# Fallow Tillage Effects on Evaporation and Seedzone Water Content in a Dry Summer Climate

J. E. HAMMEL, R. I. PAPENDICK, AND G. S. CAMPBELL

## ABSTRACT

**Early fall establishment of winter wheat is essential for high yields and erosion control following fallow in the Pacific Northwest, USA. Successful early stands require tillage management that minimizes water loss from the seedzone during the relatively dry summer. A numerical model was developed to investigate the dynamic aspects of heat and water flow in the upper soil layers with emphasis on effects of conventional tillage fallow and no-till (chemical fallow) on these processes and ultimately the seedzone water content. After 15-day simulations used to verify the model with measured values of meteorological variables and soil properties, predicted and measured water content distributions across the seedzone generally agreed within 0.01 kg/kg. Exclusion of thermally induced vapor flow from the model gave a higher predicted evaporative water loss and poorer agreement between measured and simulated soil water content distributions. The values for characterizing the atmospheric boundary layer resistance were also critical for correctly simulating soil temperature and water content profiles. Evaporative water loss for no-till over a 90-day simulation period using measured meteorological inputs was 70% higher than for conventional tillage, resulting in water contents in the seedzone that were too low for successful stand establishment under chemical fallow. Thus, for the climatic conditions and the soil studied here, some tillage may be necessary for retention of adequate seedzone water for early fall–winter wheat establishment.**

***Additional Index Words:*** **finite difference, soil heat, soil water, no-till chemical-fallow.**

Hammel, J. E., R. I. Papendick, and G. S. Campbell. 1981. Fallow tillage effects on evaporation and seedzone water content in a dry summer climate. Soil Sci. Soc. Am. J. 45:1016–1022.

SEEDZONE WATER CONTENT is a critical factor for successful early fall stand establishment of winter wheat (*Triticum aestivum* L.) after fallow in dryland areas of the Pacific Northwest. Adequate fall growth is essential for both maximum yields and surface cover for erosion control. Summers are dry in the Northwest, and early stands depend on carryover of moisture from the previous winter's precipitation. The moisture for germination must be retained no deeper than 12 to 15 cm below the surface for deep furrow planting. Thus, type of tillage and crop residue management which will influence evaporation and moisture retention may have a considerable effect on the success of the following crop.

Tillage using either a sweep cultivator or disk in early spring, followed by three to five rodweedings to establish soil mulch, has traditionally been considered the best fallow management for conserving seedzone moisture. Papendick et al. (1973) concluded that, under dryland conditions of eastern Washington, seedzone water is best conserved with a soil mulch having maximum combined resistance to liquid water and heat flow overlying a seedzone with good capillary continuity with underlying soil.

Currently, there is considerable interest in no-till (chemical fallow) or reduced tillage as alternative systems to improve erosion control of fallow soils and to conserve energy. Limited research indicates that water loss from the upper soil layers may be greater from no-till or reduced tillage fallow than with conventional methods. For example, Oveson and Appleby (1971) and Lindstrom et al. (1974) showed that, although oversummer loss of water from chemical and conventional tillage fallow was approximately equal, less moisture was retained in the surface 15 cm of silt loam soils with chemical fallow.

Water in the surface layers of a field soil moves in response to temperature and water potential gradients (Jackson et al., 1973). A tilled soil layer may conserve subsoil water during extended dry periods by slowing or preventing capillary flow to the surface where it would be lost by evaporation (Benoit and Kirkham, 1963; Fortier, 1907; Hillel, 1968; McCall, 1925; Papendick et al., 1973; Penman, 1941). A tilled, dry soil layer also decreases heat flux into the subsoil (van Duin, 1956; Willis and Raney, 1971). The thermal and hydraulic properties of soil are also controlled by texture, which may influence water loss for different tillage and residue management methods.

Solutions to the flow equation to obtain water content distribution usually assume isothermal soil conditions (Gardner, 1959; Hanks et al., 1969; Staple, 1971). Philip (1957) concluded from theory that temperature-induced moisture flow was relatively unimportant except at very low water contents. However, calculations by Jackson et al. (1974) showed that under field conditions thermally induced vapor flow can be significant at intermediate soil water contents.

During the summer, most water loss from fallow soil in the Northwest occurs across a dry layer which is 10 cm or more in thickness. Large diurnal temperature variations occur in the upper 15 cm of soil, which may influence vapor flow across the dry layer (Papendick et al., 1973). Thus it would appear that concurrent heat and water transport must be considered to model water flow under these conditions.

The soils of the Northwest dryland areas vary widely in texture ranging from sands to silt loams. Similarly, temperature and precipitation are highly variable over relatively short distances, both in long-term average values and between years. Thus, without some type of predictive method, it is extremely difficult to determine the extent to which reduced tillage or chemical fallow is feasible and where it is not, even under the most favorable conditions.

The objective of this research was to predict the influence of fallow tillage on evaporative water loss and soil water distribution within the wheat seedzone. A model of combined heat and water transport in soil was used for these predictions. Data from field experiments with conventional

[1] Contribution from the SEA–AR–USDA and Dep. of Agron. and Soils, Washington State Univ., in cooperation with the Coll. Agric. Res. Center, Washington State Univ. Scientific Paper no. 5886 of the latter. Received 13 Apr. 1981. Approved 17 June 1981.

[2] Assistant Professor of Soils, Dep. of Agron., Univ. of Georgia, Athens, GA 30602 (formerly Graduate Student, Dep. of Agron. and Soils, Washington State Univ., Pullman); and Soil Scientists, USDA–SEA–AR and Washington State Univ., Pullman, WA 99164, respectively.

tillage and no-till fallow were used for calibration and verification.

