Soil & Tillage Research 110 (2010) 221–229

Contents lists available at ScienceDirect

Soil & Tillage Research

journal homepage: www.elsevier.com/locate/still

Tillage effects on soil water redistribution and bare soil evaporation throughout a season

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ARTICLE INFO

Article history:
Received 28 January 2010
Received in revised form 10 June 2010
Accepted 29 July 2010

Keywords:
Tillage
Evaporation
Soil water
Infiltration

ABSTRACT

Tillage-induced changes in soil properties are difficult to predict, yet can influence evaporation, infiltration and how water is redistributed within the profile after precipitation. We evaluated the effects of sweep tillage (ST) on near surface soil water dynamics as compared with an untilled (UT) soil during a 7-month period. Plots were established in a fallow field devoid of residue under stubble–mulch tillage management on a clay loam soil. Soil water contents were monitored using time-domain reflectometry at 0.05–0.3 m and using a neutron moisture gage to a depth of 2.3 m. Soil temperature and net radiation was also monitored. During a 114-day period from April through July, tillage with a sweep (0.07–0.1 m) significantly decreased net water storage above 0.3 m soil depth by an average of 12 mm ($P = 0.002$) as compared with UT plots. After tillage, soil water contents at 0.05 and 0.1 m were significantly ($P < 0.05$) lower in ST plots, even following repeated precipitation events. Water contents at soil depths $\geq$0.2 m were not influenced by tillage. Cumulative 3-day evaporation following precipitation events averaged 3.1 mm greater under ST compared with UT ($P < 0.014$). After extended dry periods, evaporation rates were similar among both treatments ($\sim$0.3 mm d^{-1}) despite the greater near-surface water contents of UT plots. Although ST plots exhibited 19 mm greater cumulative evaporation from July through October, this was offset by 26 mm greater infiltration compared with UT. A more advanced surface crust development and greater initial water contents were likely responsible for lower cumulative infiltration of UT compared with ST plots. Immediately after tillage, cumulative daily net radiation averaged 22% greater for ST compared with UT surfaces and these differences diminished with time. Increased evaporation under tillage was likely a result of enhanced vapor flow near the surface and greater absorption of radiation by a tilled surface with reduced albedo.

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1. Introduction

Tillage-induced changes in soil properties are difficult to predict, yet can influence infiltration, redistribution of water within the profile, subsequent evaporation rates, and water availability to crops. The influence of tillage on soil hydraulic properties and infiltration are not always consistent across location and soils. Initially, tillage may have a positive influence on infiltration (Messing and Jarvis, 1993) but this effect is usually transitory and usually leads to a decline in infiltration rates on tilled surfaces as a result of reconsolidation and aggregate disintegration after repeated rainstorms (Moret and Arrúe, 2007). Jones et al. (1994) demonstrated that runoff averaged 56% greater on no tillage (NT) compared with stubble–mulch tillage (ST) watersheds under a winter wheat (*Triticum aestivum* L.)–sorghum (*Sorghum bicolor* (L.) Moench)–fallow rotation. Dryland residue accumulation on the NT surface was insufficient to prevent the formation of a soil crust, primarily during fallow after sorghum, which was destroyed by tillage operations. Maintaining adequate residue is often difficult in semiarid regions with high evaporative demand relative to seasonal precipitation, limited residue production, and rapid decomposition rates (Unger et al., 2006). Under such conditions, residue cover even under NT may decline to less than 30% during fallow (e.g. Lampurlanés and Cantero-Martínez, 2006) resulting in near bare soil conditions.

Exposure of moist soil to the atmosphere by tillage can initially accelerate evaporative losses during the initial few days after tillage (Unger and Cassel, 1991). Good and Smika (1976) demonstrated that sweep tillage operation reduced soil water contents by 2.3 mm after the first day and a total of 3.6 mm by the fourth day after tillage. Hatfield et al. (2001) measured soil water evaporative fluxes of 10–12 mm in central Iowa following cultivation whereas evaporative fluxes from no-tillage fields totaled <2 mm for the same 3-d period. Long-term evaporation

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0167-1987/$ – see front matter. Published by Elsevier B.V.
doi:10.1016/j.still.2010.07.015

measurements usually show a persistence of greater average soil water contents near the surface for NT as compared with recently tilled soils (Smika, 1976; Zhai et al., 1990). In Eastern Washington state, USA, evaporative water loss under NT during dry summer months has been shown to result in lower soil water contents deeper in the profile (0.2–0.3 m) compared with tillage (Lindstrom et al., 1974; Hammel et al., 1981; Schillinger and Bolton, 1993). However, spring tillage was shown to affect only the near surface soil water content profile and not cumulative soil water depletion during the summer fallow period (Lindstrom et al., 1974). Undoubtedly, differences in the precipitation pattern, potential evapotranspiration, and soils influence the mechanisms governing infiltration, redistribution and evaporation and, consequently, the overall effects of tillage on soil water storage. Because previous studies have typically been confounded by the presence of different residue amounts, differences in evaporation among tillage treatments do not necessarily reflect differences in physical properties and related hydraulic properties.

In arid and semiarid environments, most evaporation occurs as a second stage process whereby water fluxes are limited by a soil surface resistance (Brutsaert and Chen, 1995; Suleiman and Ritchie, 2003). This resistance manifests itself as an evaporation front where both vapor and liquid transport contribute to total water flux and the phase change from liquid to vapor occurs below the soil surface (Saravanapavan and Salvucci, 2000; Grifoll et al., 2005). Under these conditions vapor transport can play a key role in mass and energy flows, and can account for half of the energy (Wescot and Wierenga, 1974; Cahill and Parlange, 1998) and water (Rose, 1968a,b; Jackson, 1973) flux near the soil surface. Experimental evidence of Rose (1968a,b) and Jackson (1973) demonstrated that the direction of vapor flux oscillates in response to the diurnal temperature gradient, moving downward during the daytime and upward at night. This is manifested in sinusoidal variations in soil water content very near (e.g. < 50 mm) the soil surface.

