

# On water vapor transport in field soils

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**Abstract.** Measurements of soil volumetric moisture content and temperature were made at 2, 4, 7, 10, and 15 cm below the surface of a bare field soil, over a 1-week period at 20-min intervals. The conductive heat and liquid moisture fluxes were calculated for the soil layer 7–10 cm below the surface, and the water vapor flux was then determined from both the energy transfer and mass transfer equations. Water vapor flux in this layer transported a significant amount of the total energy flux (up to 50%) and an appreciable amount of the total moisture flux (up to 25%). There was reasonable agreement between the water vapor flux calculated by the mass transfer equation and the vapor flux calculated by the energy equation.

## 1. Introduction

For at least 80 years it has been recognized that the movement of moisture and heat in the soil are coupled [Bouyoucos, 1915]. The total heat flux in the soil occurs not only from simple conduction but also from water movement in both the vapor and liquid states. Likewise, temperature gradients can drive mass transfer. Conceptually, the coupling of the heat and mass transfer equations can be seen as largely resulting from the water vapor flux. The movement of moisture from one location in the soil to another by evaporation and the subsequent recondensation can contribute significantly to the net moisture movement in the soil. Additionally, because of the large value of the latent energy of vaporization of water, the water vapor transports significant energy when it evaporates and condenses.

Various authors have examined the significance and magnitude of the water vapor flux as it affects either the mass or energy balances in experimental studies (see Table 1 for summary results). One of the first field-scale tests of the coupled effects of soil heat and moisture transport was done by Rose [1968a, b]. Rose was interested in water vapor transport driven by temperature gradients and looked only at the mass balance. In his equation for the conservation of mass the only unknown term was the thermally driven vapor flux. Using the measured values of the other fluxes and the net change in moisture content, Rose solved for the thermal vapor flux, and found that the amount of water transported through a soil layer as vapor was on the same order of magnitude as the increase or decrease of the volumetric moisture content in that layer. Another study that looked at role of vapor flux in the soil moisture balance was done by Jackson and coworkers [Jackson, 1973; Jackson *et al.*, 1974]; they performed a similar experiment of combined measurements of soil temperature and volumetric moisture content in a field soil. Their calculated water vapor fluxes were on the same order as Rose's [1968b]. Monji *et al.* [1990] also found large values of water vapor transport due to temperature gradients.

A study which looked at the effect of vapor flux on the energy balance was that of Westcot and Wierenga [1974]. In their combined modified-field experiment and computer

model they calculated that heat transported by vapor flux was on the same order as heat flux by conduction and accounted for 40–60% of total heat flux in the top 2 cm of the soil and up to 20–25% of total heat flux at a depth of 25 cm. When the heat transport by vapor flow was not included, the soil temperature was underestimated at the middle of the day.

In the present study we examine the transport of water in a bare field soil (Yolo silt loam) using subsurface measurements of soil temperature and volumetric moisture content. Thermal conductivity and liquid water diffusivity are both calculated from well-established previous results for the soil used in the experiment. We show that there exists for these field experimental conditions a significant amount (40–60%) of heat flux due to vapor transport. The contribution of the water vapor flux to the total moisture flux is less (10–30%) but still significant. Unlike previous studies, which have looked at either mass or energy transport, we compare the vapor flux computed from the residual of the energy equation to the vapor flux computed from the mass equation. Reasonable agreement between the two calculated time series of water vapor flux is found.

## 2. Theoretical Considerations

### 2.1. Energy Equation

When the heat transported by mass transfer is considered, Fourier's law for the heat flux density in soil is amended to become

$$\mathbf{q}_h = -\lambda \nabla T + \rho L \mathbf{q}_v + c_{pL} \rho (T - T_0) \mathbf{q}_m \quad (1)$$

where  $\mathbf{q}_h$  is the heat flux density [ $\text{W/m}^2$ ],  $\lambda$  is the thermal conductivity [ $\text{W/m K}$ ],  $T$  is the temperature [ $\text{K}$ ],  $\rho$  is the density of water [ $\text{kg/m}^3$ ],  $L$  is the latent heat of vaporization [ $\sim 2.45 \times 10^6 \text{ J/kg}$ ],  $c_{pL}$  is the specific heat of liquid water [4182  $\text{J/kg K}$ ],  $T_0$  is an arbitrary reference temperature [ $\text{K}$ ], which was taken to be 25°C,  $\mathbf{q}_v$  is the water vapor flux, and  $\mathbf{q}_m$  is the total moisture flux [both in units of  $\text{m/s}$ ], which is simply equal to the sum of the vapor flux  $\mathbf{q}_v$  and the liquid water flux  $\mathbf{q}_l$  [de Vries, 1958]. The right-hand side of (1) consists of a Fourier conduction term ( $-\lambda \nabla T$ ) and two terms that express heat transported by mass transfer. The first,  $\rho L \mathbf{q}_v$ , is the heat transported by the evaporation of water vapor in one place and its recondensation at another. The second term,  $c_{pL} \rho (T - T_0) \mathbf{q}_m$ , is the transport of heat energy by the flow of water of one

