrequent calibrations. Accordingly, we adjusted the mean CO₂ measured by the open-path system to match the time-dependent Globalview estimate for Wendover, Utah [GLOBALVIEW-CO₂, 2008] prior to calculating fluxes. Corrections were examined for LI-7500 self-heating [Burba *et al.*, 2008] but led to a systematic bias of CO₂ fluxes toward more respiratory values, with results nearly identical to Figure 3 of Wohlfahrt *et al.* [2008]. Since the comparison with closed-path fluxes was much better without the self-heating corrections, they were not applied. The closed-path system was designed remove T fluctuations before the gas was measured (and verified by measuring T at 10 Hz at the IRGA outlet), and thus the sensible heat correction was not required for this analyzer. The closed-path system was calibrated hourly using World Meteorological Organization (WMO)-traceable compressed gas standards. The CO₂ measured with the calibrated closed-path IRGA during the study never deviated from the Wendover Globalview value by more than $15 \mu\text{mol mol}^{-1}$, even at night.

[13] Following these corrections, the CO₂ and ET fluxes from the closed-path and open-path systems compared favorably (Table 2), generally within 15% of one another. There were time periods of excellent agreement, and time periods of poor agreement. These results are quite similar to those of the long-term comparison of Haslwanter *et al.* [2009]. Although our fluxes are considerably smaller than theirs, the results in Table 2 and the energy balance analysis shown later provide confidence in the suitability of all the corrections. Here CO₂ and ET fluxes are reported using the closed-path data when available and the open-path data at other times. The CO₂ mole fraction was measured only at the flux measurement height (not as a vertical profile). Because diurnal variation of CO₂ at this site was nearly always $<10 \mu\text{mol mol}^{-1}$ and measurements were made within 2 m of the ground, storage fluxes were negligible, and we assume that the net CO₂ flux represents net ecosystem exchange (NEE) of CO₂.

2.4. Gap Filling and Flux Partitioning

[14] Significant gaps exist in the data because of various problems. Gaps prevented meaningful calculation of annual sums of NEE or ET, and thus we focus on year to year comparisons, rather than annual sums, to better understand the ecological processes involved. To compare total carbon

gain and total ET during the spring growth period of different years, we filled gaps in CO₂ and ET flux following Reichstein *et al.* [2005]. Years 2001 and 2002 were not analyzed because of gaps during the spring growth period and lack of soil moisture depth profile data to aid with interpretation. A total of 28%, 55%, and 38% of hourly periods were gap-filled in 2003, 2006, and 2007, respectively, in the 210 day period analyzed (days 40 to 250). Similar flux data coverage of around 50% is usually achieved in studies over forests [e.g., Krishnan *et al.*, 2008; Urbanski *et al.*, 2007]. With these large gaps, our sums of total carbon gain or ET are estimates, which we use for functional comparisons between years, and a comparison with the intensive biometric analysis in 2003 (discussed in section 2.5).

[15] The method of Scott *et al.* [2006b] was used to partition NEE into gross ecosystem productivity (GEP) and total ecosystem respiration (R_{eco}), which is based entirely on measured NEE. Gap-filled data and those excluded because of low turbulence or other quality assurance criteria were not used in the partitioning. Weekly means of nighttime NEE were used to estimate weekly mean R_{eco} , and this was subtracted from weekly mean daytime NEE to estimate weekly mean GEP. Standard errors of the weekly means are presented as a measure of uncertainty.

[16] Net carbon gain during the productive spring period (beginning on day 40 and continuing until NEE became positive each year) was assessed using cumulative sums of gap-filled hourly NEE data. Errors in this sum were assessed in two ways: first by using u^* thresholds of 0.1 to 0.4 m s^{-1} in steps of 0.05 m s^{-1} , and second by assigning a systematic bias to NEE on the basis of measured variability in weekly NEE as follows. Over the full data set, the range of values for the weekly standard error of NEE was 0.0 to $0.13 \mu\text{mol m}^{-2} \text{s}^{-1}$ ($n = 207$ weeks, mean = $0.03 \mu\text{mol m}^{-2} \text{s}^{-1}$). Ninety-five percent of these standard errors were below $0.07 \mu\text{mol m}^{-2} \text{s}^{-1}$. We assumed that NEE was systematically too low or too high by this value (a worst-case scenario) and calculate the cumulative carbon gain in each case using a u^* threshold of 0.1 m s^{-1} .

2.5. Vegetation Cover and Carbon Stock Assessment

[17] Vegetation cover near the tower was quantified in April 2003. Along each of six compass directions, line transects were sampled every 10 m to a distance of 200 m from the tower (20 lines per compass direction). Each line transect was 20 m long and perpendicular to the compass direction. Plant species and horizontal spatial extent along the line were recorded. Additional vegetation surveys were conducted at the site during 2001–2007 [Belnap *et al.*, 2009].

[18] Full carbon stocks were measured three times in 2003: (1) early spring before net CO₂ uptake began, (2) late May when peak photosynthetic biomass was expected, and (3) late September. Harvests of aboveground material in $20 \text{ m} \times 1 \text{ m}$ plots were conducted on day of year (DOY) 99–100, 147–149, and 267–268. Nodes were located along the 210° and 330° compass directions at locations ranging 110–200 m from the tower in 10 m increments. The plots were located 2 m to the right (default) or left (alternate) of each node, angled toward the tower 45° from the node. The alternate location was used for 3 plots to avoid large shrubs

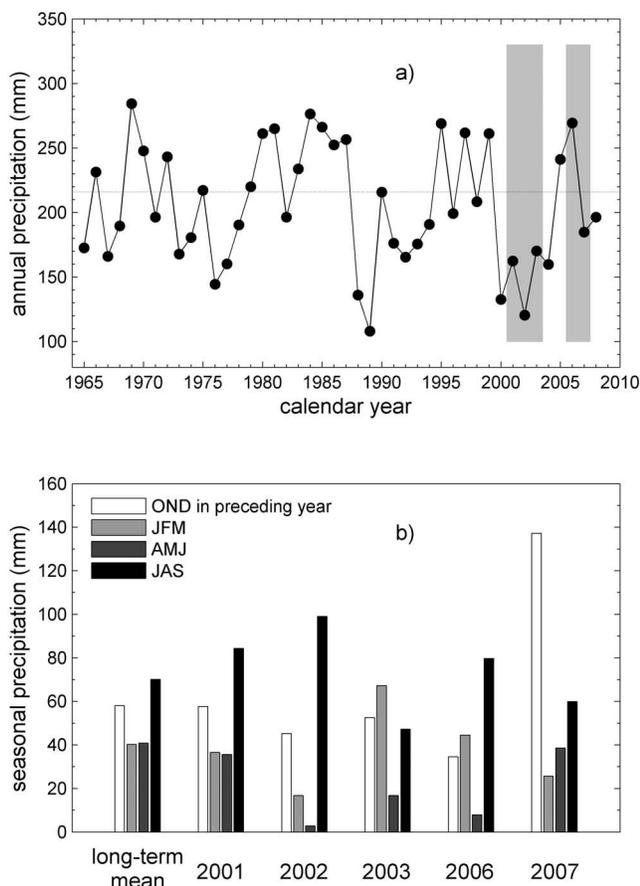


Figure 2. (a) Annual calendar year precipitation in the study area during 1965–2008. The years covered by the present study are highlighted in gray boxes. (b) Precipitation during 3 month periods, showing the long-term mean (1965–2008) and each of the study years. Note that fall precipitation (October–December (OND)) is shown for the preceding year. JFM, January–March; AMJ, April–June; JAS, July–September. Data from Western Regional Climate Center, <http://www.wrcc.dri.edu/>, for the site “Canyonlands, The Needle,” which is ~10 km from Corral Pocket.

