Reprinted from SOIL SCIENCE SOCIETY OF AMERICA PROCEEDINGS
Vol. 30, No. 4, July–August 1966, pages 428–433
677 South Segoe Road, Madison, Wisconsin 53711 USA

Soil Moisture Transport due to Thermal Gradients: Practical Aspects

J. W. Cary

Purchased by the U. S. Department of Agriculture for Official Use.

Soil Moisture Transport Due to Thermal Gradients: Practical Aspects[1]

J. W. CARY[2]

ABSTRACT

The present status of knowledge concerning thermally induced transport of moisture in soil is reviewed. This includes the various theories on mechanisms of transport and experimental data showing the magnitude of moisture flow in various porous materials. Wherever possible these data were chosen to show the relative importance of thermal versus head-type flow and some general trends are noted.

Simple equations are developed to describe the thermally induced moisture flow near the soil surface which arises from transient thermal gradients produced by the diurnal temperature cycle. Calculations of the moisture flux over 2 ten-hour time intervals for a typical field situation indicate that thermal water transport should be considered whenever moisture, salt, or heat fluxes are being studied in the soil's surface layers.

The upward flow of soil water against a moisture content gradient in the winter is also considered. A sample calculation of the amount of thermally driven moisture was made using data available in the literature. The result suggested that the thermal moisture flow was too small to account for the net movement of soil moisture into the frost zone.

THE ROOT ZONE of soils under natural field conditions is subjected to continuous temperature changes. The thermal gradients produced by these temperature changes cause soil moisture to be transferred from warmer to cooler areas. The thermally induced moisture flow may significantly affect the net transfer of the soil water and nutrients by changing the moisture content gradients and the capillary conductivity, in addition to the direct effects of mass transfer. It is the intent here to indicate the relative importance of thermal moisture flow as compared to moisture flow under head gradients, to present sample calculations of the magnitude of thermal moisture flow and to suggest areas where practical use of this information can be made.

Proposed Mechanisms of Thermal Moisture Transport

Thermal moisture transport may be thought of as the moisture flux through soil which arises solely due to a temperature gradient. It is the flow of moisture which would occur even in the absence of all moisture content, osmotic, electrical and pressure gradients. Thermal gradients and the associated moisture transfer will, however, create changes in moisture contents, pressures, etc., the effects of which must be accounted for in the final analysis of net moisture flow.

Moisture flows through soil from warmer to cooler areas in both the vapor and the liquid phases. Flow in the vapor phase is thought to be primarily a molecular diffusion process. When the vapor pressure *gradient* in the soil is determined by the temperature only, and not by osmotic or moisture content changes, the vapor transport may be described by the relation

$$J_{\text{vap}} = -\beta \frac{DpH}{R^2T^3}\frac{dT}{dz} \qquad [1]$$

where the meanings of the symbols are defined in the appendix. Equation [1] may be derived from the theory of thermodynamics of irreversible process (3) or by combining Fick's first law and the Clausius-Clapeyron equation (2). As the soil's gas phase expands and contracts under the

[1] Contribution from the Soil and Water Conservation Research Division, ARS, USDA, in cooperation with the Dep. of Water Science & Engineering, University of California, Davis. Presented at the annual meeting of the Soil Science Society of America, Columbus, Ohio, Nov. 1, 1965, as part of the Symposium on "Practical Applications of Recent Advances in Knowledge of Flow of Water in Soils." Received Nov. 16, 1965. Approved April 6, 1966.

[2] Soil Scientist, USDA, Snake River Conservation Research Center, Kimberly (Twin Falls), Idaho.

