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Long-term measurements of CO_2 flux and evapotranspiration in a Chihuahuan desert grassland

P. Mielnick[a,*], W.A. Dugas[a], K. Mitchell[b], K. Havstad[b]

[a]*Blackland Research and Extension Center, Texas Agricultural Experiment Station, 720 East Blackland Road, Temple, TX 76502, USA*
[b]*US Department of Agriculture, Agricultural Research Service, Las Cruces, NM 88003, USA*

Abstract

We measured CO_2 and evapotranspiration (ET) fluxes above a Chihuahuan desert grassland from 1996 through 2001. Averaged across six years, this ecosystem was a source (positive flux) of CO_2 in every month. Over that period, sustained periods of carbon uptake (negative flux) were rare. Averaged across all months, we estimated that CO_2 was being released into the atmosphere at a rate of $+1.5\,g\,CO_2\,m^{-2}\,d^{-1}$ ($+0.40\,g\,C\,m^{-2}\,d^{-1}$) or $+533\,g\,CO_2\,m^{-2}\,yr^{-1}$ ($+145.3\,g\,C\,m^{-2}\,yr^{-1}$). Average ET was $299\,mm\,yr^{-1}$ with a monthly range of $0.15\,mm\,d^{-1}$ in December to $2.15\,mm\,d^{-1}$ in August.
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1. Introduction

Arid and semi-arid lands cover much of the western and southwestern US (Huenneke et al., 2002) and encompass considerable areas of the rest of the globe.

*Corresponding author. Blackland Research Centre, 808 E. Blackland Road, Temple, USA. Fax: +1-254-770-6561.

E-mail address: mielnick@brc.tamus.edu (P. Mielnick).

Globally, desert shrublands comprise a land area about equal in magnitude to tropical rain forests ($\sim 18 \times 10^8$ ha) and about 20% greater in area than cultivated lands (Schlesinger, 1991). In addition, desertification of marginal agricultural land continues, a process that may be accelerated by future changes in climate (Schlesinger et al., 1990; Fredrickson et al., 1998). Thus, present and future arid ecosystems comprise a large and growing portion of the Earth's terrestrial areas.

Typically, arid ecosystems have vegetative communities with low-leaf areas, above-ground biomass, and net productivity (Schlesinger, 1991). In the US, vegetation in a typical arid community is often a mixture of C_3 shrubs and C_4 grasses. However, encroachment of C_3 shrubs is gradually converting C_4 grasslands into C_3 shrublands as a result of overgrazing, fire suppression, increasing atmospheric CO_2 favoring C_3 over C_4 plants, or long-term decadal increases in wintertime precipitation (Neilson, 1986; Gibbens et al., 1996; Polley et al., 1996; Brown et al., 1997; Johnson et al., 2000; Peters, 2000). The complexity of these ecosystems suggests that multiple drivers, rather than a single determinant factor, are involved in the vegetative transition. Differences in vegetation types in arid ecosystems, however, do not significantly affect evapotranspiration (ET) rates (Dugas et al., 1996), likely because water limitations dominate this process as opposed to differences in community structure, process, or function. In the study by Dugas et al. (1996), ET rates were essentially equal for three plant communities on sandy soils dominated by either *Bouteloua eriopoda* (black grama), *Prosopis glandulosa* Torr. var. *glandulosa* (honey mesquite), or *Larrea tridentata* (DC.) Cov.) (creosotebush).

Given the water-limited nature of arid ecosystems (Schlesinger et al., 1990), vegetative productivity and species diversity are generally precipitation driven (Neilson, 1986; Gosz, 1993; Peters, 2000) and can be influenced by episodic events of extreme precipitation that recharge deeper soil layers for use by deep-rooted plants such as shrubs or by smaller, more frequent rains that tend to enhance growth of the more shallow-rooted grasses (Gibbens and Lenz, 2001). The productivity and species community of arid biomes also are influenced by precipitation timing. Variability of within-season precipitation can control the robustness of each plant growth stage and thus the ability of a given species to disperse and establish in that season (Peters, 2000). Annual precipitation variability also can amplify the effect of larger regional, long-term climate variations (Neilson, 1986; Gosz, 1993).