that were not generally representative. Samples were separated into the following categories: green *Hilaria* foliage, senesced *Hilaria* foliage, other plants (see Table 1), surface litter (dead organic material), and animal dung (cattle and rabbit). Mineral soil was visible at the surface (no organic horizon was present). All underground rhizomes attached to living *Hilaria* plants were harvested, usually to a maximum of 7–10 cm depth. Fine roots were clipped from the rhizomes and discarded. Soil cores were collected at ten or more locations selected randomly. Cores were collected in 10 cm increments using a bucket auger on DOY 65–66, 149, and 269, to depths of 1 m, 30 cm, and 30 cm, respectively. Soil bulk density was assessed in 3 separate hand-excavated pits to avoid compaction by the bucket auger. Samples from the harvests and soil cores were dried at 70°C and stored in an oven until analysis. Aboveground samples were analyzed for carbon content for all samples collected ($n = 20$, except when categories were absent on a plot). Soil samples were selected at random from the cores,

and $n = 7$ samples processed for depths 0–30 cm, and $n = 3$ or 4 for deeper samples. Soil was sieved to 2 mm, then separated by hand into these classes: (1) fine roots (<2 mm), (2) decaying organic material, highly visible in the red soils, (3) remaining soil after categories 1 and 2 were removed. In most other soil types, category 2 would likely be included as part of the organic component of the bulk soil, because it would not be visible. This was primarily made up of recognizable leaf and root bits that were black and decomposing. Category 3 was analyzed for total carbon content both before and after acid washing to remove carbonates, providing an assessment of both organic and inorganic carbon content in the mineral soil. All samples were ground then analyzed for percentage of carbon content on an elemental analyzer (EA 1108, Carlo Erba, Italy) coupled to an isotope ratio mass spectrometer (DeltaS; ThermoFinnigan, Waltham, Massachusetts).

3. Results

[19] The drought at Corral Pocket began in 1999 and lasted 5 consecutive years, with annual precipitation between 56% and 79% of the long-term mean (Figure 2a). The first 3 years of this study were conducted during the height of the drought, and the last 2 years experienced above (125%) and below (85%) normal precipitation, respectively. Precipitation was fairly evenly distributed on average throughout the year (Figure 2b), with roughly 45% falling in fall and winter (October through March), and 32% during the monsoon months (July through September). Relative amounts of seasonal precipitation varied during the study years, with some 3 month periods receiving less than 10 mm total, and October–December precipitation in late 2006 more than double the long-term mean for that period. Monsoon rainfall was slightly above normal during 2001, 2002, and 2006, and below normal for 2003 and 2007.

[20] Soil moisture at all depths followed a strongly seasonal pattern, with more water in the soil profile in fall and winter, and moisture generally decreasing from the spring through midsummer (Figure 3). Monsoon rains typically wet only the soils near the surface with little infiltration. Fall and winter soil moisture recharge varied across the years, leading to considerable variation in deep soil moisture during the spring. Contrasting the 60 cm depth in early 2006 and 2007 shows that fall rains recharged deep soil moisture in 2006, but not in 2005 (data not shown, but the 60 cm depth was very dry in early 2006). Soil moisture in midsummer was low in all years, both during the drought (2001–2003) and afterward (2006–2007), except near the surface when it rained.

[21] Measured turbulent energy fluxes accounted for 73% to 86% of the measured available energy annually (Table 3). Lack of energy balance closure is usually observed at flux tower sites, and our results are well within the range of reported closure [Wilson *et al.*, 2002]. Heat storage in the ground was a large and highly variable component of the energy balance at Corral Pocket (data not shown), which is common in arid regions. Better closure could have perhaps been achieved with more thorough characterization of the soil heat flux and soil heat storage [Kustas *et al.*, 2000]. However, the annual energy balance ratio (Table 3), which

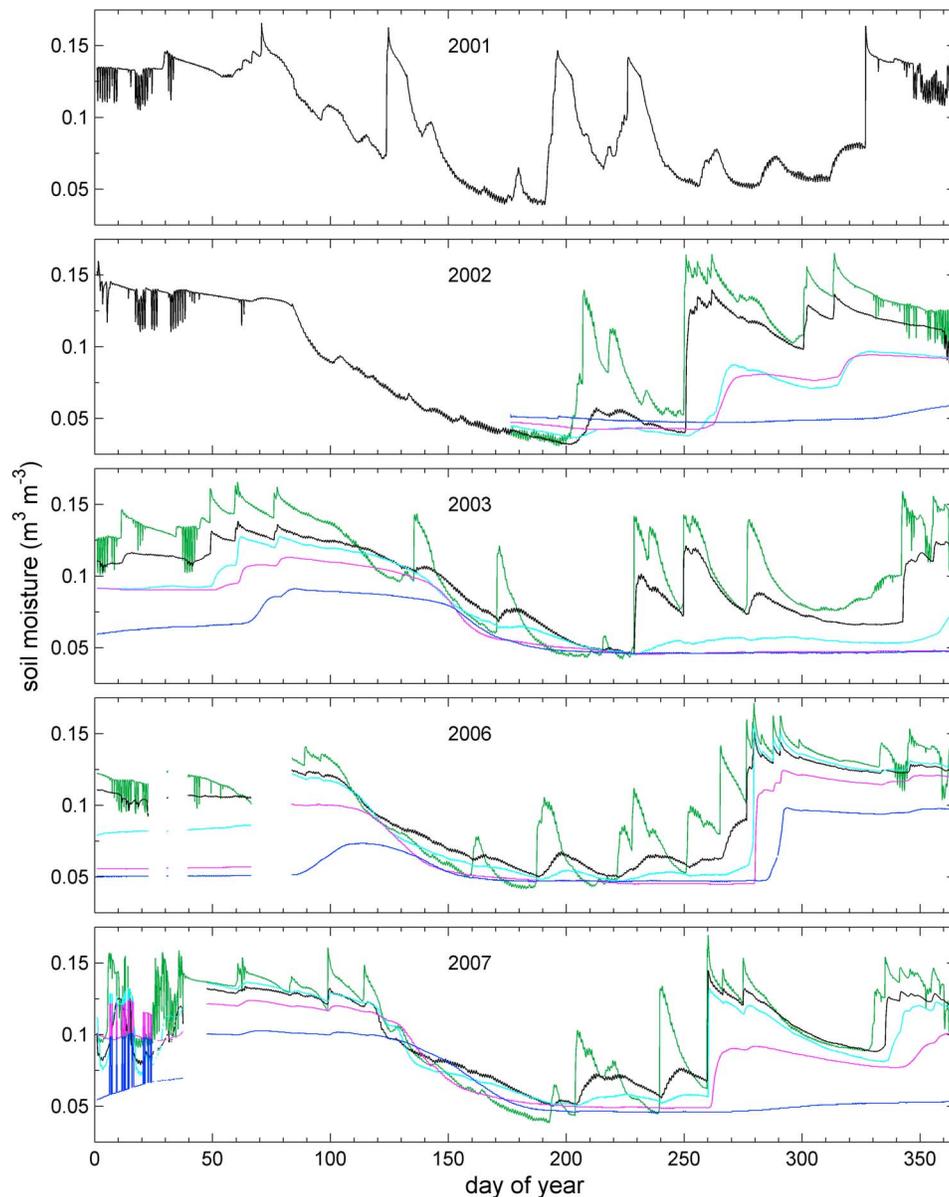


Figure 3. Volumetric soil moisture during each year at 5 (green), 10 (black), 20 (light blue), 40 (pink), and 60 cm (dark blue) depth. Noisy data in winter periods are due to freezing.

minimizes the importance of storage, was also well within the range reported at other sites [Wilson *et al.*, 2002].

