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# Bare-Soil Evaporation Under Semiarid Field Conditions

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## ABSTRACT

**Bare-soil evaporation is an important component of the water balance in semiarid systems. However, little is known quantitatively about the influence of soil texture on bare-soil evaporation. We hypothesized that soil texture would have a great influence on both the temporal dynamics of bare-soil evaporation as well as on the depth to which evaporation-influenced soil water content throughout a 51-d simulated drought period. We measured soil water in lysimeters filled with three different soils. We measured daily evaporation gravimetrically and estimated evaporation rates using the energy balance method of Ben-Asher et al. We estimated soil water content at depths of 3.8 cm, 11.4 cm, 19.0 cm, 26.6 cm, and 34.2 cm (Layers 1–5, respectively) with time domain reflectometry rods and with soil cores. Bare-soil evaporation during the first 15 d of the experiment was ≈25% higher in the silt loam than the sand loam, and ≈42% higher than the clay loam. By Day 30, bare-soil evaporation in all soils was ≈0.5 mm d$^{-1}$. Soil water content decreased in all five lysimeters layers, and it was related to time and depth ($r^2$ values ranged from 0.60 to 0.95). The slope describing change in soil water content was greatest in the top 3.8-cm layer in all soil types (−0.52 in the clay loam, −0.32 in the silt loam, and −0.21 in the sand loam). At 14 d, bare-soil evaporation had the greatest influence on the upper 4.62 cm in the sand loam and the upper 7.18 cm in the clay and silt loams. At 51 d bare-soil evaporation had the greatest influence on the upper 7.36 cm in the sand loam, the upper 9.79 cm in the silt loam and the upper 14.1 cm in the clay loam.**

TWO CHARACTERISTICS are critical in defining semiarid grasslands. The first is that the evaporative demand of the atmosphere exceeds the water supply at all temporal scales (Noy-Meir, 1973; Sala et al., 1992). The second is that plant cover is less than 50%, resulting in bare-soil evaporation being an important feature of these ecosystems (Burke et al., 1998). The overwhelming dryness of these environments makes understanding soil water dynamics critical to understanding ecosystem structure and function. Soil water influences small and large scale patterns of vegetation (Sala et al., 1981; Sala and Lauenroth, 1982; Barnes and Harrison, 1982; Lauenroth et al., 1994), net primary production (Lauenroth, 1979; Sala et al., 1988; Lauenroth and Sala, 1992), and soil organic matter pools and nutrient cycling (Parton et al., 1987; Burke et al., 1989). The amount and importance of bare ground in these ecosystems results in bare-soil evaporation being a critical control on the daily, weekly, monthly, and annual water balance.

An important tool used to explore the dynamics of vegetation patterns, net primary production, soil organic matter pools, and nutrient cycling in semiarid regions is ecosystem level simulation modeling (Parton et al., 1987; Sala et al., 1988; Lauenroth et al., 1994). Critical to the behavior of such models is an appropriate representation of soil water balance and critical to an appropriate simulation of soil water balance is data from which parameters can be estimated. The limited availability of such data prompted us to undertake a bare-soil evaporation experiment.

Soil water content at any point in time results from the difference between inputs in the form of precipitation and run on, and losses such as runoff, drainage, evaporation, and transpiration. In the semiarid conditions of the shortgrass steppe, deep drainage, runoff, and run on are infrequent and can be safely ignored (Sala et al., 1992). The major routes of water loss from the soil are bare-soil evaporation and transpiration.

Bare-soil evaporation and transpiration are tightly linked and are often discussed together as evapotranspiration, yet each contributes a variable portion to a water budget. Under semiarid conditions, bare-soil evaporation is a significant component of the overall water balance of a region (Noy-Meir, 1973). Despite the importance of bare-soil evaporation, there have been few studies that have quantified bare-soil evaporation in semiarid systems (Paruelo et al., 1991; Daamen et al., 1993; Wallace et al., 1993; Yunusa et al., 1994).

Bare-soil evaporation rates are controlled not only by

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**Abbreviations:** CPER, Central Plains Experimental Range; TDR, time domain reflectometry.

atmospheric demand, but also by conductive properties intrinsic to the soil (Jury et al., 1991; van de Griend and Owe, 1994). Fine grained soils hold more water than coarse soils, and consequently, evaporation losses are greater than in coarse soils (Noy-Meir, 1973). While coarsely textured soils dry out more quickly (Jalota and Prihar, 1986), fine textured soils with small pore spaces are thought to support a longer period of upward flux through capillary conduction before pore spaces dry out. Once pore spaces dry, water loss occurs in the form of vapor diffusion. Vapor diffusion requires more energy input than capillary conduction (van de Griend and Owe, 1994).

The purpose of this paper is to evaluate the effects of soil texture on bare-soil evaporation from lysimeters in a semiarid grassland under field conditions. The specific objectives of this research were (i) to quantify bare-soil evaporation rates in a range of soil types beyond the length of any drought event (i.e., length of time between precipitation events) one would reasonably expect to observe in the field, (ii) to evaluate the energy balance method for estimating evaporation (Ben-Asher et al., 1983; Ben-Asher et al., 1984), (iii) to compare rates of bare-soil water loss from different depths in a soil profile in a variety of soil types, and (iv) to estimate the effective depth to which bare-soil evaporation influences soil water in a variety of soil types.

Lysimeters have been shown to be an accurate way to quantify water loss (Boast and Robertson, 1982; Boast, 1986). Using minilysimeters to quantify the change in soil water has an advantage to in situ techniques because minilysimeters allow for the manipulation of soil texture, allowing us to compare soils of different textures while minimizing factors than might confound the results (i.e., differences in vegetation). In order to minimize wall effects and to maximize the length of time we could run the experiment, we constructed minilysimeters with over five times the surface area, an order of magnitude more volume, and which were 30% longer than the microlysimeters used by Evett et al. (1995). Minilysimeters also allowed us to use several different techniques to measure water loss simultaneously. In this experiment, we estimated evaporation gravimetrically and with an energy balance method. We quantified soil water at discrete points in time and depth with time domain reflectometry (TDR) and gravimetrically by taking soil cores.