In arid regions, seasonal distribution of precipitation is a critical factor in CO_2 uptake and ET fluxes. Carbon dioxide fluxes related to primary productivity (i.e. uptake) are usually assumed to be minimal in arid areas because of lack of precipitation to support substantive vegetative growth (Schlesinger, 1991). In addition to soil microbial and root activity, a potential carbon source in arid ecosystems normally not accounted for in most estimates of carbon exchange is dissolution of the soil carbonate ($CaCO_3$) pool following precipitation. Globally, the soil holds the third largest pool of carbon (750–950 Pg C) (Monger and Martinez-Rios, 2001) and may be an active participant in short-term net CO_2 fluxes, particularly in arid areas with large amounts of soil carbonates (Emmerich, 2003). Few long-term studies of carbon and water flux are available for arid ecosystems.

The objective of this research was to quantify CO_2 and ET fluxes above a Chihuahuan desert grassland on the Jornada Experimental Range (JER) in southern New Mexico and to relate these fluxes to precipitation and vegetation characteristics at the site.

2. Materials and methods

2.1. Site description

Fluxes of CO_2 and ET were measured from 1996 through 2001 in a 30-ha grassland exclosure (fenced to keep out large, but not small, herbivores) at the JER (1320 m elevation, 32.60°N, 106.75°W). The JER is located about 40 km northeast of Las Cruces, New Mexico and is managed by the USDA, Agricultural Research Service. The climate at this location is arid. Long-term (1915–1997) average annual precipitation is 272 mm, with 58% of this falling from July through September (Table 1). During the six years of our study, annual precipitation ranged from 177 mm in 2001 to 337 mm in 1997 (Table 1). Temperatures reach their maximum in June and are minimum in January. The length of the effective growing season, i.e. when both precipitation and temperatures are favorable for plant growth, is primarily controlled by precipitation timing and typically extends from July through September (Paulsen and Ares, 1962).

Soil type is a Typic Haplargid, fine loamy, mixed, thermic soil consisting of 13% clay, 12% silt and 75% sand in the top 0–0.3 m with an average bulk density of

Table 1
Monthly precipitation totals (mm) for 1996 through 2001 and the long-term monthly average precipitation (1915–1997) at the study site. Days of year for each month are shown in parentheses

Month	1996[a]	1997	1998	1999	2000	2001	Long-term average
Jan (1–31)	3.8	20.6	2.5	7.9	0.0	16.0	18
Feb (32–59)	0.0	18.8	13.0	0.0	0.0	15.5	10
Mar (60–90)	0.0	13.0	9.9	7.6	11.2	0.5	6
Apr (91–120)	13.5	4.6	1.5	1.5	2.0	2.3	5
May (121–151)	0.0	12.4	0.0	1.5	0.3	10.2	10
Jun (152–181)	49.8	40.1	3.0	61.0	123.4	3.3	20
Jul (182–212)	24.6	48.3	57.2	34.5	18.0	14.5	72
Aug (213–243)	41.1	86.1	25.4	58.7	18.8	34.0	52
Sep (244–273)	58.4	34.3	9.9	33.5	0.3	66.8	34
Oct (274–304)	18.5	18.0	55.4	25.1	61.7	0.0	15
Nov (305–334)	3.8	11.7	16.8	0.0	32.3	5.8	12
Dec (335–365)	0.0	29.5	4.1	11.4	2.5	7.9	19
Totals	214	337	199	243	271	177	272

[a]Leap year: add 1 day to last day of February to 31 Dec.

$1.45\,\mathrm{g\,cm\,cm^{-3}}$. Soil organic carbon averages $<1\%$ in the top $0.3\,\mathrm{m}$. There is an underlying caliche ($CaCO_3$) layer about 0.3–$0.5\,\mathrm{m}$ below the soil surface and numerous calcareous outcrops in the area.

Vegetation within the exclosure was dominated by the C_4 perennial, black grama grass (*B. eriopoda*), other perennial grasses and small populations of annual grasses and forbs, perennial forbs, and shrubs (Gibbens et al., 1996). Subdominate species included yucca (*Yucca elata* Engelm.) and mesquite (*P. glandulosa* Torr. Var. *glandulosa*). Vegetation in the exclosure was typical of historic black grama grasslands of this region. However, due to the gradual encroachment of C_3 shrubs in surrounding areas with sandy soils and the periodic removal of mesquite individuals within the exclosure, the vegetation in the exclosure is less similar to current dominant plant communities on sandy soils at the JER. There is a gradual transition between the grass-dominated vegetation community within the exclosure and surrounding shrub-dominated vegetation community.