[22] Evapotranspiration was highly variable, peaking briefly above 8 mm d^{-1} during or shortly after rain events (Figure 4), and becoming nearly negligible after long dry periods. Peak ET was typically $2\text{--}4 \text{ mm d}^{-1}$ between moisture pulses, depending on the year. Weekly mean ET fluxes were not considerably lower during drought than in other years. Consistent seasonal patterns in ET, such as those observed in forests or more humid grasslands, were not obvious, although in 2003 and 2007 there were extended periods in the spring with higher ET than the other years. Total ET measured by eddy covariance between day 40 and day 250 was 122, 80, and 170 mm, in 2003, 2006, and 2007, respectively. For comparison, precipitation during the 1 year period ending with the monsoon in each of these years was

184, 167, and 261 mm. Hence, during days 40–250, the grassland lost 66%, 48%, and 65% of the previous year's precipitation to ET in each year. On the basis of the ET fluxes shown in Figure 4 and the low soil moisture at all depths in summer (near day 200, Figure 3), we presume that the remainder of annual precipitation was lost to ET during the rest of the year.

[23] On the basis of CO_2 fluxes (discussed below), we defined periods representing the time of spring canopy growth, soil dry-down, and the monsoon. The relation between ET (as evaporative fraction) and soil moisture was examined in each period. Evaporative fraction was strongly dependent on soil moisture (Figure 5) during some periods. When the soil was drying, $\text{LE}/(H + \text{LE})$ declined linearly as soil moisture decreased regardless of depth considered. During the canopy growth period, at high soil moisture,

Table 3. Energy Balance Closure During Each Year of the Study^a

Year	Slope	Intercept	r^2	n (hours)	EBR
2002	0.86	-6.6	0.81	3449	0.76
2003	0.86	-6.6	0.80	4054	0.79
2006	0.73	-4.4	0.73	4481	0.68
2007	0.80	-3.1	0.81	4813	0.77

^aThe slope, intercept, and r^2 are the parameters for a regression of hourly turbulent fluxes ($H + \lambda E$) versus hourly available energy ($R_n - G - S$), with n total hours where all data were available. Soil moisture and temperature were not measured at an appropriate depth relative to the soil heat flux plate to calculate S in 2001. EBR is the annual energy balance ratio after Wilson *et al.* [2002], where $EBR = \Sigma(H + \lambda E) / \Sigma(R_n - G - S)$ with each quantity summed over 1 year. H , sensible heat flux; λE , latent heat flux; R_n , net radiation; G , soil heat flux; S , soil heat storage.

$LE/(H + LE)$ was negatively correlated with soil moisture. During the monsoon, $LE/(H + LE)$ was related to soil moisture at 5 cm depth. Notably, evaporative fraction during the monsoon was not related to soil moisture deeper in the profile, highlighting the shallow depth of infiltration of monsoon rains (Figure 3).

[24] Observed CO_2 fluxes were generally small compared to other grasslands, with weekly mean NEE during periods of maximum uptake peaking below $-1 \mu\text{mol m}^{-2} \text{s}^{-1}$ and weekly mean NEE during the monsoon period peaking at about $0.4 \mu\text{mol m}^{-2} \text{s}^{-1}$ (Figure 6a). Springtime GEP (Figure 6b) during drought years was maximally $-1.5 \mu\text{mol m}^{-2} \text{s}^{-1}$, much lower than the same period in postdrought years (as large as -2.5 to $-2.7 \mu\text{mol m}^{-2} \text{s}^{-1}$). The number

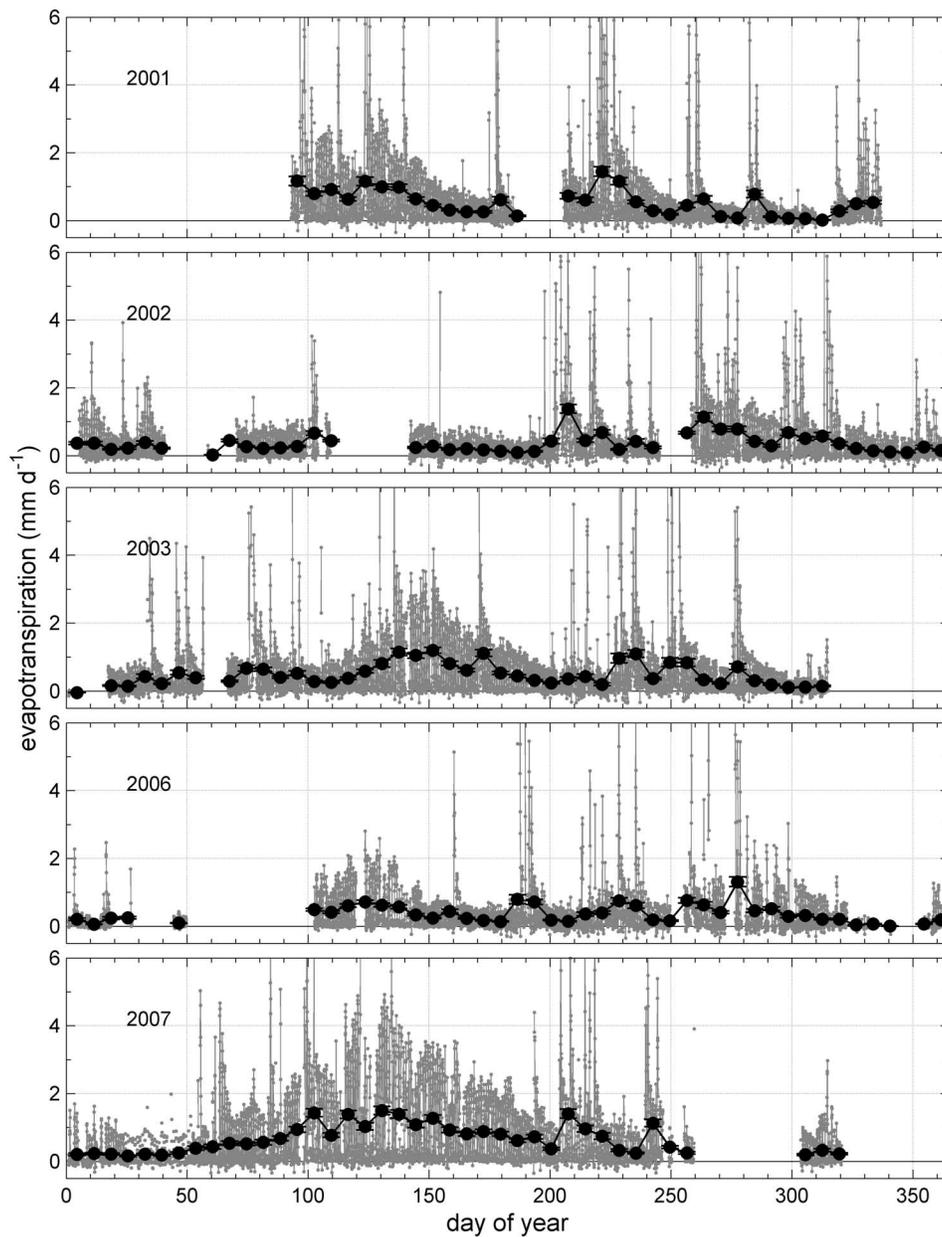


Figure 4. Hourly water vapor flux during each year. Weekly means and standard errors are also shown (black circles).