**Table 1.** Previous Studies of the Magnitude of Water Vapor Flux, Either in the Moisture Balance or the Energy Balance

Reference	Maximum Magnitude of Heat Change Due to Vapor Flow, W/m <sup>2</sup>	Maximum Magnitude of Moisture Change Due to Vapor Flow, cm/s	Depth of Observation, cm
Cary [1965]	...	$2 \times 10^{-6}$	column
Rose [1968b]	...	2 to $6 \times 10^{-6}$ , 2 to $4 \times 10^{-6}$	1-3 3-12
Jackson <i>et al.</i> [1974]	...	2 to $7 \times 10^{-6}$	0.5, 1
Westcot and Wierenga [1974]	80 70 20 10	...	0.95 1.5 5 11
This study	40 to 60	$7 \times 10^{-6}$	7-10

temperature to a region of a different temperature. The reference temperature  $T_0$  is needed because energy is not an absolute quantity but rather is described as the difference from some base level set at zero energy.

Equation (1) can be used to solve for the water vapor flux if the total heat flux density  $\mathbf{q}_h$ , the conductive flux density  $\lambda \nabla T$ , and the liquid water flux  $\mathbf{q}_l$  are known. Note that certain terms in (1) are more important than others. For our experimental conditions the net heat flux density and the conductive flux density have magnitudes on the order of  $10^1$  W/m<sup>2</sup>, while the term describing heat energy transported by liquid flow ( $c_{pl} \rho \nabla T \mathbf{q}_m$ ) is on the order of  $10^{-1}$  W/m<sup>2</sup>. It is apparent, then, that in order to use the heat equation to determine the vapor flux, an accurate estimation of the thermal conductivity must be available.

Unlike heat capacity, which can be calculated as the sum of the heat capacities of the different soil components (water, air, quartz, and other solids) weighted by their respective fraction of the soil, the thermal conductivity of soils is a function of the geometrical arrangement of the phases in the soil matrix. Estimation of the thermal conductivity of soils with varying moisture content has often been done with the method proposed by *de Vries* [1963]. This relationship is

$$\lambda = \frac{\sum_{i=0}^n k_i X_i \lambda_i}{\sum_{i=0}^n k_i X_i} \quad (2)$$

where  $X_i$  is the volume fraction of the  $i$ th phase (quartz, other solids, organics, water, or air) and  $\lambda_i$  is the thermal conductivity of the  $i$ th phase. The subscript 0 is reserved for the continuous phase, which is considered to be water for a "wet" soil and air for a "dry" one. The value of  $k_0$  is defined to be 1, and the value of  $k_i$  for  $i$  greater than 0 is given by

$$k_i = \frac{1}{3} \sum_{j=1}^3 \left[ 1 + \left( \frac{\lambda_i}{\lambda_0} - 1 \right) g_j \right]^{-1} \quad (3)$$

where  $g_j$  is a shape factor, with  $g_1 + g_2 + g_3 = 1$ . *Wierenga et al.* [1969] derived soil-specific formulae for the shape factors  $g_j$  for application of the *de Vries* formula for thermal conductivity for the Yolo silt loam; these values are used herein.

## 2.2. Mass Equation

The total moisture flux is given by

$$\mathbf{q}_m = \mathbf{q}_l + \mathbf{q}_v \quad (4)$$

where the liquid water flux is defined as

$$\mathbf{q}_l = -D_{ol} \nabla \theta - D_{Tl} \nabla T - K \quad (5)$$

where  $\theta$  is the volumetric moisture content,  $D_{ol}$  is the isothermal liquid diffusivity [m<sup>2</sup>/s],  $D_{Tl}$  is the thermal liquid diffusivity [m<sup>2</sup>/s K] and  $K$  is the unsaturated hydraulic conductivity [m/s] [*Philip and de Vries*, 1957]. The isothermal liquid diffusivity is simply the diffusivity which appears in Richards' equation:

$$D_{ol} = K \frac{\partial \psi}{\partial \theta} \quad (6)$$

where  $\psi$  is the matric potential [m]. The thermal liquid diffusivity arises from the flux because of changes in surface tension driven by changes in temperature:

$$D_{Tl} = K \gamma \psi \quad (7)$$

where  $\gamma$  is the relative change in surface tension  $\sigma$  with respect to temperature:

$$\gamma = \frac{1}{\sigma} \frac{d\sigma}{dT}$$

The functional dependency of  $K$ ,  $\psi$ , and  $\partial\psi/\partial\theta$  on  $\theta$  is defined with a van Genuchten-type equation for the  $\psi(\theta)$  relationship, while for hydraulic conductivity the Brooks and Corey relationship for  $K(\theta)$  is used. These are

$$\frac{\theta - \theta_r}{\theta_s - \theta_r} = \left[ \frac{1}{1 + (\alpha\psi)^n} \right]^m \quad (8)$$

$$K = K_{sat} \left( \frac{\theta - \theta_r}{\theta_s - \theta_r} \right)^{\eta_{BC}}, \quad (9)$$

respectively. This combination was chosen based on the analysis of *Fuentes et al.* [1992], which found that for heavy soils such as the Yolo silt loam, only this combination of matric potential and hydraulic conductivity functions gave acceptable results. The needed fitting coefficients for the matric potential and hydraulic conductivity relationships were also taken from *Fuentes et al.* [1992] and are shown in Table 2. It should be

**Table 2.** Values of the Empirical Coefficients Used in Equations (8) and (9)

Variable	Value
$m$	0.0995
$n$	2.221
$\alpha, \text{cm}^{-1}$	0.0517
$\theta_s$	0.495
$\theta_r$	0
$K_{\text{sat}}, \text{cm/s}$	1.23e-5
$\eta_{BC}^*$	9.143

\*Brooks and Corey parameter.

noted that the value of  $\theta_r = 0$  is not a physical value but is instead a fitting parameter.

