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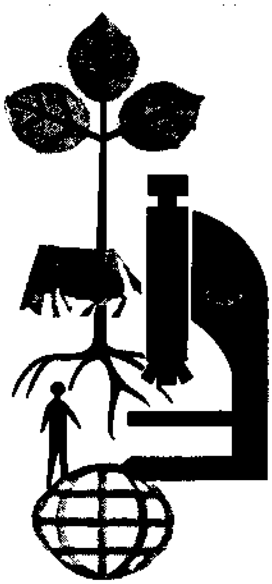

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• CHAPTER 5

The Influence of Soil Crusts on Heat and Water Storage

John W. Cary and Daniel D. Evans

There is a continuous exchange of water, air, and heat between the soil and atmosphere. This has a profound influence on weather, plant growth, and ground water storage. Increased advective energy transport and air turbulence are often consequences of heat exchanged between the soil and the atmosphere. The release of heat from a firmly packed soil which makes crop plants less apt to be frozen than those growing on a loose, recently cultivated soil is another example. The interchange of water between soil and atmosphere also profoundly affects our environment. If water reaches the soil surface faster than it can be absorbed, runoff and flooding occur. Water absorption by the soil is a basic requirement for ground water recharge and, consequently, necessary for the continued flow of all springs and wells. Taken from this veiwpoint, the exchange of heat and water through soil crusts is of quite general interest.

Heat and Water Vapor Flux through Soil Crusts

The important variables

Most of the energy arriving at the soil surface comes from the sun in the form of electromagnetic radiation. Once a portion of its radiation is absorbed by the surface and is converted to kinetic energy, it may be transported into the soil by molecular thermal conduction or as latent heat, the process being described as

$$J_H = -\lambda \nabla T + L J_v \quad [5\text{–}1]$$

where

$$J_v = -D \nabla C + CV \quad [5\text{–}2]$$

and J_H is heat flux, J_v vapor flux, D the apparent diffusion coefficient of water vapor in soil air, C the concentration of water vapor in the soil air, V the average net velocity of air through the soil pores, L the latent heat of vaporization, T the temperature, and λ the molecular thermal conductivity.

The relative magnitudes of each of the two terms in equation 5-1 can be estimated. Consider, for example, a bare surface receiving radiant energy from the sun such that a gradient of 5° C/cm occurs in the surface, as reported by Rose (1968). Under these conditions, a reasonable value for the thermal conductivity is 1 to 3 mcal(sec cm °C)$^{-1}$, giving a molecular conduction heat flux of 0.3 to 0.9 cal (cm^2 min)$^{-1}$ for the first term in equation 5-1. Assuming that the soil near the surface is damp enough (wetter than -15 bars) so that the relative humidity is near 100%, the vapor flow resulting from diffusion may be calculated for the second term in equation 5-1 as

$$J_v = \beta\,(1.56 \times 10^{-5} T^2 + 2.72 \times 10^{-3})\, \nabla T \quad [5\text{–}3]$$

where β is taken as 2 (Cary, 1966) and J_v = mm H_2O/hr, assuming the average temperature, T, to be 35° C. Multiplying by the heat of vaporization, one gets about 0.2 cal$(cm^2 sec)^{-1}$. Even very light wind above the soil surface can cause convective transfer of air in the soil surface layers, doubling the diffusive water vapor flux and causing values of LJ_v to be as large as 0.4 cal $(cm^2 min)^{-1}$ (Scotter and Raats, 1969).

The concentration of water vapor in the air is almost entirely controlled by the temperature when the soil moisture tension is less than 25 bars. Under drier conditions, concentration of water vapor is jointly controlled by the temperature and the amount of water and soluble salt in the soil. The effective diffusion coefficient of water vapor through the soil depends on the cross-sectional area of pore space in the soil and is inversely proportional to the path length or tortuosity of these pore spaces.

The molecular thermal conductivity λ increases as moisture content or bulk density increase. The thermal conductivity may be measured or calculated from the volume fractions of the soil's constituents and appropriate factors given by DeVries (1963).

The amount of shortwave radiant energy absorbed by the soil and converted to kinetic energy depends on the soil surface reflectance. Energy absorption is favored by large particles, rough surfaces, high moisture contents, and by dark surface colors. The radiant energy absorbed at the soil surface raises the temperature of the soil particles. These particles, in turn, transfer heat to the air around them and conduct heat downward into the cooler soil mass.

As heat moves downward and the soil temperature rises, vapor pressure of the absorbed water rises and it begins to diffuse in the vapor phase both up toward the lower vapor pressures in the air above the surface and down toward the lower vapor pressures in the cooler soil below. The upward diffusion rate will generally be less than 1 mm H_2O/day, while the downward diffusion rate may run about half this amount (Cary, 1967a).

Gentle winds or convective heat transfer in the air above the surface cause turbulence and pressure changes in the first few cm of soil which create viscuous flow of gas in the open soil pores. This increases the transfer of water vapor and, consequently, the flow of latent heat. Increasing the density of the soil reduces the viscuous flow of air and latent heat since viscuous flow is proportional to the square of the pore radius so long as the flow is laminar and the pore area is constant. On the other hand, increasing the density increases the thermal conductivity of the soil and increases the flow of heat by conduction.