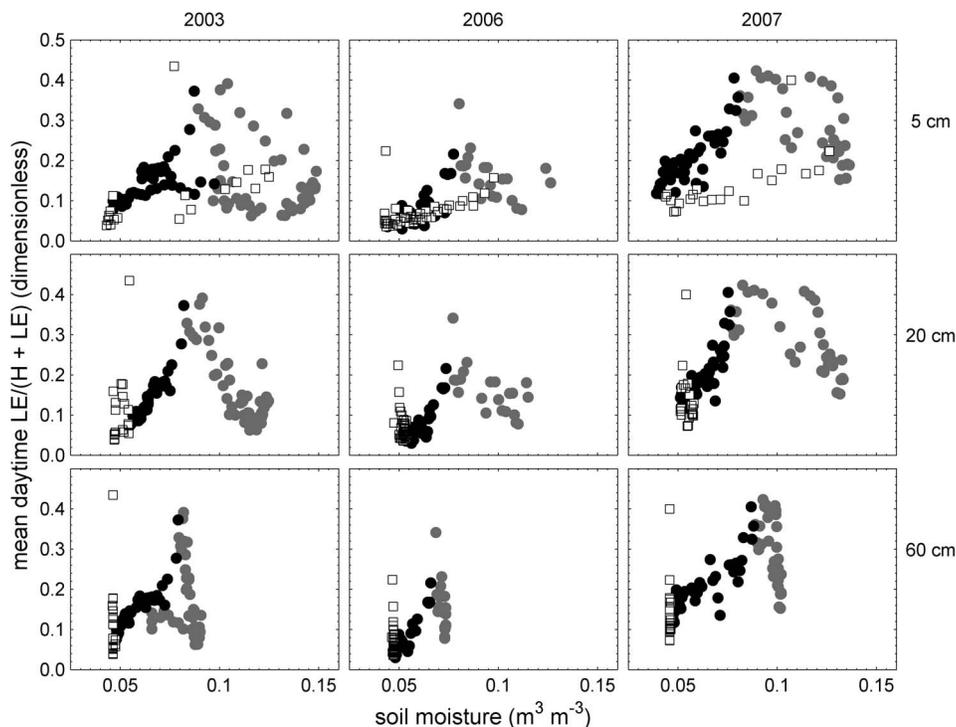


Figure 5. Daytime mean (0900 to 1500 local time) evaporative fraction ($LE/(H + LE)$), where LE is latent heat flux and H is sensible heat flux) as a function of soil moisture at different depths (rows) during different years (columns). Periods within 3 days after a rain event have been excluded and depart from the relationships shown. Data are plotted separately for periods of canopy growth (gray circles), soil dry-down and canopy senescence (black circles), and monsoon rain (open squares). Exact time periods differ for each year (canopy growth, 2003 days 50–150, 2006 days 100–130, 2007 days 75–140; dry-down 2003 days 150–200, 2006 days 130–180, 2007 days 140–200; monsoon days 200–250 in all years).

of weeks with substantial GEP fluxes in springtime was smaller during drought than afterward. GEP during the monsoon period and during periods of fall uptake was low, $-1 \mu\text{mol m}^{-2} \text{s}^{-1}$ or less. R_{eco} was also decreased during springtime in drought years relative to postdrought (Figure 6c), but this difference was much smaller than the GEP difference (scales in Figure 6 differ). Ecosystem respiration was related to temperature, and this relation differed seasonally (Figure 7). During the spring (April–June), R_{eco} increased as temperature warmed, then decreased as the soil dried, leading to a subtle inverted U shape when plotted versus temperature (Figure 7). Monsoon rains enhanced R_{eco} in all years (Figure 6), and weeks with more rain had higher respiration, regardless of whether they followed long dry periods or not (Figure 7). GEP also differed between seasons and was related to soil temperature and radiation (Figure 7). GEP was highest during April–June in all years.

[25] The length of the period of significant CO_2 uptake differed interannually (Figure 8). During 2003 and 2007, the grassland experienced high GEP and net CO_2 uptake from day 90 until DOY 190 or so. In contrast, the CO_2 uptake period of 2006 ended a full month earlier, by DOY 160. Following the CO_2 uptake period, there were short pulsed periods of respiratory CO_2 release during the monsoon rains in each year (Figures 6 and 8). Total gap-filled cumulative CO_2 exchange between day 40 and peak of

uptake (local minima on the curves in Figure 8 (bottom)) was 42 ± 11 , 38 ± 8.7 , and $66 \pm 11 \text{ g C m}^{-2}$, in 2003, 2006, and 2007, respectively (Figure 8 (bottom) symbols). The use of different u^* thresholds to exclude stable periods led to negligible variability in the cumulative sum (range of 42.1 to 42.6 for 2003, Figure 8). Applying the systematic bias to NEE on the basis of the 95% level of uncertainty as described in section 2 changed the cumulative sum for 2003 by $\pm 11 \text{ g C m}^{-2}$ (uppermost and lowermost blue lines, Figure 8 (bottom)).

[26] The length of the CO_2 uptake period, and the total amount of carbon gained (both shown in Figure 8), were strongly related to deep soil moisture (Figure 9). In late winter, there was considerably more moisture in the total soil profile in 2007 than in 2006, with 2003 intermediate between them (Figure 9). This pattern matches the ranking of carbon gain between the 3 years (Figure 8) and total ET discussed earlier: the year with the most soil moisture available in the profile (2007) had the largest net CO_2 uptake and total ET in spring, and vice versa. Evidence of drawdown of soil moisture was apparent by midspring, except in 2003 following winter rains (days 50–80) that penetrated deep in the profile (Figure 9). Monsoon rains in all years wet only the surface of the soil, and there were some small differences in summer deep soil moisture apparent between years. During the height of the drought,

