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THE DRYING OF SOIL: THERMAL REGIMES AND AMBIENT PRESSURES

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SUMMARY

Evaporation from a loam soil was studied in the laboratory during the falling-rate stage of drying. The primary variables were temperatures regimes and ambient pressures. The data suggested that: (*a*) there is a significant relation between the soil temperature and the vapor pressure of water in the air above the surface; (*b*) the water vapor transfer through the dry layer at the surface is accomplished in part by viscous air flow; (*c*) diurnal thermal gradients may cause a significant movement of water near the soil surface during the falling-rate stage of drying. Consequently, the conclusion follows that the isothermal "soil-moisture diffusion theory" cannot be expected to describe drying during time intervals of less than 24 hours. Because of the periodic nature of the thermally induced transfer phenomena, the "diffusion theory" will give curves which approximate cumulative drying over long time periods, though such relations must be considered as semi-empirical.

INTRODUCTION

The importance of evaporation from soil can hardly be overemphasized when one considers the vast arid regions on earth. A large amount of work and study concerning this process has been conducted; however, because of the complexity of the soil and the continuously changing variables associated with the atmosphere, much remains to be learned. The literature concerning the drying of porous media has been reviewed by WIEGAND and TAYLOR in 1961, and in a thesis recently completed by COVEY (1965). It is now generally believed that so long as the soil surface is moist, the rate of evaporation is controlled principally by atmospheric conditions. However, as soon as the surface dries, the evaporation rate decreases sharply and may not appear to be directly related to conditions above the surface.

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Experimental studies indicate that when the soil surface is dry, further evaporation occurs principally from a plane of "drying sites" located below the surface where dry soil grades into moist. These drying sites are always subjected to thermal gradients (WIEGAND and TAYLOR, 1962) as well as osmotic and moisture-content gradients. This complex process of drying (often referred to as the falling-rate state) has been studied analytically by PHILIP (1957), GARDNER and HILLEL (1962), and HANKS and GARDNER (1965). These theoreticians have tended to conclude that for most practical applications, evaporation may be treated as an isothermal process. This opinion has led to a soil-moisture "diffusion equation" type of analysis. Because of the variety of solutions available for this equation, it can be fitted to a number of differently shaped curves. Theorerical diffusion drying curves based on water content–conductivity and water content–suction relations (HANKS and GARDNER, 1965) do bear a good resemblance to field observations; see, for instance, those published by WIEGAND (1962). However, Wiegand was able to analyze his data with a simple empirical relation between time and moisture content.

Most of the carefully controlled laboratory experimental work on the drying of soil has been carried out under steady-state environmental conditions, rather than under a diurnally fluctuating system as occurs in the field. The results of these laboratory studies have formed much of the basis for analytical analysis. WIEGAND and TAYLOR (1962), studying evaporation from soil columns under closely controlled conditions, found that it was impossible to dry soil columns isothermally because of the latent heat flow associated with the evaporation of water vapor. In their conclusions, the suggestion was made that isothermal analysis of the phenomena may not be adequate. KING and SCHLEUSENER (1961) did make a study of soil drying under a diurnally changing laboratory environment; however, they were primarily looking for hysteresis effects on moisture flow. GARDNER and HANKS (1966) included a fluctuating temperature regime in recent laboratory evaporation study, but presented only a cursory analysis of the results.

Because theoretical studies of the falling-rate stage have been developing with the constant temperature assumption, emphasizing only the moisture-flow properties of the soil, it appeared to be worthwhile to collect some data to determine the applicability of the isothermal assumption. In particular one might wonder about the effect of the cyclic potential evaporation changes at the soil surface and the thermally induced water transport in the soil. Because thermally induced moisture flow is predominately in the vapor phase under the drier soil conditions, ambient pressure can be used as a tool to study the importance of vapor diffusion in relation to other variables affecting the drying rate (JACKSON, 1965).