Although the relative change in surface tension with respect to temperature  $\gamma$  is not constant [Hopmans and Dane, 1986], it is generally assumed that the variation is small and the value of  $-2.09 \times 10^{-3} \text{ }^{\circ}\text{C}^{-1}$  given by Philip and de Vries [1957] was used. Note that for the Yolo silt loam, the liquid water flux due to thermal effects is calculated to be 3 orders of magnitude smaller than the diffusive flux due to moisture gradients for the moisture contents seen in this experiment. Any realistic change in  $\gamma$  is unlikely to affect our results significantly.

The theory of Philip and de Vries [1957] uses the thermodynamic relationship between water vapor density  $\rho_v$ , matric potential, and temperature [Edlefsen and Anderson, 1943]

$$\rho_v = \rho_{vs} \exp \left( \frac{\psi g}{RT} \right) \quad (10)$$

where  $\rho_{vs}$  is the saturated water vapor density at a given temperature,  $g$  is the acceleration due to gravity, and  $R$  is the gas constant. This relationship expressing the dependence of vapor density on matric potential and temperature allowed them to break the vapor flux into an isothermal part driven by the moisture content gradient and a thermal part driven by the temperature gradient,

$$\mathbf{q}_v = -D_{\theta v} \nabla \theta - D_{T v} \nabla T \quad (11)$$

where  $D_{\theta v}$  and  $D_{T v}$  are the appropriate diffusivities. The expressions derived by Philip and de Vries [1957] for the vapor diffusion coefficients are

$$D_{\theta v} = \frac{\alpha a D_a v g \rho_v}{RT \rho_l} \frac{\partial \psi}{\partial \theta} \quad (12)$$

and

$$D_{T v} = \eta \alpha a D_a v \frac{d \rho_{vs}}{dT} \quad (13)$$

where  $a$  is the volumetric air content;  $\alpha$  is a tortuosity factor;  $D_a$  is the diffusivity of water vapor in still air;  $v$  is a mass-flow factor taken to be 1; and  $g$ ,  $R$ , and  $\rho_{vs}$  are defined above. The term  $\eta$  (which is entirely different than the  $\eta_{BC}$  in (9), the Brooks and Corey equation) is a factor that accounts for (1) additional pore space available for the water vapor to move through, owing to the ability of water vapor to condense on one side of a water-saturated pore and evaporate on the other, and (2) the enhanced temperature gradient in individual pores relative to the bulk temperature gradient. For Philip and de Vries [1957]  $\eta$  is given by

$$\eta = \frac{a + f(\theta)}{\alpha a} \frac{(\nabla T)_a}{\nabla T}$$

where  $f$  is a moisture-dependent function that determines how much of the porosity is available for vapor transport, and  $(\nabla T)_a$  is the temperature gradient across the air-filled pores. The values for  $f$  suggested by Philip and de Vries [1957] are empirical, and likewise the value of  $(\nabla T)_a$  cannot be measured, so empirical values of  $\zeta = (\nabla T)_a / (\nabla T)$  must be used. Models for the enhancement factor  $\eta$  have been developed [Jury and Letey, 1979; Cary, 1979] but have been found not to agree well with measured values [Cass et al., 1984].

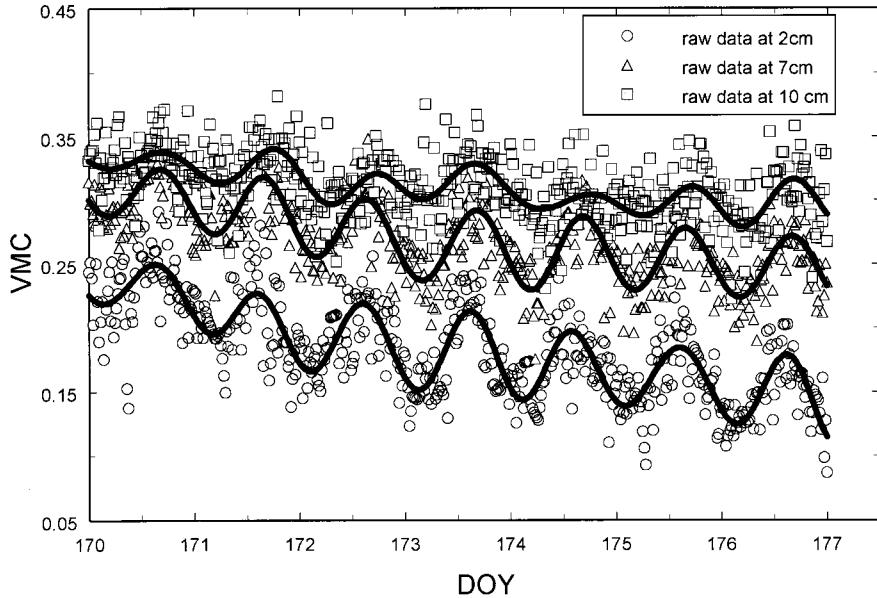
An alternate formulation for (11), which has been proposed by Milly [1982], uses matric potential  $\psi$  as the independent variable instead of  $\theta$ . The use of matric potential as the independent variable removes some of the difficulties which arise if the soil is not homogeneous, or if hysteresis is significant (Philip and de Vries' model has the assumption that there is no hysteresis.) Since our measurements were of soil moisture and not of soil matric potential, we did not consider the matric potential formulation. The transformation from  $\theta$  to  $\psi$  would have incorporated additional uncertainty.