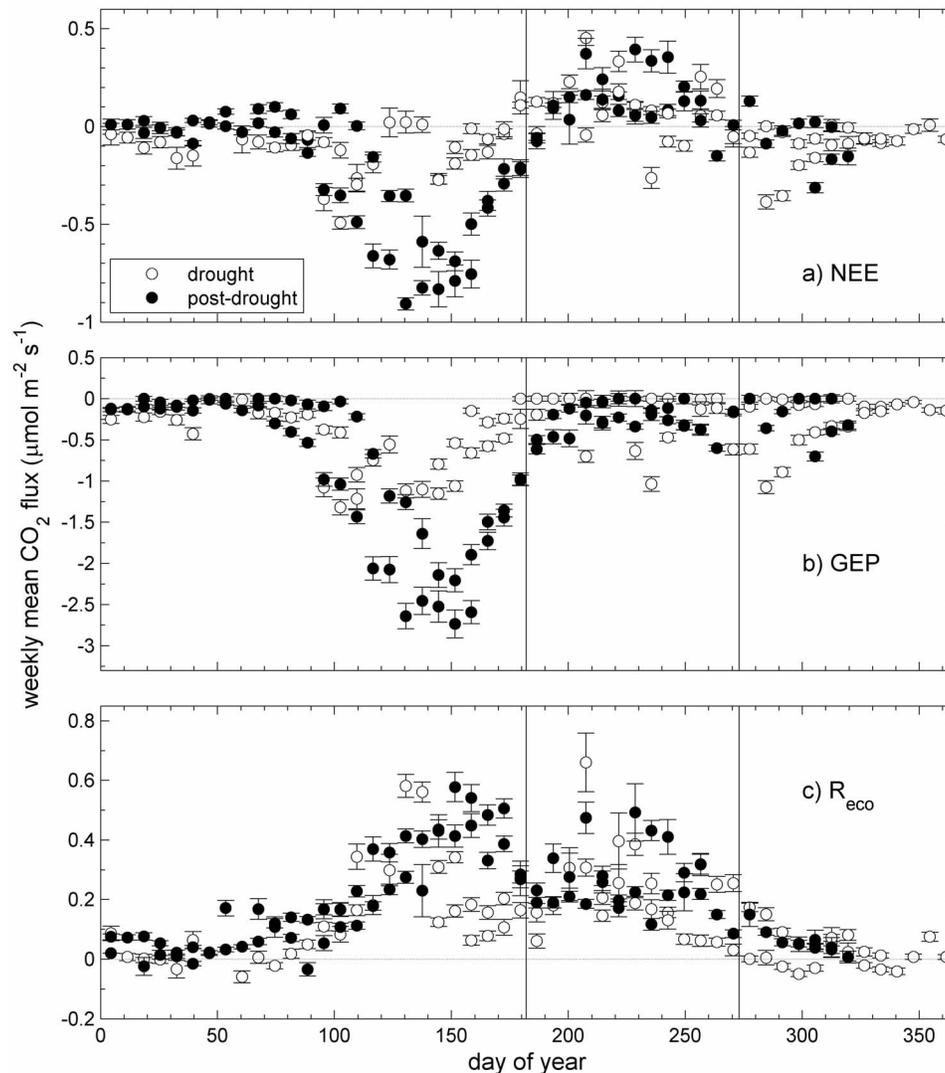


Figure 6. Weekly mean (a) NEE, (b) GEP, and (c) R_{eco} . Note that the scales on the ordinate axis differ in each panel. Drought years (2001 and 2002) and the two most productive postdrought years (2003 and 2007) are shown. Error bars are standard errors of the mean. Vertical lines show the July, August, September monsoon period. Negative values indicate CO_2 uptake, positive values indicate CO_2 loss.

2002, soil moisture deeper in the profile was the lowest observed during all the years of the study (Figure 9).

[27] Measured carbon stocks at Corral Pocket were dominated by organic rather than inorganic forms (Table 4). However, most of the inorganic carbon at the site probably resides in spatially variable caliche horizons [Schlesinger, 1982] that occur roughly 60–100 cm below the surface; these were not sampled quantitatively, but we noted this horizon when digging soil pits and coring. During 2003 live plant material above and belowground increased from 117 ± 20 to 229 ± 42 g C m^{-2} (not accounting for changes in roots below 30 cm). Decaying organic matter in the top 30 cm varied 138 ± 30 to 176 ± 36 g C m^{-2} , with most of it found in the top 10 cm (data not shown). There was little surface plant litter present (maximally 19 ± 3 g C m^{-2}). During the spring uptake period of 2003, the grassland gained 72 ± 55 g C m^{-2} in live plant material (as shoots, roots, and rhizomes).

4. Discussion

[28] Precipitation in desert regions is a major driver of ecosystem productivity. The C_4 -dominated grasslands of the warm Chihuahuan and Sonoran deserts are highly dependent on summer rain during the North American monsoon. In contrast, the Great Basin receives almost entirely winter precipitation, and the Colorado Plateau receives an intermediate amount of summer rain. This distinction in winter and summer moisture has long been recognized as a principal factor influencing the distribution of plants in the arid western United States [Comstock and Ehleringer, 1992].

[29] Soil moisture at Corral Pocket is recharged in fall and winter with rain and snow (Figure 3). Seasonal patterns of soil moisture were similar in drought years and in favorable years, exhibiting a general pattern of higher moisture in the late winter and spring, and dry conditions through most of the soil profile in the summer (Figure 3). This seasonal

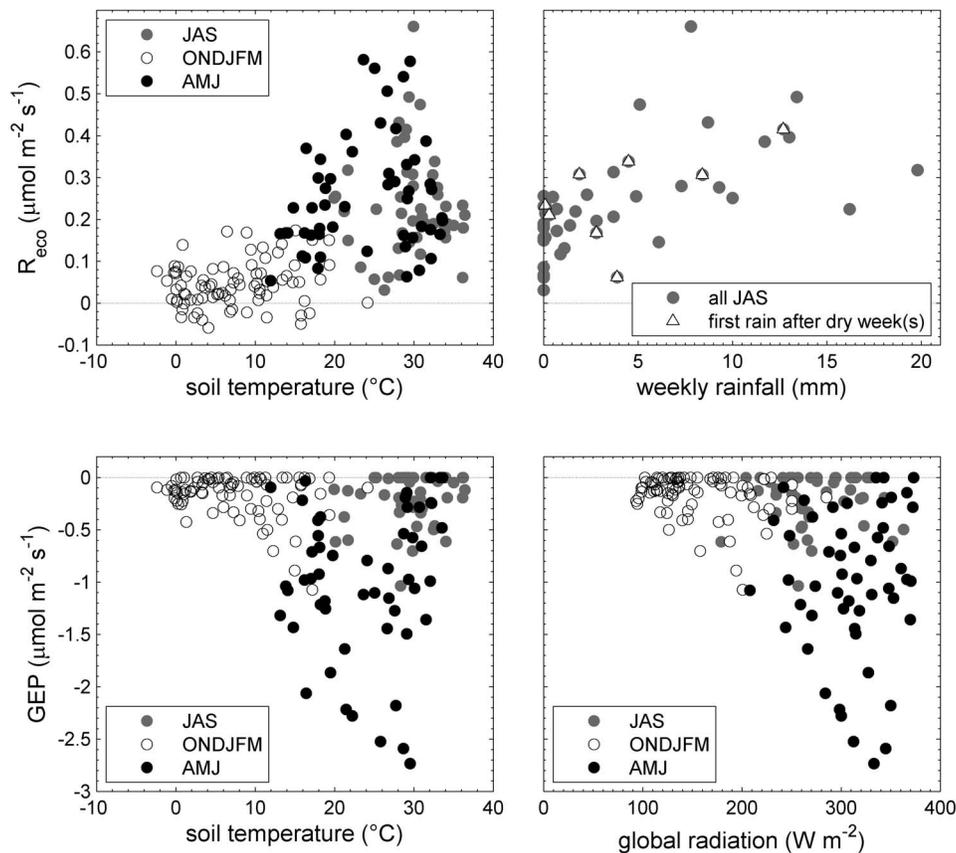


Figure 7. Relations between (top left) R_{eco} or (bottom left) GEP and soil temperature and (bottom right) GEP versus global solar radiation. Data are shown separately for the summer monsoon period (JAS, July–September), fall and winter (ONDJFM, October–March), and the period of spring productivity (AMJ, April–June). (top right) Relation between R_{eco} and total weekly rainfall during the summer monsoon period only. Weeks with rain that followed one or more dry weeks (no rainfall) are highlighted with open triangles. All values are weekly means, and all years are shown. Typical uncertainties can be inferred from Figure 6 (omitted here for clarity).