## METHODS

### Site Description

We carried out experiments at the Shortgrass Steppe Long Term Ecological Research (LTER) site, located on the northern Colorado piedmont (40°49′ N 104°46′ W). The shortgrass steppe region is an area of semiarid grasslands located within the western portion of the central grassland region (Lauenroth and Milchunas, 1992). The Central Plains Experimental Range (CPER), which includes the LTER, receives an average of 321 mm (SD = 98) of precipitation, falling on ≈60 days out of the year (Lauenroth and Sala, 1992; Lauenroth and Burke, 1995). It has a mean annual temperature of 8.6°C, a mean annual potential evapotranspiration of 1300 mm, and an average net aboveground primary production of 97 g $m^{-2}$ (Lauenroth and Sala, 1992; Lauenroth and Burke, 1995).

### Historical Precipitation Record

We gathered a 54-yr precipitation record for the CPER and calculated the number of days between precipitation events: a drought event. Since we were interested in drying time and did not want to underestimate drought periods, we treated missing data as days with no measurable precipitation. We split the precipitation record into growing and nongrowing periods, based upon Sims et al. (1978), who reported the thermal potential growing season of the region to be from mid-April to mid-October. We calculated the frequency distribution of the number of consecutive days without precipitation.

### Minilysimeter Preparation

We selected three soils: a clay loam (Ustollic Haplargid), a sandy loam (Aridic Argiustoll) and a silt loam (Ustic Torrifluvent) for our bare-soil evaporation study. We thoroughly mixed and sieved each soil through a 0.25-cm screen, removing as much visible organic material as possible. We used 15 minilysimeter replicates for each of the three soil textures for a total of 45 minilysimeters.

The minilysimeters were constructed out of 20-cm-diam. by 40-cm-deep PVC pipe. They had solid PVC bottoms drilled with 50 1-cm holes and were lined with wire screen. We filled the minilysimeters with soil in 7-cm layers and inserted pairs of TDR wave guides horizontally through 5 pairs of predrilled holes spaced on 7.6-cm centers. Horizontal waveguides have been used frequently as a method to resolve small depth increments (Seyfried, 1993; Vanclooster et al., 1995; Gupte et al., 1996; Wu et al., 1997). Waveguide material was 0.32-cm-diam. stainless steel rod, which is a commonly used material (Seyfried, 1993; Radcliffe et al., 1996; Gupte et al., 1996). We filled the first 3.8 cm of soil up to the level of the first pair of TDR waveguide holes, then packed the soil down with a flat-bottom soil tamper. We then installed a TDR probe and adjusted its geometry to parallel before placing the next 7 cm of soil on top of it. We used epoxy to fix the probes in place.

After all the minilysimeters were filled with soil, we added water until it was draining out of the bottom of each lysimeter. We covered the tops of the lysimeters with aluminum foil and allowed them to sit undisturbed until water stopped draining (≈48 h). We randomly placed the 45 minilysimeters in a prepared trench (the trench was on level ground and oriented north to south). The top of each minilysimeter was flush with the soil surface and fit tightly between two minilysimeters on either side and the walls of the trench. The soil and the vegetation surrounding the trench were undisturbed. We let the lysimeters equilibrate to soil temperature for an additional 48 h (for a total drainage time of 96 h) before we removed the covers and recorded the first measurements.

### Bare-Soil Evaporation

Once the lysimeters were in place and their covers removed, we allowed them to begin drying. We left the lysimeters exposed to the environment at all times except during precipitation events. We covered the lysimeters with a plastic tarp for a 12-h period five times during the 51-d experiment. The first rainfall event occurred 5-d after we removed the covers.

We used four approaches to estimate bare-soil evaporation.

First, we estimated evaporation rates for whole minilysimeters gravimetrically. We calculated bare-soil evaporation as the change in weight over time. One gram of water corresponded to 0.03-mm depth equivalent of water from the minilysimeters.

Second, we estimated bare-soil evaporation for the whole profile with infrared thermometry. We used a hand-held Minolta Cyclops Compact 3 infrared thermometer (Minolta, Ramsey, NJ) for measuring surface temperature and calculated evaporation via the energy balance method described by (Ben-Asher et al., 1983). The Ben-Asher method calculates evaporation using the difference between dry and drying soil surface temperatures.

$$E = S(T_d - T_e) \quad [1]$$

where $E$ is evaporation (J m$^{-2}$ d$^{-1}$), $T_d$ is the surface temperature of dry soil (K), and $T_e$ is the surface temperature of evaporating soil (K). $S$ is represented by the following equation:

$$S = 8.7\left(\frac{\rho C_p}{r} + 4\varepsilon\sigma T^3\right) \quad [2]$$

where $\rho$ is air density (kg m$^{-3}$), $C_p$ is the specific heat of air (J kg$^{-1}$ K$^{-1}$), $r$ is the resistance to heat transport in hours per meter ($r = 0.035u^{-0.96}$), $u$ is wind speed in m s$^{-1}$, $\varepsilon$ is bare soil surface emissivity (0.95), $\sigma$ is the Stefan-Boltzmann constant ($2.04 \times 10^{-4}$ J h$^{-1}$ m$^{-2}$ K$^{-4}$), and $T$ is the average surface temperature of dry and evaporating soil (K). We converted J m$^{-2}$ d$^{-1}$ into mm d$^{-1}$ with the conversion $2.4 \times 10^6$ J m$^{-2}$ per 1 mm of evaporation per day. We measured temperatures between 1230 and 1330 h (daylight time).