### 3. Experiment

The data used for estimation of the diffusivities were collected at the Campbell Tract research field of the University of California at Davis. The soil of this field is the Yolo silt loam. Platinum resistance temperature detectors (PRTD) were installed at five depths: 2, 4, 7, 10, and 15 cm. Along side the PRTDs, time domain reflectometry (TDR) probes were inserted horizontally into the soil. These TDR probes were 30 cm long and were of the three-prong design discussed by Hennimovaara [1993]. The TDR waveforms and soil temperatures were measured every 20 min, and saved to a computer. The waveforms were later analyzed using the calibration presented by Dasberg and Hopmans [1992] for Yolo silt loam. This calibration was chosen after tests in the summer of 1994 with the same TDR set. In these tests it was also found that there were no calibration problems with the probes placed at the 2-cm depth.

The soil surface was kept bare of vegetation for the time period analyzed. The field was initially irrigated by approximately 4 cm of water by sprinkler irrigation on day of year (DOY) 164. Subsequent to the irrigation, an unexpected rain-storm took place on DOY 166, which although it delayed the dry-down of the soil, did provide the benefit of insuring that the soil was more uniformly wetted. The data used for the analysis in this paper are taken from DOYs 170–177, since that was an uninterrupted period of hot, cloudless days, which provided maximal evaporative forcing.

Because of noise inherent in the interpretation of TDR waveforms, it was deemed necessary to smooth the time series of soil moisture. This was accomplished by low-pass filtering at a bandwidth of  $1.82 \text{ day}^{-1}$  [Press et al., 1992]. As can be seen in Figure 1, which presents the results of this smoothing for the soil moisture time series, this procedure effectively removed the noise while preserving the essential shape of the diurnal variation in moisture content. The temperature measurements, which are shown in Figure 2, are much less affected by noise; in addition, certain features in the soil temperature time series, such as the sharp change in soil temperature when the sun rises (or when the temperature wave caused by the heating from sunrise reaches a



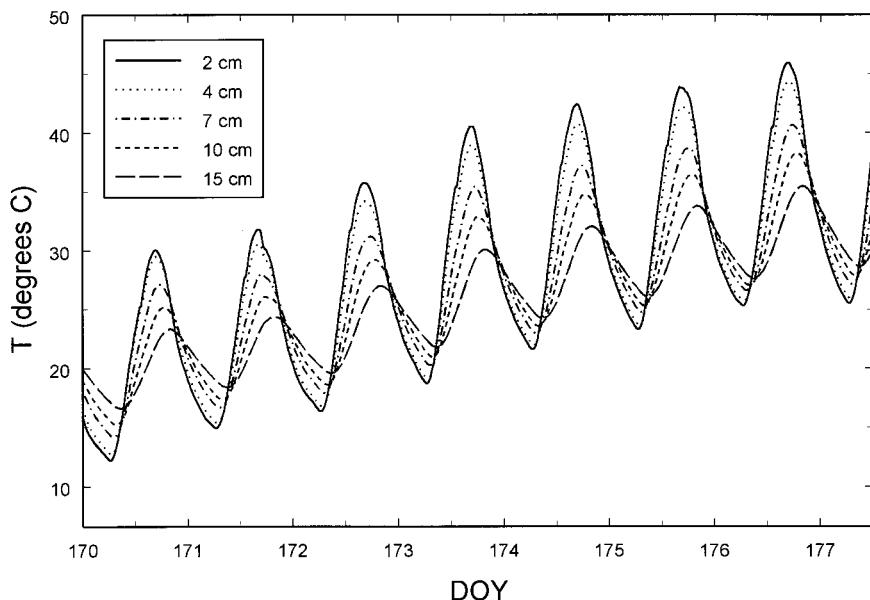
**Figure 1.** Raw moisture content measured by TDR, along with the smoothed data used in this study.

given depth) would be removed by smoothing, so no smoothing was done to these time series.