During the nighttime there is a net loss of longwave energy from the soil surface as it radiates to the sky. This causes an upward soil heat flux. In general, on a clear night the air will be warmer than the soil surface and so will lose heat to the soil. The upward thermal gradient in the soil also creates an upward vapor pressure gradient, causing water vapor diffusion from deeper soil toward the surface. A packed soil surface, or one with a thick crust, favors nighttime heat loss. Gradwell (1963) has reported that a bare soil with a bulk density of 0.8 lost 40 to 50 cal/cm^2 during the dark, while denser soil, 1.2 gm/cm^3, lost 58 to 89 cal/cm^2.

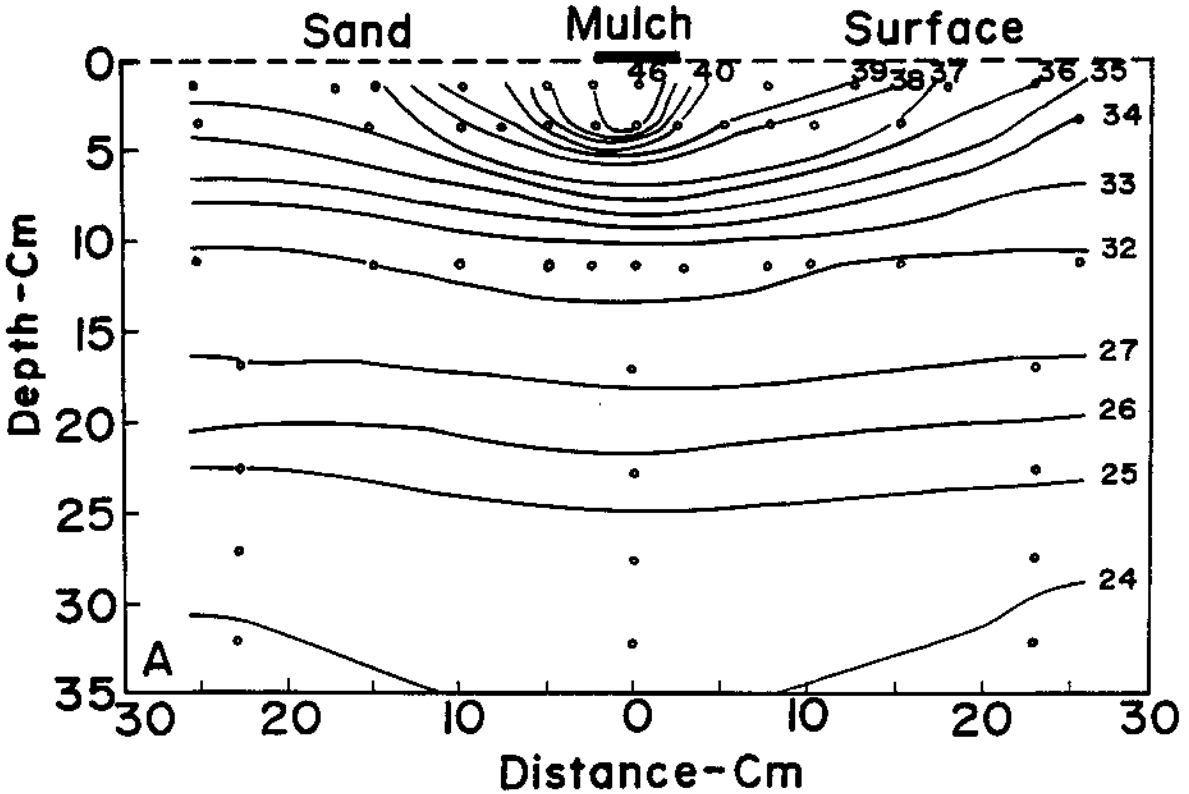
Modifying heat and vapor flow through crusts

Artificial layers such as asphalt, plastic, or gravel mulches, will strongly affect the transfer of energy between the soil and atmosphere as well as decrease crusting. Miller (1968), working at Prosser, Washington, found that an asphalt mulch sprayed in 30-cm wide strips over rows of sweet corn seed raised the daytime soil temperatures at a depth of 6.4 cm by as much as 3° C. However, nighttime temperatures on all treatments at this depth approached the same minimum. Measurements made in southern Idaho of net radiation 1.5 meters above silt loam soil with 8-cm-wide asphalt strips sprayed on 48-cm center rows showed no detectable difference from the smooth check plots.