Third, we estimated water loss from specific layers within the profile from soil cores. We removed a 38-cm core (1.5 cm internal diam.) from five texture replicates at 7 d, 13 d, and at 51 d. We separated the 7-d cores into 7-cm sections representing five layers. We divided the 13-d and 51-d cores into 2-cm sections representing 18 layers. We placed each section in a ziploc bag and stored them on ice for transportation to the lab. After placing samples in foil pans, we weighed each sample on an electronic balance (±0.001 g) and then oven-dried each at 105°C for 24 h before reweighing and calculating soil water content based on the change in sample weight.

For the final method for determining water content, we measured volumetric soil water content ($\theta_v$) from specific depths with TDR. Time domain reflectometry relates volumetric soil water content (cm$^3$ $H_2O$ cm$^{-3}$ soil) to the apparent dielectric constant ($K_a$) of the soil (Topp et al., 1980). We used a Tektronix 1502B metallic cable tester (Tektronix, Beaverton, OR) for TDR measurements. We found the instantaneous rates of change in soil water content from the first derivative of the natural log transformed TDR data. The first derivative for the natural log transformed regression lines is

$$\frac{d\theta_v}{dt} = bkt^{b-1} \quad [3]$$

where $b$ is the slope of the least squares regression line for natural log transformed soil water content ($\theta_v$) vs. natural log transformed time, $k$ is the constant, and $t$ is time in days.

### Depth of the Influence of Bare-soil Evaporation

To estimate the effective depth to which bare-soil evaporation influenced water availability, we used the soil water content from 13- and 51-d cores. We identified the soil depth at which the slope of the change in soil water content with depth shifted from large to small (Hartigan, 1994). These analyses were based on 18 observations of soil water content along a 38-cm profile. We did not use the 7-d core data because the number of layers was insufficient.

We used the TDR reading on Day 0 as an estimate of initial water holding capacity from five randomly chosen minilysimeters of each soil type. We calculated bulk density and performed texture analysis on each soil type (Bouyoucous, 1936).

**Table 1. Physical properties for soil types used in minilysimeters. Values reported as means ($n$ = 5); standard errors given in parentheses.**

| Soil type | Clay | Silt | Sand | Bulk density | Field capacity† |
|---|---|---|---|---|---|
| | % | | | g cm$^{-3}$ | cm$^3$ cm$^{-3}$ |
| Clay loam | 24 (0.49) | 24 (0.26) | 52 (0.66) | 1.43 (0.024) | 0.317 (0.006) |
| Sand loam | 13 (0.89) | 18 (0.45) | 68 (0.40) | 1.41 (0.040) | 0.264 (0.005) |
| Silt loam | 18 (0.62) | 39 (0.32) | 43 (0.63) | 1.22 (0.042) | 0.349 (0.006) |

† Water holding capacity at time zero measured by TDR.

## RESULTS

### Soil Physical Properties

The three soils were classified as sandy clay loam, sandy loam, and silt loam (Soil Survey Staff, 1951) (Table 1). Sandy loam and clay loam were packed to nearly equal bulk density, while the silt loam was packed ≈15% less dense than the sandy loam and clay loam treatments. Initial soil water content was highest in silt loam and lowest in the sandy loam (Table 1).

### Length of Dry Periods on the Central Plains Experimental Range

There was no significant difference in the mean length of drought periods between the growing and nongrowing season. The mean number of days for a drought event during the growing season was 6.8 (standard error = 0.3) and the mean number of days for a drought event for the nongrowing season was 7.1 (standard error = 0.4). As the number of days without precipitation increased, so did the probability a drought event would be of shorter duration: {Cumulative probability = $0.98 \times [1 - \exp(-0.19 \times \text{d without precipitation})]$}. In the last 54 years on the CPER, ≈90% of all drought

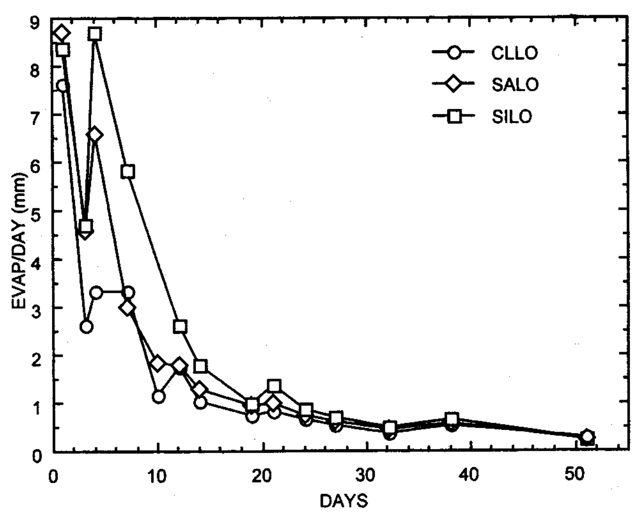


**Fig. 1. Daily evaporation rates in minilysimeters measured gravimetrically for 51 d for clay loam (CLLO), silt loam (SILO), and sand loam (SALO).**

**Table 2. Regression of daily evaporation measured by the energy balance method described by Ben-Asher et al. (1983) vs. gravimetric estimates. Results reported for combined soil types and split by individual soil type.**

| | Least squares regression | | | | |
|---|---|---|---|---|---|
| Texture | $n$ | Slope | Intercept | $r^2$ | $P$-value |
| All | 374 | 0.87† | 0.23‡ | 0.89 | <0.0001 |
| Clay loam | 125 | 0.92† | 0.13‡ | 0.94 | <0.0001 |
| Sand loam | 124 | 0.88† | 0.25 | 0.89 | <0.0001 |
| Silt loam | 125 | 0.85† | 0.27 | 0.85 | <0.0001 |

† Not significantly different from 0.
‡ Not significantly different from 1.

events were <15 d in length, and nearly all of the drought events in the last 54 years on the CPER were <30 d in length. The probability that a longer drought period than the 51 d duration of our experiment would occur on the CPER was nearly zero.