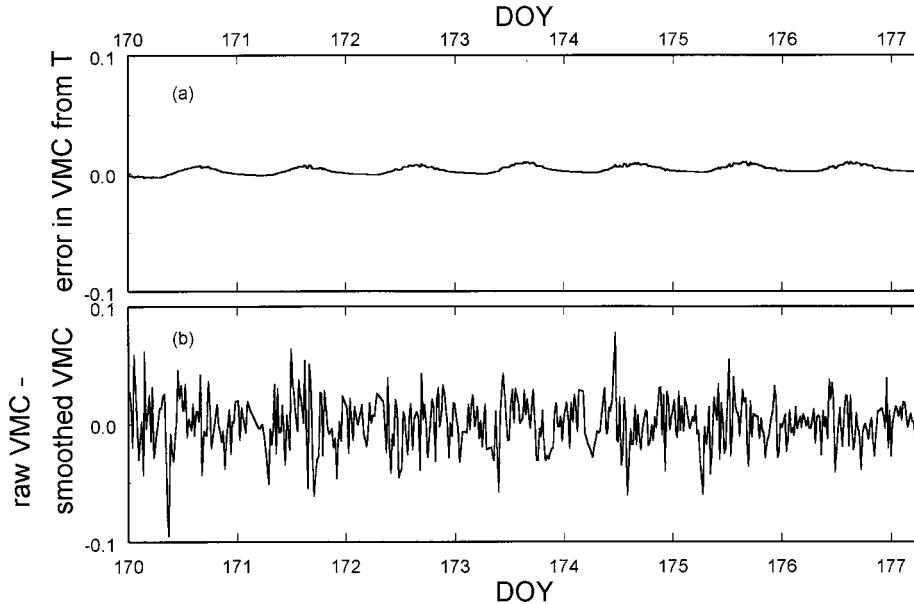
Some authors have proposed a temperature dependence of TDR soil moisture measurements, because of the temperature dependence of the dielectric constant of water. However, this correction seems to be insignificant for our measurements. Figure 3 presents the error due to temperature changes using the method of *Pepin et al.* [1995] with a reference temperature of 20°C, along with the noise removed by the low-pass filtering for the raw data taken at the 2-cm depth. As can be seen, the magnitude of the temperature correction is much less than the probable measurement noise itself. We have therefore ignored temperature effects on TDR measurements in our analysis and use the simple low-pass-filtered soil moisture time series.

#### 4. Analysis

The goal of our analysis was to invert (1) and (4) independently to obtain two different time series of  $\mathbf{q}_v$  for comparison. Ideally, if (1) and (4) are correct models of the heat and moisture transport and the values of  $\lambda$  and  $\mathbf{q}_v$  are known, then the two inversions should yield the same values of  $\mathbf{q}_v$ . In reality, even if the hydraulic and thermal properties as functions of moisture content were known perfectly, errors in the measurement of temperature and moisture content would introduce some disagreement between the two. We use the measured values of temperature and moisture content to compare the residuals from inverting the two transport equations, which in both cases should yield the vapor flux.



**Figure 2.** Temperature at listed depths, measured by platinum resistance temperature detectors.



**Figure 3.** (a) The change in the TDR moisture contents due to the temperature correction proposed by Pepin *et al.* [1995] and (b) the noise removed by the low-pass filtering, both for the 2-cm data.

The equation for conservation of energy can be written as

$$C \frac{\partial T}{\partial t} = -\nabla \cdot \mathbf{q}_h \quad (14)$$

where  $C$  is the heat capacity of the soil and  $t$  is time. With the assumption of one-dimensional vertical heat flow, this equation can be integrated over the layer of interest, yielding

$$\begin{aligned} \int_{7\text{cm}}^{10\text{cm}} C \frac{\partial T}{\partial t} dz &= \left( -\lambda \frac{\partial T}{\partial z} \right)_{10} - \left( -\lambda \frac{\partial T}{\partial z} \right)_7 \\ &+ (\rho(L + c_p(T - T_0))q_v)_{10} - (\rho(L + c_p(T - T_0))q_v)_7 \\ &+ (\rho c_p(T - T_0)q_l)_{10} - (\rho c_p(T - T_0)q_l)_7 \end{aligned} \quad (15)$$

where the left-hand side is net change in heat energy, the first two terms of the right-hand side are change due to conduction, terms three and four are change due to vapor flux, and the last two terms are change due to liquid flux. The derivatives of  $T$  were approximated by finite difference formulations. Formulae for the derivatives at 7 and 10 cm were developed using truncated Taylor series expansions around the points of interest; because measurements were available at three or more points, the resulting approximations for the first derivatives are second-order accurate. For the integral of the time rate of change, a polynomial was fit to  $C(\partial T / \partial t)$  as a function of depth, using the time series of temperature at the five measured depths, and then integrated between the given limits. This approach was taken (instead of simply averaging the values of  $C(\partial T / \partial t)$  at 7 and 10 cm and multiplying by 3 cm) after inspection of the profile of  $C(\partial T / \partial t)$  indicated that for some time periods there was significant curvature in the profile (which would cause error in the averaging approach). Although these time periods were very few in number, the integral approach was chosen for greater accuracy then.