Table 1—Comparisons of moisture flow due to thermal gradients and due to head gradients in various porous materials. All thermal moisture flow is from warm to cool. Other similar data may be found in references 5, 6, 11, 12, and 14

Type of porous material	Moisture content or suction in bars	Thermal flow, mm H_2O/day per °C/cm	cm H_2O head equiv. to 1C gradient	Mean temperature	Type of solution flowing	Phase of flow	Source of data
Sintered glass, 10⁴A mean pore diameter	saturated	0	0	25	distilled water	liquid	13
Du Pont 600 cellophane	saturated	0.1*	—	40	double distilled water	liquid	17
Cellulose acetate 27A mean pore diameter	saturated	—	7.0	20	distilled water	liquid	7
Millipore filter 100A mean pore diameter	saturated	—	0.09	20	distilled water	liquid	7
Wyoming bentonite paste 18%	saturated	0.1†	4.0†	35	0.1N NaCl	liquid	‡
Yolo loam soil $\rho_B = 1.4$	0.06	2.0	5.5	35	0.01N $CaSO_4$	liquid + vapor	‡
Houston Black Clay	0.66	0.7	—	38	distilled water	liquid + vapor	‡
Houston Black Clay	1.4	0.8	—	37	distilled water	liquid + vapor	‡
Millville silt loam $\rho_B = 1.5$	saturated	—	0.07	25	water	liquid	20
Millville silt loam	2.2	—	1.4×10^3	25	water	vapor + liquid	19
Millville silt loam	126	—	8.3×10^4	25	water	vapor	19
Columbia loam $\rho_B = 1.2$	0.07	1.8	3	19	distilled water	20% vap. 80% liq.	2
Columbia loam $\rho_B = 1.2$	0.24	0.9	13	8	distilled water	33% vap. 67% liq.	2
Columbia loam $\rho_B = 1.2$	0.24	2.0	14	33	distilled water	55% vap. 45% liq.	2
Columbia loam $\rho_B = 1.2$	0.45	1.6	250	25	distilled water	vapor + liquid	2
Stable air	100% porosity	0.4	8×10^4	27	distilled water	vapor	4

* Cross sectional area estimated as 4 cm² and membrane thickness at 0.1 cm.
† Results changed with aging.
‡ Previously unpublished data obtained by the author in experiments similar to those reported in (2).

influence of the diurnal thermal wave passing through the root zone, some mass flow of vapor is initiated. However, any moisture transport resulting from mass air flow is generally either neglected or inadvertently included in the experimental evaluation of β.

The flow of moisture in thin films or through small capillaries under the influence of a thermal gradient is not well understood. In general the flow occurs from warm to cool; however, there have been some reports of flow from cool to warm, primarily associated with organic-type membranes (8, 13). There are at least four possible reasons why water flows in the liquid phase under the influence of a thermal gradient. Since the surface tension of water against air increases as the temperature drops, moisture in unsaturated soil could flow from warm to cool under the influence of a surface tension gradient. Soil moisture suction also increases as the temperature drops, which could contribute to the moisture flow. This is the basis of the thermally induced liquid flow equation developed by Philip and de Vries (16). Deryaguin and Melnikova (8) also developed an analysis based on surface tension gradient. However, as a second possibility they felt there might be some flow from cool to warm resulting from the difference in specific heat content between the liquid layer adsorbed on the solid surface and the specific heat content of the bulk of the liquid in the pores. Since moisture flows from warm to cool in systems with no air-water interface, there must be mechanisms of flow other than surface tension gradients. This has led to a proposal that transfer results from a net motion generated by random kinetic energy changes associated with the hydrogen bond distribution which develops under a thermal gradient (2). A fourth possibility is that the flow results from thermally induced osmotic gradients. Most dissolved salts will spontaneously diffuse through a solution from warmer areas into cooler areas (Soret effect). For example, suppose a soil solution with 50 meq/liter of salt is subjected to a thermal gradient of 1C/cm. If the Soret coefficient was an 2×10^{-3} C^{-1}, the solution could eventually develop a steady-state concentration gradient of 0.25 meq/liter cm^{-1} (15, 18). If the thermal field held this osmotic gradient so that it acted as a perfect semipermeable membrane, the equivalent pressure gradient on the soil moisture would be about 10 cm of H_2O cm^{-1} which compares with some of the observed values listed in Table 1. There is a possibility that this type of phenomena may have been observed in fritted glass plates (13).