2.2. CO₂ flux measurements

Surface CO_2 and ET fluxes are commonly measured using one of two micrometeorological methods: the eddy covariance method, a direct measurement of flux (Baldocchi et al., 1996; Valentini et al., 1996; Meyers, 2001; Suyker and Verma, 2001) and the Bowen Ratio energy balance method, an indirect or calculated flux measurement (Dugas, 1993; Dugas et al., 1999; Angell et al., 2001; Frank and Dugas, 2001; Sims and Bradford, 2001; Frank, 2002). Both methods have been successfully used in various ecosystems.

Twenty-minute, above-canopy CO_2 fluxes were calculated from Bowen ratio/ energy balance measurements (Dugas, 1993) made for various periods in 1996 through 2001 (Table 2). Fluxes calculated using the Bowen ratio/energy balance method have been shown to be similar to daytime soil CO_2 fluxes measured on a bare soil using a soil respiration chamber (Dugas, 1993), daytime canopy fluxes calculated from leaf gas exchange measurements for three C_4 grasses (Dugas et al., 1997) and mesquite savannas (Ansley et al., 2002), daytime canopy fluxes calculated from small canopy chamber measurements above sagebrush (*Artemisia tripartita* Rydb. spp.

Table 2
First and last day of flux measurements and number of days of valid flux measurements for each year of the study

Year	First day	Last day	Number of data days
1996	176 (25 June)	365 (31 Dec.)	132
1997	1 (1 Jan.)	270 (27 Sept.)	236
1998	76 (17 Mar.)	240 (28 Aug.)	87
1999	71 (12 Mar.)	365 (31 Dec.)	230
2000	1 (1 Jan.)	300 (27 Oct.)	295
2001	1 (1 Jan.)	310 (6 Nov.)	263

Rupicola Beetle) (Angell et al., 2002), daytime canopy fluxes measured by several methods above sorghum (*Sorghum bicolor* L.) (Twine et al., 2000), and daytime and night-time tallgrass prairie fluxes measured using eddy covariance instrumentation (Dugas et al., 2001).

In this study, Bowen ratios were calculated from 20-minute average temperature and humidity gradients measured every 2 s at 2 and 4 m above the soil. Sensible heat flux was calculated from the Bowen ratio, average net radiation, and soil heat flux (Tanner, 1960). The latter two variables were measured using a net radiometer (Model Q7.1, REBS, Seattle, WA) at 1.5 m height, three soil heat flux plates (Model HFT-3, REBS), and three soil temperature sensors. Yearly calibration of the net radiometer was conducted above a grass surface using a laboratory standard. Turbulent diffusivity, assumed equal for heat, water vapor, and CO_2, was calculated (Tanner, 1960) using the sensible heat flux and measured temperature gradient. CO_2 fluxes, corrected for vapor density differences at the two heights (Webb et al., 1980), were calculated as a product of the turbulent diffusivity and the average CO_2 gradient that was measured along with the humidity gradient with an infrared gas analyzer (Model 6262, LiCor, Inc., Lincoln, NE, USA). The analyzer was calibrated every three months using an external canister of a CO_2 gas at a known concentration. Daily CO_2 fluxes were calculated from 20-minute values. Daily averages for each month were calculated using all available days for that month during the six-year period (negative CO_2 flux = uptake).

When sensible heat flux was opposite to the temperature gradient sign (Ohmura, 1982), turbulent diffusivity was calculated using wind speed at 3.4 m, atmospheric stability, and canopy height (Dugas et al., 1999). This alternative method for calculating diffusivity was used about 23% of the time—almost entirely at night when fluxes were small. When Bowen ratios approached −1, typically at dawn and dusk when fluxes were near zero, diffusivity calculations became unstable and 20-minute fluxes were linearly interpolated. This occurred less than 3% of the time. Sensor instrumentation was located near the center of the grassland exclosure. Fetch was ca. 0.5 km when winds were from the south/southwest direction (typical for the growing season) and > 1 km when winds were from the north or east. If more than 5% of a single day's 20-minute fluxes were missing, that day was deleted from the analysis. Overall, a total of 1243 days were used in our analysis (Table 2).