## Gravimetric and Energy Balance Bare-Soil Evaporation Estimates

Daily evaporation rates measured gravimetrically indicated that rates for minilysimeters were as high as 7.5 to 9 mm d$^{-1}$ in the early stage of the experiment, but then they dropped off rapidly (Fig. 1). Cool and cloudy conditions resulted in evaporation being ≈50% lower in all textures on Day 3 (Fig. 1). The silt loam lost >4 mm less water on Day 3 than it did on Day 4. The difference between Days 3 and 4 in the sand loam was almost 2 mm d$^{-1}$, and the clay loam lost almost 1 mm more water on Day 4 than it did on Day 3. For the first 6 d of the experiment, evaporation rates were highest in the silt loam (around 8.5–6.0 mm d$^{-1}$), followed by the sandy loam (around 8.75–4.0 mm d$^{-1}$), and then clay loam (around 7.5–4.5 mm d$^{-1}$) (Fig. 1). By Day 30, daily evaporation in all soil treatments was practically zero (Fig. 1).

Estimates of daily water loss by evaporation using the energy balance method and the gravimetric method were in good agreement ($y = 0.26 + 0.87x$, $r^2 = 0.89$, $P < 0.001$). The relationships for each texture were positive and significant ($P < 0.001$) (Table 2). In each case, the slope was not significantly different from 1.0 and the intercept was not significantly different from zero. Daily water loss ranged from >10 mm to very near zero.

## Soil Water through Time

Soil water content was strongly related to time and depth in the soil profile and decreased in all five layers through time (Table 3). At the beginning of the experiment, soil water was evenly distributed in minilysimeter profiles in all soil types (Fig. 2a, 2b, and 2c).

In the clay loam, soil water declined rapidly in the top layer (Fig. 2a). The steep decrease in soil water content in the 3.8-cm layer continued until approximately Day 20. In the remaining four layers, soil water content decreased less rapidly in the first 20 d of the experiment than did soil water content of the top layer. By Day 51, the decrease in soil water in clay loam appeared to stop in all layers. The greatest difference in soil water content between layers (approximately a seven fold difference) occurred at around 26 d into the experiment.

While soil water content decreased ≈20% in the silt loam during the first 4 d of the experiment (Fig. 2b), the greatest decrease (≈35%) occurred in the top layer (3.8 cm) between Day 0 and Day 4. Soil water continued to decrease (≈50% from Day 0) up to Day 40. Soil water content among layers in the silt loam reached a maximum difference of around 35% by Day 6, after which it remained fairly constant (Fig. 2b).

Soil water content in the top layer of the sand loam decreased ≈44% in the first 4 d of the experiment. The next three layers showed a similar pattern, but with only a 14% decrease in soil water (Fig. 2c). The 34.2-cm layer did not indicate a rapid decrease in soil water in the early stage of the experiment. At Day 51, soil water content in the sandy loam was still decreasing. The greatest difference in soil water content among layers was ≈32% and occurred on Day 6.

The relationship between soil water content and time was significant and negative for all of the soils (Table 3). Analysis of the natural log transformed soil water data showed the steepest slopes occurred in the top layer, with the fastest change in soil water through time occurring at the 3.8-cm layer in the clay loam. The next greatest change in soil water through time occurred in the silt loam at the 3.8-cm layer, followed by the sandy loam at the 3.8-cm layer (Table 3). The differences in water loss rates among the layers fell off drastically below the 3.8-cm layer.

## Daily Evaporation Rates

The rate of water loss per day calculated from TDR data varied with depth, time, and soil type (Fig. 3a, 3b and 3c). Water loss was high on Day 1 (2.2 mm d$^{-1}$ in the 3.8-cm layer of the clay loam) and decreased quickly in the early stage of the experiment. By the third week, the change in water content was nearly zero at all depths

**Table 3. Change in water content by layer measured by time domain reflectometry in minilysimeters on the Central Plains Experimental Range in northcentral Colorado. Slopes were calculated from regressions of natural log transformed volumetric soil water content and natural log transformed time.**

| | Clay loam | | | Sand loam | | | Silt loam | | |
|---|---|---|---|---|---|---|---|---|---|
| Depth | Slope† | $r^2$ | $P$-value | Slope† | $r^2$ | $P$-value | Slope† | $r^2$ | $P$-value |
| cm | | | | | | | | | |
| 3.8 | −0.52 (0.03) | 0.90 | <0.0001 | −0.21 (0.02) | 0.75 | <0.0001 | −0.32 (0.02) | 0.95 | <0.0001 |
| 11.4 | −0.14 (0.02) | 0.70 | <0.0001 | −0.15 (0.01) | 0.82 | <0.0001 | −0.19 (0.01) | 0.89 | <0.0001 |
| 19.0 | −0.12 (0.01) | 0.74 | <0.0001 | −0.14 (0.02) | 0.76 | <0.0001 | −0.16 (0.01) | 0.91 | <0.0001 |
| 26.6 | −0.09 (0.01) | 0.68 | <0.0001 | −0.13 (0.01) | 0.78 | <0.0001 | −0.16 (0.01) | 0.91 | <0.0001 |
| 34.2 | −0.02 (0.02) | 0.83 | <0.0001 | −0.11 (0.02) | 0.63 | <0.0001 | −0.15 (0.01) | 0.84 | <0.0001 |

† Slope of ln q (cm$^3$ cm$^{-3}$) vs. ln time (d); l standard errors in parentheses.

in all soil types. While the greatest difference among layers was between Layer 1 and Layer 2, the range of water content change varied among depths and was different in the different soil types.

In clay loam, the daily water loss on Day 1 in the 3.8-cm layer was almost twice daily water loss in the remaining layers. By Day 10, daily water loss in clay loam was nearly zero in all layers. While daily water loss was highest in the top layer on Day 1 relative to the lower layers in the minilysimeter profile, by Day 4 it was approximately equal throughout the profile (Fig. 3a).