pattern is common in semiarid regions, including cold desert grasslands and shrublands of the United States [Fernandez et al., 2006; Obrist et al., 2004; Prater and DeLucia, 2006], Mediterranean grasslands [Aires et al., 2008; Xu and Baldocchi, 2004], Mojave Desert shrublands [Wohlfahrt et al., 2008], short-statured semiarid conifer forests [Breshears et al., 2005; Ivans et al., 2006; West et al., 2007], ponderosa pine forests [Thomas et al., 2009], and subalpine forests [Hu et al., 2010]. Notably, seasonal soil moisture patterns in the warm Chihuahuan and Sonoran deserts are different, being sometimes higher in summer than winter, because of the strong influence of the NAM [Kurc and Small, 2007; Scott et al., 2004, 2009].

[30] The overall pattern of moisture near the soil surface (5–10 cm) did not differ substantially during drought and favorable years at Corral Pocket (Figure 3). Deep soil moisture was different between years, particularly in fall and winter (Figure 9). Fall recharge of deep soil moisture was high in 2006 (Figure 3), leading to a productive growing season in 2007 (Figures 6 and 8). In contrast, fall soil moisture recharge in late 2005 (not shown) and early 2006 (Figure 3) was limited, and deep moisture at the start of the growing season in 2006 was low.

[31] The drought of the early 2000s significantly impacted grassland gas exchange. In the midst of the drought, 2001 and 2002 had low calendar year precipitation, and NEE and GEP were substantially reduced, especially in the peak spring growing season (Figure 6). R_{eco} was reduced during drought but less so than GEP. Seasonal patterns of ET were similar during the years, with exceptions noted below. Rain events in all years led to pulses of ET that lasted days to weeks (Figure 4). The pulsed nature of ET at Corral Pocket contrasts sharply with the typically smoother seasonal patterns of ET in forests [Amiro et al., 2006; Wilson and Baldocchi, 2000], more humid grasslands [Meyers, 2001; Wever et al., 2002], and grasslands of Mediterranean climates [Baldocchi et al., 2004], but is consistent with studies from other semiarid grasslands, shrublands, and forests [Ivans et al., 2006; Kurc and Small, 2004; Mielnick et al., 2005; Scott et al., 2006a]. Precipitation in arid regions arrives in pulses, and ecosystem CO_2 and water fluxes in arid biomes respond in a pulsed fashion [Jenerette et al., 2008; Potts et al., 2006a; Williams and Albertson, 2004; Williams et al., 2009]. The growing season periods of 2003 and 2007 at Corral Pocket had higher ET for sustained periods compared to the other years (Figure 4). This coincided with the

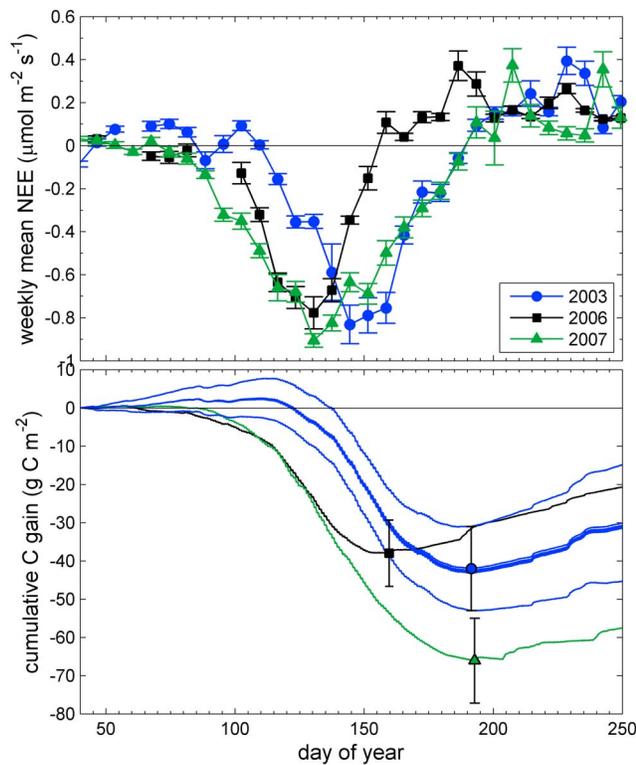


Figure 8. (top) Weekly net ecosystem exchange of CO_2 (means and standard errors) during 2003, 2006, and 2007, days 40–250 (mid-February to early September). (bottom) Cumulative ecosystem carbon gain using gap-filled NEE. The cumulative carbon gain for 2003 is shown calculated with (1) varying u^* thresholds between 0.1 and 0.4 m s^{-1} (tight cluster of blue lines) and (2) maximum and minimum errors assigned systematically to NEE (uppermost and lowermost blue lines) as described in section 2. The symbols and error bars represent the cumulative sum of NEE for each year using a u^* threshold of 0.1 m s^{-1} (symbols) and the maximum and minimum errors on NEE (error bars). Lines shown for 2006 and 2007 show only the cumulative sum using the base u^* threshold of 0.1 m s^{-1} , and the error bars have the same meaning as for 2003 (results were similar to 2003).

prevalence of deep soil moisture (Figures 3 and 9) and hence is likely due to transpiration rather than soil evaporation.

[32] Evaporative fraction, expressed as the ratio of evapotranspiration to available energy, or in our case $\text{LE}/(H + \text{LE})$, is often a function of soil moisture, particularly when moisture is limiting [Denmead and Shaw, 1962; Kelliher et al., 1993; Kurc and Small, 2004]. At Corral Pocket, $\text{LE}/(H + \text{LE})$ was strongly dependent on soil moisture during the soil dry-down period, regardless of depth considered or year (Figure 5). At higher soil moisture, the relationship between evaporative fraction and soil moisture usually saturates [Baldocchi et al., 2004; Kelliher et al., 1993]. Corral Pocket has low plant cover (Table 1), and the grass canopy is nearly fully removed by cattle grazing each year in the late fall or winter. The spring period of highest soil moisture coincided with the regrowth of the perennial bunchgrasses and evaporative fraction was inversely related to soil moisture at higher soil moisture as

the canopy developed (Figure 5), again regardless of depth or year. During the monsoon rains, $\text{LE}/(H + \text{LE})$ was linearly related to near-surface soil moisture (5 cm depth, Figure 6), but not to moisture at greater depth. While grasses are generally more shallowly rooted than shrubs, most cold desert grasses do not have much root biomass above 10 cm depth in summer when soils are hot [Abbott et al., 1991; Belnap and Phillips, 2001; Peek et al., 2005; Schwinning et al., 2003; West et al., 1972]. These results suggest that the majority of the ET flux during the monsoon at Corral Pocket was due to soil evaporation and not plant transpiration, in agreement with observations in the Chihuahuan desert [Kurc and Small, 2004].