In view of the uncertainties concerning the mechanism of thermally induced liquid phase flow, one must resort to a phenomenological equation to describe this component of flow, i.e.,

$$J_l = -\frac{KQ}{aT}\frac{dT}{dz} \qquad [2]$$

(2) where the symbols are defined in the appendix. While equations [1] and [2] were developed primarily from the general theory of irreversible thermodynamics, Taylor and Cary (21) have shown that these equations are identical in form to the equations developed by Philip and de Vries (16). In fact, most investigators have used similar relations.

Relative Moisture Flows Arising from Thermal and Head Gradients

Quantitative data on the magnitude of thermally induced flow are still inadequate. However, a few results from various sources are tabulated in Table 1. From these data one can see that a significant flow of water may occur through small pores from warm to cool, even at saturation. Compared to the flux arising from head gradients, this flow becomes relatively more important as the hydraulic conductivity decreases. In unsaturated soils the relative importance of thermally induced flow rises rapidly as the moisture content decreases. The decrease in moisture content is accompanied by a decrease in the thermal liquid moisture flow and by

an increase in the thermal vapor moisture flow. At any given moisture content the thermal vapor phase moisture flux increases faster with rising mean temperatures than does the liquid phase flux because of the exponential dependence of vapor pressure on temperature.

Application of the Flux Equations

Temperature changes in the soil have been conveniently classified as annual and diurnal. Annual changes in the average soil temperature may be significant to depths far beneath the ordinary root zone. Daily temperature changes are commonly significant to a depth of 20 or 30 cm. For example, some measurements made by Carson and Moses (1) at Argonne, Illionis, are summarized in Table 2. This type of data suggests that there are also two types of thermally induced moisture flows in soil which should be considered, i. e., moisture flow near the soil surface which may change direction every 12 hours and the longterm thermal moisture flow deep within the soil profile which changes direction only once every 6 months.

Analysis of the diurnally induced moisture transfer requires some knowledge of time-dependent thermal gradients under the soil's surface and the numerical values of the thermal water flow coefficients. Typical temperature waves at 13- and 15-cm depths of one soil are shown in Fig. 1. The thermal gradient at 14 cm may be estimated at any given time by dividing the temperature difference between the two curves by 2 cm. The amount of liquid or vapor transfer may be estimated by dividing the curves up into short enough time intervals so that average values of dT/dz, D, p, H, and T may be used for steady-state solutions of equations [1] and [2]. For example, in Fig. 1 between 11 AM and noon the average value of dT/dz was 0.43 and the average value of T was 295.2. However, between noon and 1 PM $dT/dz \sim 0.50$, $T \sim 296.5$ and D, p, K, and H would have also increased as a result of their temperature dependences.

When detailed soil temperatures are not available or when less accurate estimates are satisfactory, the following method may be used. One may note that the temperature variations shown in Fig. 1 are approximately sine curves. As pointed out by van Wijk and de Vries (22), this is generally the case. Suppose, therefore, the temperature at any depth in the soil may be described by

$$\phi = (b - cz) + \phi_0 \exp - (z/d) \sin (\omega t - z/d) \qquad [3]$$

where the symbols are described in the appendix. Then $d\phi/dz$ (i. e. dT/dz) can be written as $f(t)$ where the function may be evaluated from equation [3]. Moreover, plotting DpH/R^2T^3 against ϕ, one finds a quadratic relation so that equation [1] may be written as

$$J_v = -\beta\,(1.56 \times 10^{-5}\,\phi^2 + 2.72 \times 10^{-3})(d\phi/dz) \qquad [4]$$

or

$$J_v = -\beta\, g(t)\, f(t) \qquad [5]$$

where $g(t)$ may be evaluated from equations [3] and [4]. Since J_v = mm H_2O/hour or dy_v/dt, equation [5] may be written as

$$y_v = -\beta \int_{t_1}^{t_2} g(t)\, f(t)\, dt \qquad [6]$$

taking β independent of time where y_v is the vapor flow in mm of water.