EXPERIMENT

The experimental work was done in the laboratory under controlled conditions. An insulated lucite cylinder 18 cm in diameter and 75 cm deep was used to hold a Columbia loam soil sample. A resilient layer of plastic sheet and plastic foam was used to line the inside cylinder wall. This maintained a seal between the vertical walls and the soil as drying took place. An airtight lid was placed over the top of the column so that experiments could be done at different ambient pressures. A 12-cm air space was left between the top of the soil and the cylinder's lid. A $7^1/_2$-W light bulb was mounted inside this space to supply heat and radiant energy to the soil's surface. The soil was dried with an air stream that had been passed first through a column of silica gel to remove the moisture, and then through a heat exchanger to bring it to the desired temperature. The air left the chamber through a lithium-chloride relative humidity cell, an air-flow meter, and finally passed into another silica-gel column to remove the moisture absorbed from the soil's surface. The humidity sensor and the thermistors in the soil profile were connected to a recorder. Thus, as the drying proceeded, the soil's temperature profile and the relative humidity of the air leaving the system were continuously monitored. Temperatures were measured at 0, 1.1, 3.5, 7.5, 13.5, 21.5, and 40 cm below the surface. The air flow was adjusted to a constant 2 l/min so that the relative humidity meter could be calibrated to record the instantaneous evaporation rate. Air-flow velocities across the soil surface had an average velocity of less than 1 m/min.

The heat exchanger controlling the temperature of the air stream and the $7^1/_2$-W light in the air chamber above the soil were programmed as desired to give either diurnally changing soil-surface temperatures or various steady-state energy supplies to the soil surface. The whole experimental system was contained in a controlled temperature box at $24 \pm 1°$ C.

With the exceptions of a few short-term pressure tests, all observations were made on the same large soil column. The soil was packed into the cylinder to a bulk density of 1.15 ± 0.04 g/cm^3 and wet to a depth of approximately 60 cm. The sample was then put through a series of wetting and drying cycles over a period of about 6 months while the monitoring equipment was calibrated and perfected, thus giving the soil a chance to settle and stabilize.

The soil was then subjected to nine different drying conditions. In order to apply these nine treatments to the single soil column so that they would be comparable, the following procedure was used. The surface 3 or 4 cm of soil was allowed to become dry, while the soil below 4 cm remained at approximately 10% moisture to nearly the bottom of the column. At this time 7.5 mm of water was added to the surface and allowed to infiltrate for 2 days without evaporation. A 10-cm core for soil moisture was taken from the perimeter of the column with a probe 1.5 cm in diameter. One of the nine environmental conditions was then chosen and evaporation allowed to take place until 7.5 mm of water had been removed from the soil. Moisture samples were again taken, the surface rewet, allowed to stand for 2 days and another treatment applied, etc.

The experimental treatments were:

(*1*) Continuous heating, that is the air was passed into the sample at a constant temperature of 30°C and the $7^1/_2$-W bulb in the chamber above the soil was left burning continuously.

(*2*) Continuous cooling, that is the air was passed into the soil chamber after having been brought to 0°C and no light was used.

(*3*) An alternating temperature pattern was created by 12 h of heating and 12 h of cooling, as described in treatments *1* and *2*.

(*4*) A heating and cooling cycle was developed by heating the surface for 12 h as described in *1*, after which the surface was covered with a thin plastic film and cool air passed over it as described in *2*. The principal difference between this treatment and treatment *3* was that no evaporation was allowed to occur during the 12-h cooling period.

(*5*) A heating and cooling cycle was used in which the first 12 h of the cycle was heating as described in *1*, followed by 2 h of cooling as described in *2*, then the surface was covered with plastic film and packed with ice for 8 h, followed then by 2 h of the cooling treatment *2* before the heating cycle was again resumed.

(*6*) Treatment *3*, with an ambient pressure of 1.53 bar and the soil surface bare.

(*7*) Treatment *3*, with an ambient pressure of 0.65 bar and the soil surface bare.

(*8*) Treatment *3*, with an ambient pressure of 1 bar, and the soil surface covered with a clear plastic sheet containing small 1-cm slits such that the porosity was approximately 15%.