2.3. Evapotranspiration

Twenty-minute values of ET were calculated from measurements of net radiation, soil heat flux, and the Bowen ratio (Tanner, 1960). The same interpolative criteria that were used for CO_2 fluxes (see above) were applied to ET fluxes when Bowen ratios approached –1. Daily ET totals were calculated from the 20-minute values.

2.4. Ancillary measurements

Precipitation was continuously measured at the study site. Biomass and leaf area were measured from four, l-m^2 quadrats located about 5 m apart within the

Table 3
Average above-ground biomass ($g\,m^{-2}$) for grasses, forbs, shrubs, and litter at the study site from 1998 through 2001

Vegetation component	Above-ground biomass ($g\,m^{-2}$)			
	1998	1999	2000	2001
Grasses	68.2 ($\pm$12.9)	75.7 ($\pm$20.2)	125.1 ($\pm$20.8)	75.3 ($\pm$17.9)
Forbs	3.3 ($\pm$1.2)	22.0 ($\pm$5.4)	21.3 ($\pm$8.5)	0.7 ($\pm$0.3)
Shrubs	1.2 ($\pm$1.3)	0.4 ($\pm$0.4)	0.2 ($\pm$0.4)	0.8 ($\pm$0.5)
Litter	56.4 ($\pm$12.1)	88.9 ($\pm$76.3)	76.3 ($\pm$16.9)	84.7 ($\pm$10.6)
Total	129.1 ($\pm$6.9)	187.0 ($\pm$25.6)	223.0 ($\pm$11.65)	161.5 ($\pm$7.3)

Values in parenthesis are standard errors calculated from 4 replicates, spring and fall sampling periods combined.

exclosure. Above-ground biomass was measured twice per year for four years on May 15 and November 6 in 1998, June 30 and October 27 in 1999, July 14 and November 8 in 2000, and July 25 and November 26 in 2001 (Table 3). Leaf area was measured in the spring and fall in 1998, 1999, and 2000 using a LI-2000 belt-driven leaf area meter (LiCor, Lincoln, Nebraska) or an LI-3000 scanner. Leaf area indices ranged from 0.25 to 1.0.

2.5. Statistical analysis

As is the norm for most flux measurements, one micrometeorological station was used in this study site. This precluded standard, agronomic statistical analysis using the traditional multiple replicates over space. We characterized the temporal variability of flux measurements (i.e. standard error) by replicating over the six years for each month of flux measurements. For vegetation, variability was assessed using the standard error calculated from four spatial replicates.

3. Results and discussion

3.1. Vegetation

Over the four-years (1998–2001), vegetation in the exclosure was dominated by grass species (Table 3), especially the perennial C_4 black grama grass *B. eriopoda*. The largest above-ground biomass was measured in 2000, a year with the highest precipitation during June and July compared to the other six years (Table 1). In 2000, grasses responded to the increased precipitation with a 60% increase in biomass when compared to the previous year (Table 3). Summer rainfall, as opposed to winter or spring precipitation, enhances plant growth from flowering to dispersal, especially for *B. eriopoda*, a native grass species that is particularly responsive to summer precipitation, and which constitutes the dominant vegetation at the site

(Neilson, 1986; Peters, 2000). The annual sum of above-ground biomass in the four years it was measured was linearly related to the total yearly precipitation (slope = 0.898, $r^2 = 0.71$, results not shown). Average above-ground biomass values from our study were similar to those reported by Gibbens et al. (1996) for a *B. eriopoda* community and Huenneke et al. (2002) for a Chihuahuan desert grassland.