Daily water loss in silt loam followed the same pat-

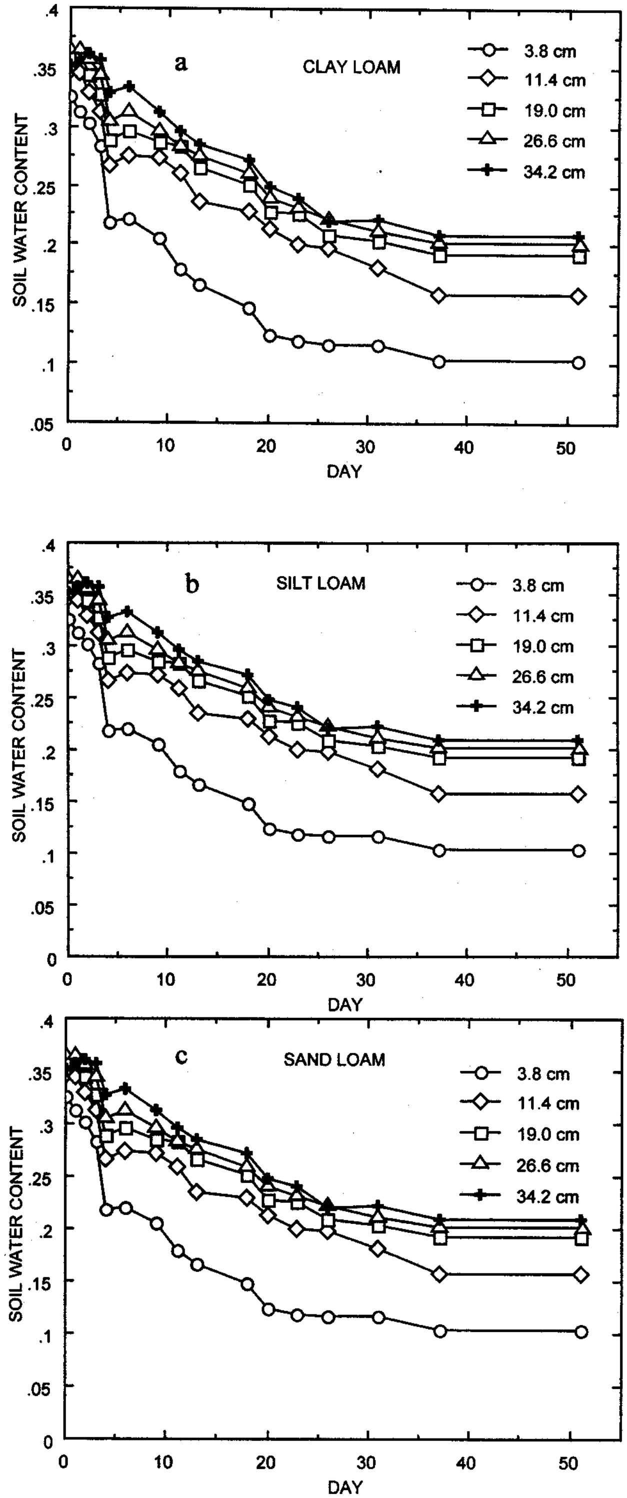


**Fig. 2. The change in volumetric soil water content over time by layer measured with time domain reflectometry in minilysimeters on the Central Plains Experimental Range in northcentral Colorado for three different soil textures: (a) clay loam, (b) silt loam and (c) sand loam.**