Early on it was recognized that estimation of near surface soil water contents at high temporal resolution using time domain reflectometry (TDR) could be used to monitor changes in soil water content and aid in the understanding of infiltration and evaporation processes (Zhai et al., 1990; Evett et al., 1993; Plauborg, 1995; Cahill and Parlange, 1998). Evett et al. (1993) showed that vertical arrays (0.0–0.4 m) of horizontally installed TDR probes in conjunction with neutron scattering at deeper depths could be used to estimate daily change in soil water storage to within 0.7 mm. A limitation with electromagnetic measurements such as TDR is that the apparent permittivity of soil can be strongly temperature dependent, especially in fine-textured soils (Or and Wraith, 1999; Schwartz et al., 2009a). Consequently, temperature corrections are required to properly determine the actual diurnal changes in soil water content under field conditions.

We monitored near-surface soil water and temperature dynamics during fallow on untilled (UT) and periodically sweep-tilled (ST) field plots to examine tillage effects on infiltration and evaporation in the absence of residue. *In situ* monitoring of soil water has the advantage of integrating the precipitation and evaporation history and gradual changes in hydraulic properties on the aggregate response of the system which is manifested as soil water storage.

2. Materials and methods

The study was established in a fallow field on a Pullman clay loam (Fine, mixed, superactive, thermic Torrertic Paleustolls) that was previously under stubble–mulch tillage management. The Bt horizon (0.15–0.75 m) in this field was homogeneous with respect to bulk density (1.41 Mg m^{-3}) and clay content (50.7%) (Schwartz et al., 2008). The field was kept weed free and devoid of residue throughout the study period. Intensive tillage operations were necessary to break-up a plow pan and permit installation of time domain reflectometry (TDR) probes. In September 2004, the entire field was tilled using a para-plow (Tye Co., Lockney, TX)[1] to a depth of 0.25–0.30 m followed by a chisel chopper (BJM Co., Hereford, TX), rotary hoe, and 0.3-m sweeps. During the following month, three plots with TDR and thermocouple instrumentation were established in each of the four parallel strips with alternating tillage treatments imposed the following year. Type-T thermocouples and 200-mm trifilar TDR probes were installed horizontally in the 12 plots at soil depths of 0.05, 0.1, 0.15, 0.2, and 0.3 m accessed through small ($0.25\ m \times 0.35\ m \times 0.35\ m$) excavated pits. Waveforms were acquired using a metallic cable tester (Tektronix, Inc., Beaverton, OR, model 1502C) and processed by a computer running the TACQ software (Evett, 2000a,b). Interconnects between the cable tester and TDR probes consisted of 12 m of RG8/U (Belden 9913), two 16-port coaxial multiplexers (Dynamax, Inc., Houston, TX, model TR-200; Evett, 1998) and 4 m of RG 58A/U (Alpha 9058AC) 50-Ohm coaxial. Waveforms from each of the probes were acquired at half-hourly or hourly intervals and soil temperatures were recorded at 5-min intervals.

On 7 April, 20 May, and 21 July, 2005, (ST) tillage plots were tilled to a depth of 0.07–0.1 m using a plow with two 0.9-m sweeps. Only a single strip was tilled for the 7 April tillage operation. The other two plots were untilled (UT) throughout the remainder of the year. Prior to tillage, TDR probes and thermocouples at 0.05 and 0.1 m depth were excavated and removed. Probes and thermocouples were reinstalled in the same location a few hours after tillage. Soil bulk densities of the surface 0.0–0.05 and 0.05–0.1-m depth increments were determined using extracted soil cores before and after tillage and periodically throughout the study. Measurements were centered between wheel traffic/tracks. Soil water contents were also monitored using a neutron moisture gage (Campbell Pacific Nuclear International, model 503DR, Martinez, CA) at three locations in each of the four plots from 0.1 to 2.3 m depth in 0.2 m increments at weekly intervals. The gage was previously calibrated *in situ* on the Pullman soil at Bushland, TX. Ambient air temperature, relative humidity, wind velocity, and global radiation (LICOR Biosciences, model LI-200SA pyranometer, Lincoln, NE) sensors were deployed at 2 m in the field interior during the study. Net radiation (REBS model Q7.1, Bellevue, WA) was measured at 1 m above one tilled and one untilled strip and corrected for wind velocity effects. Precipitation depth was recorded every 0.25 h with a tipping bucket rain gage (Texas Electronics, model TR-525M, Dallas, TX). Reference evapotranspiration (ET_0) was determined with the ASCE equations (Allen et al., 2005) for a short grass reference crop using meteorological data collected at the site.

The complex permittivity model of Schwartz et al. (2009a,b) was used to estimate water content from measurements of apparent permittivity and soil temperature. Bulk electrical conductivity was estimated with a power law model (Schwartz et al., 2009a) using the fitted parameters for the Pullman soil (Schwartz et al., 2009b). Water contents at 0.3 m estimated using the neutron probe can be compared with TDR measurements by integrating TDR water contents with depth across a sphere of influence with a radius of 0.15 m. This radius corresponds to a sphere that contains approximately 80% of the response for a water content of 0.35 $m^3\ m^{-3}$ (Kristensen, 1973; Ølgaard, 1965). With these assumptions, treatment averaged TDR water contents for this radius differed by only -0.001 to 0.021 $m^3\ m^{-3}$ with neutron

[1] The mention of trade names of commercial products in this article is solely for the purpose of providing specific information and does not imply recommendation or endorsement by the U.S. Department of Agriculture.

probe measurements throughout the months of July through October.

Soil water storage within a 0.6-m control volume was estimated by integrating linearly interpolated TDR (0.0–0.3 m) and neutron probe (0.5 and 0.7 m) water contents with depth. Water content at the surface (0.0–0.05 m) was assumed equal to the TDR water content at 0.05 m. Drainage out of the control volume was estimated using the calculated Darcy flux with the gradient based on the measured soil water contents at 0.5 and 0.7 m and the unsaturated hydraulic conductivity based on the interpolated water content at 0.6 m. The van Genuchten–Mualem model (van Genuchten, 1980) with parameters fitted using the iterative method of Schwartz et al. (2008) was used to describe the soil hydraulic properties in the 0.5–0.7 m layer for this particular field. Potential errors in the calculated drainage fluxes are approximately 0.1 mm d^{-1} and includes both uncertainties in calculated change in storage as well as the parameterized drainage model (Schwartz et al., 2008). Evaporation and cumulative infiltration were calculated from the change in soil water storage (0.0–0.6 m) with time using the procedures of Schwartz et al. (2008). Using this method, increases in soil water storage during and several hours after precipitation events are used to calculate cumulative infiltration. Increase in storage for several hours after precipitation has ceased represents drainage of any detention storage plus any near surface (<30 mm) soil water into the underlying soil sensed by the uppermost TDR probe. Because of water balance requirements, uncertainties in cumulative infiltration and evaporation are equivalent and totaled ±5 mm in a month with 103 mm of precipitation (Schwartz et al., 2008). Evaporation immediately after a tillage event was calculated from the change in water storage in the 0.0–0.1 m depth increment.