Integration of equation [6] shows the quantity of water vapor (in mm of liquid) which will move across the soil plane of depth z, between times t_1 and t_2 will be

$$y_v = 1.56 \times 10^{-5} \frac{\beta}{\omega} [Ax - B \cos x - C \sin x \cos x + E \sin x + F \sin^2 x]_{t_1}^{t_2} \qquad [7]$$

where the coeficients are defined in the appendix and the third power term has been neglected.

In a similar manner, one finds that plotting K/aT against ϕ leads from equation [2] to

$$J_l = -\overline{KQ}\,(2.65\,\phi + 77)(d\phi/dz) \qquad [8]$$

where the capillary conductivity has been assumed to be inversely proportional to the viscosity of water. The thermally induced liquid phase flow under the diurnal thermal wave then becomes

$$y_l = \frac{\overline{KQ}}{\omega} [Gx - I \cos x - M \sin x \cos x + N \sin x + M \sin^2 x]_{t_1}^{t_2} \qquad [9]$$

where the coefficients are defined in the appendix.

Equations [4] and [8] may be used under steady-state conditions. The inclusion of the empirical temperature

Table 2—Soil temperature changes at Argonne, Illinois (1)

Depth	Mean annual variation in temperature °C	Maximum daily variation in temperature °C
1 cm	25	12
10 cm	24	9
20 cm	23	3
50 cm	22	0.5
10 feet	8	0
29 feet	1	0

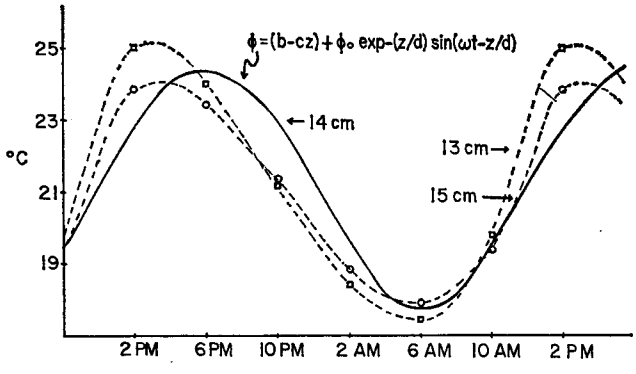

Fig. 1—Soil temperatures at 13- and 15-cm depths in July under a grass cover as replotted from the data of Carson and Moses (1). The smooth curve is the temperature at 14 cm predicted by equation [3] with the surface amplitude taken as 6.8C, average surface temperature 22.8C, damping depth 20 cm and c = 0.13. Note the relative displacement of the measured temperature curves both horizontally and vertically. These changes, in addition to the decreasing amplitude, markedly affect the thermal gradient.

functions do not cause errors greater than 5% so long as $0C < \phi < 40C$. Other errors may arise, if the vapor pressure *gradient* is affected by the soil moisture content (i.e. the moisture tension at the depth z, must be in the plant growth range); or if it is affected by osmotic gradients (i.e. the soil solution must not be highly saline). It is also assumed that $\overline{KQ}$ and β are independent of time at depth z, and that the temperature dependence of K is inversely proportional to that of the viscosity of water.

Equations [1], [4], and [7] assume an ambient pressure of 1.0 bar. If they are to be used at high altitudes the results may be corrected by multiplying by the ratio of sea level pressure divided by atmospheric pressure (3).