(*9*) Treatment *3*, with an ambient pressure of 1 bar, and with the surface covered with an aluminum foil sheet with a pattern of slits identical to those in the plastic cover described in *8*.

The cumulative evaporation resulting from these treatments is summarized in Table I and II. Typical maximum and minimum temparature profiles which were developed in the soil by the first four thermal regime treatments are shown in Fig.1. The temperature extremes at the surface were comparable to those which occur under natural field conditions. Shown in Fig.2 are the initial and final soil moisture contents for experimental treatments *1* through *5*. Since they are reasonably similar, it was felt that the differences between the treatments were not caused by different initial soil conditions. Moisture contents associated with experimental tests *6* through *9* were similar in shape to the ones shown in Fig.2, except they were approximately 2% higher.

In addition to treatments *1* through *9*, several short-term pressure experiments were conducted to study specifically the relation between temperature and vapor diffusion. The same soil and equipment were used in these tests, except the length of the soil column was shortened to 10 cm so that when the ambient pressure was changed there would be less viscous air flow through the soil profile due to the compression or expansion of air. An air-flow control plate was also used in these tests. This consisted of a lucite disc laid on the soil's surface such that incoming air was forced to flow in under the perimeter of the disc and out through a port in the center. The experimental procedure was essentially the same as described in treatment *3*, except that during various stages of drying, changes in ambient pressure were made over 1 or 2-h periods. These short-term changes served to illustrate some of the interrelated phenomena between soil temperature and vapor transport.

TABLE I

EFFECTS OF TEMPERATURE REGIMES ON EVAPORATION, MAXIMUM SOIL TEMPERATURE AND VAPOR FLUX

Treatment[1]	*Cumulative evaporation (mm H_2O), time (days)*							*Maximum soil temperature after 3 days (°C) at soil depths (cm)*			*Moisture loss on the 4th day (mm H_2O), period*		*Vapor flux across the 5-cm depth on the 4th day (mm H_2O), period*	
	1	*2*	*3*	*4*	*5*	*6*	*7*	*1.1*	*3.5*	*6.5*	*heating*	*cooling*	*heating*	*cooling*
1	2.5	3.9	5.0	5.9	6.6	7.4	-	37.4	32.1	30.6	0.9	-	—1.0	-
2	1.7	3.1	4.1	4.9	5.6	6.2	6.8	18.2	19.4	20.2	-	0.8	-	0.43
3	2.2	3.5	4.4	5.3	6.1	6.8	7.5	34.6	29.4	28.2	0.45	0.35	—0.24	0.15
4	1.8	3.1	3.8	4.5	5.0	5.6	6.1	35.3	30.2	29.3	0.7	-	—0.30	0.14
5	2.2	3.5	4.5	5.2	5.9	6.5	7.0	33.3	28.3	26.9	0.55	0.15	—0.34	0.52

[1] Treatment *1*: Continuous heating. Treatment *2*: Continuous cooling. Treatment *3*: Alternating heating and cooling. Treatment *4*: Alternating heating and cooling with cooling 100% rh. Treatment *5*: Alternating heating and ice with ice 100% rh.

TABLE II

EFFECTS OF AMBIENT PRESSURE AND MULCHING ON EVAPORATION DURING DIURNAL TEMPERATURE CYCLES

Treatment[1]	*Cumulative evaporation (mm H_2O), time (days)*							*Maximum soil temperature after 3 days (°C) at soils depths (cm)*			*Moisture loss on the 4th day (mm H_2O), period*		*Vapor flux across the 5-cm depth on the 4th day (mm H_2O), period*	
	1	*2*	*3*	*4*	*5*	*6*	*7*	*1.1*	*3.5*	*6.5*	*heating*	*cooling*	*heating*	*cooling*
6	1.3	2.6	3.2	3.9	4.7	5.4	6.0	35.2	31.4	29.7	0.46	0.24	—0.37	0.16
7	2.3	4.3	5.5	6.7	7.8	8.8	-	34.3	29.8	28.7	0.65	0.55	—0.45	0.31
8	1.5	2.9	4.2	5.2	6.1	7.0	7.8	35.3	31.1	30.0	0.55	0.45	—0.36	0.19
9	1.2	2.4	3.6	4.5	5.4	6.2	7.1	29.2	27.9	27.3	0.46	0.44	—0.08	0.15

[1] Treatment *6*: Ambient pressure 1.53 bar. Treatment *7*: Ambient pressure 0.65 bar. Treatment *8*: Clear plastic mulch. Treatment *9*: Aluminum foil mulch.