Autocorrelation statistics (SAS, 2008) were carried out to evaluate the spatial dependency of soil water contents at all depths and change in water storage after a precipitation event using data acquired in March prior to any sweep tillage. In all cases, Moran's I was close to zero and non-significant ($P > 0.20$) indicating a random spatial pattern for the plots. Evett et al. (2009) also notes that Pullman soils exhibit minimal spatial variability and are relatively uniform across a field with respect to their water content and soil water storage. In the absence of spatial autocorrelation, we used a completely randomized design (two-sample *t*-test) to evaluate differences in means which leads to tests of significance with powers equivalent to other statistical tests (Legendre et al., 2004). Given the potential that mean water contents variances may differ with tillage, we used Welch's (unequal variance) *t*-test (Rasch et al., 2009) to evaluate significant differences with respect to mean water contents, temperatures, and changes in soil water storage. Because cumulative infiltration and evaporation represent changes in storage at the plot level, variances and appropriate statistics could be calculated for these statistical comparisons using the six instrumented plots in each tillage treatment.

Soil water flow, heat flow, and vapor transport were simulated for day of year (DOY 208–210) with Hydrus 1D – Ver. 4.12 (Šimůnek et al., 2009) to evaluate the response of soil temperatures to measured net radiation which was specified as the surface boundary for tilled and untilled plots. This time period was selected for simulation because the soil was dry and evaporation was small so that temperature differences between treatments would arise primarily because of differences in net radiation rather than latent heat exchange. Initial soil water contents and temperatures were also specified for each tillage treatment whereas other meteorological measurements were acquired from the weather station. Vapor transport by diffusion with an enhancement factor was included in the simulation (Saito et al., 2006). Volumetric heat capacity of the solid phase was set equivalent to 2.39 MJ m^{-3} K^{-1} and the thermal conductivity function of Campbell (1985) was used to simulate heat transport. The Mualem–van Genuchten function (van Genuchten, 1980) was used to describe soil hydraulic properties. Parameters were based on the soil water retention measurements of Moroke (2002) with saturated and residual θ set to 0.45 and 0.0 m^3 m^{-3}, respectively, and shape parameters $n = 1.23$ and $a = 4.8$ m^{-1}. A saturated conductivity of 30 mm d^{-1} was used in the simulation (Unger and Pringle, 1981). Because soil water content of the surface 0.05 m was small (<0.09 m^3 m^{-3}) during this simulation, vapor transport dominated and cumulative evaporation was insensitive to saturated conductivity (see, for instance, Boulet et al., 1997).

3. Results and discussion

3.1. Soil water contents

The time series of mean soil water contents after the final tillage operation on 21 and 22 July (DOY 202 and 203) and extending through mid-October (DOY 290) illustrate the combined effects of 103 mm of precipitation in August and the prolonged dry-down period thereafter for each tillage treatment (Fig. 1). Water contents at 0.1 and 0.05 m in UT plots significantly exceeded water contents in ST plots except during and immediately following precipitation (Fig. 1). Using the iterative procedure of Schwartz et al. (2008), cumulative fluxes for the 0–0.6 m depth increment (Fig. 2) were approximated based on changes in soil water contents with time. From DOY 194 to 290, changes in storage were dominated by infiltration and evaporation fluxes. Cumulative drainage was a small component of the change in storage and averaged 0.17 mm d^{-1} during the month of August under both tillage treatments. When prorated on the basis of rainfall during an average year with a mean precipitation depth of 475 mm, estimated mean annual drainage rate is 24.3 mm for this fallow field. Scanlon et al. (2008) reported somewhat lower drainage fluxes (9–20 mm $year^{-1}$) based on a chloride mass balance study for the Pullman soil in an adjacent field. Lower drainage fluxes would be expected for the chloride balance study because the field was under a wheat–sorghum–fallow rotation rather than exclusively fallow.

Differences between ST and UT plots were not reflected in most bulk density measurements in the 0- to 0.10-m soil depth increment throughout the season (Fig. 3). Although tillage reduced bulk density, field variability and an intense precipitation event on DOY 162 (71 mm during a 7.5 h period) masked temporal patterns and treatment differences. Alletto and Coquet (2009) showed similar trends in that tillage effects were spatially variable, transient and difficult to capture. Nonetheless, they noted that near saturated conductivities were still influenced by tillage even after differences in bulk density had disappeared.

From DOY 94 to DOY 208, soil water storage (0.0–0.6 m) declined by 25 and 37 mm under UT and ST, respectively. Soil disturbance due to tillage were largely responsible for the 12 mm difference ($P = 0.002$) in stored water between treatments on DOY 208. Differences in stored water were a result of significantly ($P < 0.05$) greater water contents at 0.05 and 0.1 m soil depths for UT compared with ST (Fig. 4). Decreases in measured soil water content at 0.05 and 0.1 m depth were evident within several hours after tillage (Fig. 5). A portion of the water content reductions can be ascribed to an abrupt decrease in bulk density upon tillage. The corrected change in volumetric water content $\Delta\theta_c$ resulting from both evaporation and a change in bulk density can be written as

$$\Delta\theta_c = \theta_i\left(\frac{\rho_{bt}}{\rho_{bi}}\right) - \theta_t \tag{1}$$