Kowsar et al. (1969) have studied the effect of petroleum mulch on soil water content. In addition to increased soil temperatures, they found an increase in water content several centimeters below the mulch. This was evidently caused by surface sealing and a downward flux of water vapor away from the warm soil-asphalt interface. Qashu and Evans (1967) studied the effect of a black granular mulch on soil temperatures and water distribution. These treatments had a profound effect on afternoon soil temperatures (Figure 5-1), and consequently increased the heat exchange between the soil and atmosphere. Field observations showed that the soil 2.5 cm below this granular-type mulch retained more water than the control. A friable soil layer formed under the mulch, while a hard crust was formed where no mulch was present. Evidently the vapor transfer upward through the mulch was less than the net upward transfer through the dry soil crust. Because of the higher soil temperatures, the vapor pressure gradient toward the surface should have been greater under the mulch. However, the air velocity term in equation 5-2 may have been enough smaller to account for the net decrease. There was considerable cracking of the natural soil crust which could have encouraged convective transfer of the water vapor. It is also possible that differences in hydraulic conductivity were involved.

Bresler and Kemper (1970), working with columns of soil in the laboratory, demonstrated differences in drying rates which were associated with soil crusts. Soil columns were wetted by flooding, artificial rain, and

Figure 5-1. The effect of the shape of 2 black granular mulches on soil temperatures in the field. Figure A has a 5-cm wide mulch, and Figure B a 10-cm V-shaped mulch (Qashu and Evans, 1967).



Sand
Mulch
Surface
Depth-Cm
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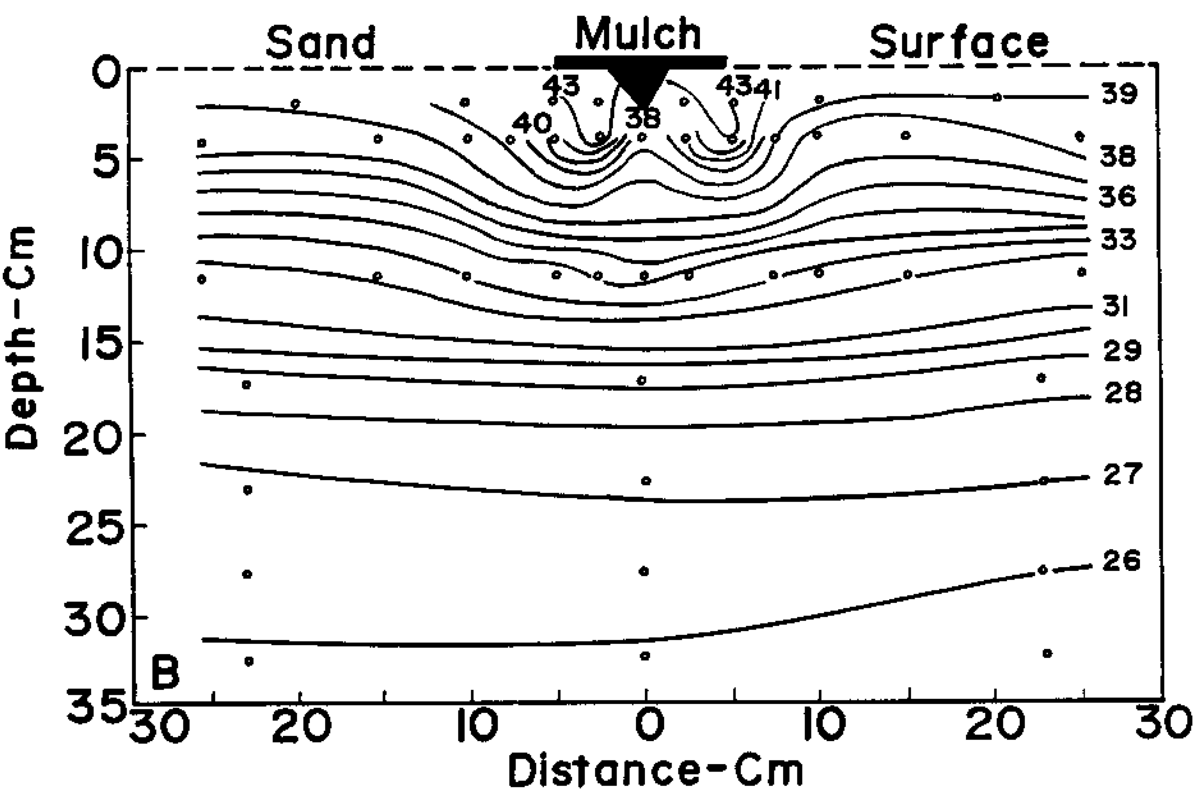

Sand
Mulch
Surface
Depth-Cm
Distance-Cm
B
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rain with NaC1 added at the surface. The columns wetted by rain formed the most severe crusts and also showed lower evaporation rates during the next few days as drying proceeded. Because the pores in the crust were generally smaller than those just below it, hydraulic conductivity was higher in the crust than at the crust-soil interface during the initial stage of drying as the larger pores at the interface began to empty. Consequently, the crust-soil interface was unable to transmit enough water to meet the evaporative demand, causing the pores in the crust to empty and dry rapidly. The dry crust then helped insulate the moist soil from the incoming heat necessary to cause evaporation.