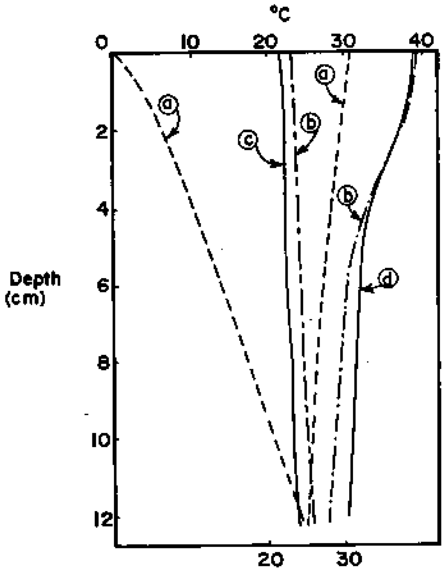

Fig.1. These curves show the typical temperature extremes in the soil profile under the various treatments. Curves *a* are the extremes reached with 8 h of ice and 12 h of heating; curves *b* are the extremes reached with 12 h of heating with the air and the light, followed by 12 h of cooling with the air near 0°C; curve *c* is the steady run with continuous cooling by air; and curve *d* is the steady-state continuous heating test.

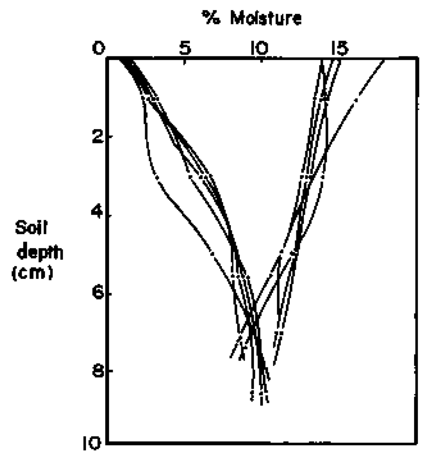

Fig.2. The initial and final soil moisture contents for the five thermal regime tests. The one low moisture curve and the one high moisture curve resulted from the continuous heat treatment.

THEORY

After the surface is dry, it is obvious that water vapor contributing to evaporation must be transported through the dry layer. This transport has been generally believed to be a molecular diffusion-type of flow, so that the instantaneous rate of drying may be given by the relation:

$$J_e = -\,\bar{D}\,\frac{P_0}{P}\,\frac{\mathrm{d}c}{\mathrm{d}z} \tag{1}$$

where J_e is the drying rate, $\bar{D}$ an effective diffusion coefficient at pressure P_0, P_0 the normal atmospheric pressure, P ambient pressure, c concentration of water vapor, and z the thickness of the dry layer, which is the distance between the drying sites in this soil and the viscous flow air currents above the soil surface. In most cases, $\mathrm{d}c/\mathrm{d}z$ will be determined on the one side of the dry layer by the vapor pressure of the air above the soil surface and on the other side by the temperature of the moisture film on the drying sites. Under conditions of high salinity, the vapor pressure of the drying sites may also be affected by osmotic stress interacting with the evaporation rate (CARY, 1965). Vapor pressures throughout the dry soil layer are, of course, affected jointly by temperature, osmotic stress and adsorptive soil forces (CARY et al., 1964). However, in this layer, the vapor-pressure gradients are controlled primarily by the drying sites on the bottom and by the air at the surface in accordance with eq.1. Consequently, the value of D at each point in the dry layer is the physical quantity which controls the absolute value of vapor pressures at each point. The moisture content, and to a lesser extent the temperature, are forced to adjust themselves accordingly in the dry layer.