[33] Gross ecosystem productivity occurred primarily in the spring (Figure 6), although there was some limited uptake in fall of 2002 following a strong monsoon season (Figure 6). Maximal hourly rates (not shown) of spring and fall net CO_2 uptake were 4 and $2 \mu\text{mol m}^{-2} \text{ s}^{-1}$, respectively, which are at the lower end of published NEE for water-limited ecosystems [Fu et al., 2009; Hastings et al., 2005;

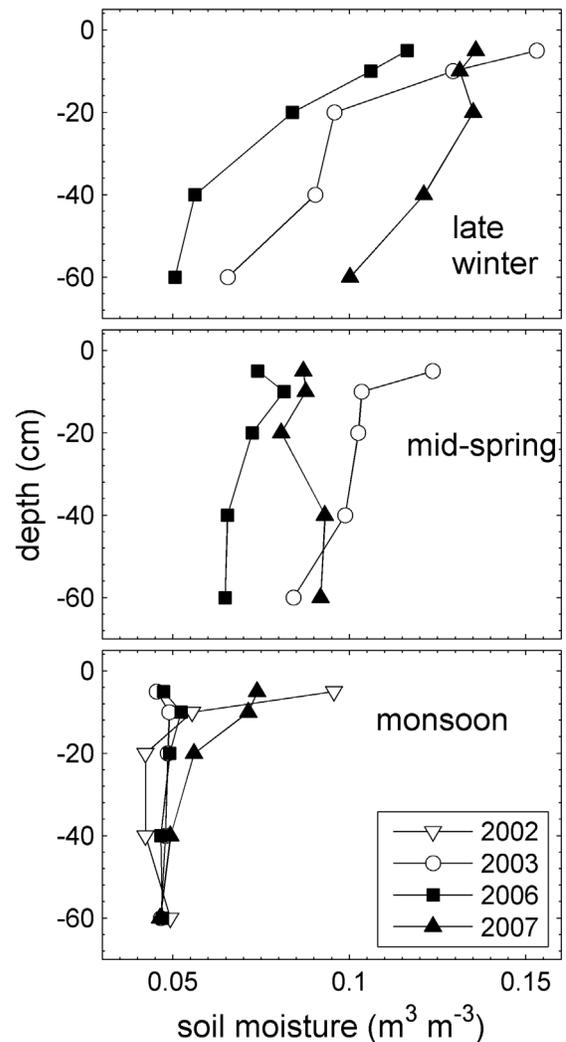


Figure 9. Soil moisture depth profiles during late winter (day of year (DOY) 50, 19 February), mid-spring (DOY 135, 15 May), and during the monsoon period (DOY 213, 1 August) in different years.

Table 4. Carbon Stocks at Corral Pocket During Early Spring, the Period of Maximum C Uptake, and Autumn in 2003^a

Depth	Category ^b	Time Period		
		Early Spring	Maximum C Uptake	Autumn
0–30 cm	aboveground green plant (a)	4.5 ± 1.2	12.2 ± 1.1	12.7 ± 2.8
	aboveground senesced plant (b)	3.3 ± 0.5	3.0 ± 0.6	5.3 ± 0.7
	surface litter (c)	12.9 ± 1.5	15.3 ± 2.1	18.8 ± 2.9
	dung (d)	16.4 ± 6.0	21.8 ± 10.7	4.3 ± 2.8
	roots (e)	95.6 ± 16.6	159.7 ± 31.5	191.9 ± 36.8
	<i>Hilaria</i> rhizome (f)	13.3 ± 1.3	13.6 ± 1.8	19.0 ± 2.2
	decaying organic matter (g)	137.6 ± 29.6	141.0 ± 37.1	176.4 ± 35.6
	organic bulk (h)	0.8 ± 0.1	1.3 ± 0.2	0.9 ± 0.1
	inorganic bulk (i)	4.8 ± 0.3	4.7 ± 0.4	5.1 ± 0.4
	30–100 cm	roots (j)	84.2 ± 20.6	nd
decaying organic matter (k)		23.5 ± 9.6	nd	nd
organic bulk (l)		3.6 ± 1.1	nd	nd
inorganic bulk (m)		28.0 ± 3.3	nd	nd
total aboveground plant (a + b)		7.8 ± 1.7	15.2 ± 1.7	18.0 ± 3.4
total belowground plant (0–30 cm) (e + f)		108.9 ± 17.9	173.3 ± 33.3	210.9 ± 39.0
total plant (roots above 30 cm) (a + b + e + f)		116.7 ± 19.6	188.5 ± 35.0	228.9 ± 42.4
total plant (all roots) (a + b + e + f + j)		200.9 ± 40.2	nd	nd
total soil organic (c + d + g + h + k + l)		194.8 ± 47.9	nd	nd
total soil inorganic (no caliche) (i + m)		32.8 ± 3.6	nd	nd

^aValues are means and standard errors. Units are g C m⁻² ground area. nd, no data.

^bEach category is labeled with a letter, and sums of different categories are indicated at the bottom. “No caliche” indicates that the measured soil inorganic pool did not include the caliche horizon.

Potts *et al.*, 2006b; Scott *et al.*, 2004, 2006a, 2006b, 2009; Veenendaal *et al.*, 2004; Wohlfahrt *et al.*, 2008; Xu and Baldocchi, 2004]. Soil respiration rates for grasslands and shrublands within a few kilometers of Corral Pocket are in the range 0–1.5 $\mu\text{mol m}^{-2} \text{s}^{-1}$ for most of the year [Fernandez *et al.*, 2006], with excursions as high as 6 $\mu\text{mol m}^{-2} \text{s}^{-1}$ following moisture addition [Fernandez *et al.*, 2006; Schaeffer and Evans, 2005]. The lower range is consistent with respiration at Corral Pocket as we did not observe respiratory CO₂ hourly exchange rates above $\sim 2 \mu\text{mol m}^{-2} \text{s}^{-1}$ following moisture pulses in the 5 year record (hourly data not shown).

[34] Weekly R_{eco} was maximally 0.6 $\mu\text{mol m}^{-2} \text{s}^{-1}$ and strongly related to temperature, but the relationship differed across the seasons (Figure 7). Respiration was highest during the spring, increasing as the vegetation canopy grew (Figure 6) and reflecting increased autotrophic respiration and increased leaf area. As the soil dried, respiration decreased (Figure 6) despite higher soil temperature, and the decrease was coincident with the seasonal senescence of the bunchgrasses. Weeks with more summer rain had higher ecosystem respiration (Figure 7), in agreement with manipulative studies that have found the magnitude of the soil respiratory response is correlated with the amount of rainfall received [Munson *et al.*, 2010; Sponseller, 2007]. Weekly R_{eco} following dry antecedent conditions was not higher than R_{eco} following weeks with rain as is sometimes observed for soil respiration [Austin *et al.*, 2004; Cable *et al.*, 2008]. This phenomenon may have been masked by the weekly averaging to obtain R_{eco} from NEE.