3.2. Daily CO$_2$ flux

Daily CO_2 flux varied markedly within and between years because of differences in radiation and precipitation (Fig. 1), the latter varying by nearly a factor of two between years (Table 1). Both seasonal timing and total amount of precipitation received during active stages of plant growth (June–September) affected the CO_2 flux response. During much of the year, daily CO_2 fluxes were near zero or slightly positive (i.e. CO_2 efflux), especially when precipitation amounts were small (e.g. days 300–365 in 1996 and 1999, days 1–140 in 1997, and days 1–151 in 2000). However, large positive CO_2 fluxes were often measured in association with precipitation during the growing season (e.g. days 175 in 1996, day 190 in 1997, days 202–207 in 1998, days 160–165 in 1999, days 152 and 153 in 2000, days 256 and 257 in 2001). During winter and spring, the magnitude of positive fluxes associated with precipitation (e.g. days 1–150 in 1997) was relatively small compared to the large positive fluxes in summer when soil temperatures were warmer. A portion of the large positive CO_2 effluxes in the summer can be attributed to the increased soil microbial and root activity in warmer soils as soil CO_2 efflux from those sources is strongly related to soil temperature and moisture (Mielnick and Dugas, 2000). However, abiotic sources of CO_2 may also have been a factor. Dissolution of calcium carbonate, a mineral present in large amounts within soils and surface outcrops in the area (Monger and Gallegos, 2000), may have contributed to the pulse of CO_2 that frequently followed precipitation. Emmerich (2003) found a similar pattern of rainfall followed by CO_2 efflux in southeastern Arizona. He measured CO_2 fluxes over the dormant winter season when the contribution to CO_2 efflux by plants, roots, and soil microbial populations were assumed to be negligible due to senescent vegetation and cool soil temperatures, and attributed CO_2 efflux following precipitation during these wintertime periods to dissolution of soil carbonate stores at those sites. Across the six-years, the maximum daily positive CO_2 flux in our study was about $+40\,g\,CO_2\,m^{-2}\,d^{-1}$ with the exception of the wet fall of 1997 when fluxes reached a maximum of $+50\,g\,CO_2\,m^{-2}\,d^{-1}$.

Negative daily CO_2 fluxes (indicating CO_2 uptake) were primarily confined to the July–September growing season (days 182–273) and were usually short in duration. In the wet year of 2000, however, the duration of negative fluxes lasted about 70 days, likely due to 50% higher precipitation in June and July (142 mm) as compared to the other 6 years of this study (Table 1). Uptake during July 2000 averaged $-16\,g$ $CO_2\,m^{-2}\,d^{-1}$, the only month within the six-year period when this ecosystem was a large sink for CO_2 (Fig. 2).