The vapor pressure of water increases sharply with temperature and the relative humidity in moist soil is essentially 100%. Thus, as the surface is warmed, a vapor-pressure gradient develops from the drying sites toward the atmosphere. Because this warming creates thermal gradients, a vapor-pressure gradient also develops from the drying sites down toward the underlining cool, moist soil. Assuming there is no viscous transport of water vapor, it has previously been shown (CARY, 1966) that the amount of vapor flux in a moist, low salt soil caused by thermal gradients may be estimated by the simple relation:

$$J_v = \beta\,\frac{P_0}{P}\,(1.56 \cdot 10^{-5}\,\varphi^2 + 2.72 \cdot 10^{-3})\,\frac{\mathrm{d}\varphi}{\mathrm{d}z} \tag{2}$$

where J_v is the vapor flux in mm/h, and φ is soil temperature in degrees C. Previous measurements made on this same soil (CARY, 1964) showed β to have a value of approximately 5 at the 10–12% moisture content. Since the thermal gradients could be estimated from plots of the continuous temperature measurements, it was possible to use eq.2 to study the effect of thermally induced vapor flow on the evaporation rates observed during the nine environmental treatments. The soil was known to have a moisture content of 7.6% at 15 bar of suction (i.e., about 98% relative humidity). Consequently, eq.2 was applicable when the moisture content was above this level.

DISCUSSION

An example of the type of data produced by the experiment is shown in Fig.3. The temperature waves became more sinusoidal as they moved into the greater soil depths. It is apparent that as the soil surface dried, greater amounts of incoming energy were utilized in raising the soil temperature. Curve *c* in this figure illustrates a typical response of the drying rate to an abrupt change in ambient pressure, and the apparent reversibility of the response.

The results of the treatments have been summarized in Table I and II. The first section of these tables shows the cumulative water loss during the first 7 days of drying. Duplication of treatments suggested that the experimental error associated with the cumulative evaporation data is about $\pm$ 5% in each table. However, because the moisture content of the soil in the surface 12 cm decreased by about 2% over the 10-month period of data collection, an uncertainty of $\pm$ 10% should be considered when comparing the results between tables. The moisture contents, and consequently the evaporation rates, were relatively higher for runs reported in Table II. It should be noted that in most cases the soil surface was dry 1 day after the experiment was begun, thus the drying time was largely in the falling-rate stage. The second part of Table I and II shows the maximum temperatures achieved by each treatment at the end of the third day's evaporation. These temperatures are of particular interest because the vapor-pressure gradient in eq.1 depends on the temperature of the drying sites. The third section of the tables shows the water loss on the fourth day during the warming period and during the cooling period. Associated with these water losses are the data in the fourth section, which show the amount of thermally induced vapor flow at the 5-cm depth during the fourth day, as calculated from eq.2. The negative sign indicates a downward vapor flux.

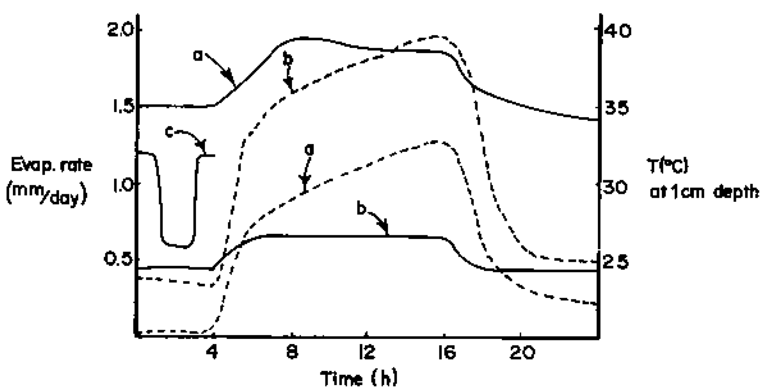

Fig.3. Typical evaporation rates and soil temperatures at the 1 cm depth during treatment *3*. Dashed curves are temperature. *a* was near the start as the surface was just becoming dry. *b* was near the end of the cycle when the surface was dry to about 4 cm. *c* is a typical evaporation-rate curve caused by changing the ambient pressure from 0,65 bar to 1.53 bar and back to 0.65.