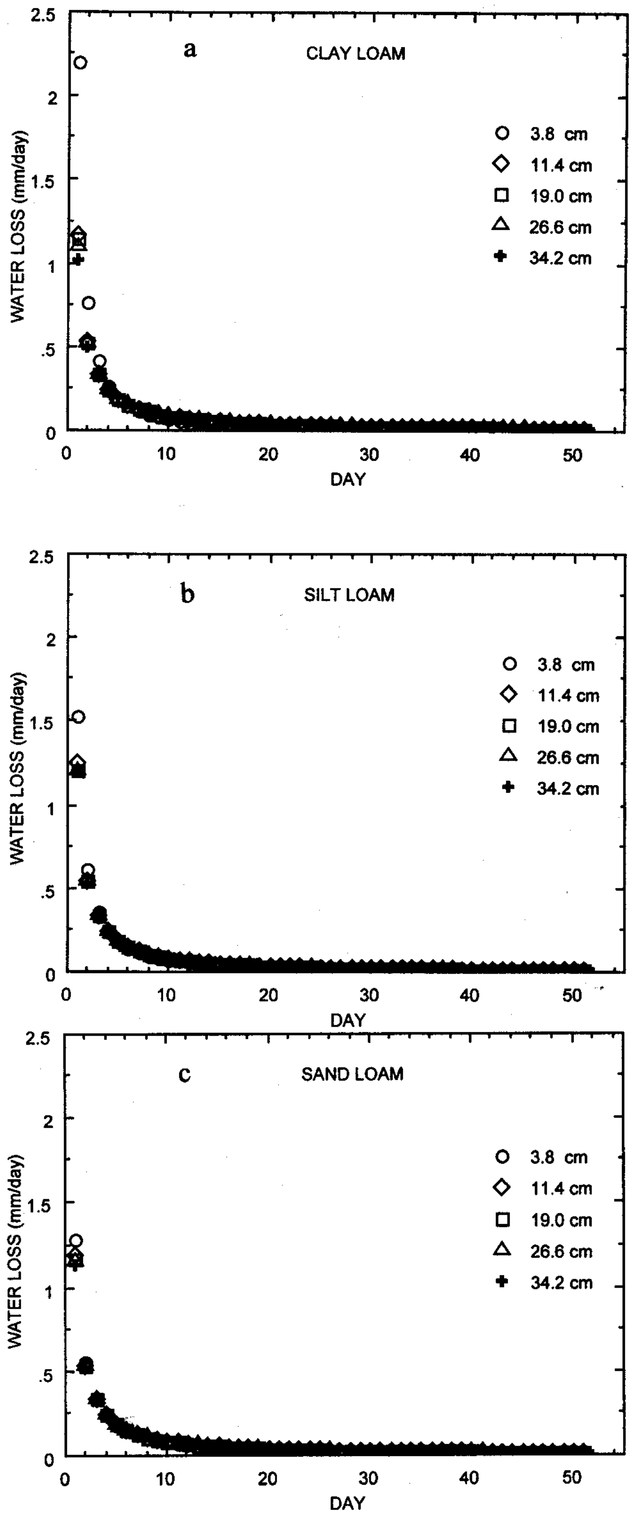


**Fig. 3. Water loss over time from minilysimeters on the Central Plains Experimental Range in northcentral Colorado. Data is from three different soil textures for 51 d by layer: (a) clay loam, (b) silt loam, and (c) sand loam.**

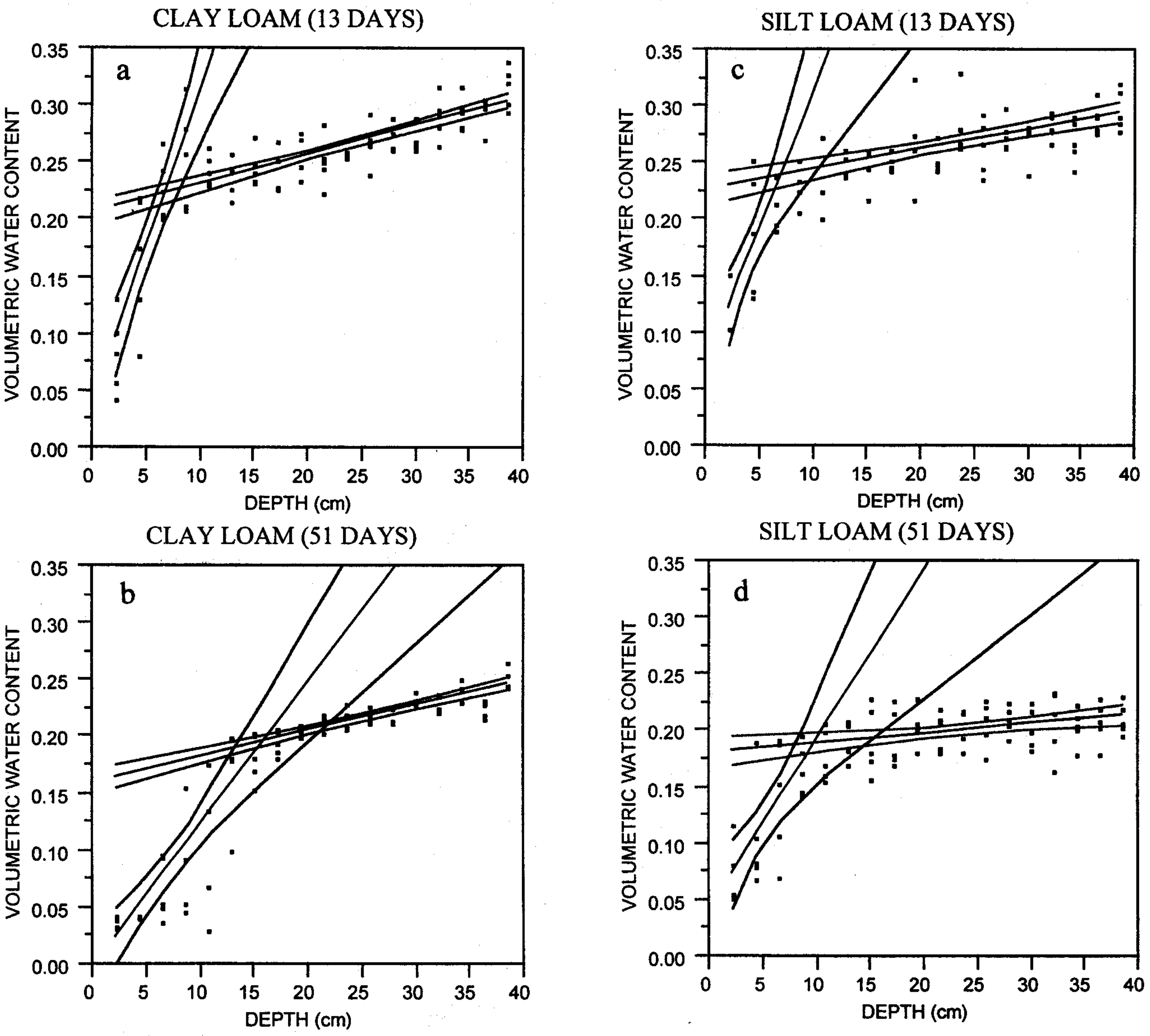


**Fig. 4. Continued on next page.**

tern, with the highest rate occurring in the 3.8-cm layer on Day 1 (1.5 mm $d^{-1}$). The difference between the 3.8-cm layer and the remaining layers increased nearly 50%, but from Day 2 on, it remained approximately uniform. By Day 20, daily water loss in silt loam had slowed to around zero in all layers (Fig. 3b).

In the sand loam, daily water loss was 1.25 mm $d^{-1}$ in the 3.8-cm layer on Day 1 (Fig. 3c). However, by Day 2, daily water loss was ≈0.5 mm $d^{-1}$ and nearly equal in all layers. By Day 20 daily water loss approached zero in the entire profile.

## Change Point Analysis

The depth at which bare-soil evaporation was most influential was related to time and texture (Fig. 4a, 4b, 4c, 4d, 4e and 4f). On both Day 14 and Day 51, soil water content increased with depth in the profile (Fig. 4a, 4b, 4c, 4d, 4e and 4f). The change in soil water with depth decreased at a certain depth. The portion of the profile located above this depth was the portion of the profile most influenced by evaporation.