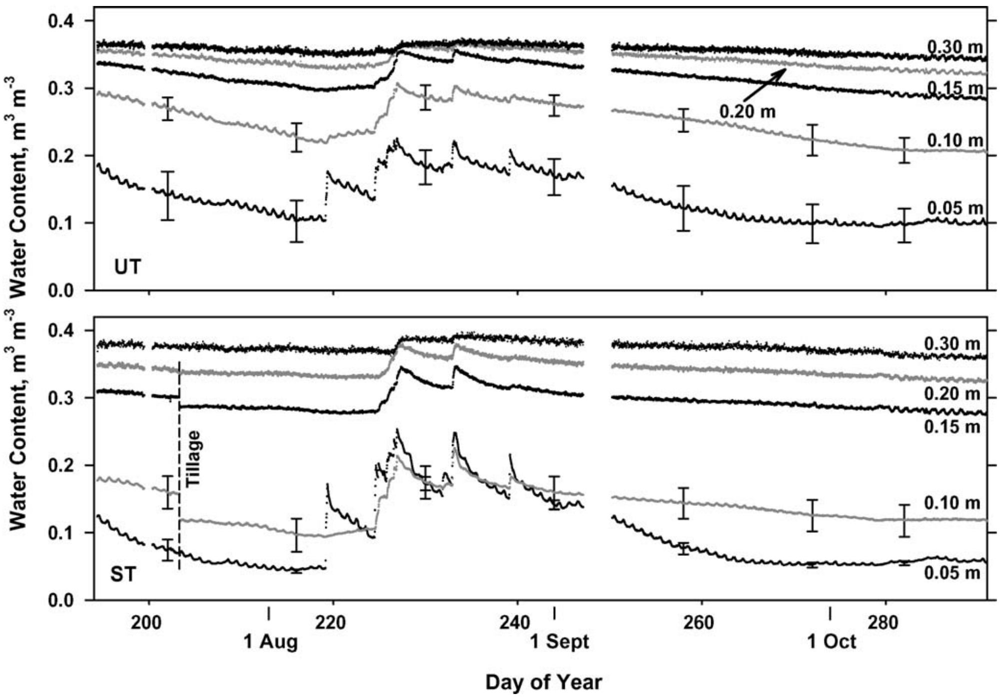

Fig. 1. Mean soil water contents for UT and ST plots after the final tillage operation. Error bars are 95% confidence intervals.

where θ_i and θ_t are volumetric water contents (m^3 m^{-3}) before and after tillage, respectively and ρ_{bi} and ρ_{bt} are soil bulk densities (Mg m^{-3}) before and after tillage, respectively. Accounting for changes in bulk density, average soil water depletion for the 0 to 0.1-cm depth increment during the day of tillage ranged from 0.8 to 7.6 mm and decreased with increasing number of tillage passes (Table 1). The ratio of daily tillage-induced evaporation to ET_0 declined with each additional tillage operation because of increasingly dryer soil conditions. A small but abrupt decline in estimated soil water content was also observed at 0.15 m on some plots during the final tillage operation (Fig. 1). This implies a small decline in bulk density rather than water content. Loosening of the soil below the tillage depth could have been caused by disturbance of large aggregates or peds located at the shear plane. Because of

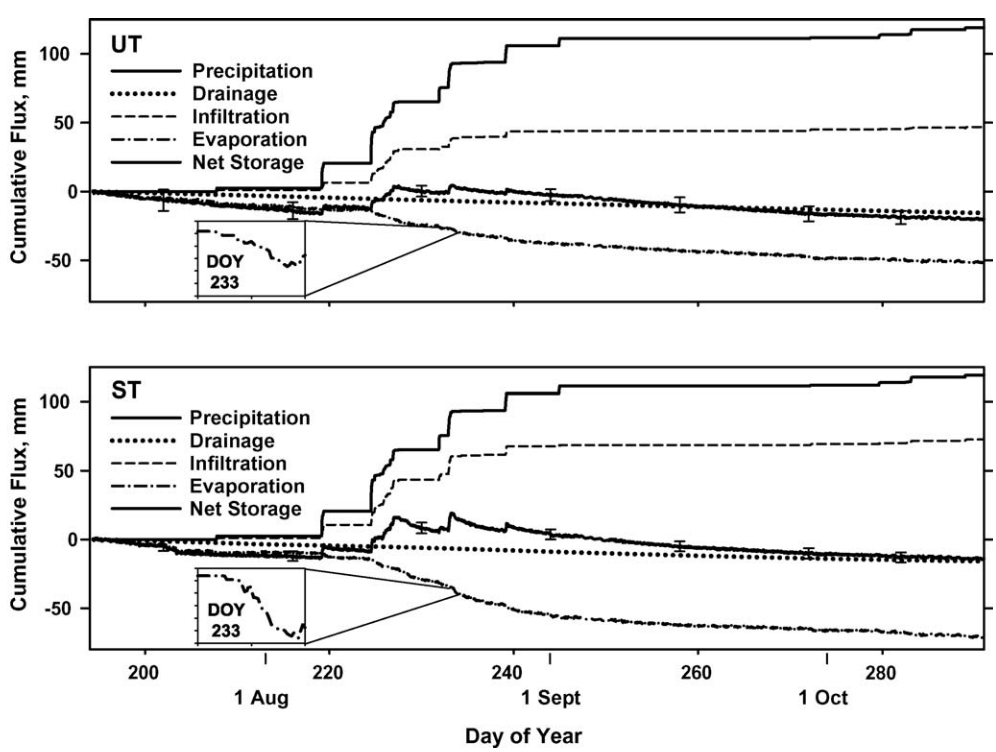

Fig. 2. Mean soil water balance and estimated cumulative fluxes within the 0- to 0.6-m control volume for the UT and ST plots. Insets show cumulative evaporation on DOY 233 for each respective treatment. Error bars are 95% confidence intervals delineating the variability of storage calculations.