One's first impression from the data in Table I produced by treatments *1*, *2* nd *3* is that in the falling-rate stage the evaporation rates were correlated to the net amount of energy supplied to the soil surface, as suggested by COVEY (1966). One may note, though, that during the fourth day evaporation from the continuous cooling environment was almost as great as the evaporation from the continuous warming treatment in spite of the obviously large difference in soil temperatures.[1] Inspection of the data on the thermally induced vapor flux past the 5-cm depth sheds some light on the interpretation. In the continuous heat treatment, evaporation from the drying sites was close to 2 mm/day, but about half of this was diffusing down into the deeper depths of the moist soil while the other half was diffusing upward to escape to the atmosphere. On the other hand, for the continuous cooling treatment less than 0.4 mm/day was evaporating from the drying sites, but due to the upward thermal gradient there was an evaporation and diffusion of 0.4 mm/day up toward the surface past the 5-cm depth, with the latent heat being supplied from the heat capacity of the soil. This thermally induced flow made up about half of the net soil-water loss. One would suspect from this that the drying pattern of soil should be different under continuous heating when compared to continuous cooling. In fact, the curve in Fig.2 that is obviously drier than the other four was obtained at the end of the continuous heat cycle and was a result of downward thermal water-vapor flow. It was further noted that this extremely dry surface caused a slower infiltration and redistribution of the following irrigation. This shows in Fig. 2 as the slightly different curve in the set on the right. These observations alone could make one hesitant to extrapolate isothermal theory and steady-state laboratory data to soil-drying problems in the field.

The fifth and sixth column in Table I show results of two environments which re more analogous to actual field conditions. Having the surface covered during the cooling period was comparable to having high nighttime relative humidities so that water loss was small. It is apparent that the cumulative evaporation was greater when the surface was severely cooled during the night. If one compares the cumulative water loss with respect to the time the surface was exposed for drying (e.g., after 3 days of treatment *5* and 4 days of treatment *4*) the net drying was nearly identical, even though the total energy supplied by the ice system was much less than the net energy supplied by the alternating warm and cool system. The high rate of evaporation sustained by the ice treatment was aided by the 0.2 mm/day net upward thermal vapor transfer as shown at the right of Table I.

The data in Table II summarize the results from treatments *6* through *9*. These environments were basically variations of treatment *3*. The principal effect of an increase in ambient pressure is to reduce the rate of molecular water-vapor diffusion in the pore space of the soil. While the ambient pressure was important in determining the water loss, the differences were not related by an inversely proportional

[1] Because of the design of the system, similar drying rates necessarily mean that the average vapor pressure of the air in the chamber above the soil was also similar.

constant to the ambient pressure as suggested by eq.1. Note that the maximum soil temperatures on the third day were higher when the ambient pressure was 1.53 ba This was true throughout the experiment. The explanation is, of course, that as evaporation was reduced by the lowered vapor-diffusion coefficient, more of the energy supplied to the soil surface was used in warming the soil. The increase in soil temperature raised the vapor pressure of the soil moisture and thus increased the vapor-pressure gradient toward the surface. Consequently, the rate of water loss was somewhat less than inversely proportional to the ambient pressure because of the increase in dc/dz.