The effects of a crust on water infiltration and its subsequent return flow toward the surface upon drying may cause significant secondary changes in evaporation through salt redistribution (Quyyum and Kemper, 1962). The water-holding capacity and hydraulic conductivity of many soils is dependent on the type and amount of salt present (Rasmussen and McNeal, 1973). Another possible effect is the concentration of salt in the air-water interface during evaporation. Evaporation rates as low as 0.5 mm/day can more than double the amount of salt at the sites of evaporation in wet soil (Cary, 1965), and the concentration may increase by an order of magnitude with greater evaporation rates. Though increasing the salt concentration decreases vapor pressure, resulting in decreased evaporation under isothermal conditions, conditions in the field are not isothermal, and when the sun's energy is not used for evaporation more heat goes to warming the soil. As the soil warms, the vapor pressure rises, and so the effect of salt accumulation on vapor pressure tends to be reduced under real field conditions. Another effect of salt, and perhaps its most important, involves the reflection of light. If, upon drying, the crust surface is covered with enough crystalline salt to give it a lighter color, the reflectance will increase and both heat and water vapor flow through the crust will decrease (Cary, 1967a).

It is possible to form a variety of crusts on the same soil by different management treatments. Cary (unpublished data) formed 4 different soil surfaces on the Portneuf silt loam by flooding, subirrigating, sprinkling, and dry mulching the soil in replicated 10-liter containers. During a period of 60 days in the field, differences in water evaporation between the four treatments were less than 10% and so within the experimental uncertainty. The average daily evaporation rate is compared to the average U. S. Weather Bureau pan evaporation in Figure 5-2. At the end of the drying period, the surface-sprinkled treatment had a dense surface crust 3 to 5 mm thick overlaying several centimeters of very dry friable soil. The treatment that had been wet only by subbing had no detectable crust and differed from the dry mulched treatment only in a denser surface. The treatment that had been flooded had a massive hard crust on the top 2 cm which had cracks about 0.5 cm wide, but only 2 cm deep.

It appears, assuming identical initial moisture conditions, soil crusts may have negligible effects (i.e., less than 10%) on water loss from the soil under field conditions, provided there is no difference in color and provided the crusts do not penetrate or crack more than 2 or 3 cm below the soil surface. However, in the event of restricted initial water infiltration, deep cracking, or dust mulches 10 cm or more deep, differences in drying will occur (Adams, et al. 1969; Papendick et al. 1973).

The effect of natural soil crusts on heat transfer was also studied on the silt loam soil at the Snake River Conservation Research Center in southern Idaho. Basins approximately 3 m^2 were formed on a field plot, and different soil crusts developed on them as shown in Figure 5-3. Soil temperatures were measured on a clear, calm afternoon in September with air temperatures in the 70's. There was no significant difference in temperatures at the 10 cm depth, but average surface temperatures during the middle of the afternoon were:

Plot No.	1	2	3	4	5
Temperature °F	94	87	89	84	90

The high temperature on plot No. 1 was caused by the low thermal conductivity of the rototilled surface. Its bulk density was only 1.04 g/cm^3. The low surface temperature of plot 4 was caused by a light-colored layer of silt which had been deposited on the surface by the irrigation water. The surface crust bulk density of all plots except No. 1 ranged from 1.5 to 1.7 g/cm^3; significant differences could not be measured because of the experimental error in volume measurements.

The crust on plot 5 was most severe. Because of extreme cracking, the soil below it had dried out more than under the other crusts. As plot 5 dried, it tended to form two layers of crust. The upper one, about 1 cm thick, tended to peel away from the lower layer of hard, massive material 4 to 6 cm thick, which showed a definite vesicular structure similar to Figure 1-1.

The crust on plot 4 was about 4 cm thick and had developed some vesicular structure 1 cm below the surface. This vesicular structure did not develop in the crusts formed by sprinkler irrigation and natural rainfall. The only obvious difference between the ponding of water on plots 2 and 4 was that the water ponded on plot 4 contained suspended silt and clay, whereas that on plot 2 remained clear. The least severe crust was formed by natural rainfall. It should be noted, too, in Figure 5-3 that the severe soil crusts reduced emergence of weeds which could ultimately have profound effects on heat and water transfer between the soil and the atmosphere.

The density of a dry soil surface is important in determining heat exchange (Allmaras et al., 1972). An example of this effect was shown with net radiation and