The heat energy transported by the liquid flux was at all times very small, at least an order of magnitude less than the

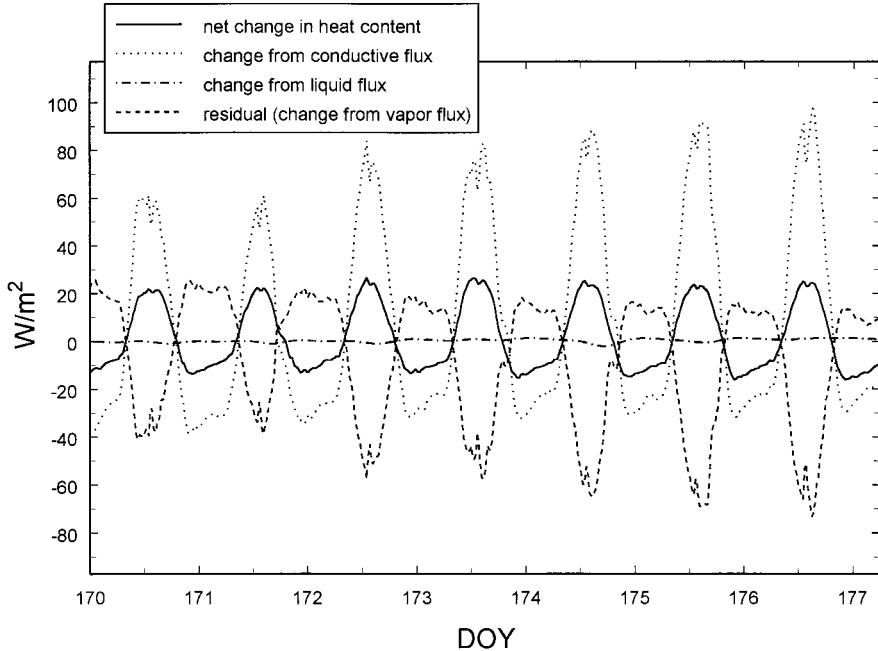
conductive heat flux. Time series of the four sets of terms in (15) (left-hand side, terms one and two, terms three and four, and terms five and six) are plotted in Figure 4. The large positive values of the change due to the conductive flux during the day indicate that the layer is heating up from the downward flux of heat energy. The net time rate of change in heat content in the 7- to 10-cm layer does not match the change due to conductive flux. The residual is the change due to vapor flux. Its large negative mean value indicates that significant heat is exiting the layer in the form of water vapor.

A similar analysis can be done for the mass flux. The mass balance equation can be integrated over the soil layer

$$\int_{7\text{cm}}^{10\text{cm}} \frac{\partial \theta}{\partial t} dz = (q_l)_{10} - (q_l)_7 + (q_v)_{10} - (q_v)_7 \quad (16)$$

where the left-hand side is the net change in moisture content, the first two terms of the right-hand side are change due to liquid flux, and the last two terms are change due to vapor flux. Since the net change in moisture content in the layer and the liquid water flux can be calculated with finite differences, the residual is again due to the vapor flux, this time in mass units. The results for the finite difference approximations of the left-hand side and the first two terms of the right-hand side of (16) along with the residual which is taken to be the vapor flux are shown in Figure 5. We see again that the mean residual is negative, indicating that water vapor must be leaving the layer, in order for the mass equation to close.

The time series of vapor flux calculated by the mass equation and vapor flux calculated by the energy equation are compared in Figure 6. As can be seen, there is in general good agreement between the two sets of values of water vapor flux. A number of factors could account for the what appears to be phase shift on the rising limb of the daily cycle, and the somewhat greater magnitude of the vapor flux derived from the mass balance. Since both vapor fluxes are calculated as residuals, they accumulate the errors in the measurements, the parameter esti-



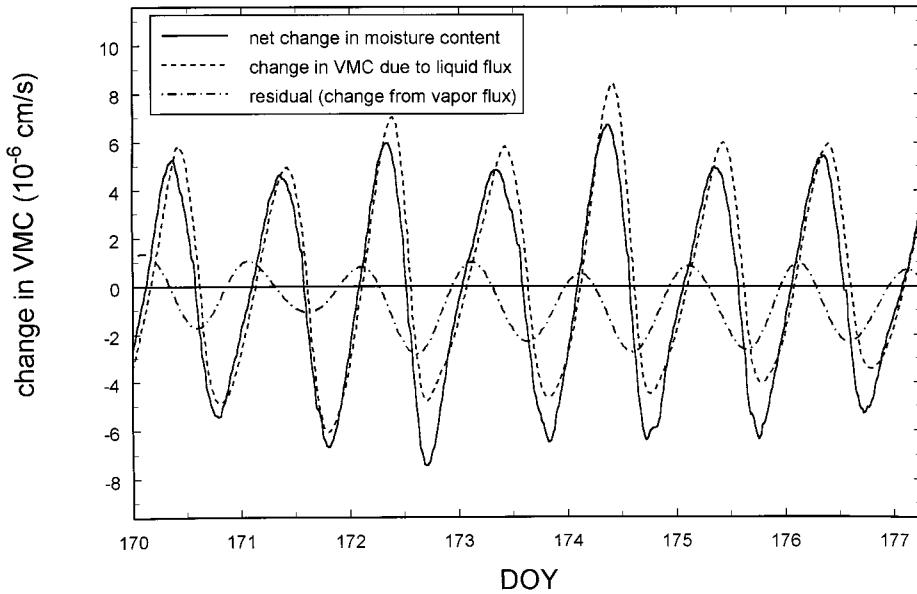
**Figure 4.** The net change in the heat content of the 7- to 10-cm soil layer, the amount of that change due to the difference between heat in through the top and out the bottom, the amount due to the heat transported by liquid water, and the residual, which is the heat transported out of the layer by vapor.