## THE MATHEMATICAL MODEL

The differential equation for one-dimensional water flow in soil for combined moisture and temperature gradients can be written (Philip and DeVries, 1957):

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left[(D_{\theta\mathrm{liq}}+D_{\theta\mathrm{vap}})\frac{\partial\theta}{\partial z} + (D_{T\mathrm{liq}}+D_{T\mathrm{vap}})\frac{\partial T}{\partial z}\right] + \frac{\partial k}{\partial z}. \quad [1]$$

The equation for heat flow is:

$$\rho_s c_s \frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(K_T\frac{\partial T}{\partial z} + LD_{\theta\mathrm{vap}}\frac{\partial\theta}{\partial z}\right). \quad [2]$$

In Eq. [1] and [2], $\theta$ is the water content, $t$ is time, $z$ is depth, $D_{\theta\mathrm{liq}}$ is the isothermal liquid water diffusivity, $D_{\theta\mathrm{vap}}$ is the isothermal water vapor diffusivity, $D_{T\mathrm{liq}}$ is the nonisothermal liquid water diffusivity, $D_{T\mathrm{vap}}$ is the nonisothermal water vapor diffusivity, $T$ is the temperature, $k$ is the unsaturated hydraulic conductivity, $\rho_s$ is the soil bulk density, $c_s$ is the soil specific heat, $K_T$ is the apparent thermal conductivity of the soil, and $L$ is the latent heat of vaporization of water.

Since the soil water content during the dry period of fallow is usually below field capacity (often $< -1$ bar water potential), we assume gravitational flow is negligible. Calculations by Jackson et al. (1974) show that thermally induced, liquid water flow is also very small compared with other components at these potentials and can be neglected. Equation [1] then becomes:

$$\frac{\partial\theta}{\partial t} = \frac{\partial}{\partial z}\left[(D_{\theta\mathrm{liq}}+D_{\theta\mathrm{vap}})\frac{\partial\theta}{\partial z} + D_{T\mathrm{vap}}\frac{\partial T}{\partial z}\right]. \quad [3]$$

The heat and soil water flow equations (Eq. [2] and [3]) can be solved using the Crank–Nicolson time-centered finite difference scheme (Crank, 1956). If $\phi$ represents temperature or water content, then the finite difference form of these two equations is (Hammel, 1979):

$$\begin{aligned}
&\frac{-K_{i-1/2}^{j+1/2}}{(z_i-z_{i-1})}\phi_{i-1}^{j+1} \\
&+\left[\frac{2\Delta z_i C_i}{\Delta t} + \frac{K_{i-1/2}^{j+1/2}}{(z_i-z_{i-1})} + \frac{K_{i+1/2}^{j+1/2}}{(z_{i+1}-z_i)}\right]\phi_i^{j+1} \\
&-\frac{K_{i+1/2}^{j+1/2}}{(z_{i+1}-z_i)}\phi_{i+1}^{j+1} = \frac{K_{i-1/2}^{j+1/2}}{(z_i-z_{i-1})}\phi_{i-1}^{j} \\
&+\left[\frac{2\Delta z_i C_i}{\Delta t} - \frac{K_{i-1/2}^{j+1/2}}{(z_i-z_{i-1})} - \frac{K_{i+1/2}^{j+1/2}}{(z_{i+1}-z_i)}\right]\phi_i^{j} \\
&+\frac{K_{i+1/2}^{j+1/2}}{(z_{i+1}-z_i)}\phi_{i+1}^{j} + 2U_i^{j+1/2}, \quad [4]
\end{aligned}$$

where the superscript $j$ and subscript $i$ denote the time and depth increments, respectively, $K$ is the soil water diffusivity or thermal conductivity, $C$ is the heat or water capacity, $U$ is the source term, $\Delta t$ is the time increment, $z_i$ is the depth of node $i$, and $\Delta z_i$ is the thickness of the $i$th soil layer.

When Eq. [4] is written for each of $n$ depths in the soil, a set of $n$ algebraic equations with $n+2$ unknowns is obtained. The unknowns are the terms with superscript $j+1$ on the left-hand side of Eq. [4]. The coefficients of $\phi$ form a tridiagonal matrix. All coefficients of Eq. [4] are known. Therefore, if appropriate boundary conditions are supplied, the matrix can be solved for the $\phi^{j+1}$ terms.

The source term $U$ is used to incorporate thermally induced vapor transport in the water flow equations and isothermal latent heat transport in the heat flow equations. Thermally induced vapor flow is calculated from:

$$\begin{aligned}
U_{\theta i} = {} & D_{\substack{T\mathrm{vap}\\ i+1/2}}^{j+1/2}(T_{i+1}^{j+1/2} - T_i^{j+1/2})/(z_{i+1}-z_i) \\
& - D_{\substack{T\mathrm{vap}\\ i-1/2}}^{j+1/2}(T_i^{j+1/2} - T_{i-1}^{j+1/2})/(z_i-z_{i-1}), \quad [5]
\end{aligned}$$

and latent heat transfer by isothermal vapor flow from:

$$\begin{aligned}
U_{Ti} = L\Bigg[ & D_{\substack{\theta\mathrm{vap}\\ i+1/2}}^{j+1/2}(\theta_{i+1}^{j} - \theta_i^{j})/(z_{i+1}-z_i) \\
& - D_{\substack{\theta\mathrm{vap}\\ i-1/2}}^{j+1/2}(\theta_i^{j} - \theta_{i-1}^{j})/(z_i-z_{i-1})\Bigg]. \quad [6]
\end{aligned}$$

The latent heat transfered by isothermal vapor flow is assumed to change slowly enough so that an estimate based on the water content at the start of the time step adequately represents the value at the center of the time step.

The solutions for soil temperature and water content require specific initial and boundary conditions at the upper and lower boundaries. At the lower boundary, temperature and water content are assumed constant. The upper boundary condition for the soil temperature equation is the average air temperature. The convective conductance for heat for the top element (above the soil surface is $\rho c_p/r$, where $\rho c_p$ is the volumetric specific heat of air and $r$ is the atmospheric boundary layer resistance. The boundary layer resistance, $r$, is calculated from (Campbell, 1977):

$$r = \frac{(\ln z/z_m + \psi)^2}{k^2 u}, \quad [7]$$

where $k$ is von Karman's constant (0.4), $u$ is the mean windspeed at height $z$, $z_m$ is the roughness length for momentum, and $\psi$ is a stability correction for the wind profile. In Eq. [7] it is assumed that the exchange surface for heat and water vapor is at the soil surface and that the momentum, heat, and water vapor roughness parameters are equal.