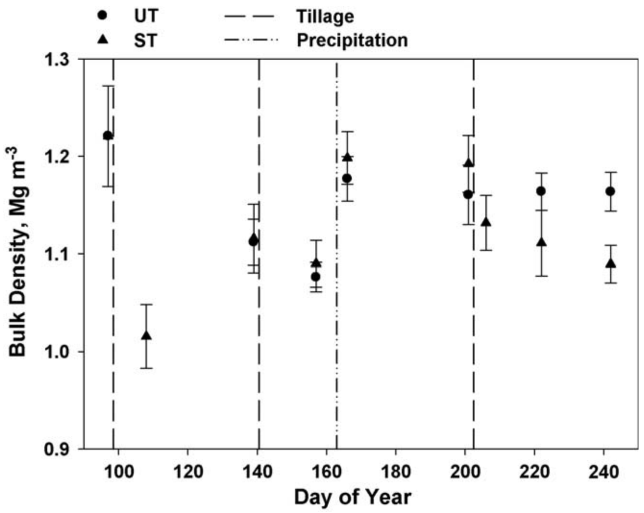

Fig. 3. Average bulk density of UT and ST plots for the 0- to 0.10-m soil depth increment. Error bars are 95% confidence intervals.

the existing dry soil conditions, soil aggregates would have had high shear strength and therefore resistant to disintegration and deformation (Munkholm and Kay, 2002).

3.2. Evaporation, precipitation, and soil water storage after tillage

Immediately after precipitation events on DOY 227 and 233, we observed steep daytime declines in soil water storage on ST plots indicative of high evaporation rates (Fig. 2, inset). Daily cumulative evaporation from ST plots totaled 3.5 and 3.4 mm on DOY 227 and 233, respectively, which represent 87 and 120% of ET_0 and reflective of rates associated with bare soil evaporation under energy-limiting conditions (Mutziger et al., 2005). In contrast, daytime evaporative losses observed for UT plots during these two days were 2.2 and 1.9 mm d^{-1}, respectively, or approximately 59% of ST losses. These results suggest that UT plots quickly transitioned into a soil-limiting evaporation stage (with a non-negligible soil surface resistance to vapor transport) compared with ST despite similar initial soil water contents.

Because of random errors in soil water content measurements, soil water storage errors can approach 1 mm and become a large

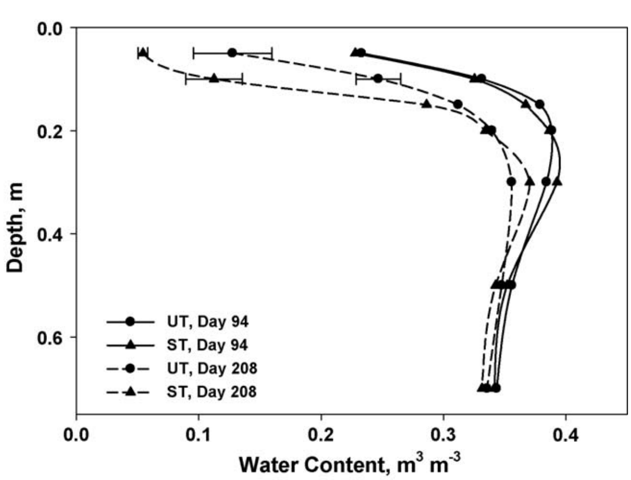

Fig. 4. Average soil water contents (n = 6) prior to tillage (day 94) and shortly after final tillage (day 208). Water contents were estimated using TDR (0–0.3 m) and a neutron gage (0.5 and 0.7 m). Error bars (shown only for significantly different treatment comparisons) are 95% confidence intervals.

Table 1
Soil water depletion of the 0.0–0.1 m soil layer during the day of tillage for subsequent tillage passes[a]

Tillage pass	DOY	θ_i[b] ($m^3\,m^{-3}$)	Corrected depletion		E/ET_0
			Mean (mm)	SE[c] (mm)	
1	141	0.173	7.6	2.0	0.83
2	140	0.131	2.7	1.8	0.37
2	202	0.100	1.0	1.2	0.11
3	203	0.083	0.8	0.5	0.09

[a] Initial and final water contents measurements were determined in the morning prior to tillage and at midnight on the day of tillage.
[b] θ_i is the weighted average of initial water contents for the 0.0–0.1 m depth increment.
[c] SE is the standard error of the mean change in soil water storage.

component of daily evaporation. Relative storage errors will decline as the time period over which cumulative fluxes are calculated, consequently, assessment of evaporation rates are more meaningful when evaluated throughout several days rather than on a daily basis. Cumulative 3-day evaporation based on the change in storage following precipitation events averaged 3.1 mm greater under ST compared with UT ($P < 0.015$; Table 2). For these time periods, daily evaporation rates averaged 0.8 for UT and 1.8 mm d^{-1} for ST. These rates are less than 40% of ET_0 and indicative of evaporation limited by a significant soil surface resistance to vapor transport. Mean evaporation after extended 10-day drying periods in September ranged from 0.2 to 0.4 mm d^{-1}. At the end of this period (DOY 262–271), evaporation under UT was significantly ($P < 0.01$) greater than ST. This may be a result of the greater near surface water content on UT (0.111 $m^3\,m^{-3}$) compared with ST plots (0.064 $m^3\,m^{-3}$). Throughout the entire observation period (DOY 194–290) cumulative evaporation under ST totaled 70.7 mm compared with 51.3 mm for UT.

Most evaporation in arid-regions occurs under conditions whereby vapor and liquid fluxes are limited by soil surface resistance (Brutsaert and Chen, 1995; Suleiman and Ritchie, 2003). Based on the 3 months of observations in Fig. 2, this is the case for our observations, with the energy-limiting evaporation phase occurring on only 2 or 3 days out of over 3 months of observations from July through October. Measurements and simulations of Bittelli et al. (2008) demonstrated that after stage 1 evaporation, aerodynamic resistance is relatively small compared with soil resistance to vapor flow. Consequently, differences in surface roughness between tillage treatments that would influence sensible heat and vapor transport above the soil surface would be expected to have a minor effect on observed differences in seasonal evaporation in this study.

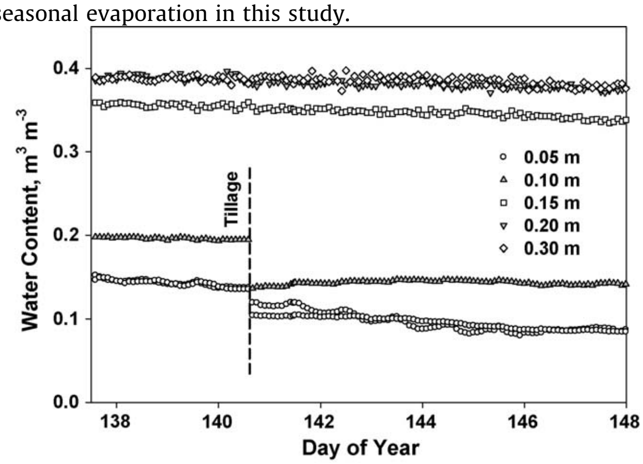

Fig. 5. Soil water contents before and after the second tillage operation on plot 7B on day 140. Two TDR probes were installed at 0.05 m to check the variability of estimated water contents in the tilled layer.