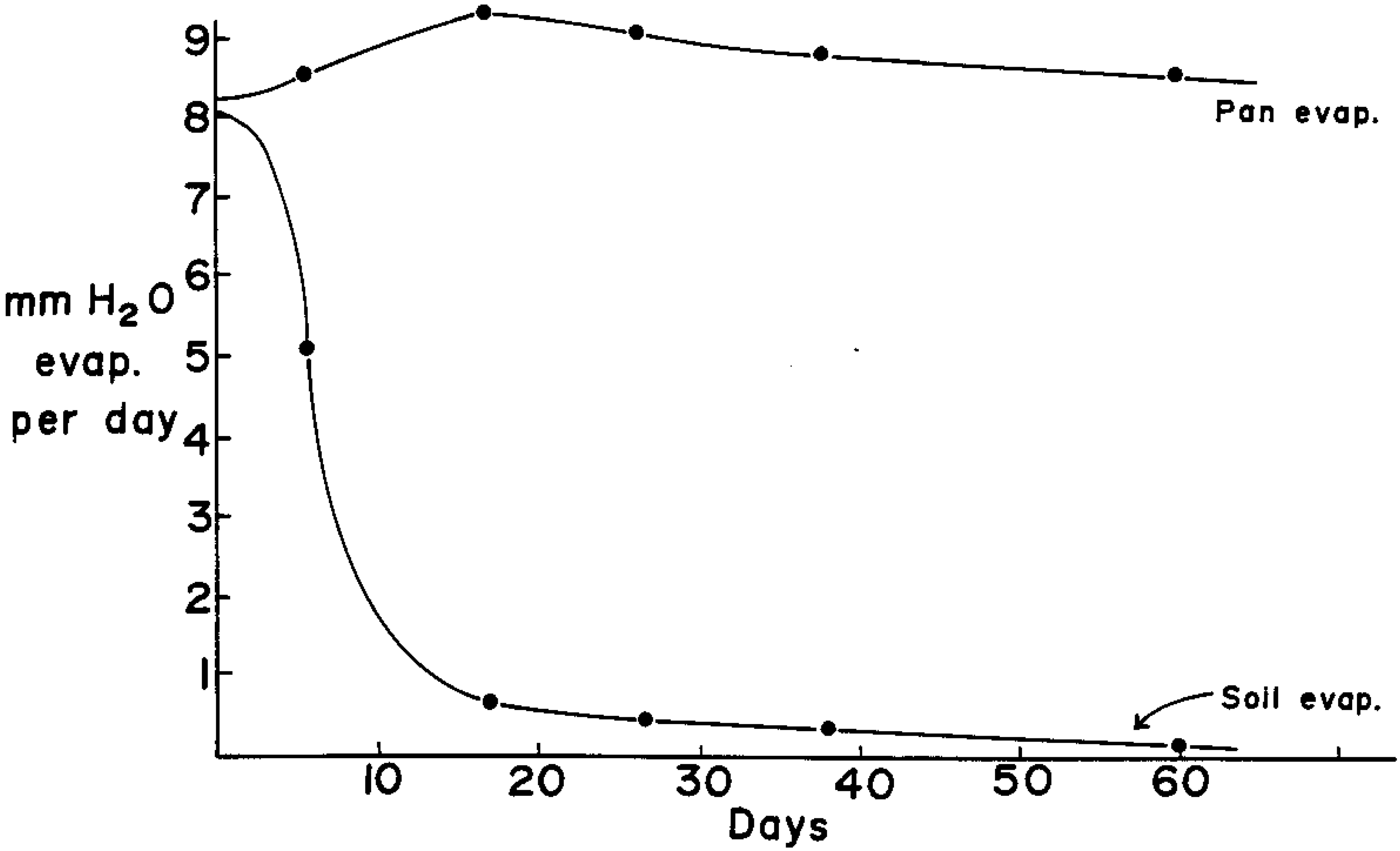


Figure 5-2. Comparison of evaporation from bare soil to a free water surface.

soil temperature measurements on the silt loam soil in southern Idaho (Cary and Wright, unpublished data). During a warm afternoon in June, the net radiation was 2 or 3 % less over a smoothly packed plot than over the granular hand-raked surface of the control. Daytime soil temperatures at the 6-cm depth were about 2.5° C warmer under the packed surface than under the control. This temperature increase disappeared during the night because of the greater upward heat flux described by Gradwell (1963). Infrared measurements of afternoon surface temperatures showed about the same trend as the 6-cm soil temperatures.The higher soil temperatures in the packed plot resulted from a greater downward conduction of energy from the soil surface. The lower net radiation of the packed plot suggests that it had a significantly lower sensible heat flux to the air than did the rougher check surface.

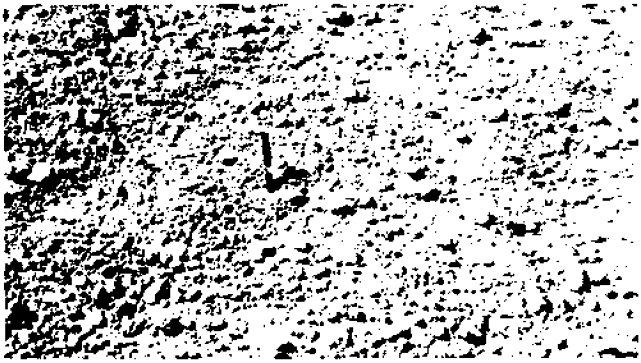
It appears from these observations, that one should not expect a soil crust to affect the heat flux across the soil-atmosphere interface by more than a few percent unless there is a large increase in bulk density, an obvious change in surface color, deep cracking to encourage greater evaporation of soil moisture, or a reduction in plant cover.