Although total yearly precipitation in 1997 exceeded that of 2000 (Table 1), precipitation was more evenly distributed throughout the year in 1997 (21% greater
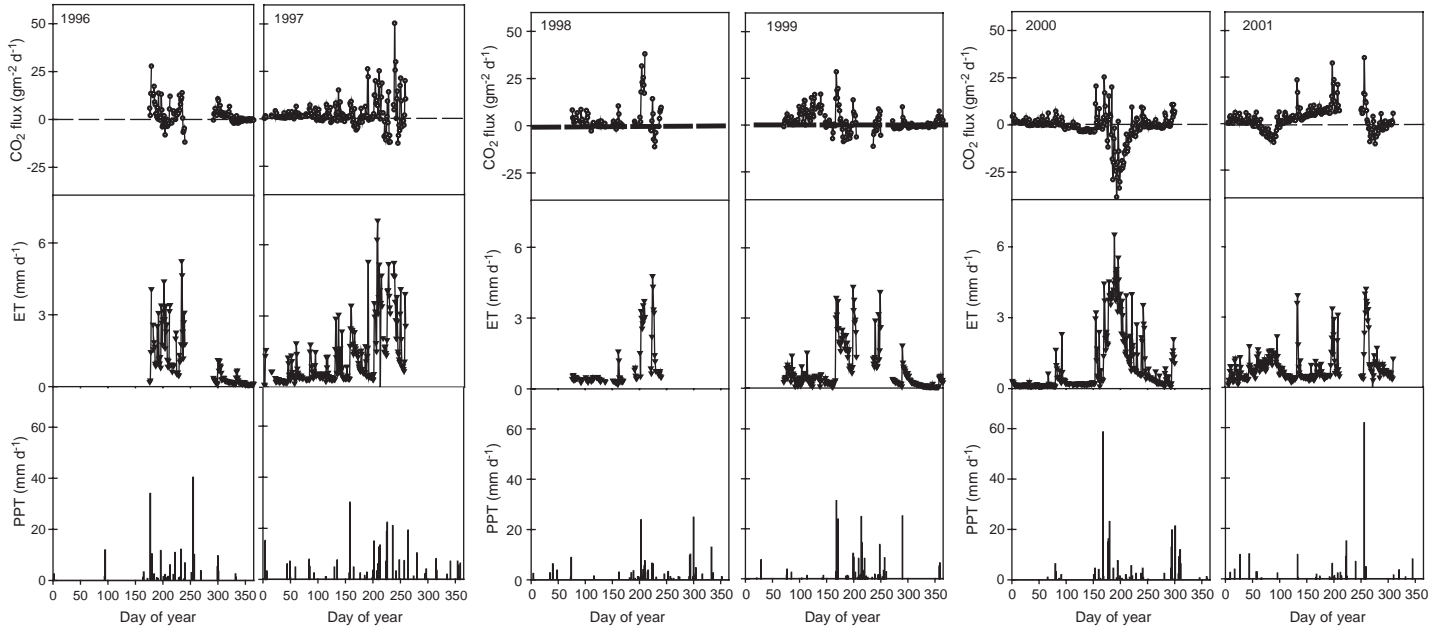
Fig. 1. Daily total CO_2 flux, evapotranspiration (ET), and precipitation (PPT) for 1996 through 2001. (Negative CO_2 flux indicates uptake).
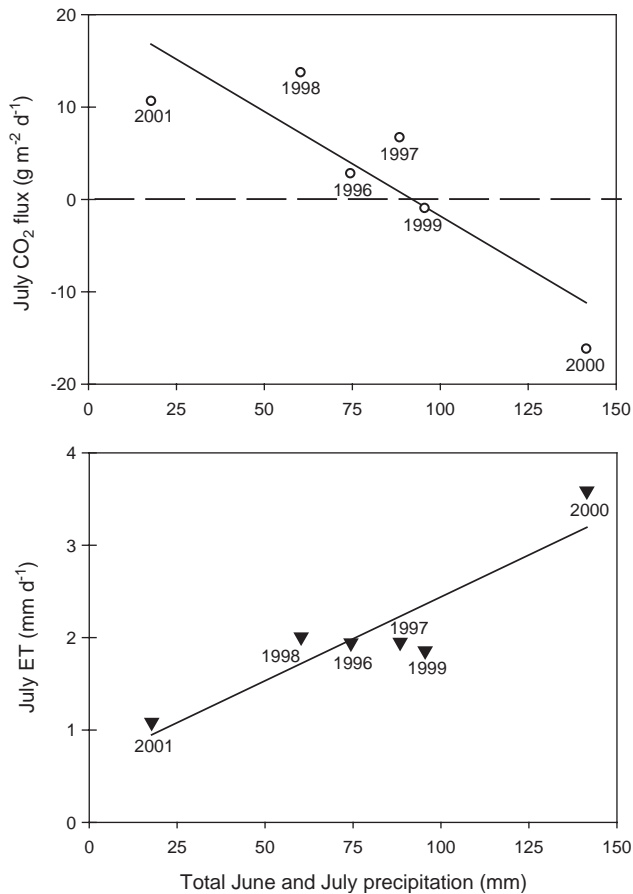
Fig. 2. Average daily July CO_2 flux and evapotranspiration (ET) versus total June and July precipitation for 1996 through 2001. ($r^2 = 0.75$ for CO_2 and 0.83 for ET.)

precipitation days) as compared to an uneven distribution pattern in 2000. In 2000, the majority of the precipitation was concentrated during the summer growing season with the single largest rainfall of the year (59 mm) falling on day 168 (mid-June) in the middle of the active growth period. Most of the remainder of that year's precipitation fell between days 150 and 250 (June–early September). Summer rains are important because they fall at a time when many grass species are at a critical stage in their life cycle and are also photosynthesizing at maximum capacity and are thus fully able to benefit from the additional moisture (Neilson, 1986; Gosz, 1993; Peters, 2000). The higher CO_2 uptake in July 2000 (Fig. 1) resulted in a 50–100% greater above-ground grass biomass in 2000 as compared to 1998, 1999, and 2001 (Table 3), however, there was no significant relationship between the annual CO_2 fluxes and the annual sum of above-ground biomass over four years of this study.

A possible reason for this may be the contribution of the inorganic component to net CO_2 flux that was not related to plant biomass accumulation and long periods of near zero fluxes interspersed with a few days of large positive fluxes related to precipitation.