Table 2
Evaporation (E) and corresponding standard errors (SE) for selected time periods with no precipitation.

DOY[a]	UT			ST						
	θ_i[b] ($m^3\,m^{-3}$)	Cumulative E		θ_i[b] ($m^3\,m^{-3}$)	Cumulative E		Daily E			P c
		Mean (mm)	SE (mm)		Mean (mm)	SE (mm)	UT ($mm\,d^{-1}$)	ST ($mm\,d^{-1}$)	ET_0[d] ($mm\,d^{-1}$)	
228–230	0.200	2.4	0.6	0.217	6.8	0.7	0.8	2.3	5.6	0.001
234–236	0.196	3.4	0.6	0.195	6.8	0.9	1.1	2.3	5.9	0.015
240–242	0.185	1.0	0.4	0.177	2.6	0.2	0.3	0.9	6.1	0.004
252–261	0.142	3.9	0.3	0.108	4.2	0.3	0.4	0.4	5.8	0.783
262–271	0.111	2.7	0.2	0.064	1.7	0.2	0.3	0.2	5.8	0.002

[a] DOY 228–230, 234–236, and 240–242 represent drying periods immediately following precipitation.
[b] Initial average soil water content at 0.05 m.
[c] P-value for treatment mean differences based on Welch's t-test.
[d] Reference evapotranspiration based on the ASCE equations for a short grass reference crop (Allen et al., 2005).

The Pullman soil is highly susceptible to soil crust formation and sealing, especially in the absence of residues (Unger, 1984). A crust was present on the untilled plots throughout the entire year whereas crusts were destroyed on ST plots in conjunction with tillage events. A thick crust approximately 10 mm thick developed on the entire field as a result of the intense rainfall event on 12 June. Because the final tillage operation was carried out on DOY 203, all evaporation comparisons (Table 2) reflect a well-developed crust on UT and an emerging crust on ST that developed in response to August precipitation.

Within the top 0–30 mm of soil, the fractions of liquid and vapor fluxes contributing to total evaporation can change rapidly under conditions of high evaporative drying (Grifoll et al., 2005). Under these conditions, water vapor fluxes may equal or exceed liquid fluxes (Rose, 1968a,b; Jackson, 1973). The creation of greater and more connected pore space in this region, as with tillage, will obviously facilitate a greater gas phase transport potential. Modification of this zone by tillage will also destroy the surface crust or seal associated with this soil (Jones et al., 1994). In the case of a crusted soil, the crust may behave as a buffer zone that dries out quickly during the first few hours of the evaporation process. In contrast, evaporation from a soil without a surface crust that has uniform unsaturated conductivities near the surface will tend to reduce soil water contents over a greater depth (Assouline, 2004). This rapid drying of the surface crust may result in reduced evaporation rates at later times, as was observed by Bresler and Kemper (1970). Reduced vapor transport may come about by a modification of the vapor concentration gradient near (0–20 mm) the surface (e.g. Grifoll et al., 2005) due to the presence of a crust at the surface with a high bulk density and tortuous pores space. In addition, vapor concentration gradients near the surface will also be influenced by the temperature gradients which are dissimilar under UT and ST surfaces. These mechanisms may explain our observations of evaporation rates under UT. It also may explain the larger water contents near the surface under UT. The rough surface produced by tillage may also give rise to a larger effective surface area for evaporation thereby decreasing surface resistance to transport (Holmes et al., 1960).

Tillage increased infiltrated depth of rainfall by 21 mm compared with UT for five short duration (<12 h) precipitation events (Table 3). In two cases, mean infiltration depths were significantly ($P < 0.05$) greater under ST compared with UT, even though greater variability of water content measurements reduced the precision with which change in storage calculations could be estimated for these short time periods. Based on the calculated change in storage values during precipitation events, it was apparent that some instrumented plots were near or under topographic microlows which resulted in increased cumulative infiltration. This was especially evident for no tillage where, for example, on DOY 224 cumulative infiltration ranged from 1.8 to 13.9 mm. The field scale variability in cumulative infiltration is therefore being captured, although a larger number of instrumented plots would be necessary to detect differences in cumulative infiltration for some of the precipitation events in Table 3.

Consistently lower infiltrations rates under no tillage for the Pullman soil has also been reported by Jones et al. (1994) and Baumhardt and Jones (2002). Despite relatively high residue cover (57–86%) under no-tillage, surface crusts were able to form and decrease cumulative infiltration compared with sweep-tilled soils (Jones et al., 1994). With respect to the current study, a well-developed crust was present on UT plots whereas ST plots lacked crust development (DOY's 151 and 219) or a crust that gradually developed in response to August precipitation (DOY's 224, 233, and 239). Lower cumulative infiltration under UT in this study is likely a result of greater antecedent soil water contents and also a more advanced crust development compared with ST, especially for the first two precipitation events after tillage. Zhai et al. (1990) also demonstrated greater recharge under conventional tillage com-

Table 3
Cumulative infiltration (I) and corresponding standard errors (SE) for selected precipitation events with durations less than 12 h.

DOY	Precip. (mm)	UT			ST			P[a]
		θ_i[b] ($m^3\,m^{-3}$)	Cumulative I		θ_i[b] ($m^3\,m^{-3}$)	Cumulative I		
			Mean (mm)	SE (mm)		Mean (mm)	SE (mm)	
151	19.2	0.182	6.0	0.6	0.077	9.9	1.2	0.046
219	18.1	0.102	5.5	0.9	0.046	9.6	0.3	0.135
224	26.0	0.135	9.0	1.9	0.092	12.7	0.9	0.012
233	17.5	0.182	6.1	1.0	0.169	13.2	1.7	0.057
239	12.1	0.171	3.9	0.7	0.150	6.1	0.4	0.052
Total	92.9		30.5			51.5		

[a] P-value for treatment mean differences based on Welch's t-test.
[b] Average antecedent soil water content at 0.05 m.