Transport of Liquid Water through Soil Crusts

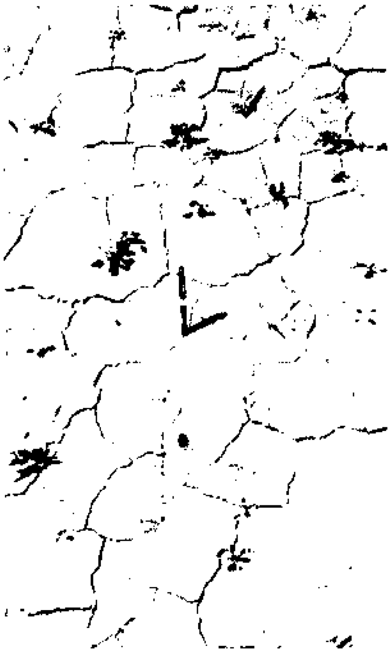
Theory

A description of the classical theory of water flow in soil has been reviewed by Miller and Klute (1967). The movement of water within a soil is described by a

Figure 5-3. Crusts formed for the heat exchange study: 1. Dry mulch. 2. Sprinkled with ponding. 3. Gentle rain without ponding. 4. Surface irrigated with ponding. 5. Ponded and puddled.

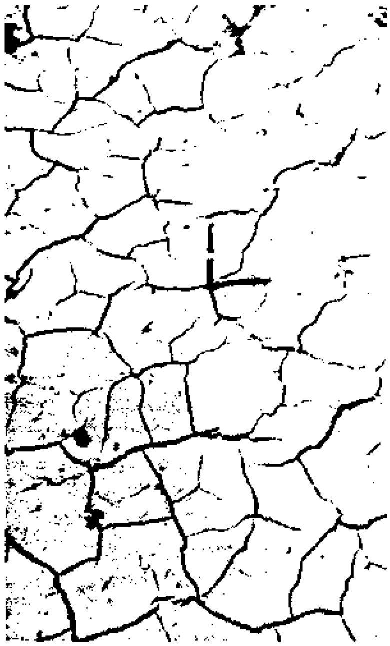


PLOT 1



PLOT 2

PLOT 3



PLOT 4

PLOT 5

Darcy-type equation which, for the vertical direction only, is

$$J = K(dS/dz + 1) \qquad [5\text{–}4]$$

where J is the volume of flow per unit area per unit time, K is the hydraulic conductivity and a function of the water content, and dS/dz is the soil suction gradient in the z direction. Combining equation 5-4 with the conservation of mass principle for a small element of soil volume gives:

$$\frac{\partial\,(K\ \partial S/\partial z)}{\partial z} + \frac{\partial K}{\partial z} = \frac{\partial \theta}{\partial t} \qquad [5\text{–}5]$$

where θ is the moisture content.

Equation 5-5 has been solved by various techniques for simple cases to obtain the soil water content or soil suction as a function of depth and time, or the infiltration or evaporation rate as a function of time. The usual assumption is made that the soil is homogeneous and isotropic within the flow region under consideration. Philip (1957) solved equation 5-4 for a uniform soil and initial water content to obtain an equation which approximates the accumulated water intake I by the soil at various times after water has been continuously applied at the soil surface for $t > 0$. The solution was

$$I = At^{1/2} + Bt \qquad [5\text{–}6]$$

where A and B are soil parameters. Equation 5-5 has not been solved analytically for layered or crusted soils, though numerical solutions have been tested for flow into layered soils (Miller and Klute, 1967).

Hillel and Gardner (1969) considered the effect of a crust on the steady-state infiltration of water, that is, at a time when the wet front had reached a sufficient depth so that the infiltration rate was approaching a constant value. Under these conditions

$$J_c = J_u \qquad [5\text{-}7]$$

or

$$K_c[(dS/dz)_c + 1] = K_u[(dS/dz)_u + 1] \qquad [5\text{-}8]$$

where the subscripts c and u refer to the crust and subcrust, respectively. The suction gradient in the subcrust goes to zero under stead conditions, so that

$$J = K_u = K_c \frac{S_c - S_o + L_c}{L_c} \qquad [5\text{-}9]$$

where S_o and S_c are the soil suctions at the soil surface and the bottom of the crust, respectively, and L_c is the thickness of the crust. Assuming that $S_o = 0$, $L_c << S_c$ and S_c does not exceed the air entry value of the crust (i.e., K_c is constant and equal to the saturated conductivity), then

$$\frac{K_u}{S_c} = \frac{K_c}{L_c} \qquad [5\text{-}10]$$

or

$$J = S_c \frac{K_c}{L_c} \qquad [5\text{-}11]$$

Equation 5-11 then predicts that the steady-state infiltration rate will be less for thicker crusts and for those with lower conductivities.

In general, the theory indicates that the hydraulic properties of the crust and the subcrust interact to cause a steady infiltration rate and moisture profile. The suction in the subcrust adjusts to a constant value, creating a suction gradient across the crust sufficient to make the flow rate through the crust identical to the flow rate below the crust. Hillel and Gardner obtained laboratory results which agree reasonably well with theoretical predictions.

Of more general interest is infiltration of water under transient conditions, i.e., when the infiltration rate and soil moisture profile are changing with time. Hillel and Gardner, in a later paper (1970), examined the effects of a soil crust for transient conditions using an approach proposed by Green and Ampt (1911) and later by Philip (1957). The pertinent assumptions are: (1) there is a constant effective suction at the wetting front; and (2) there is a constant water content profile and hydraulic conductivity in the subcrust above the wetting front. These assumptions simplify the flow equation to a form amenable to analytical solution.