The clear plastic mulch covering was not particularly effective in conserving soil moisture. Basically, a mulch may conserve soil moisture in two ways: (*1*) by increasing the resistance to vapor flow between the soil and the atmosphere; and (*2*) by reducing the rate of energy exchange between the atmosphere and the soil. While the plastic mulch may have increased the resistance to vapor transport across the soil–atmosphere interface, any beneficial effect was offset by the plastic's negative insulation properties, i.e., soil temperatures were high. On the other hand, the aluminum foil mulch cover had the same effect on transfer of vapor, but it was a much better energy reflector judging from the low soil tempertarures. In fact, temperatures near the surface were so low that there was a net upward thermal flux of water vapor. This surface cooling resulted from a latent heat loss responding to the low vapor pressures of the air entering the chamber above the soil. This dry air forced the moisture to diffuse out of the soil into the atmosphere, but since the aluminum foil was a good reflector, the soil was forced to supply some of the energy for evaporation from its heat-capacity reservoir. The thermally induced vapor-flow data also show that the insulation from the diurnal thermal wave significantly reduced the amount of cyclic vapor flux across the 5-cm depth, which could have a numbe of ramifications both in the areas of salt transfer and plant–water relationships (CARY; 1966).

Comparison of the cumulative drying under an ambient pressure of 1.53 bar with drying under the aluminum foil mulch suggests that controlling the water-vapor diffusion coefficient was more effective in reducing evaporation than was controlling the exchange of atmospheric energy with the soil's surface. It might be expected that this would be true whenever the soil is warm and the air above it is very dry—thereby producing a favorable gradient for upward vapor flux. On the other hand, when the vapor pressure in the air above the soil has some fixed minimum value, say 10 mm Hg, then controlling soil temperatures such that the moisture at the drying sites also has a vapor pressure no greater than 10 mm would stop drying, no matter what the effective vapor-diffusion coefficient was. Under this type of condition, there would be a transition point where insulation of the soil surface from the sun's energy would become a more important factor in controlling evaporation than would be the reduction of the vapor-transfer coefficient in the soil.

Table III summarizes the results of four short-term experiments in which the ambient pressure was abruptly changed for 1 or 2 h. The evaporation responded

TABLE III

FECT OF AMBIENT PRESSURE ON THE SOIL'S TEMPERATURE AND RATE OF DRYING

Soil moisture	*Ambient pressure (bar)*	*Vapor pressure of air (mm Hg)*	*Evaporation rate (mm/day)*	*Temperature of the soil (°C)*				*Ratio of evaporation (rates at high and low ambient pressure)*	
				surface	*1 cm*	*3 cm*	*10 cm*	*predicted*	*observed*
Surface moist,	1.45	9.5	1.20	20.0	19.3	20.4	21.8	0.98	0.60
cooling cycle	0.4	13.8	2.00	19.4	18.2	20.3	21.8		
Surface beginning to dry,	1.45	8.8	1.13	28.8	27.6	28.1	26.5	0.35	0.66
heating cycle	0.4	12.2	1.70	27.8	26.9	28.2	27.2		
Surface dry to 4 cm,	1.45	2.1	0.30	24.5	24.4	24.1	25.0	0.30	0.54
cooling cycle	0.4	3.8	0.55	24.2	24.1	24.0	24.5		
Surface dry to 4 cm,	1.45	3.3	0.43	–	35.7	32.5	29.8	0.30	0.48
heating cycle	0.4	5.9	0.89	–	33.0	31.3	29.5		

quickly to changes in the ambient pressure as shown in Fig.3. Equally interesting are the abrupt responses of soil temperature to the changes in ambient pressure. As the ambient pressure was reduced, the diffusion rate of water vapor increased and more heat was required for the vaporization of water; consequently soil temperatures decreased.

The temperature distribution near the end of a cooling cycle when the surface was dry to the 4-cm depth is of particular interest. Even though the air had been ntering the chamber at 0° C for nearly 10 h, the principal cooling was occurring not from the surface downward, but from the drying sites somewhere between the 3 and 10-cm depths. Because the air entering the chamber was very dry, latent heat was supplied by the soil near the drying sites. In this case, the cooling of the soil profile was as much a function of vapor-pressure gradients as it was of the energy balance at the soil-atmosphere interface. This is an observation of the same phenomenon reported by WIEGAND and TAYLOR (1962). It serves to emphasize that classical Fourier-type heat-flow solutions are not applicable to some of the detailed heat-transfer problems which arise in moist soil.