Across all years of this study, July CO_2 fluxes were linearly related to June and July cumulative precipitation (Fig. 2, $r^2 = 0.75$). Thus, timing as well as total amounts of precipitation appeared to control the CO_2 flux response. Negative CO_2 fluxes were sometimes preceded by a period of large positive fluxes that occurred in association with the first substantial precipitation during the growing season (e.g. days 200–205 in 1998, days 170–180 in 1999, and days 152–170 in 2000). This initial, short-lived, positive pulse of CO_2 followed by a longer period of negative (uptake) fluxes may have been influenced by soil $CaCO_3$ dissolution (positive flux) from initial precipitation (Emmerich, 2003) followed by a period of CO_2 uptake (negative flux) due to enhanced photosynthesis. Maximum daily negative flux (e.g. uptake) for all years was about $-11\,\mathrm{g}\ CO_2\,\mathrm{m}^{-2}\,\mathrm{d}^{-1}$ except for 2000 when it approached—$40\,\mathrm{g}$ $CO_2\,\mathrm{m}^{-2}\,\mathrm{d}^{-1}$.

3.3. Daily evapotranspiration

Maximum daily ET rates occurred during the growing season from July through September (days 182–273) when precipitation and radiation were high (Fig. 1). Rates of ET were similar to those reported by Dugas et al. (1996) for the same site. In this study, maximum daily ET was about $5\,\mathrm{mm}\,\mathrm{d}^{-1}$ or less for 1996, 1998, 1999, and 2001, and was about $7\,\mathrm{mm}\,\mathrm{d}^{-1}$ in 1997 and 2000. Rates of ET in July also were related to the cumulative June and July precipitation (Fig. 2, $r^2 = 0.83$) (note the fit of the regression line was heavily influenced by the 2000 and 2001 data). During the dormant period, large ET rates were often associated with large positive CO_2 fluxes, both of which were associated with precipitation (Fig. 1). When followed by a period without precipitation, ET rates after a day with at least $10\,\mathrm{mm}\,\mathrm{d}^{-1}$ of rain showed the classic pattern of steadily declining ET over the following week or two (e.g. days 300–365 in 1996, days 290–350 in 1999, days 91–150 in 2000, and days 260–310 in 2001).

3.4. Six-year average CO_2 flux and ET

Averaged across the six years of our study, this ecosystem was a source for CO_2 in every month (Fig. 3). Average daily fluxes varied from $+0.39\,\mathrm{g}\ CO_2\,\mathrm{m}^{-2}\,\mathrm{d}^{-1}$ in December to $+3.36\,\mathrm{g}\ CO_2\,\mathrm{m}^{-2}\,\mathrm{d}^{-1}$ in June. The number of days that fluxes were measured, expressed as a percentage of the total number of available days, varied from 85% in July to 33% in December (Fig. 3).

Precipitation events in arid ecosystems lead to a variety of responses. Relatively large amounts of precipitation falling during the growing season can stimulate both CO_2 uptake by photosynthesis and CO_2 loss from plant and soil respiration. In soils with a large inorganic carbon pool, a potential source of CO_2 efflux may come from soil carbonate dissolution (Emmerich, 2003). Precipitation received during dormant
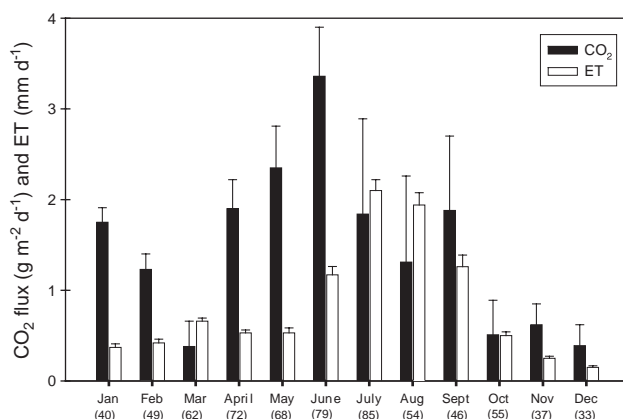
Fig. 3. Six-year averages of daily CO_2 flux and evapotranspiration (ET) for each month. Numbers in parentheses, in percent, represent the number of days in a month that measurements were made. Standard error bars represent between-year variability for each month.

or inactive stages of plant growth, however, typically only stimulate a CO_2 loss. In the sparsely vegetated and water-limited ecosystem of the Chihuahuan Desert grassland, CO_2 fluxes were seasonally variable and highly dependent on precipitation timing (Fig. 1). During much of the year, daily fluxes were near zero with brief periods of large positive CO_2 fluxes and high ET rates following precipitation.