The profile correction factor, $\psi$, is (Businger, 1975):

$$\psi = 4.7\zeta, \qquad \zeta > 0,$$

or

$$\psi = -\ln\{[1+(1-16\zeta)^{1/2}]/2\}, \qquad \zeta < 0, \quad [8]$$

where atmospheric stability, $\zeta$, is (Campbell, 1977):

$$\zeta = -\frac{kzgH}{\rho c_p T_{ak} u^{*3}}. \quad [9]$$

In Eq. [9], $g$ is the gravitational acceleration, $H$ is the flux density of sensible heat, $T_a$ is the air temperature at height $z$, and $u^*$ is the friction velocity. The terms $H$ and $u^*$ are approximated by:

$$H = \rho c_p (T_s^{j+1/2} - T_a^{j+1/2})/r^{j+1/2},$$

and

$$u^* = ku^{j+1/2}/(\ln z/z_H + \psi^{j+1/2}), \quad [10]$$

where $T_s$ is the soil surface temperature and $z_H$ is the roughness

parameter for sensible heat exchange. The stability index, $\zeta$, and the profile correction, $\psi$, are obtained by iteration.

Part of the energy input at the soil surface can be accounted for using the isothermal net radiation, $R_{ni}$ (Monteith, 1973).

$$R_{ni}=(1-r_s)S_t+\epsilon_A\sigma T_{ak}^4-\epsilon_s\sigma T_{ak}^4, \quad [11]$$

where $r_s$ is the albedo, $S_t$ is the total shortwave irradiance, $T_{ak}$ is air temperature, $\epsilon_A$ and $\epsilon_s$ are the air and soil longwave emissivities, respectively, and $\sigma$ is the Stefan–Boltzmann constant. The surface source term $U_{T1}$ is $R_{ni}$ minus the latent heat loss. The remainder of the radiant energy exchange at the soil surface (difference between $R_{ni}$ and net radiation) is accounted for by adding a radiative conductance term to the convective conductance giving the total conductance between surface and air as:

$$K_{Ti}=\rho c_p/r+4\epsilon_s\sigma T^{*3}, \quad [12]$$

where $T^*$ is the average of the soil surface and air temperatures.

The boundary condition for evaporation is provided by specifying an equivalent water content and conductance such that the transport through the boundary element is equal to the actual evaporation from the soil. The actual evaporation from the soil is given by (Campbell, 1977):

$$E=(\rho_{vs}-\rho_{va})/r\rho_w, \quad [13]$$

where $\rho_{vs}$ and $\rho_{va}$ are vapor concentrations at heights zero (surface) and $z$ (atmosphere), respectively, and $\rho_w$ is the density of liquid water. The vapor concentrations can be rewritten as:

$$\rho_{vs}=h_s\rho_{vs}',$$

and

$$\rho_{va}=h_{ai}\rho_{vs}', \quad [14]$$

where $h_s$ is the humidity of the surface soil (a function of water content), $\rho_{vs}'$ is the saturation vapor concentration at surface temperature, and $h_{ai}$ is the relative humidity the atmosphere would have if its temperature were that of the soil surface. The humidity difference can be converted to a water content difference through multiplication by $d\theta/dh$, the slope of the humidity–water content function. The surface conductance, $K_{\theta 1}$, is therefore:

$$K_{\theta 1}=\frac{\rho_{vs}'}{\rho_w r}\frac{dh}{d\theta}, \quad [15]$$

and the water content boundary condition is:

$$\theta_a=\theta_s-(h_s-h_{ai})d\theta/dh. \quad [16]$$

In Eq. [15] and [16], $h$ is the atmosphere relative humidity, $\theta_a$ is an equivalent water content of the atmosphere, and $\theta_s$ is the equivalent water content of the atmosphere at the soil surface.

The isothermal and nonisothermal vapor diffusivities for soil (Philip and DeVries, 1957),

$$D_{\theta vap}=\alpha AD_{va}\rho_v\frac{\partial h}{\partial\theta} \quad [17]$$

and

$$D_{Tvap}=\alpha AD_{va}h\frac{\partial\rho_v}{\partial T}, \quad [18]$$

(mass flow factor assumed equal to unity) are obtained using a relationship where relative humidity is expressed as a function of water content (Fink and Jackson, 1973):

$$h=[1+(\theta/a)^{-1/c}]^{1/B}. \quad [19]$$

In Eq. [17], [18], and [19], $\alpha$ is the soil tortuosity, $A$ is the soil porosity, $\rho_v$ is the saturation water vapor density of soil air, $D_{va}$ is the water vapor diffusivity in air, and $a$, $B$, and $c$ are constants. By differentiating Eq. [19] and substituting into Eq. [17] (Kimball et al., 1976), the complete isothermal vapor coefficient becomes:

$$D_{\theta vap}=\alpha AD_{va}\rho_v'(aBc)^{-1}h^{1+c}(h^{-c}-1)^{1-B}, \quad [20]$$

where $\rho_v'$ is the saturation vapor density of the atmosphere and is given by this empirical formula (Teten, 1930):

$$\rho_v'=1.323\exp[17.27T/(237.3+T)]/(273.16+T). \quad [21]$$

Substitution of a relationship for the slope of the saturation vapor curve, $\partial\rho_v'/\partial T$ (Fuchs et al., 1978), into Eq. [18] yields the complete nonisothermal vapor coefficient:

$$D_{Tvap}=\alpha AD_{va}h\rho_v'[5307/(T+273)-1]/(T+273). \quad [22]$$

The relationship used to obtain the vapor diffusion coefficient in air as a function of temperature is (Monteith, 1973):

$$D_{va}=2.12\times10^{-5}(1+0.007T). \quad [23]$$

Philip and DeVries (1957) showed that vapor fluxes in a thermal gradient are larger than those predicted using Eq. [22] and proposed the use of an enhancement factor. Jury and Letey (1979) reexamined the data and proposed modifications to the enhancement model. Both see enhancement as the result of two conditions. One is that the local, microscopic temperature gradient across a soil pore is larger than the macroscopic temperature gradient in the soil. The other is that water, having crossed a pore under the influence of the microscopic gradient, condenses, moves quickly through the liquid phase to the next pore, evaporates, and moves to the next liquid island. Thus, if the resistance to flow in the liquid phase is small compared to that in the vapor phase, vapor moves through soil at a rate determined by the microscopic temperature gradient. In dry soil, the resistance for liquid flow is much larger than for vapor flow, so we reasoned that there should be no enhancement in the dry tillage layer that we wanted to model. Comparisons of model predictions with field data indicated that seedzone drying was substantially overpredicted using the enhancement model of Jury and Letey (1979). We therefore chose to follow Kimball et al. (1976) and ignore enhancement.