[35] GEP was related to temperature and radiation (Figure 7) but also to soil moisture. Cold temperature in fall and winter was associated with low GEP, and T was probably the most limiting factor for productivity during this period. During spring, when moisture was available, GEP was high (Figures 6 and 7), as was radiation. Separation of the importance of these covarying controls on GEP is difficult, but it appears that moisture is among the most important.

The drought years of 2001 and 2002 exhibited minimal CO₂ uptake during the spring, but 2003 and especially 2007 had high GEP in spring (Figure 6). Spring productivity was dependent on deep soil moisture, particularly in late winter (Figures 8 and 9). The year 2007 had the highest amount of moisture in the soil profile (Figure 9), the highest carbon uptake (Figure 8), and the highest cumulative ET of all years (170 mm during days 40–250), coinciding with relatively high plant cover [Belnap *et al.*, 2009]. In contrast, 2006 had much lower soil moisture, lower plant cover, lower carbon uptake, lower cumulative ET (80 mm), and a shorter growing season than 2007. Carbon uptake in 2003 was more similar to 2006 than 2007, despite deep soil moisture recharge by late winter rains (Figures 3, 8, and 9). Deep moisture was not measured during late winter in 2001 and 2002, but from the seasonal pattern of limited CO₂ uptake (Figure 6) and limited ET (Figure 4), combined with the low precipitation in those years (Figure 2), we infer that deep soil moisture was low during the drought. Plant cover was very low in 2001 [Belnap *et al.*, 2009].

[36] The magnitude of CO₂ exchange determined by eddy covariance compares favorably with repeated observations of carbon stocks in 2003, although the flux-based estimate was lower. Flux measurements indicated $42 \pm 11 \text{ g C m}^{-2}$ of net uptake between days 40 and 190 (local minimum in Figure 8 (bottom)). Live plant stocks above 30 cm soil depth increased by $72 \pm 55 \text{ g C m}^{-2}$ between early spring and the period of maximum CO₂ uptake (Table 4). The latter is likely an underestimate as we did not characterize changes in deep root biomass, which may become more important as soils dry in summer [Peek *et al.*, 2005]. Given the large data gaps and the unlikely ability of the gap-filling routine to adequately model response to moisture pulses, uncertainty in the cumulative CO₂ flux estimate is difficult to quantify. Energy balance closure was generally good (Table 3), which provides some level of confidence in the turbulent fluxes on average. Varying the u^* threshold for acceptance of flux data under highly stable conditions only altered the cumulative

carbon gain by $<1 \text{ g C m}^{-2}$. Applying a fairly high systematic bias to NEE altered the cumulative sum for 2003 by $\pm 11 \text{ g C m}^{-2}$ (Figure 8 (bottom)). Measurements of NEE were highly consistent within a given week, with fairly low coefficient of variation (compare the error bars to the means in Figure 6a), which indicates that the uncertainty in NEE was considerably smaller than the measured values. The relative ranking of carbon gain between years (Figure 8) is likely correct (e.g., 2003 and 2006 were similar, 2007 was higher).

[37] Calendar year precipitation alone does not explain the difference in carbon uptake between years. The year 2006 had the highest annual precipitation postdrought (Figure 2a) but lower net CO_2 uptake (Figures 6 and 8) and lower ET (Figure 4) compared to 2007, which had much lower total annual precipitation (Figure 2a). Most of the precipitation during 2006 fell during October–December (Figure 2b), resulting in deep soil moisture recharge and a highly productive 2007 growing season (Figures 3 and 6). Hence, in the cold deserts of North America, the seasonality of precipitation is a better predictor of grassland productivity than the total annual amount. Precipitation during the fall and winter (October–March) is considerably more important for plant activity than spring or summer moisture, and the growing season (April–June) is separated in time from the precipitation that drives it. This result contrasts with many other world grasslands. Productivity of tallgrass and shortgrass prairies of the Great Plains [Flanagan et al., 2002; Heisler-White et al., 2008; Suyker et al., 2003] and warm deserts [Emmerich and Verdugo, 2008; Mielnick et al., 2005; Scott et al., 2006b, 2009] of North America and grasslands of Mongolia [Fu et al., 2009; Li et al., 2005] is primarily dependent on summer precipitation and occurs during the summer growing season. Mediterranean grasslands are dependent on winter rain, and the growing season is concurrent with the winter precipitation [Aires et al., 2008; Xu and Baldocchi, 2004].

[38] The dominant grasses *Hilaria jamesii* (C_4) and *Stipa hymenoides* (C_3) are capable of responding to rain events as small as 6 mm with enhanced photosynthesis and transpiration [Schwinning et al., 2003]. However, productivity and plant water status of cold desert plants are more limited by winter drought than by summer drought [Schwinning et al., 2005a, 2005b]. Seasonal changes in relative physiological activity of C_3 and C_4 grasses have been demonstrated at the whole ecosystem level in the Great Plains [Lai et al., 2006; Still et al., 2003]. We hypothesized that both C_3 and C_4 grasses at Corral Pocket would respond to monsoon rains with net CO_2 uptake, as has been shown for C_4 -dominated ecosystems of Arizona and New Mexico during the NAM [Emmerich and Verdugo, 2008; Mielnick et al., 2005; Scott et al., 2006b, 2009]. We only rarely observed net uptake of CO_2 in summer (Figure 6), even with above-average monsoon rains in 2001 and 2002 (Figure 2b). Despite the abundance of C_4 grasses, this grassland did not respond to summer rain with substantial CO_2 uptake, refuting our hypothesis.

[39] Climate change is already occurring in the western United States, with evidence that the length of the frost-free season has increased, spring is warmer and earlier, snowmelt and runoff are earlier, and a greater fraction of spring precipitation falls as rain rather than snow [Cayan et al.,

2001; Knowles et al., 2006; McCabe and Clark, 2005; Mote et al., 2005; Mote, 2006; Stewart et al., 2005]. There is broad consensus among climate models that more arid conditions will be prevalent in the west in the next century [Seager et al., 2007]. Increased aridity is likely to mean generally drier future soil conditions in the region [Wang, 2005], perhaps with increased frequency of heat waves and larger summer precipitation events [Diffenbaugh et al., 2005]. Because CO_2 fluxes in cold desert grasslands are highly dependent on soil moisture, drier soils will lead to lower net CO_2 uptake in the future, as we observed during the drought. However, the seasonality of soil moisture will modulate any effect of total moisture on CO_2 fluxes. The North American monsoon is not well reproduced by the current generation of general circulation models [Lin et al., 2008], and predictions about its future spatial extent are weak at best. While important for the warm deserts, the NAM may continue to be unimportant for cold desert grasslands, because of their dependence on winter precipitation. If future aridity is exhibited as decreased winter precipitation, the productivity of these grasslands will decrease. However, at least one study predicts more winter precipitation in the Colorado River basin over the next century [Christensen and Lettenmaier, 2007], which on its own would enhance productivity of the cold desert grasslands. Such predictions are confounded by other ecological factors common in the deserts of North America such as soil surface disturbance by grazing, fire, or recreation [Belnap et al., 2009; Iverson et al., 1981; Neff et al., 2005; Obrist et al., 2004], or invasion by annual grasses [Evans et al., 2001; Novak and Mack, 2001; Sperry et al., 2006].

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