Rates of ET, which are primarily controlled by radiation, precipitation, and leaf area, followed an expected seasonal trend of lower rates in the winter months and higher rates during the growing season, depending upon precipitation rates (Fig. 3). Six-year averages of ET ranged from $0.15\,\mathrm{mm\,d^{-1}}$ in December to $2.15\,\mathrm{mm\,d^{-1}}$ in August.

Variability across individual months was relatively large for CO_2, but less for ET (Fig. 3). Absolute variability in CO_2 flux tended to be higher during the growing season (June–September) and lower during late winter (January and February), although the highest coefficient of variation (CV) for a single month was in March ($>900\%$). High variability in CO_2 flux can be influenced by several factors. First, the amplitude of the daily CO_2 flux measurement can be large and can be either positive or negative depending on temperature, radiation, precipitation, and/or plant growth stage. As observed in our six-year study, precipitation and biomass varied substantially among months and years (Tables 1 and 3). Second, dissolution of soil carbonates following precipitation may add another source of variability to CO_2 fluxes (Emmerich, 2003). Finally, soil respiration, a component of net CO_2 flux, is spatially and temporally heterogenous and highly dependent upon soil temperature and soil moisture (Mielnick and Dugas, 2000).

The coefficient of variation for ET was highest in May (116%) and lowest in March (66%). The lower variability of ET as compared to CO_2 is related to the fact that ET is always greater than zero, as compared to the measurement of CO_2 flux, which can be either positive or negative.

4. Conclusions

Our study quantified six years of CO_2 and water vapor fluxes in the Chihuahuan Desert at the JER in southern New Mexico. Six-year average monthly CO_2 and water vapor fluxes were generally small and positive in every month (CO_2 source). Averaged monthly ET fluxes ranged from $0.15\,mm\,d^{-1}$ in December to $2.10\,mm\,d^{-1}$ in July, averaging $299.3\,mm\,yr^{-1}$ (Fig. 3). Averaged across six years, CO_2 fluxes ranged from $+0.39\,g\,CO_2\,m^{-2}\,d^{-1}$ ($+0.12\,g\,C\,m^{-2}\,d^{-1}$) in December to $+3.36\,g$ $CO_2\,m^{-2}\,d^{-1}$ ($+0.92\,g\,C\,m^{-2}\,d^{-1}$) in June (Fig. 3).

Sustained periods of CO_2 uptake during any year were rare and occurred only when a substantial amount of precipitation fell at a critical plant growth stage (i.e. July 2000). Precipitation received during dormant conditions stimulated an outward flux of CO_2. Averaged across all months in the six-year measurement period, we estimated that CO_2 was being released into the atmosphere at a rate of $+1.5\,g$ $CO_2\,m^{-2}\,d^{-1}$ ($+0.40\,g\,C\,m^{-2}\,d^{-1}$) or $+533\,g\,CO_2\,m^{-2}\,yr^{-1}$ ($+145.3\,g\,C\,m^{-2}\,yr^{-1}$). This is slightly greater than the carbon losses reported by Emmerich (2003) ranging from $126\,g\,C\,m^{-2}\,yr^{-1}$ to $144\,g\,C\,m^{-2}\,yr^{-1}$ at two southeastern Arizona sites, but less than a short-term (91 day) carbon loss of $+155\,g\,C\,m^{-2}$ reported by Meyers (2001) during a drought year in an Oklahoma grassland.

Although carbon fluxes in this desert ecosystem were generally small, the area encompassed by the JER and the Chihuahuan Desert Rangeland Research Center is large, exceeding 100,000 ha (Huenneke et al., 2002). The study site had periodic shrub removal that was not necessarily representative of other, Chihuahuan desert plant communities with a similar soil type. However, Dugas et al. (1996) reported no differences in ET fluxes over three vegetatively different desert communities with similar sandy soils, likely because precipitation and radiation, rather than leaf area, were the predominant drivers in flux response. If we assume that ET and CO_2 fluxes are coupled because each is responding to the same dominant environmental factors of a water-limited ecosystem, than carbon fluxes in our study likely are comparable to those of other arid ecosystems. Given that arid ecosystems comprise a large portion of the globe, substantial areas may be serving as a source of carbon for atmospheric CO_2.

Acknowledgements

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