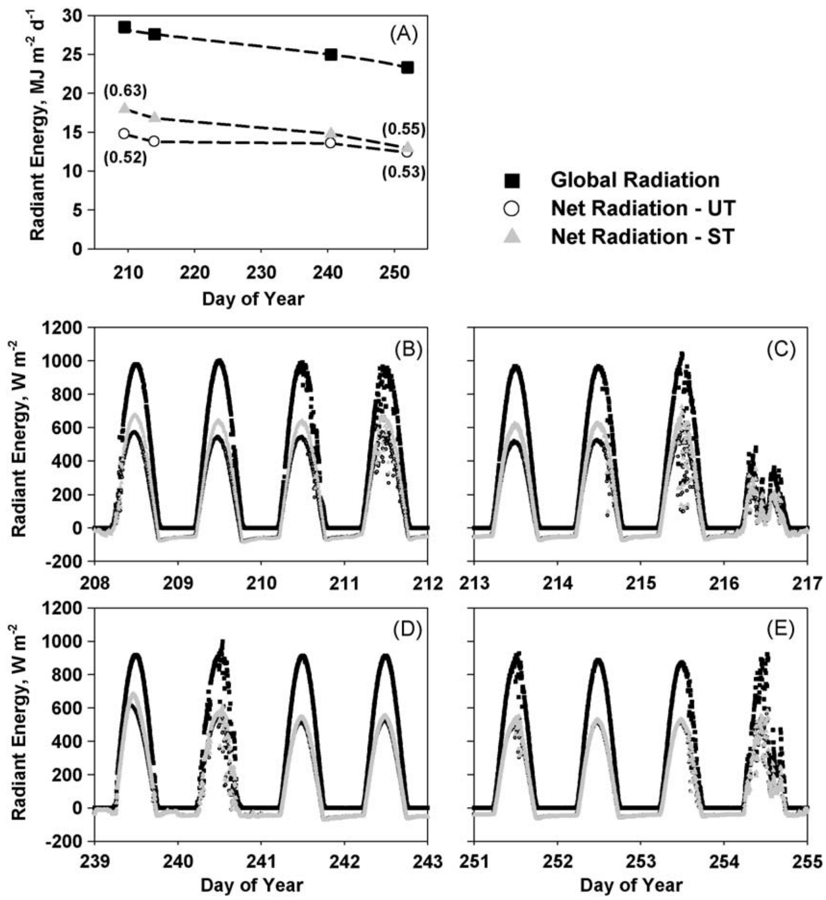

Fig. 6. Global and net radiation for selected days in 2005. Daily radiant energy (A) is the mean of predominately clear sky days for each 4-day period shown in plots B through E. Numbers in parenthesis in plot A denote the fraction of net relative to global radiation. Net radiometers were switched between plots on DOYs 208–212 and 213–217 to check for instrumental bias.

pared with no tillage. However, they attributed these differences to greater interception of precipitation by residues for these small (<10 mm) storm events.

The proportions of cumulative precipitation that infiltrated based on soil water storage calculations were relatively small (30–75%) for these low intensity storms (<25 mm h^{-1}). Using a rainfall simulator at an intensity of 48 mm h^{-1} for 1 h, Jones et al. (1994) and Baumhardt and Jones (2002) report a cumulative infiltration as a fraction of precipitation ranging from 40 to 84% for the Pullman soil under similar conditions. Lower cumulative infiltration in the current study would be expected because of the absence of surface residue. However, an incomplete accounting of the storage and subsequent evaporation in the surface ~30 mm of soil above the uppermost TDR probe could also explain these deviations. Because this additional, but unaccounted for, infiltration is offset by an equivalent depth of evaporation, these errors do not impact soil water storage over the long-term. Cumulative infiltration depths reported in Table 3 are therefore better interpreted as effective values reflecting precipitation less evaporation occurring during the event and immediately thereafter.

3.3. Net radiation and soil temperature

Average daily (sunrise to sunset) net radiation of the tilled surface on DOY 208 to 211 was 17.9 MJ m^{-2} d^{-1}, significantly ($P < 0.001$) greater than the 14.7 MJ m^{-2} d^{-1} of the untilled surface (Fig. 6). The differences between ST and UT integrated over a 24-h period was the approximately equivalent to the daytime (sunrise to sunset) value (3.2 MJ m^{-2} d^{-1}), signifying that nighttime upward longwave radiation was of similar magnitude for both surfaces. Differences in net radiation diminished with time after tillage (Fig. 6) and by DOY 251–254, were 0.5 MJ m^{-2} d^{-1} and non-significant ($P = 0.601$). The proportion of net radiation to total global radiation under the ST surface declined from 0.63 on DOY 208 to 0.55 on DOY 251, with similar water contents at 0.05 m on these 2 days, implying that the reflectance properties of the tilled surface was changing with time. In contrast, the proportion of net radiation to total global radiation for the UT surface was stable throughout this time period (Fig. 6). A lower albedo under tillage is likely a result of a greater random roughness (Allmaras et al., 1977; Potter et al., 1987; Oguntunde et al., 2006) and clay enrichment associated with the crusted UT surface (Ben-Dor et al., 2003). Tillage-induced roughness decreases with increasing cumulative rainfall amounts whereas no-tillage surfaces will remain stable with respect to random roughness (Unger, 1984; Zobeck and Onstad, 1987; Mwendera and Feyen, 1994). Consequently, increasing similarities between net radiation on UT and ST surfaces with time probably result from the gradual surface aggregate destruction and reconsolidation of tilled plots and the concurrent rise in albedo that, by DOY 252, was similar to that of the untilled field.