Equations [7] and [9] may be used to calculate the thermal vapor and liquid moisture flux due to the transient temperature gradients occurring diurnally near the soil surface. These relations require the additional assumption that the soil's temperatures be described by equation [3]. While this assumption is not completely fulfilled under most natural conditions (22), equations [7] and [9] are still useful for making practical estimates of moisture flow in the absence of detailed soil temperature data. As an example, equation [3] was used to predict the temperature at the 14-cm depth of the soil as shown in Fig. 1. Equation [7] then predicted that the downward vapor flux across the 14-cm depth between 9:30 AM and 8:30 PM was 0.033 β mm H_2O. On the other hand, dividing the measured temperature curves in Fig. 1 into 10 time intervals and using average hourly values in equation [1] the vapor flow was found to be 0.045 β mm H_2O.

A Practical Example and Its Implications

Some insight into the significance of thermally induced moisture flow may be gained by considering an example. Suppose that we have a loam soil with the surface 1 or 2 cm quite dry but grading into moist conditions with a ½ bar suction at 4 cm, and into a ⅓ bar suction at the 30-cm depth. Suppose also that atmospheric conditions are such that the average surface temperature is 26C, but changes sinusoidally with an amplitude of 9C. The soil might likely have a damping depth of 13 cm and an average temperature at the 10-cm depth of 25 C (which would mean $c = 0.1$). Reasonable values at the 4- to 10-cm depth would be $\beta = 5$, $\overline{K} = 5 \times 10^{-4}$ mm/hr and $\overline{Q} = 8 \times 10^{-2}$ cal/g.[3] Using these numbers, equations [7] and [9] provide the calculated thermal moisture flows shown in Table 3. The flow due to the suction gradient was estimated by a steady-state calculation

$$J_h = -K\frac{dh}{dz} = -(5 \times 10^{-4})\left(-\frac{200}{25}\right) \qquad [10]$$

assuming that the suction changed linearly from ⅓ bar to ½ bar between the 30-cm and the 5-cm depth.

The results listed in Table 3 have several interesting implications. Consider for example the drying of a bare soil.

[3] Some experimental values of β have been given in reference (2) and others may be easily calculated from data given in references (5, 12, 16) and equation [4]. Values of Q may be found in reference (2). The product KQ for several saturated systems may be obtained by using equation [8] and portions of the data summarized in Table 1.

Table 3—Vapor and liquid moisture flux arising from a diurnal temperature cycle under one typical set of conditions

Time period, hours	Soil depth, z_1 cm	Thermally induced flow of moisture through the plane of z_1
Warming period $t = 0$ to $t = 10$	4	0.52 mm downward (vapor phase)
Warming period $t = 0$ to $t = 10$		0.05 mm downward (liquid phase)
Cooling period $t = 12$ to $t = 22$		0.22 mm upward (vapor phase)
Warming period $t = 0$ to $t = 10$	8	0.36 mm downward (vapor phase)
24-hour period	8	0.1 mm upward due to moisture suction gradient (liquid phase, see equat on 10)

Under the conditions specified the daily evaporation of soil moisture would be in the neighborhood of 1.25 mm of H_2O. At the 4-cm depth, a downward net daily vapor flux of 0.3 mm occurred. This means that of the total water vaporized in the drying zone at about the 3-cm depth only 70 or 80% escaped into the atmosphere, while the 0.5-mm downward vapor flux carried with it 30 cal. cm^{-2} of heat. It is also apparent that about one-third of the downward vapor flow from the surface recondensed before it reached the 8-cm depth. Thus much of the downward vapor flux could be returned to the surface by capillary flow and consequently would act as a salt pump serving to move plant nutrients from the root zone toward the surface. If the soil solution contained 50 meq of salts/liter and no salt sieving occurred, then the thermally induced vapor flow could be responsible for moving up to 175 meq of salt/m^2 into the soil's surface layer over a 1-week interval. This would be in addition to that which moves up as moisture and is lost to the atmosphere (9). The ability of a thermal gradient to indirectly cause the transfer of significant amounts of soluble salts from cool soil regions into warm regions has been well demonstrated (11). This type of transport may be much greater than the Soret type thermal salt diffusion mentioned previously.