The liquid water diffusivity is calculated from this relationship (Gardner, 1959):

$$D_{\theta liq}=ae^{b\theta}, \quad [24]$$

where $a$ and $b$ are constants derived from soil characteristics.

The thermal conductivity of the soil, $K_s$, is calculated assuming it is a linear function of water content:

$$K_s=K_o+\eta\theta, \quad [25]$$

where $K_o$ is the oven-dry thermal conductivity, and $\eta$ is a proportionality constant. Equation [25] does not follow DeVries (1963) theory, but it is consistent with the data of Kimball et al. (1976) and with the few measurements we had on the soils used in this study.

The apparent thermal conductivity of the soil is calculated from

$$K_T=K_s+LD_{Tvap}. \quad [26]$$

For finite difference calculations, the diffusivities and conductivities must be appropriate averages over each element that makes up the solution network. Average values for $D_{\theta vap}$, $D_{Tvap}$, and $K_T$ were computed using the average water content and temperature of the nodes bounding each element. The liquid "conductance" for each element was obtained by integrating the steady-state flow equation over each element using Eq. [24] to obtain

$$K_{i-1/2}^{j+1/2} = a\left[\frac{(\exp b\theta_i^j - \exp b\theta_{i-1}^j)}{b(\theta_i^j - \theta_{i-1}^j)}\right]. \quad [27]$$

## MATERIALS AND METHODS

Conventional tillage and no-till fallow treatments were established in the spring of 1978, on wheat stubble, approximately 2 metric tons/ha, near Lacrosse, Wash. (average annual precipitation, 32 cm), on a Walla Walla silt loam (coarse-silty, mixed, mesic, typic Haploxeroll). Conventional tillage consisted of sweep cultivation to a depth of 15 cm in late March followed by three rodweedings spaced during the spring and summer to control weeds and to establish a loose soil mulch approximately 11 cm in depth and leaving essentailly a smooth, bare surface. With no-till, the soil remained partially covered with straw, and weed growth was prevented with herbicides. The fallow treatments were established on 3- by 7-m plots located over a ridgetop position in the field and arranged in a randomized, block design with four replications. Weeds had been completely controlled with herbicide applications on all plots during the 1977 to 1978 winter.

The experimental site was instrumented for measurement of total shortwave irradiance, windspeed, and air, wet bulb, and soil temperatures. Shortwave irradiance was measured with a pyranometer that was calibrated against an Eppley Precision Pyranometer;[3] windspeed was measured with a fast response 3-cup anemometer (Fritschen and Hinshaw, 1972), wet bulb temperature with a ceramic wick psychrometer, and air temperature with a shielded thermocouple positioned beside the wet bulb sensor. Both the psychrometer and the thermocouple were shielded from the sun with a 25-cm diam aluminum plate covered with reflective aluminized mylar. Shortwave irradiance and air and wet bulb temperature were measured at a height of 1 m, and windspeed was measured at 2 m. Soil temperatures were measured on adjacent plots with copper–constantan thermocouples in one replication of each treatment. Three thermocouples in parallel and spaced approximately 1 m apart were positioned at 2.5-, 5-, 10-, 15-, 30-, and 60-cm depths in the no-till and at 2.5-, 5-, 10-, 15-, 20-, 30-, 45- and 60-cm depths in the conventional tillage treatment.

The above measurements were initiated in early July and continued through most of August. Integrated values of all were recorded hourly with a battery-powered digital recorder (Model CR5, Campbell Scientific, Inc., Logan, Utah).[3] In addition, soil surface temperatures were measured hourly on one replication of both treatments with an infrared thermometer from 2000 hours on 6 August to 1800 hours on 7 August.

Soil cores 5 cm in diam by 60 cm in depth, one per plot for each treatment, were obtained on 25 July and again on 9 August. The cores were sectioned into 2-cm increments, and water content was determined gravimetrically.

The moisture characteristic curve and isothermal water diffusivity were determined in the laboratory. Thin-wall aluminum cylinders, 10 cm in diam and 20 cm in length, were pressed into no-till soil, two per plot, until the top was nearly flush with the soil surface. Water was added until cores were saturated, after which they were sealed on top with plastic bags and left in place for 2 days. The cores were then removed from the field, sealed in plastic bags, stored at 22°C for 1 week, after which evaporation was allowed to proceed at room temperature for 25 days. Each core was weighed daily. Soil water diffusivity was then estimated using a method similar to that of Gardner and Gardner (1969). Two additional cores were exposed to the atmosphere for 1 week, after which they were sectioned for measurement of gravimetric water content and water potential by thermocouple psychrometry.

The model for simultaneous vertical heat and water flow was programmed in HPL for a Hewlett Packard 9825A calculator[3] with 6,660 byte memory. The simulations were run using 11 depth nodes and one-hour time steps. Computation time was 1.7 min for a 1-day simulation. Node depths for the simulation were at 0, 1, 3, 5, 7, 9, 11, 15, 25, 45, and 65 cm.

The model was used to simulate soil moisture and temperature over a short (15 day) and a long (90 day) time period for conventional tillage and no-till conditions. The short simulations were used for model validation; the long simulations were used to indicate tillage effects on evaporative loss and soil water content over a longer part of the dry fallow season. For the 15-day simulation, measured soil water content and soil temperature were used as initial conditions, and atmospheric data were used for boundary conditions. These simulations used inputs of hourly integrated shortwave irradiance, air temperature, and windspeed.

For the 90-day simulations, solar radiation was computed form (Campbell, 1977):

$$S_t = 0.5S_p^o\,(1+\tau^m)/m \quad [28]$$

where $S_p^o$ is the solar constant, $\tau$ is the atmospheric transmission factor (0.8 for our simulations), and $m$ is the airmass number, given by

$$m = [\sin\lambda \sin S + \cos\lambda \cos S \cos 15(t-12)]^{-1}. \quad [29]$$

In Eq. [29], $\lambda$ is the latitude (47°), and $S$ is the solar declination.