mates, and the discretization. Obviously, an error in the magnitude of the hydraulic and/or thermal parameters could lead to the greater magnitude of the mass balance vapor flux. The fact that the mass balance vapor flux often increases later in the evening than the energy balance vapor flux also is due to the accumulation of errors, especially to the parameter estimate. The filtering of the moisture time series may also cause some of the divergence between the two vapor fluxes. The high-frequency components which were truncated from

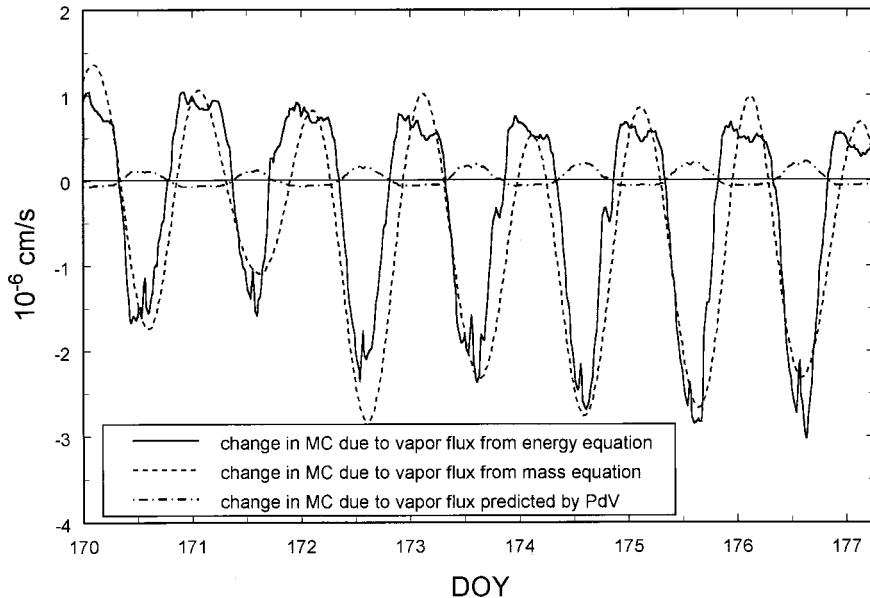
the moisture time series will have the greatest impact at regions where the flux time series is changing direction, so the removal of the signal along with the noise leads to the greatest disagreement there.

## 5. Discussion

It is apparent that if the values for thermal conductivity given by the shape factors of *Westcot and Wierenga* [1974] are



**Figure 5.** The net change in the moisture content of the 7- to 10-cm soil layer, the amount of that change due to the difference in liquid water flux through the upper and lower boundaries, and the residual, which is water removed from the layer by vapor flux.



**Figure 6.** Comparison of the changes in the moisture content in the 7- to 10-cm soil layer due to vapor transport calculated from the energy equation and the moisture equation. Positive values indicate a net transport of vapor to the soil layer; negative values indicate that more vapor is leaving the layer than entering. Also shown is the vapor flux through this layer calculated with the *Philip and de Vries* [1957] expressions for the vapor diffusion coefficients (equations (12) and (13)).

used, the effect of the water vapor flux on the total heat flux is significant. This is in contrast to the findings of *Kimball et al.* [1976], who continued the analysis of Jackson's group's [Jackson, 1973; Jackson *et al.*, 1974] soil temperature and moisture content data. *Kimball et al.* [1976] found that by adjusting the shape factors for Adelanto silt loam, they were able to match well the measured change in heat content and conductive flux, and they concluded thereby that vapor flux was negligible. This result would suggest that the large discrepancy between the net change in heat content and the change due to conductive flux could be because the Wierenga values for the shape factors for Yolo silt loam yield thermal conductivities are too large, which in turn overestimate the heat fluxes through the top and bottom of our layer. However, in order to match the net change in heat content with the change in heat content due to the conductive flux, it would be necessary to reduce the thermal conductivities on average by a factor of 3. This adjustment would in turn make the thermal conductivity of the soil less than that of pure water when saturated, which should not be the case, since the thermal conductivity of the solids is greater than pure water. We conclude therefore that the difference between the net change in heat in the soil layer and the change due to conductive flux cannot be due solely to inaccuracy in the determination of thermal conductivity.

An examination of Figure 5 shows that the mass equation too cannot be easily adjusted to remove the residual which is attributed to the net vapor flux. The residual largely arises from the phase difference between the total change in moisture content and the change due to the liquid water flux. One obvious means of reducing the residual would be to make the diffusion coefficient  $D_{0l}$  somewhat smaller during the early morning and somewhat larger during the late afternoon. Since the moisture contents at a given depth at these times are often roughly the same (see Figure 1), this sort of adjustment is not consistent. Rather than use the isothermal liquid diffusion

coefficient as a tuning parameter therefore, we believe that the use of the well-validated results of *Fuentes et al.* [1992] for the Yolo silt loam is justified.