Table 3 shows that the upward flow of soil moisture past the 4-cm depth was about 0.22 mm per night or 220 ml of water/m^2. This moisture, if totally recondensed in the soil above 4 cm, would cause an increase in moisture content of < 0.5% in a soil where the $\rho_B = 1.2$. This appears to be a negligibly small amount so far as rewetting the soil surface or supplying moisture for evapotranspiration is concerned. However, the vapor will not necessarily recondense uniuniformly throughout the soil. A good portion of it might condense directly on plant roots since they provide a good heat conduction path to the cooler nighttime atmosphere. While a few milliliters of water gained by the plant in this manner is not significant in transpiration it could be a factor in the re-establishment of the plant turgidity during the night and also contribute to the guttation phenomenon sometimes observed in the early morning. Measurements of plant root temperatures relative to soil particle temperatures are needed to ascertain the importance of such a phenomenon.

Another interesting aspect of the calculations shown in Table 3 is that the net thermal moisture flux is downward, which is the general case during the spring and summer months when the soil as a whole is warming. This does not mean that the upward flow of vapor during the night is

not important. Since the drying of soil develops as a dry front moving into the soil[4] it follows that much of the vapor moving downward in the daytime will originate from a plane of drying sites because this is where the energy from the atmosphere meets the soil moisture and so becomes available for conversion to latent heat. However, vapor moved upward during the night period will not necessarily recondense entirely on these drying sites. If the vapor pressure of the night air above the surface is high, the upward flowing vapor will recondense throughout the surface layers depending on their moisture content, osmotic pressure, temperature, and ability to dissipate the 580 cal/g of latent heat liberated by the condensation. Even though this upward flux of vapor is small, it may condense in such a way as to replenish critical regions of the soil moisture film and so increase the capillary conductivity. This could result in a significant increase in liquid phase capillary flow toward the surface in the early morning hours. The atmosphere's relative humidity during the cooling period could interact similarly in the re-establishment of soil moisture film contact with plant roots.

The rate of vapor transfer is, of course, directly proportional to the diffusion coefficient of water vapor. If the soil contained a dense or compacted layer at the 8-cm depth the value of β at 8 cm would have been reduced accordingly. If, furthermore, this layer did not change the temperature patterns in the profile the downward vapor flux at 8 cm would have been smaller than 0.36 mm, and consequently more vapor would have been forced to condense between the 4- and 8-cm depths. This type of phenomena has been observed qualitatively in laboratory experiments conducted by Matthes and Bowen (14). This could prove to have interesting implications in seedbed preparation and in studies of the salt distribution in soil profiles in arid regions.

Moisture Flow into the Freezing Zone in the Winter

One of the most striking soil moisture flow phenomenon, which develops concurrently with a significant thermal gradient, occurs during the winter months as moisture moves up into the frost zone. For example, in one case observed by Willis et al. (23) the water table dropped more than 1 m during the winter because of the flow of moisture up into the frost zone in the surface layers. This phenomenon is doubly important because it may be associated with "frost heaving" and significant damage to small plants. Ferguson et al. (10) have also shown that water moves upward *against* its own moisture content gradient during the winter months. Their data suggest that on one plot, which was moist to a depth of 94 inches in the fall, the soil between the 30-inch and 72-inch depth lost about 25 mm of water to the frost zone between 12-20-62 and 3-20-63. The average thermal gradient during that period was 0.033C cm^{-1} at the 30-inch depth. Based on the assumption that the soil had an average $\overline{K}$ = 0.2 mm/day and $\overline{Q}$ = 0.04 cal/g, equation [8] predicts that there should have been about 2 mm H_2O thermally induced to flow upward in the liquid phase during the 3-month period. Taking β = 5, the vapor flow should have been about 1 mm of H_2O. In spit of the uncertainties involved in choosing average values of β, $\overline{K}$, and $\overline{Q}$, the predicted net flux of 3 mm of H_2O is obviously too small to account for the upward flow of moisture. Further investigation of the transport mechanisms associated with the soil-ice interfaces are needed.