Clay loam and silt loam soils dried 35% deeper than the sandy loam at 13 d (Table 4). The inflection point in the rate of change of soil water content was deeper in all soils at 51 d than at 14 d. The depth at which bare-soil evaporation decreased soil water content to the greatest degree in the clay loam was 20% deeper at 51 d than in the silt loam (Fig. 4b and 4h). The change point for bare-soil evaporation was >40% deeper in the clay loam profile than in the sandy loam at 51 d (Fig. 4b and 4f). The shallowest change point at both 14 and 51 d was in the sandy loam.

# DISCUSSION

## Length of Dry Periods

The controlled drying cycle in this experiment had a duration comparable to a substantial drought. A 51-d

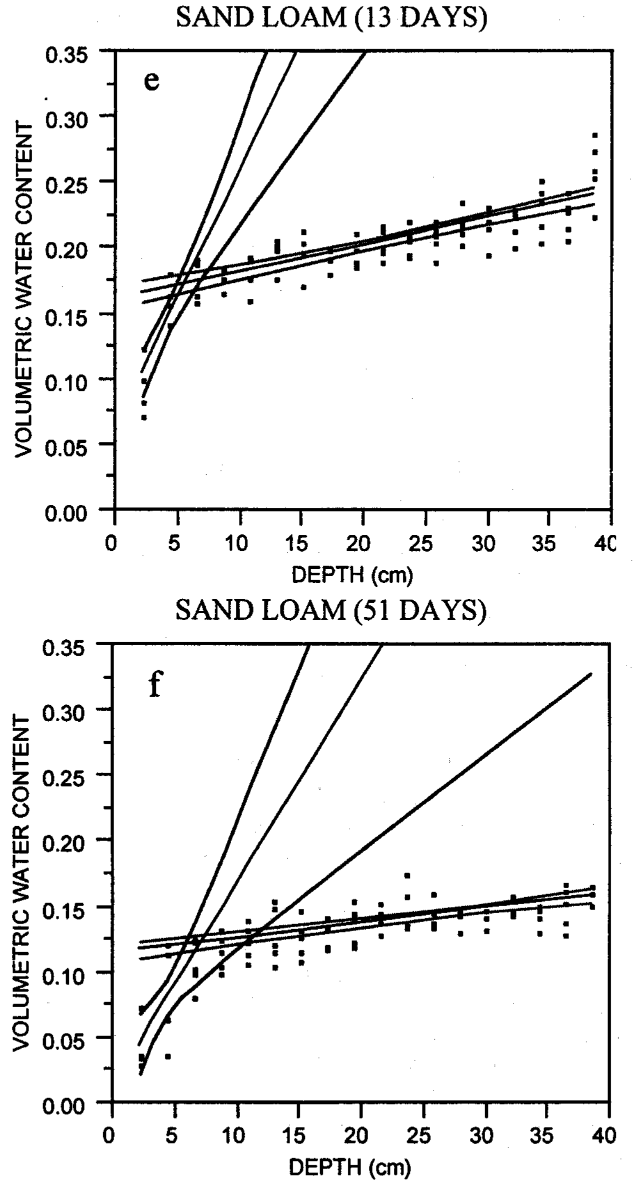


**Fig. 4. Soil water content (θ) vs. depth (cm) calculated from core data. Intersection of two least squares regression lines represents the lower limit of the depth over which bare-soil evaporation had the greatest influence on soil water content in minilysimeters for three soil types and at two time periods: (a) clay loam at 13 d, (b) clay loam at 51 d, (c) silt loam at 13 d, (d) silt loam at 51 d, (e) sand loam at 13 d, and (f) sand loam at 51 d. Curved lines represent 0.95 confidence intervals.**

drought event would be a rare occurrence on the CPER and the study site would have a low probability of observing a longer period without precipitation {Cumulative probability = 0.98 × [1 − exp(−0.19 × d without precipitation)]}. Sala et al. (1992) reported that daily average precipitation increased during the warm months and was highest in late spring. Our results agree with those findings that indicate that the increased precipitation observed at the CPER during the growing season is due to greater precipitation events and not a decrease in the number of days between precipitation events.

Surface resistance to evaporation increases rapidly as soil surface water content decreases (Wallace, 1995). Increasing resistance to evaporation, coupled with the fact that 75% of all drying periods are <8 d and 90% are <15 d, shifts the focus on the bare-soil evaporation component of the water budget on the CPER to a relatively short time frame (particularly efforts to model the water budget on the CPER).

Since the mean number of days between precipitation events on the CPER is ≈6 d during the growing season, a single growing season should average ≈27 short drought events with an intervening precipitation event wetting the soil between drought events. Nonetheless, the fact that two of the soils continued to lose water through the end of a 51-d drying period (albeit at a very low rate) indicated that during a prolonged period without precipitation, it is likely that some soils would continue to dry down, leaving a substantial water deficit.

These finding are important to ecosystem level modeling applications on the CPER and other semiarid regions. They imply that while prolonged drought events may exceed 51 d in length, a drying period length of 15 d will cover ≈90% of the periods without a water recharge on the CPER. Also, simulation models run on the CPER can use the same drying period parameters in both the growing and nongrowing seasons.

## Gravimetric and Energy Balance Comparison

We did not observe significant intratexture bias in comparing the energy balance–estimated evaporation to gravimetric evaporation estimates. Previously, the Ben-Asher et al. (1984) model had been reported to overestimate evaporation from microlysimeters (Paruelo et al., 1991); however, overestimation may be a particular problem at low evaporation rates (Ben-Asher et al., 1983). Since this experiment began with minilysimeters at or near field capacity, the relatively large number of high evaporation observations may have contributed to the good agreement with gravimetric measurements. The agreement between the energy balance method of the Ben-Asher et al. (1984) model and the physical weighing minilysimeters to estimate water loss supports the assumption that the minilysimeters were losing water out of their tops only. Since the energy balance method used infrared thermometry to estimate a surface phenomenon, drainage loss out of the bottoms would have created disagreement between the two methods.