Eq.1 was utilized in the analysis of the data given in Table III. Assuming that D was independent of ambient pressure, eq.1 may be used to describe the ratio of evaporation rates resulting from changes in pressure as:

$$\frac{J_h}{J_l} = \frac{P_l \, \Delta p_h}{P_h \, \Delta p_l} \tag{3}$$

where h and l denote the high and low ambient pressures respectively. From the temperature profiles of the soil and the apparent depth of the drying layer, the vapor pressure of the drying sites was estimated. The mean vapor pressures of the air above the sample were measured continuously and are listed in the table. Ratios calculated from eq.3 are shown, along with the observed changes in evaporation rate whic resulted from the ambient pressure decrease. Except in the case of a moist soi. surface, the predicted ratios were less than the observed ratios. This suggests that the pressure dependence of moisture transfer to the surface was not as great as that predicted by eq.1. This would be true if some of the vapor transfer through the dry soil layer was by a viscous type of gas flow rather than by molecular diffusion. If one uses the pressure changes which are induced by the diurnal thermal wave to calculate viscous flow of air in soil pores in accordance with Poisseuille's law, it is difficult to believe that the vapor transfer would be by a mechanism other than diffusion. On the other hand, it may be that there were sufficient pressure changes induced either by the air flow across the surface or by large microscopic thermal discontinuities in the soil pores to cause viscous soil air flow to develop to the same order of magnitude as molecular diffusion of water vapor. This would be in accord with observations made by HANKS and WOODRUFF (1958), PEARSON et al. (1965), and by FARRELL et al. (1966), who have suggested that the movement of soil gas is affected by pressure changes associated with turbulent air flow above the surface. This effect likely damps out very quickly with increasing soil depth.

CONCLUSIONS

Because of the exponential relation between temperature and the vapor pressure of water, thermal gradients will produce significant fluxes of water vapor in a moist soil profile which is subjected to natural diurnal temperature changes. The loss of water from soil is dependent on these thermal gradients and soil temperatures. Moreover, soil temperatures are dependent on the vapor pressure of the air abov the surface as well as on the energy exchange at the soil–atmosphere interface. Because of the temperature dependence expressed in eq.2, the diurnal thermal wave will cause a net daily downward vapor flux unless the cooling conditions are severe.

The transfer of water vapor upward through the dry layer to the soil surface is not stricktly a molecular diffusion process. It appears that even low wind velocities over the surface produce a viscous soil air flow of a magnitude equal to that of molecular diffusion. This effect, plus thermal vapor transport, provides ample reason to avoid steady-state laboratory studies when extrapolation to field conditions is desired.

Loss of soil moisture may be reduced in several ways. One is to reduce the transfer coefficient of water vapor between the moist soil and the atmosphere. Another is to insulate the soil from the energy supply of the atmosphere. The most effective method depends in part on the vapor pressure of water in the air above the soil's surface. If the vapor pressure in the air is relatively high, drying might best be reduced by insulating the soil from incoming energy so that it does not warm to a point where its moisture's vapor pressure is higher than that of the air at the surface. On the other hand, if the atmosphere is very dry, evaporation control may require a reduction in the coefficient of transfer for water vapor to the soil surface. Because this transfer is in part viscous gas flow, it could prove beneficial to consider the pore radius distribution in the dry layer or surface mulch, as well as the overall porosity.

The drying of soil is basically a heat-flow problem with some of the parameters affecting the heat flow controlled by the simultaneous transport of soil moisture. Isothermal drying exists in theory only, never under practical field conditions. It appears that the isothermal soil-moisture diffusion theory cannot be expected to describe the details of soil drying under field conditions over any time interval of less than 24 h because it makes no account of the phenomena produced by the diurnal thermal wave. Because this wave has a 24-h period, there is some compensation of errors if long time intervals are considered. For this reason and because of the versatile nature of the solutions existing for this diffusion equation, it can produce curves which describe cumulative drying data over relatively long periods of time. However, such relations must be considered as semi-empirical.

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