[4] H. R. Gardner and R. J. Hanks. Evaluation of the evaporation zone by measurement of heat flux in soil. Presented at the Western Society of Soil Science meeting, Riverside, Calif., June 23, 1965.

APPENDIX

Definitions of symbols used in the text

$A = c(b - cz)^2 + 174c + \frac{1}{2}c\,[\phi_0 \exp - (z/d)]^2 + \frac{1}{d}(b - cz)[\phi_0 \exp - (z/d)]^2$

$B = \phi_0 \exp - (z/d)[2c(b - cz) + 174\,\frac{1}{d} + \frac{1}{d}(\phi_0 \exp - (z/d))^2 + \frac{1}{d}(b - cz)^2]$

$C = [\phi_0 \exp - (z/d)]^2\left[\frac{1}{2}c + \frac{1}{d}(b - cz)\right]$

D: diffusion coefficient of water vapor in air, $cm^2\ sec^{-1}$

$E = \left[\frac{1}{d}\phi_0 \exp - (z/d)\right][(b - cz)^2 + 174]$

$F = (b - cz)\frac{1}{d}[\phi_0 \exp - (z/d)]^2$

$G = 2.65c\,(b - cz) + 77c + 1.33\frac{1}{d}[\phi_0 \exp - (z/d)]$

H: heat of vaporization of water, cal. $mole^{-1}$

$I = \phi_0 \exp - (z/d)[2.65\frac{1}{d}(b - cz) + 2.65\frac{1}{d} + 77c]$

J_{vap}: thermally induced flow of vapor, moles sec^{-1} cm^{-2}

J_v: thermally induced flow of vapor, mm hr^{-1}

J_l: thermally induced liquid phase of flow of moisture, mm hr^{-1}

J_h: head induced liquid phase flow of moisture, mm hr^{-1}

K: capillary conductivity, mm hr^{-1}

$\overline{K}$: capillary conductivity, at 25C, mm hr^{-1}

$M = 1.33\frac{1}{d}[\phi_0 \exp - (z/d)]^2$

$N = 2.65\frac{1}{d}(b - cz)\,\phi_0 \exp - (z/d) + 77$

Q: liquid phase heat of transport, ergs g^{-1} (*see* reference 2)

$\overline{Q}$: liquid phase heat of transport, cal g^{-1} (*see* reference 2)

R: gas constant, cal. $°K^{-1}$ $mole^{-1}$

T: temperature, degrees Kelvin

a: acceleration of gravity, cm sec^{-2}

b: average soil temperature at the surface, °C

c: coefficient to account for changing mean temperature with depth, °C cm^{-1}

d: damping depth of the soil, cm (*see* reference 22)

h: equivalent hydraulic head, cm of water

p: vapor pressure of water, cal. cm^{-3}

t: time hours

x = $(\omega t - z/d)$, radians
y: mm of water flow
z: soil depth, cm
β: a factor accounting for pore geometry, cross sectional area of voids and microscopic temperature discontinuities, dimensionless (*see* reference 2)
ϕ: temperature, °C
ϕ_0: temperature amplitude at the soil surface, °C
ρ_B: soil bulk density, g cm^{-3}
ω: radial frequency, equals 0.262 hr^{-1}