## Soil Water through Time

The rapid initial drop in soil water content followed by a more gradual decline of soil water agrees with the Ritchie (1972) two stage model of evaporation. van de Griend and Owe (1994) described this phenomena as a result of the difference between relative low resistance to evaporation via water conduction and the relatively high resistance to evaporation via vapor diffusion. The decrease in soil water content in all five minilysimeter layers implies that bare-soil evaporation influenced soil water content at least until a depth of 38 cm in all the soils used in this experiment. This result suggests that experiments with longer drought periods will require minilysimeters of greater depth than 40 cm. The continued decline of soil water content in the late stages of the experiment suggests that while most of the water

**Table 4. Bare-soil evaporation change points for two line least squares regression for 13 and 51 d from three soil types based on core data from minilysimeters. Slopes were calculated by minimizing the sums of squares for both regression lines, then solving both equations simultaneously for depth.**

| | Clay loam | | Silt loam | | Sand loam | |
|---|---|---|---|---|---|---|
| | 13 d | 51 d | 13 d | 51 d | 13 d | 51 d |
| Slope 1† | 0.002 (0.0003) | 0.002 (0.0004) | 0.002 (0.0002) | 0.001 (0.0002) | 0.002 (0.0003) | 0.0009 (0.0003) |
| $r^2$ | 0.71 | 0.67 | 0.64 | 0.61 | 0.64 | 0.61 |
| *P*-value | <0.0001 | <0.0001 | <0.0001 | <0.0001 | <0.0001 | <0.0001 |
| Slope 2‡ | 0.030 (0.0060) | 0.012 (0.0020) | 0.024 (0.0060) | 0.0150 (0.0040) | 0.024 (0.0060) | 0.015 (0.0040) |
| $r^2$ | 0.69 | 0.67 | 0.56 | 0.57 | 0.59 | 0.59 |
| *P*-value | <0.0001 | <0.0001 | <0.0001 | <0.0012 | <0.0001 | <0.0001 |
| Change point | 7.18 cm | 14.1 cm | 4.62 cm | 7.36 cm | 7.18 cm | 9.79 cm |
| Total sums of squares | 0.08 | 0.094 | 0.042 | 0.018 | 0.046 | 0.030 |

† Slope for first least squares regression line standard errors in parentheses.
‡ Slope for second least squares regression line standard errors in parentheses.

lost to bare-soil evaporation is gone in the early stages of evaporation, bare-soil evaporation will continue through a 51-d drought period.

The rapid drying phase in the clay loam extended longer in the 3.8-cm layer than in any other of the loam soils (Fig. 2a, 2b and 2c). This conclusion was supported by the calculated slope of natural log transformed TDR data, in which the 3.8-cm layer in the clay loam had the steepest transformed change in soil water through time. The order of increasing time for the rapid phase of water loss followed by these soils agrees with the predictions of van de Griend and Owe (1994).

Starting at uniform soil water distribution at the beginning of the experiment, each of the soils in this study reached a point of maximum difference in soil water content among individual layers. As the experiment proceeded, the difference in soil water content between individual layers decreased. Variation of that point in time between soil treatments indicates that there were differences in the flux rates of water through the profile. The differences in the order of flux rates among the different layers of the profile agrees with the predictions of the van de Griend and Owe (1994) model.

## Daily Evaporation Rates

Our results suggest a more complicated process than that suggested by Jury et al. (1991) for the progression of evaporation through time. Evaporation rates in the rapid evaporation stage for the loam soils in whole minilysimeters followed the pattern suggested by Jury et al. (1991) for the first 3 d only. From Day 4 on, silt loam replaced sandy loam as the texture with the highest evaporation rates. The change in the order of evaporation rates suggests that through time, evaporation is a function of not only initial resistance to evaporation, but also of (i) the changes in resistance to evaporation in a particular soil type, (ii) the amount of energy available to drive the evaporative process, and (iii) the amount of water available to evaporate.

The van de Griend and Owe (1994) model suggests that while coarsely textured soils dry out quickly during the constant rate phase of evaporation, resistance to evaporation increases quickly as pore spaces dry out. Therefore, evaporation rates in coarse soils may be surpassed by evaporation rates in finer soils that are still evaporating via water conduction (Phillip, 1957). Total lysimeter evaporation at any one time is also a function of how resistance to water movement changes with depth in the soil column.

Water loss data measured with TDR indicated that in the first layer, the rate of water loss was greatest in the finest soil (clay loam), followed by the silt loam and the sandy loam, suggesting that individual layers lose water at different rates among soil types, and that the summation of these fluxes determines the total evaporation rate.

## Change Point Analysis

The large differences in the change in soil water through time between the top two layers indicated that bare-soil evaporation was most influential in the upper portion of the profile in the three loam textures. The differences in rates of water loss in Layers 2 through 5 were much smaller. This observation indicated that in the early portion of the evaporation experiment, most of the water lost to bare-soil evaporation was from the upper portion of the profile.

These observations correlate with the calculated change point for the three loam soils. The TDR data indicated that the change point should be between 3.8 and 11.4 cm in depth and is dependent upon the length of time bare-soil evaporation proceeds without precipitation input. The exception was the clay loam at 51 d, where the depth of maximum influence was substantially deeper.

The depths to which evaporation influenced soil water in the different soils used in this experiment agrees with the van de Griend and Owe (1994) model. The change point for clay loam at 51 d suggests that if allowed to dry long enough, the wet front for fine textured soils may continue to move down in the soil profile, limited finally by the energy available to overcome the various resistance factors (Wallace, 1995). Finally, the results of this experiment indicate that the non-zero slope in the slow phase of the change point curve means that there is slow change in water content through time across a range of soil textures to a depth of 38 cm, and that water continues to move up through the profile for at least 51 d, albeit at a low rate.

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