Wind, as a function of time of day, was estimated from a polynominal fit-to-the-field data. Hourly temperatures were interpolated from daily maximum and minimum values using a sine function. The daily values were from climatic summaries for Lacross, Wash., for 15 June to 15 Sept. 1977. Initial conditions for these simulations were $\theta = 0.16$ g/g (near field capacity), and $T = 23$°C at all nodes.

Values of the constants $a$ and $b$ in Eq. [24] for the isothermal liquid diffusivites were $4 \times 10^{-11}$ m$^2$/sec and 24, respectively, for the tillage layer ($\rho_s = 1.0$ Mg/m$^3$) and $2 \times 10^{-10}$ m/sec and 24, respectively, for the untilled soil ($\rho_s = 1.3$ Mg/m$^3$).

Values of oven-dry and saturated soil thermal conductivities were calculated from DeVries' (1963) theory. Thermal conductivities for the soil solids, water, and air were 2.93, 0.57, and 0.025 W m$^{-1}$ C$^{-1}$, respectively. Computed thermal conductivities of oven-dry tilled and untilled soil were 0.17 and 0.20 W m$^{-1}$ C$^{-1}$, respectively. For saturated soil, the conductivites for the tilled and untilled soil were 1.15 and 1.40 W m$^{-1}$ C$^{-1}$, respectively. From these, the values for $\eta$ in Eq. [25] were 1.63 and 2.36 for the tilled and untilled soil, respectively.

The soil moisture characteristic for Walla Walla silt loam (Fig 1) was used to compute the relative humidity of the soil atmosphere as a function of water content. Values of $a$, $B$, and $c$ in Eq. [19] and [20] were 0.0312, −0.191, and 1.10, respectively.

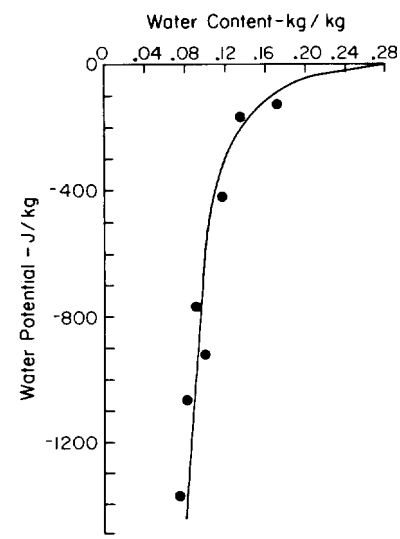


**Fig. 1—Soil water content–water potential relationship for Walla Walla silt loam.**

[3]Trade names and company names are included for the benefit of the reader and do not imply endorsement or preferential treatment of the product by the USDA.

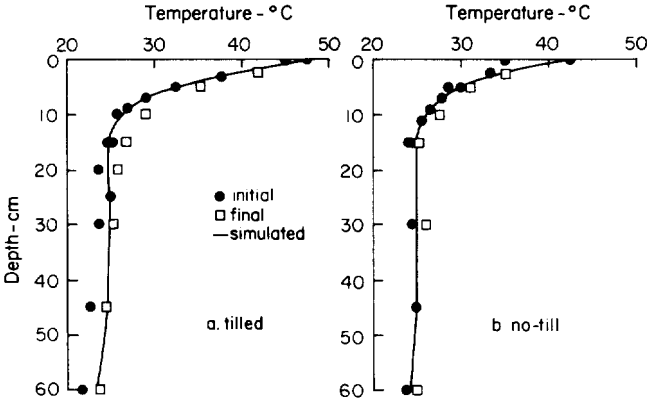


**Fig. 2—Comparison of measured field soil temperatures for different soil depths at the beginning and end of a 2-week simulation with predicted soil temperatures at the end of the 2 weeks for conventional tillage and no-till fallow.**

Other parameters used during simulations were $r_s$ (albedo) = 0.2, $z_m$ = 0.01 m for no-till, and $z_m$ = 0.001 m for conventional tillage. Measured $r_s$ was the same for both treatments apparently because of sparse crop residues and a small difference between the albedo of straw and the soil. The vapor density of the air, measured at random times with a sling psychrometer, and from hourly integrated air and wet bulb temperatures, showed a relatively constant value of 0.006 kg/m$^3$ from 25 July through 9 August, the time period used for the 15-day validation simulation. This value was used in all simulations. Soil emissivity, $\epsilon_s$, was assumed to be unity. The air emissivity, $\epsilon_A$, calculated from the vapor density formula of Brutsaert (1975), was 0.75. Porosities of the tilled and untilled soil from field measurements were 0.62 and 0.50 m$^3$/m$^3$, respectively.

## RESULTS AND DISCUSSIONS

### Soil Temperatures for the 15-day Simulation

Figure 2 shows that predicted soil temperatures at the end of the 15-day simulation period (1200 hours, 9 August) agree reasonably well with measured temperatures and give confidence that the heat flow component of the model is working satisfactorily. Predicted temperatures are generally lower than measured temperatures at most of the soil depths for both tillage treatments. The maximum difference between simulated and measured temperatures is approximately 3°C, but for most of the depth, differences are around 1°C or slightly greater. Soil temperatures below about 20 cm are approximately the same in the no-till and the conventional tillage, and simulated temperatures agree better with measured values. Above 11 cm, the measured soil temperatures are 1 to 6°C higher in the tilled soil than in the no-till at this time of the diurnal period.

Figure 3 gives a one-day comparison of measured and predicted temperatures at the soil surface, and the 15- and 30-cm depths for the two fallow treatments. In general, both the measured and predicted values show that surface temperatures are lower at night and higher during the day (by as much as 6°C) on the tilled than on no-till treatment. At the deeper depths, mainly below 15 cm, the measured soil temperatures tend to be a degree or more lower in the tilled than in the no-till treatment.

The lower thermal conductivity of the tillage layer slows net heat transport into the deeper layers of the tilled soil compared with the no-till. The higher temperature at the tilled surface results in greater radiant and convective heat losses to the atmosphere than in the untilled soil, but this is partially offset by the enhanced turbulent transport from the untilled soil, due to its larger roughness coefficient. The net result is slightly higher temperatures at lower depths in the untilled soil.