For a uniform profile and vertical infiltration, the Darcy-type equation giving the infiltration rate i at any particular time is

$$i = K\,\frac{S_f - S_o + L_f}{L_f} \qquad [5\text{-}12]$$

where S_f and L_f are the effective suction at the wetting front and the depth of the wetting front, respectively, S_o is the suction at the soil surface which may be taken as zero for a thin layer of water on the surface.

To account for the effect of a crust, we can write

$$J = K_u\,\frac{S_f - S_c + L_u}{L_u} \qquad [5\text{-}13]$$

where $S_f - S_c$ and L_u are the effective suction difference and the distance between the lower boundary of the crust and the wetting front, respectively. The equation applies only after the wet front has passed through the crust. If the crust is such that it initially saturates and remains saturated, then J is equal to the infiltration rate i and

$$S_c = \frac{i\,L_c}{K_c}\,. \qquad [5\text{-}14]$$

Also,

$$i = \frac{dI}{dt} = \triangle\,\theta\,\frac{dL_u}{dt} \qquad [5\text{-}15]$$

where I is the accumulated intake in the subcrust, and $\triangle\,\theta$ is the increase in water content between the crust and the wetting front. Combining equations 5-15, 5-14, 5-13, and integrating over depth and time gives:

$$L_f - (S_f - L_c K_u / K_c) \ln \frac{S_f + L_f}{S_f} = \frac{K_u t}{\Delta \theta} \quad [5\text{–}16]$$

Equation 5-16 cannot be solved explicitly for L_f as a function of time. However, a trial and error procedure may be used to evaluate the depth of wetting as a function of soil properties and time. Again, as the thickness of the crust or the ratio K_u/K_c increases, the infiltration rate decreases.

Application of the theory

Edwards and Larson (1969), using a numerical analysis of infiltration through a surface seal into a homogeneous soil, found results in agreement with those predicted in the preceding section. From their study, they also predicted that the suction gradient in the surface layer should increase as K_c decreases and as K_u increases. This suction gradient increase could partially offset lower infiltration rates caused by low K_c values.

The theory predicts that the thicker and denser the crust, the lower the infiltration rate will be. The effect of change in porosity (or conductivity) on infiltration rate is illustrated in Figure 5-4 by data published by Miller (1969). The wetting front moving from the silt loam soil into the sand layer is analogous to wetting front moving through a compacted dense crust and into the more porous soil below.

An increase in exchangeable Na at the soil surface leads to dispersion and plugging of the larger pores during periods of high water content. Water flow conditions then become those described by the layered soil theory. However, as soluble salts are leached from the soil surface, K_c will not be constant, but will decrease rapidly and this must be accounted for in any analysis. As pointed out in Chapter 2, Figure 2-3, calcium may be added to the soil's surface to alleviate this problem.

Though cracks are not really covered in detail in any mathematical formulation of infiltration, they are extremely important in increasing the initial entry of water into dry crusted soils (Ritchie et al., 1972). In some heavy clay soils, cracks more than 10 cm wide and over 50 cm deep may develop. Cracks of this size increase the total surface area of the soil-ätmosphere interface by 3 to 5 times (Adams et al., 1969), as well as being very effective in trapping runoff water. Even relatively small cracks such as those shown in Figure 5-3 increase infiltration during the initial period of surface wetting. Because of a lack of cracking, a high sand content in a soil surface crust may actually be detrimental to infiltration. Kemper and Noonan (1970) found that maximum runoff from rain falling on crust-prone soils occurred when the sand content was between 50 and 80%.

A soil crust may also affect the upward movement of liquid phase water in at least two ways. If the crust if severe enough to restrict the intake of water during wetting, the final moisture content of the soil immediately below it will be lower than if the intake had been normal. This will generally be true even if the infiltration time is extended so that the total net intake for the restricted and normal cases are the same. A lower moisture content below the crust will result in a lesser upward movement of liquid phase soil water in response to evaporation from the surface.

A second and more direct way in which a crust may control upward movement of soil water results from abrupt changes in pore size distribution. As water moves from a soil layer with one given pore size distribution into a layer with a different size distribution, the suction gradient will remain fairly smooth and continuous, but because of the different water-holding capacities of the different layers at a given suction, the moisture content gradient may change abruptly. In a transient unsaturated system, increasing the outflow rate requires an increased water tension gradient. This requires a decrease in moisture content. Generally, when comparing unsaturated soils, a lower water content indicates a lower liquid conductivity, and an increase in tension can cause a sharp drop in conductivity. The drop in conductivity forces even a larger change in tension gradient, which produces yet a lower conductivity, and soon the liquid flow becomes very small.