LITERATURE CITED

1. Carson, J. E., and H. Moses. 1963. The annual and diurnal heat-exchange cycles in upper layers of soil. J. App. Metrol. 2:397–406.
2. Cary, J. W. 1965. Water flux in moist soil: Thermal versus suction gradients. Soil Sci. 100:168–175.
3. Cary, J. W. 1964. An evaporation experiment and its irreversible thermodynamics. Int. J. Heat Mass Transfer. 7:531–538.
4. Cary, J. W. 1963. Onsager's relation and the non-isothermal diffusion of water vapor. J. Phys. Chem. 67:126–129.
5. Cary, J. W., and S. A. Taylor. 1962. The interactions of the simultaneous diffusion of heat and water vapor. Soil Sci. Soc. Amer. Proc. 26:413–416.
6. Cary, J. W., and S. A. Taylor. 1962. Thermally driven liquid and vapor phase transfer of water and energy in soils. Soil Sci. Soc. Amer. Proc. 26:417–420.
7. Corey, A. T., and W. D. Kemper. 1961. Concept of total potential in water and its limitations. Soil Sci. 91:299–302.
8. Deryaguin, B. V., and M. K. Melnikova. 1956. Experimental study of the migration of water through the soil under the influence of salt concentration, temperatue and moisture-gradients. Int. Congr. Soil Sci., Trans. 6th (Paris) p. 305–314.
9. Doering, E. J., R. C. Reeve, and K. R. Stockinger. 1964. Salt accumulation and salt distribution as an indicator of evaporation from fallow soils. Soil Sci. 97:312–319.
10. Ferguson, H., P. L. Brown, and D. D. Dickey. 1964. Water movement and loss under frozen soil conditions. Soil Sci. Soc. Amer. Proc. 28:700–703.
11. Gurr, C. G., T. J. Marshall, and J. T. Hutton. 1952. Movement of water in soil due to a temperature gradient. Soil Sci. 74:335–345.
12. Hanks, R. J. 1958. Water vapor transfer in dry soil. Soil Sci. Soc. Amer. Proc. 22:372–374.
13. Hutchison, H. P., I. S. Nixson, and K. G. Denbigh. 1948. The thermo-osmosis of liquids through porous materials. Discuss. Faraday Soc. no. 3. p. 86–94.
14. Matthes, R. K., and H. D. Bowen. 1963. Water vapor transfer in soil by thermal gradients and its control. Trans. Amer. Soc. Agr. Eng. 6:244–248.
15. Peyton, A. D., and J. C. R. Turner. 1962. Soret coefficients and heats of transport of some salts of alkaline earth metals in water at 25C. Faraday Soc. Trans. 58:55–59.
16. Philip, J. R., and D. A. de Vries. 1957. Moisture movement in porous materials under temperature gradients. Amer. Geophys. Union Trans. 38:222–232.
17. Rastogi, R. P., R. L. Blokhra, and R. K. Agarwal. 1964. Cross-phenomenological coefficients. Faraday Soc. Trans. 60:1386–1390.
18. Sagert, N. H., and W. G. Breck. 1961. Thermal diffusion of ammonium salts in aqueous solutions. Farraday Soc. Trans. 57:436–446.
19. Taylor, S. A. 1962. Influence of temperature upon the transfer of water in soil systems. Mededelingers, Landbouwhogeschool, Gent. Belgium. 27:535–551.
20. Taylor, S. A., and J. W. Cary. 1960. Analysis of the simultaneous flow of water and heat or electricity with the thermodynamics of irreversible processes. Int. Congr. Soil Sci., Trans. 7th (Madison, Wis.) I:80–90.
21. Taylor, S. A., and J. W. Cary. 1964. Linear equations for the simultaneous flow of matter and energy in a continuous soil system. Soil Sci. Soc. Amer. Proc. 28:167–172.
22. Van Wijk, W. R., and D. A. deVries. 1963. Periodic temperature variations in a homogeneous soil. *In* W. R. Van Wijk [ed.] Physics of Plant Environment. John Wiley & Sons, New York.
23. Willis, W. O., H. L. Parkinson, C. W. Carlson, and H. J. Haas. 1964. Water table changes and soil moisture loss under frozen conditions. Soil Sci. 98:244–248.