Analyses of tillage effects on soil temperature generally ignore the effects of surface roughness. Our simulation indicates that the roughness length may be the most important factor involved in the effect of tillage on soil temperature. The roughness lengths used in our simulations were estimates based on the heights of the roughness elements (10 cm for no-till with some stubble on surface, 1 cm for conventional tillage). When we tried $z_m$ = 0.001 m for the no-till, simulated soil temperatures exceeded measured values by more than 10°C in the surface layers and by several degrees at deeper depths. Varying other parameters in the energy budget within reasonable limits would not properly compensate for this change.

We also found that stability corrections to the boundary layer resistance ($\psi$ in Eq. [7]) are important for correct simulation of bare surface-soil temperatures. With low

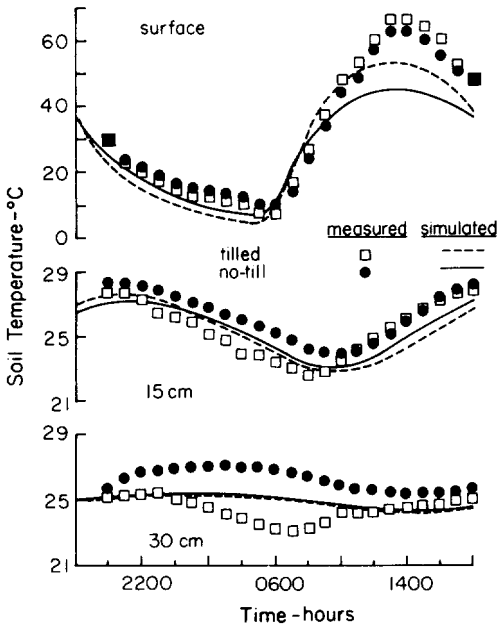


**Fig. 3—Comparison of measured and predicted field soil temperatures at 3 depths for conventional tillage and no-till from 2000 hours 6 August to 1800 hours 7 August.**

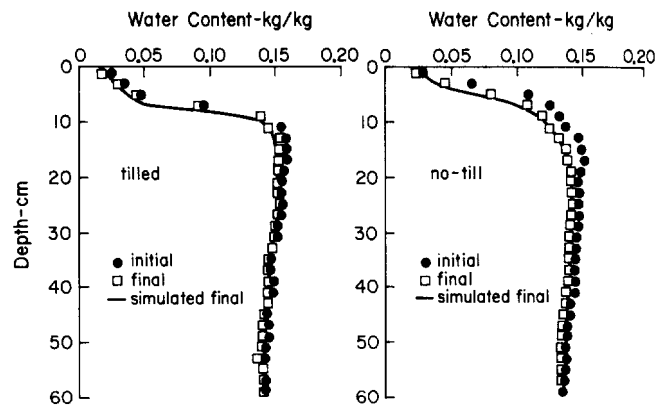


**Fig. 4—Comparison of measured field soil water contents for different soil depths at the beginning and end of a 2-week simulation with predicted water contents at the end of the 2 weeks for conventional tillage and no-till fallow.**

windspeeds, small roughness lengths, and clear skies, convective heat transfer between the soil and the atmosphere drops to almost zero at night, resulting in soil temperatures which are several degrees below air temperature. These soil surface temperatures are not correctly simulated when stability corrections are ignored.

Simulated temperatures differ from measured temperatures by <2°C at the 15- and 30-cm depths (Fig. 3). The oscillation of the measured temperature at the 30-cm depth of the conventional tillage may appear somewhat extreme when compared with the no-till. However, when considering the variability within treatments present under field conditions, the level of agreement between simulated and measured temperatures at all depths still seems reasonable for the modeling purposes here.

## Soil Water Content and Evaporation for the 15-day Simulation

Figure 4 presents measured water content profiles at the beginning of the 15-day simulation and compares measured and predicted values at the end of the period for the tilled and no-till treatments. Differences between simulated and measured water contents were generally <0.01 kg/kg for either treatment, and the agreement was not improved by adjusting the soil water diffusivities. More drying occurred in the seedzone of the no-till than in the tilled treatment. Simulated cumulative evaporative water loss for the 15-day period was 2.6 mm for conventional tillage and 4.7 mm for no-till.

The importance of thermally induced vapor flow in simulating actual field conditions was evaluated by excluding thermally induced vapor flow for the 15-day simulation on the no-till treatment. Measured and simulated water content distributions showed that simulated evaporative water loss was 13% (0.6 mm) greater without thermally induced flow than with thermally induced vapor flow included, and this resulted in poorer agreement between measured and predicted water content profiles. The largest water content difference between the two simulations was at the 5-cm depth, where it was 0.009 kg/kg. The difference between the simulations would increase with longer time and result in significant errors in predicting the seedzone water content. Thus, a major limitation of previous water flow models for our application in evaluating tillage effects on wheat germination and emergence is that they do not account for thermally induced moisture movement.

## The 90-Day Simulation of Soil Water Content and Evaporation

Both treatments were initialized with identical water content and temperature profiles. Results of the simulation for soil water content are shown in Fig. 5. At the end of the run, the water content of the seedzone (12 to 18 cm) is substantially lower with no-till than with conventional fallow tillage. This trend agrees with field measurements of Lindstrom et al. (1974) and Oveson and Appleby (1971) and with farmer observations. With no-till, the water content was 0.031 kg/kg lower than with conventional tillage in the soil directly below the tillage layer. This difference in water content corresponds to a water potential difference of about 600 J/kg for this particular soil. Seedzone water potentials in the tilled soil would be high enough for rapid germination

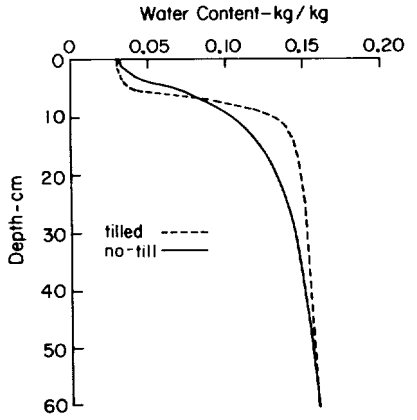


**Fig. 5.—Comparison of simulated field soil water contents at different depths for conventional tillage and no-till fallow at the end of a 90-day drying period.**