From days 208 to 216, maximum daytime temperatures at 0.05 m for ST plots averaged 2.2 °C greater than UT plots ($P = 0.0075$). Minimum temperatures of ST plots also averaged 1.1 °C lower than UT plots, although this difference was not significant. By days 251–254, differences in maximum daytime temperatures between UT and ST at 5 cm narrowed to 1.2 °C and were insignificant ($P = 0.060$). Simulation of the soil water and

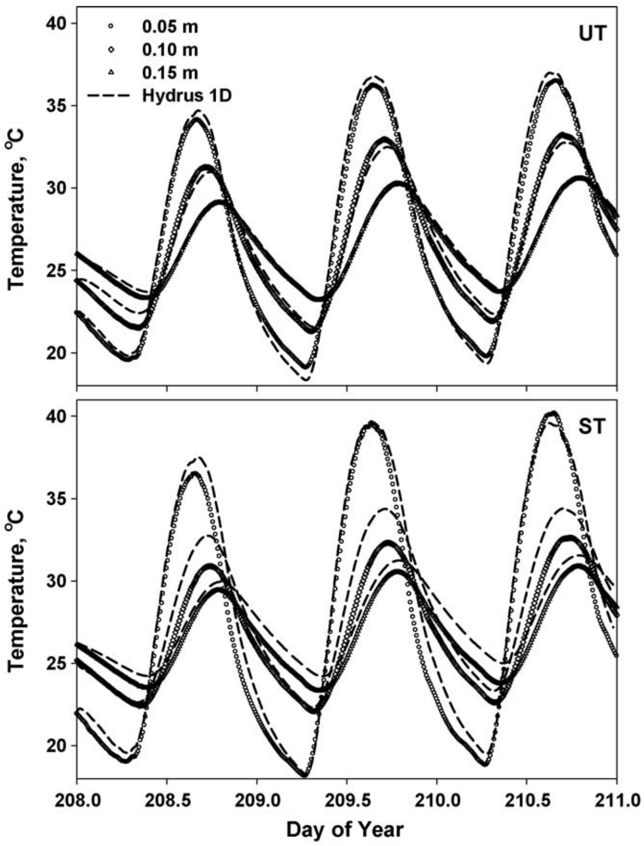

Fig. 7. Measured and simulated (Hydrus 1D) soil temperatures for UT and ST plots at three soil depths during 3 days with clear sky conditions. Simulations were based on measured net radiation (Fig. 6) and soil water contents (Fig. 1) which differed among treatments.

energy balance using Hydrus 1D for DOY 208–210 indicate that the differences in maximum temperatures between tillage treatments are a result of ST plots receiving more net radiation. Temperature simulations for these clear days closely track measured temperatures (Fig. 7) and suggest an average difference in daytime maximum temperatures between ST and UT of 2.7 °C. During this period (DOY 208–210) the upper 50 mm of the soil was dry ($\theta < 0.12\ m^3\ m^{-3}$) and latent energy associated with predicted evaporation was less than 2% of net radiation and limited by vapor transport to the surface. This suggests that latent energy exchange on these days was relatively unimportant in generating near surface temperature differences between tillage treatments.

Assuming a latent heat exchange rate of 2.4 MJ $m^{-2}\ d^{-1}$ per mm of water, the greater net radiation observed for ST following precipitation events (Table 2) only explains 53% of the additional energy consumed by evaporation under ST compared with UT (Table 2). Thus, greater net radiation under ST is only partly responsible for increased evaporation rates compared with UT. Immediately after a precipitation event on DOY 239, change in net radiation for the UT surface was asymmetric compared with both net radiation under the ST surface and global radiation. The early decline in net radiation under UT on this day suggests that albedo was increasing throughout the day as a result of near surface soil drying (Idso et al., 1974). A greater surface drying rate under UT is suggestive of the mechanisms proposed by Assouline (2004) to explain lower cumulative evaporation for sealed or crusted soils.

4. Summary and conclusions

Multiple-pass sweep tillage operations significantly influenced soil water storage as compared with untilled plots throughout a period extending from April to October. Compared with untilled surfaces, multiple tillage passes increased evaporation by 10 mm due to drying of exposed soil immediately after tillage. During a 3-month period after the final tillage operation, cumulative evaporation was 19 mm greater on tilled compared with untilled surfaces. A portion of the greater evaporation under tilled surfaces was attributed to reduced albedo and the concomitant elevated absorption of radiant energy at the surface. Greater cumulative evaporation under ST was offset by 26 mm greater infiltration compared with UT during this same 3-month period. Improved infiltration under ST likely resulted from the disintegration of the surface crust by tillage and lower initial soil water contents compared with UT.

A lower albedo for UT surfaces can account for some, but not all, of the differences in cumulative evaporation between tillage treatments. Rose (1968a) and Grifoll et al. (2005) show that in the very near surface region (0–20 mm) of the soil, there is a change in the contribution of the various transport mechanisms to net water flux. In this region, soil water content is less than about 0.12 $m^3\ m^{-3}$ and liquid water flux ceases to be an important transport mechanism. During midday with a positive upward temperature gradient, there exists a well-defined maximum in the vapor concentration near the surface (<20 mm) with steep gradients in both directions that drive vapor diffusion both towards and away from this zone. The presence of a crust at the surface with high bulk density and a tortuous pore space would decrease the effective gas phase diffusion thereby slowing vapor transport to the surface. The surface crust would also effectively decrease the vapor concentration gradient near the surface that drives the evaporation process and controls the upward liquid flux of water from below. This mechanism may explain the reduced evaporation under UT inferred from our measurements despite much greater near surface soil water contents compared with ST.

Maintaining adequate residue is often difficult in semiarid regions with high evaporative demand relative to seasonal precipitation. With low residue cover, such as after dryland sorghum, a surface crust may form in dispersive soils which can result in low water infiltration. Based on the preceding analyses, the crust may also be responsible for reduced evaporation in these fine-textured soils. In absolute terms, the difference in soil water depletion between tillage treatments is small (~10 mm) with respect to seasonal evapotranspiration in dryland sorghum (~300 mm). However, greater soil water contents near the surface in conjunction with slower drying rates after precipitation afford improved moisture conditions and a longer window of opportunity for dryland crop establishment. Favorable soil water status near the surface under no tillage can promote rapid crop establishment and root proliferation early in the growing season and lead to increased water use efficiency (Moroke et al., 2005).

Acknowledgements

We gratefully acknowledge the technical assistance of Ms. Jourdan Bell for electronics and micrologger programming and Mr. Duncan Vernon for assembly and setup of field apparatus.

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