Another good example of the effect of the discontinuity in pore size is shown by the data of Bresler and Kemper (1970) in Figure 5-5. In this experiment, columns of crust-prone soil were wet by rain with NaC1 on the surface, by flooding, and by slow infiltration through filter paper. These treatments resulted in a decreasing severity of crusts. The surfaces were allowed to begin drying, and the conductivity calculated 1, 2, and 4 hours later. The soil's resistance to water flow (1/K) is

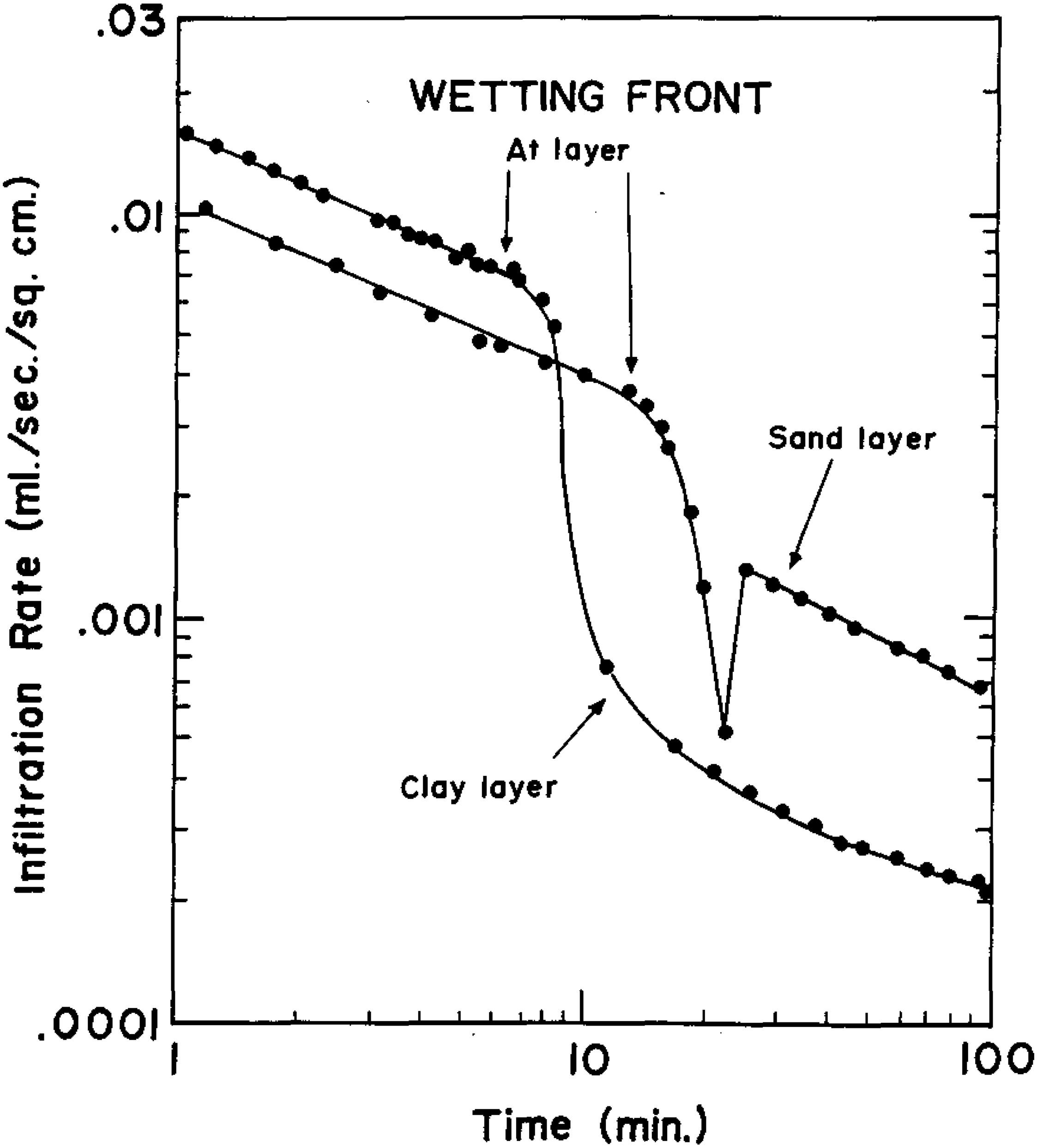


Figure 5-4. Effect of sand and clay layers on infiltration rate into Palouse silt loam as function of time (Miller, 1969).

plotted in Figure 5-5 and clearly shows the increase which developed between the crust and the underlying soil. The greater the discontinuity, the sooner liquid phase flow will approach zero at the crust-soil interface during high evaporative demands at the surface.

By way of general conclusions, one must realize that a soil crust can reduce water intake by one or two orders of magnitude, and the result may be flooding and erosion. Effects of crusts on water vapor and heat exchange between the soil and the atmosphere are more subtle, but nonetheless real. The extent and depth of cracking, soil density, surface texture, and color are important. Severe soil crusts may ultimately have the greatest effects on water and energy exchange by preventing the establishment of plants, including weeds.

Figure 5-5. The resistance of soil to water flow (1/K) as a function of soil depth for 3 soils with decreasing degrees of surface crusts. The curve parameters are hours after drying had begun (Bresler and Kemper, 1970).