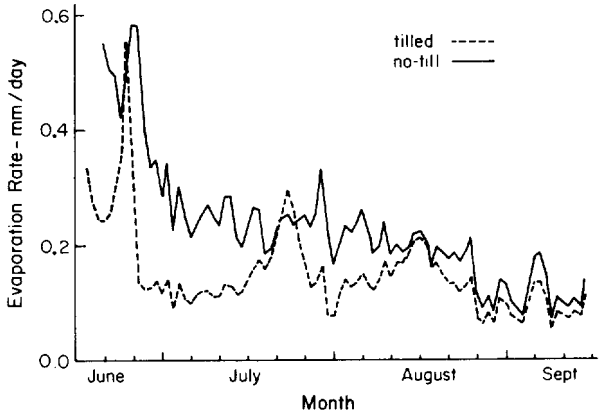


**Fig. 6.—Comparison of simulated evaporation rate for no-till and conventional tillage fallow during a 90-day simulation period.**

and emergence, while those in the no-till would slow stand establishment (Lindstrom et al., 1976).

Predicted total evaporative water losses for the 90 days were 14 and 24 mm for the conventional tillage and no-till treatments, respectively (Fig. 6). Loss rates averaged 0.16 mm/day for conventional tillage and 0.27 mm/day for no-till. These rates agree with other estimates in nearby areas (Papendick et al., 1973). Simulated evaporation from the no-till decreased steadily with time and approached the rate from conventional tillage at the end of the 90 days. The reduction in the evaporation rate on the simulated no-till treatment resulted from increased resistance to liquid flow to the surface as the shallow layers dried. The tilled soil layer greatly restricts liquid flow immediately, due to disruption of pore continuity with lower layers. This substantially reduces water loss during the time required for the no-till surface layers to dry. The lower loss rates from conventional tillage allow water lost from the seedzone to be replenished from layers below by flow due to matric potential gradients as suggested by Papendick et al. (1973).

## APPENDIX

$A =$ volume of air-filled pores, $m^3\ m^{-3}$
$c_p =$ specific heat of air, $J\ kg^{-1}\ °C^{-1}$
$c_s =$ specific heat of soil solids, $J\ kg^{-1}\ °C^{-1}$
$D_{va} =$ water vapor diffusivity in air, $m^2\ sec^{-1}$
$D_{\theta liq} =$ isothermal liquid water diffusivity, $m^2\ sec^{-1}$
$D_{\theta vap} =$ isothermal water vapor diffusivity, $m^2\ sec^{-1}$
$D_{Tliq} =$ nonisothermal liquid water diffusivity, $m^2\ sec^{-1}$

$D_{Tvap}$ = nonisothermal water vapor diffusivity, $m^2\ sec^{-1}$
$E$ = evaporative water vapor flux, $m\ sec^{-1}$
$g$ = gravitational acceleration, $m\ sec^{-2}$
$H$ = flux density of sensible heat, $W\ m^{-2}$
$h$ = atmosphere relative humidity, dimensionless
$h_{ai}$ = the relative humidity of the atmosphere when it is at the soil surface temperature, dimensionless
$h_s$ = relative humidity at the soil surface, dimensionless
$k$ = unsaturated hydraulic conductivity, $m\ sec^{-1}$
$K_o$ = oven-dry thermal conductivity, $W\ m^{-1}\ °C^{-1}$
$K_s$ = the thermal conductivity of the soil as a function of water content $W\ m^{-1}\ °C^{-1}$
$K_T$ = apparent thermal conductivity of soil, $W\ m^{-1}\ °C^{-1}$
$K_{T1}$ = total conductance for heat between soil surface and air, $W\ m^2\ K^{-1}$
$K_{\theta 1}$ = conductance of heat at the soil surface, $W\ m^{-2}\ K^{-1}$
$k$ = von Karmans constant
$L$ = latent heat of vaporization, $J\ kg^{-1}$
$m$ = air mass number, Eq. [28], dimensionless
$R_{ni}$ = isothermal net radiation, $W\ m^{-2}$
$r$ = soil surface boundary layer resistance, $sec\ m^{-1}$
$r_s$ = albedo, dimensionless
$S$ = solar declination, degrees
$S_t$ = total incoming shortwave irradiance, $W\ m^{-2}$
$S_p^o$ = solar constant, 1,360 $W\ m^{-2}$
$T$ = temperature, °C
$T^*$ = average temperature of soil surface and air, K
$T_a$ = air temperature, °C
$T_{ak}$ = air temperature, °K
$T_s$ = soil surface temperature, °C
$t$ = time, sec
$u$ = mean wind speed, $m\ sec^{-1}$
$u^*$ = friction velocity, $m\ sec^{-1}$
$z$ = vertical displacement, m
$z_H$ = roughness length for sensible heat exchange, m
$z_m$ = roughness length for momentum exchange, m
$\alpha$ = tortuosity, dimensionless
$\epsilon_A$ = sky emissivity, dimensionless
$\epsilon_s$ = soil surface emissivity, dimensionless
$\sigma$ = Stephen–Boltzman constant, $W\ m^{-2}\ K^{-4}$
$\zeta$ = atmospheric stability index, dimensionless
$\lambda$ = latitude, degrees
$\theta$ = soil water content, kg/kg
$\theta_a$ = equivalent water content of atmosphere above the soil surface, $kg\ kg^{-1}$
$\theta_s$ = equivalent water content of atmosphere at the soil surface, $kg\ kg^{-1}$
$\rho$ = density of air, $kg\ m^{-3}$
$\rho_v$ = saturation water vapor density of soil air, $kg\ m^{-3}$
$\rho_s$ = soil bulk density, $kg\ m^{-3}$
$\rho_v'$ = saturation vapor density of the air, $kg\ m^{-3}$
$\rho_{va}$ = water vapor density of air, $kg\ m^{-3}$
$\rho_{vs}$ = water vapor density of evaporating soil surface, $kg\ m^{-3}$
$\rho_{vs}'$ = saturation vapor concentration at the soil surface temperature, $kg\ m^{-3}$
$\rho_w$ = density of water, $kg\ m^{-3}$
$\psi$ = stability correction factor, dimensionless
$\tau$ = atmospheric transmission factor, dimensionless