Some comment should be made about the use of a low-pass filter to smooth the moisture time series. The use of this filter and the bandwidth chosen were both arbitrary choices made to ensure that the time derivative of the moisture content was smooth so that a reasonable comparison between the time and spatial derivatives could be made. Finite differencing tends to accentuate high-frequency noise, and the raw data of the moisture content was noisy enough that even when a 21-point moving average was used, the time derivatives had spurious spikes. As was mentioned earlier, the use of the filter may have caused some of the disagreement between the vapor fluxes calculated by the two different equations. However, there is no simple way of determining how much of the high-frequency part of the moisture content spectrum should be kept, so that the error introduced is an unavoidable result of the goal of a smooth finite difference approximation of the time derivative.

Even without the filtering, it is apparent that the moisture content at 2 cm peaks shortly after noon each day. Soil moisture peaks at this depth and time have been observed by other researchers [e.g., *Rose*, 1968a; *Jackson*, 1973; *Monji et al.*, 1990] under similar conditions. We believe that the increase in moisture content at 2 cm may be caused in part by the recondensation of water vapor from above. We are, however, unable to test this exactly because of the great difficulty of fine-scale measurements of soil moisture at the soil surface.

In Tables 3 and 4 we present some summary statistics on the changes caused by the different fluxes shown in Figures 4 and 5. From Figure 4 it can be seen that the three significant terms in the energy balance all reach their maximum magnitude around the same time each day. Table 3 shows that on average, the change of the heat content due to vapor flux is at nearly 50% of the sum of the remaining terms in the energy balance,

**Table 3.** The Average Daily Maximum Increase and Decrease of Heat Content in the 7- to 10-cm Soil Layer Due to the Different Transport Processes

	Average Daily Maximum Increase in Heat, W/m <sup>2</sup>	Average Daily Maximum Decrease in Heat, W/m <sup>2</sup>
Storage change	24.4	14.7
Conductive flux	81.0	33.3
Liquid flux	1.0	0.7
Residual (vapor flux)	19.4	57.5

during the time of maximum change. For the changes in moisture content, the changes are not so neatly in phase; however, the magnitude of the residual in Figure 5 does not change much from its maximum at the point where the other two terms are near their maximum. The vapor flux accounts for roughly 25% of the remaining two terms at the time of maximal moisture content change.

The majority of vapor flow in systems where combined heat and moisture transfer is considered has been attributed to thermal vapor flow [Rose, 1968b]. The time series of the changes in moisture content in the 7- to 10-cm layer from vapor flux using the Philip and de Vries [1957] theoretical values for the diffusion coefficients is shown in Figure 6 along with the measured time series from this study. The total theoretical vapor flux (predicted by Philip and de Vries in the caption) is essentially all thermally driven vapor flux, since the thermal term is  $10^3$  larger than the isothermally driven flux. It is apparent that although the phases of the theoretical and experimental time series are well correlated, the magnitude of change due to the theoretical vapor flux is much smaller than the measured change and, perhaps more strikingly, opposite in direction. The fact that the magnitude of the change due to the vapor flux predicted by the Philip and de Vries coefficients is less than the magnitude of the measured change is not unusual, since this result has been reported by Cass *et al.* [1984], who found that a value of 10 to 20 is needed for the enhancement factor for satisfactory match between the predicted water vapor flux and observations. However the change in sign between the theoretical change and the measured change indicates that the theoretical model may not be capturing all of the physical processes occurring. The fact that the theoretical change is positive during the daytime indicates that the moisture content in the 7- to 10-cm layer is increasing because of vapor flow at the time when there is maximal solar heating, due to the downward thermal gradient. The negative sign of the measured change indicates that the soil layer is losing moisture during this time period from the net vapor flux. This loss of soil moisture is what would be expected during the daytime, when

a large amount of evaporation can be measured from the soil surface. The consistency of the results from the mass and energy equations, the low probability that the residuals in these equations can be entirely explained away by parameter adjustment, and the intuitive expectation that during the daytime when evaporation occurs the moisture content of the soil in general will decrease, leads us to believe that the measured results in Figure 6 are an accurate picture of the effect of the soil vapor flux on the studied soil layer.

## 6. Conclusions

We have presented the experimental results of an investigation of combined heat and moisture transport in a field soil. It was found that the energy and mass balances yielded good agreement on the magnitude and direction of the water vapor flux. The theory for diffusive vapor flow described by Philip and de Vries [1957] underestimated the magnitude of vapor flux significantly and, more importantly, predicted that the net vapor flux would increase the moisture in a soil layer, when in fact the net vapor flux removed moisture from the soil layer. These differences indicate that the current theory of vapor flux is not accounting for some important physical process. The amount of water and energy transported by water vapor was found to be a significant part of the mass and energy balances. The importance of the vapor flux to the mass and energy balances is due to the large diurnal temperature fluctuations in the soil and the lack of vegetation at the site. Because the vapor fluxes were calculated as residuals in the energy and mass balances, they contain all the errors associated with measurement and parameter uncertainty.

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**Table 4.** The Average Daily Maximum Increase and Decrease of Moisture Content in the 7- to 10-cm Soil Layer Due to the Different Transport Processes

	Average Daily Maximum Increase in $\theta$ , cm/s $\times 10^{-6}$	Average Daily Maximum Decrease in $\theta$ , cm/s $\times 10^{-6}$
Storage change	5.39	6.32
Liquid flux	6.29	4.61
Residual (vapor flux)	0.89	2.47

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