## LITERATURE CITED


1. Benoit, G. R., and D. Kirkham. 1963. The effect of soil surface conditions on evaporation of soil water. Soil Sci. Soc. Am. Proc. 27:495–498.
2. Brutsaert, W. 1975. On a derivable formula for long-wave radiation from clear skies. Water Resour. Res. 11:742–744.
3. Businger, J. A. 1975. Aerodynamics of vegetated surfaces. p. 139–165. *In* D. A. DeVries and N. H. Afgan (ed.) Heat and mass transfer in the biosphere, part I. Transfer processes in the plant environment. John Wiley & Sons, Inc., New York.
4. Campbell, G. S. 1977. An introduction to environmental biophysics. Springer–Verlag, New York.
5. Crank, J. 1965. Mathematics of diffusion. Oxford University Press, London.
6. DeVries, D. A. 1963. Thermal properties of soils. p. 210–235. *In* W. R. van Wijk (ed.) Physics of plant environment. North–Holland Publishing Co., Amsterdam, The Netherlands.
7. Fink, D. H., and R. D. Jackson. 1973. An equation for describing water vapor adsorption isotherms of soils. Soil Sci. 116:256–261.
8. Fortier, S. 1907. Evaporation losses in irrigation and water requirements of crops. Bull. no. 177, USDA. U.S. Government Printing Office, Washington, DC.
9. Fritschen, L. J., and R. Hinshaw. 1972. A reed switch anemometer. J. Appl. Meteorol. 11:742–744.
10. Fuchs, M., G. S. Campbell, and R. I. Papendick. 1978. An analysis of sensible and latent heat flow in partially frozen unsaturated soil. Soil Sci. Soc. Am. J. 42:379–385.
11. Gardner, W. R. 1959. Solutions of the flow equation for the drying of soils and other porous media. Soil Sci. Soc. Am. Proc. 23:183–187.
12. Gardner, H. R., and W. R. Gardner. 1969. Relation of water application to evaporation and storage of soil water. Soil Sci. Soc. Am. Proc. 33:192–196.
13. Hammel, J. E. 1979. Modeling tillage effects on evaporation and seedzone water content during fallow in eastern Washington. Ph.D. Thesis. Wash. State Univ. Univ. Microfilms. Ann Arbor, Mich. (Diss. Abstr. 7915130).
14. Hanks, R. J., A. Klute, and E. Bresler. 1969. A numeric method for estimating infiltration, redistribution, drainage, and evaporation of water from soil. Water Resour. Res. 5:1065–1069.
15. Hillel, D. 1968. Soil water evaporation and means of minimizing it. Final technical report PL-480 Project A10-SWC-32 (mimeo) submitted to the USDA. Faculty of Agric., Volcani Inst. of Agric. Res. and Hebrew Univ. of Jerusalem, Rehovat, Israel.
16. Jackson, R. D., B. A. Kimball, R. J. Reginato, and F. S. Nakayama. 1973. Diurnal soil-water evaporation: time–depth–flux patterns. Soil Sci. Soc. Am. Proc. 37:505–509.
17. Jackson, R. D., R. J. Reginato, B. A. Kimball, and F. S. Nakayama. 1974. Diurnal soil-water evaporation: comparison of measured and calculated soil-water fluxes. Soil Sci. Soc. Am. Proc. 38:861–866.
18. Jury, W. A., and J. Letey, Jr. 1979. Water vapor movement in soil: reconciliation of theory and experiment. Soil Sci. Soc. Am. J. 43:823–827.
19. Kimball, B. A., R. D. Jackson, R. J. Reginato, F. S. Nakayama, and S. D. Idso. 1976. Comparison of field-measured and calculated soil-heat fluxes. Soil Sci. Soc. Am. J. 40:18–25.
20. Lindstrom, M. J., F. E. Koehler, and R. I. Papendick. 1974. Tillage effects on fallow water storage in the eastern Washington dryland region. Agron. J. 66:312–316.
21. Lindstrom, M. J., R. I. Papendick, and F. E. Koehler. 1976. A model to predict winter wheat emergence as affected by soil temperature, water potential, and depth of planting. Agron. J. 68:137–141.
22. McCall, M. A. 1925. The soil mulch in the absorption and retention of moisture. J. Agric. Res. 30:819–831.
23. Monteith, J. L. 1973. Principles of environmental physics. Edward Arnold Limited, London.
24. Oveson, M. M., and A. P. Appleby. 1971. Influence of tillage management in a stubble mulch fallow–winter wheat rotation with herbicide weed control. Agron. J. 63:19–20.
25. Papendick, R. I., M. J. Lindstrom, and V. L. Cochran. 1973. Soil mulch effects on seedbed temperatures and water during fallow in eastern Washington. Soil Sci. Soc. Am. Proc. 37:307–314.
26. Penman, H. L. 1941. Laboratory experiments on evaporation from fallow soil. J. Agric. Sci. 31:454–465.
27. Philip, J. R. 1957. Evaporation and moisture and heat fields in the soil. J. Meteorol. 14:354–366.
28. Philip, J. R., and D. A. DeVries. 1957. Moisture movement in porous materials under temperature gradients. Trans. Am. Geophys. Un. 38:222–232.
29. Staple, W. J. 1971. Boundary conditions and conductivities used in the isothermal model of evaporation from soil. Soil Sci. Soc. Am. Proc. 35:853–855.
30. Teten, O. 1930. Uber einige meteorologische begriffe. A Geophys. 6:297–309.
31. van Duin, R. H. A. 1956. On influence of tillage on conduction of heat, diffusion of air, and infiltration of water in soil. Reports of Agric. Invest. no. 62.7. Min. of Agric., Fisheries and Food Supply of the Netherlands, The Hague.
32. Willis, W. O., and W. A. Raney. 1971. Effects of compaction on content and transmission of heat in soils. p. 165–177. *In* K. K. Barnes et al. (ed.) Compaction of agricultural soils. Am. Soc. Agric. Eng., St. Joseph, Mich.